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► **To cite this version:**

Edwin R.C. Baynes, Dimitri Lague, Jean-Jacques Kermarrec. Supercritical river terraces generated by hydraulic and geomorphic interactions. *Geology*, 2018, 46 (6), pp.499-502. 10.1130/G40071.1 . insu-01769812

HAL Id: insu-01769812

<https://insu.hal.science/insu-01769812v1>

Submitted on 5 Jul 2018

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1 **Supercritical river terraces generated by hydraulic and geomorphic interactions**

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5

6 **ABSTRACT**

7 The alternating cycle of valley widening through lateral erosion ('strath planation') and valley
8 narrowing through vertical incision into bedrock ('strath terrace abandonment') due to
9 variations in sediment supply (Q_s) relative to river transport capacity (Q_{sc}) is a common
10 feature in many mountainous environments, yet our understanding of the mechanics of the
11 processes that drive this landscape change remains poorly quantified. Here, we used an
12 experimental and numerical study to identify the geomorphic and hydraulic controls driving
13 the response of mixed bedrock-alluvial rivers to variable sediment supply, water discharge
14 and tectonic tilting. The experimental channels exhibit a multi-stage response of channel
15 narrowing, stripping of the alluvial cover in a downstream migrating incision wave followed by
16 destabilization of the bed and development of a single vertical step in the bed profile
17 ('knickpoint') when the hydraulic conditions are supercritical. In our experiments headward
18 erosion by knickpoints is the most efficient process of strath terrace abandonment,
19 contributing the majority of the total vertical incision in a short period of time. We show
20 experimentally, that knickpoint developing under supercritical flow conditions, driving the
21 rapid response of fluvial systems to upstream perturbations in Q_s/Q_{sc} despite no base-level
22 fall. This has implications for the understanding of distributions of strath terrace ages, the
23 inference of baselevel variations from knickpoint propagation, and how landscapes respond
24 to climatic or tectonic perturbations.

25 **INTRODUCTION**

26 Past cycles of global climatic change have had consequences for the hydrological regime
27 and the pattern of sediment production and transport within landscapes (Schumm and
28 Parker, 1973; Fuller et al., 2009; Wobus et al., 2010). The balance between the flux of
29 sediment entering a mixed bedrock-alluvial river (Q_s) and the capacity of the river to
30 transport that sediment (Q_{sc}) is important in setting the landscape morphology, such as the
31 generation and preservation of flights of strath terraces above the active channel (e.g.,
32 Pazzaglia et al., 1998; Wegmann and Pazzaglia, 2002).

33 Several mechanisms have been proposed for the generation of strath ‘terraces’, including
34 climate driven oscillations in Q_s/Q_{sc} resulting in valley widening (high Q_s/Q_{sc}) or vertical
35 incision (low Q_s/Q_{sc}) (Bull, 1990; Hancock and Anderson, 2002). The presence of strath
36 terraces within a fluvial system have also been shown to result from headward incision by
37 ‘base-level knickpoints’ initiated by fault throw or tectonically driven base level fall (Finnegan,
38 2013; Finnegan and Balco, 2013), unsteady lateral river erosion (Limaye and Lamb, 2016),
39 bedrock structure (Wohl, 2008), the cutoff and migration of river meander bends (Finnegan
40 and Dietrich, 2011), sea level highstands (Merritts et al., 1994), river capture (Stokes et al.,
41 2002). The influence of tectonics in generating strath terraces is usually limited to the
42 consideration to the initiation of headward migrating incision waves (e.g., Finnegan 2013),
43 while the impact of regional landscape tilting is rarely considered. A clear identification of the
44 mechanism of strath terrace formation is important for the interpretation of the presence and
45 pattern of strath terraces within a landscape to date past climate changes (e.g., Fuller et al.,
46 2009; Stokes et al., 2017), estimate fluvial incision rates (e.g., Lavé and Avouac, 2001), or to
47 help calibrate empirical models of fluvial erosion (Finnegan, 2013). However, a universal
48 model of strath terrace formation is currently limited by an incomplete understanding of the
49 mechanics of vertical and lateral erosion in bedrock and difficulties in identifying the erosional
50 pattern during the response of channel processes to perturbations in Q_s/Q_{sc} from preserved
51 strath terrace age distributions.

52 Here, we present a physical modeling study that explores the mechanisms of channel
53 response and resulting terrace geometry to three scenarios of perturbations to Q_s/Q_{sc} . The
54 first two experiments explore the response of a channel following a sudden perturbation to
55 Q_s/Q_{sc} from the same initial conditions while the third explores the impact of a gradual
56 decrease in Q_s/Q_{sc} , through progressive tilting of the experiment. The experiments identify
57 patterns of erosional response including the resulting preserved terrace morphology, aiding
58 the interpretation of terrace age distributions in natural landscapes for purposes such as the
59 reconstruction of climate or tectonic histories (e.g., Fuller et al., 2009; Stokes et al., 2017)

60

61 **EXPERIMENTAL SETUP**

62 Experiments have been used extensively to explore the dynamics of bedrock channels,
63 including abrasion processes (e.g., Johnson and Whipple, 2007), the dynamics of bedrock
64 plucking (e.g., Dubinski and Wohl, 2013), the mechanics of waterfall erosion in homogenous
65 bedrock (e.g., Baynes et al., 2018), and landscape stability (e.g., Hasbargen and Paola,
66 2000). We used the 80 cm long, 30 cm wide and 20 cm deep Bedrock River Experimental
67 Incision Tank at the Université de Rennes (Fig. 1; S1), with an initial configuration of a thin (2

68 cm) layer of well-sorted sand (grain size = 250 μm) overlying homogenous silica paste (grain
69 size = 45 μm). The cohesive silica paste material is commonly used to represent bedrock
70 and is eroded by hydraulic shear under clear water flow in the absence of sediment (e.g.,
71 Turowski et al., 2005; Turowski et al., 2006; Turowski, 2007) and the experimental channels
72 in this study exhibit a similar hydraulic scaling of width and slope to natural bedrock rivers
73 (Baynes et al., 2018). No direct scaling between our experiments and a particular target
74 natural system is sought but the behavior of the experiment is similar to that observed in
75 natural systems, allowing the dynamics of processes and their controls to be explored
76 quantitatively (Hooke, 1968; SI Section 1).

77 Constant input water discharge, set within 2% by a regulated pump on the inflow pipe, was
78 fed into a reservoir upstream of the inlet, and sediment was added using a Gericke infinite
79 screw feeder system. The channel outlet elevation was fixed, although it was free to migrate
80 across the width of the box, and sediment was collected in a reservoir beyond the outlet to
81 avoid deposition in the channel as a result of the outlet. The same initial conditions for the
82 first two experiments were $Q = 1.5 \text{ l/min}$, $Q_s = 13.33 \text{ g/min}$, slope = 4.5° , selected after
83 previous experiments for the best combination of parameters to generate an alluvial river at
84 equilibrium conditions inset in the sand material. Channel slope, width and depth were free to
85 evolve during the course of the experiment through erosion (vertical and lateral) and
86 deposition. After 37 minutes, once the channel had reached a constant width, one parameter
87 was altered while all others kept constant to represent a sudden shift in Q_s/Q_{sc} ; experiment
88 1: Q_s reduced to 0 g/min, experiment 2: Q doubled to 3 l/min. In the third experiment, Q_s and
89 Q were maintained constant throughout but the slope was increased from an initial slope of
90 2.5° at a continuous rate of 1° per hour for 4.7 hours, gradually decreasing Q_s/Q_{sc} . A Leica
91 ScanStation 2 Terrestrial Laser Scanner (green laser footprint = 3 mm) was programmed to
92 collect point clouds of the topography every 2 min. The topography was then gridded with a
93 pixel size of 2 mm, and combined with the hydrodynamic model Floodos (Davy et al., 2017),
94 to evaluate the hydraulic conditions within the channel (e.g., Froude number) and to quantify
95 the evolution of the channel geometry (e.g. width and depth) during the experiments (SI
96 Section 2).

97 **EXPERIMENTAL RESULTS AND DISCUSSION**

98 Changing Q_s/Q_{sc} had a significant, rapid and complex impact on the initial stable state
99 (channel width: 150 mm) in the first two experiments. Following the removal of input Q_s in
100 experiment 1, an initial narrowing of the channel coincided with a downstream migrating
101 incision wave initiated at the inlet that removed the alluvial cover (Fig. 2A), exposing the
102 bedrock surface to erosion. The removal of input Q_s led to $Q_s/Q_{sc} = 0$, providing capacity for

103 the flow to entrain and strip the sediment stored in the alluvial bed. Further narrowing of the
104 channel occurred until the bed destabilized into a series of steps (spaced at 25-30 mm with
105 height of 3-5 mm, at $t = 83$ min) in the bedrock surface that evolved into a single vertical step
106 (15-20 mm in height at $t = 111$ min) that migrated upstream as a knickpoint, depositing
107 material downstream of the plunge pool (Fig. 2A). The same multi-stage process of
108 downstream stripping of alluvial cover followed by knickpoint development and headward
109 erosion also occurred in experiment 2 where Q was doubled (SI Section 3). The
110 development of steps before the generation of a single knickpoint were not observed in the
111 topographic data from experiment 2 as the channel evolved from downstream stripping to
112 headward erosion within the time period between successive laser scans ($t = 120$ s; Fig.
113 S5B). Despite this, the knickpoint formed mid-channel in the absence of any base level fall or
114 tectonic uplift.

115 The initial stages of experiment 3 (until $t \sim 100$ min) were characterized by an alluvial fan-like
116 regime near the inlet with cone-shaped deposits frequently reworked by high lateral channel
117 mobility (Fig. 2B) while the slope increased (Fig. S5C). As Q_s/Q_{sc} decreased, the flow
118 focused into a narrower single thread, triggering the onset of a downstream incision wave of
119 the alluvial cover. Subsequently, a knickpoint developed in the bedrock ($t \sim 208$ min) which
120 retreated upstream in the same manner as the other experiments (Fig. 2B; SI Section 3). In
121 each experiment, the stripping of alluvial cover by the downstream incision wave coincided
122 with a significant narrowing of the channel inset in the alluvial cover, with the narrowest width
123 occurring in bedrock when a knickpoint had formed (Fig. 2B). The channel narrowing
124 associated with the incision of both the alluvial cover and the bedrock led to the
125 abandonment of the formerly active bed at higher elevations, representing the generation of
126 strath terraces seen in natural environments. However, the rates of vertical incision and
127 terrace abandonment were not spatially or temporally constant during the experiments, with
128 the highest local rates of vertical incision associated with the downstream stripping of
129 sediment and the headward migration of the knickpoints through bedrock (Fig. 2B; SI Section
130 4). A similar contribution to total vertical incision by knickpoint migration was identified in
131 Hawai'i (Mackey et al., 2014).

132 The Floodos hydrodynamic model output (SI Section 5; Fig. S10) shows that the sequence of
133 steps, and subsequently the knickpoint, developed in experiment 1 where flow had a Froude
134 number > 1 . The Froude number, a dimensionless number defined as the ratio of the inertia
135 forces to gravitational forces, characterizes the transition from subcritical to supercritical flow
136 when $Fr > 1$. During supercritical flow, interactions between the flow and the bed are
137 inherently unstable (Yokokwa et al., 2013; Izumi et al., 2017; Scheingross and Lamb, 2017),
138 and the development of steps in the bedrock surface occur during our experiments when the

139 flow conditions exceed the threshold Froude number for supercritical flow. We therefore refer
140 to these features as 'supercritical knickpoints' to distinguish them from 'base-level
141 knickpoints'. Subsequently, we refer to the terraces generated by the passage of supercritical
142 knickpoints as 'supercritical terraces'.

143 **IMPLICATIONS FOR THE UNDERSTANDING OF NATURAL LANDSCAPES**

144 Our experiments provide direct evidence of the hydraulic processes that lead to strath
145 terrace formation following perturbations to Q_s/Q_{sc} . Wobus et al. (2010) show that climate
146 change induced incision was associated with a downstream incision wave due to changes in
147 the relative sediment supply to transport capacity, while base-level induced (e.g. river
148 capture, tectonics) incision was characterized by an upstream migrating incision wave.
149 Crucially, our experiments show that channels in homogenous substrate respond to climatic
150 perturbations through a complex multi-stage incision mechanism, containing a downstream
151 (Fig. 3A) and upstream (Fig. 3C) wave component that both generate terraces, in the
152 absence of any base level fall.

153 The multi-stage nature of the incision (Fig. 3) has important implications for studies that use
154 the age of terraces to reconstruct climatic changes (e.g., Fuller et al., 2009) or rates of fluvial
155 incision and tectonic uplift (e.g., Lavé and Avouac, 2001), due to the time transgressive
156 nature of the terraces generated by this response (Zaprowski et al., 2001). During the
157 downstream incision wave, the upstream parts of the alluvium are abandoned first, while the
158 opposite is true during the upstream incision wave, leading to spatially variable terrace ages
159 (Fig. 3D). If it is incorrectly assumed that the entire length of a terrace was abandoned at the
160 same time, the average vertical incision rate calculated from a single terrace abandonment
161 age would be misleading due to the failure to identify the erosion history of an incision wave
162 and the spatially and temporally non-uniform nature of the river incision (Finnegan et al.,
163 2014). Strategic sampling of high-resolution point ages from terraces may differentiate
164 between strath terrace formation by vertical incision or knickpoint retreat (e.g., samples
165 collected longitudinally from the same terrace surface; Baynes et al., 2015), allowing
166 accurate interpretation of the mechanisms and triggering perturbations that generated them.

167 Knickpoints in a channel are traditionally interpreted as markers of either local lithological
168 boundaries (e.g., Haviv et al., 2010), a history of base level fall or tectonic uplift rate
169 variability (e.g., Finnegan, 2013), or threshold drainage areas for channel incision (Crosby
170 and Whipple, 2006). However, our experiments indicate that the presence of a knickpoint in a
171 channel with homogenous bedrock and constant baselevel is not necessarily diagnostic of
172 any of these sources as knickpoints are generated following perturbations to Q_s/Q_{sc} (Fig. 3).
173 Under normal conditions, flow in large natural alluvial rivers where strath terraces are

174 commonly found is subcritical, but above-threshold supercritical flow conditions ($Fr > 1$) can
175 occur due to a combination of driving factors including channel narrowing with variable width
176 and increased discharge (SI Section 5). Base-level terraces are likely to have the same or a
177 similar slope to the active channel bed through parallel knickpoint retreat (Finnegan, 2013),
178 whereas terraces generated during the multi-stage response identified here are steeper than
179 the active channel (Fig. 2, 3) due to new equilibrium conditions with lower sediment supply,
180 higher discharge or continued tilting. An additional diagnostic link could be the highest
181 terrace having a decreasing abandonment age with distance downstream (Fig. 3D) that does
182 not occur during terrace formation driven by base-level fall. Thus, the morphology of terraces
183 and spatial patterns of abandonment ages can be diagnostic of the processes and forcing
184 perturbations that led to terrace generation and should be considered carefully when
185 analyzing topography for the purpose of understanding and reconstructing past tectonic or
186 climatic histories.

187 **CONCLUSIONS**

188 Landscapes are able to respond rapidly following climatic or anthropogenic perturbations that
189 switch the erosion processes from one regime to another. The transition from lateral erosion
190 and strath planation to channel narrowing, incision and strath terrace abandonment occurs
191 through a multi-stage process of downstream then upstream incision waves. The headward
192 passage of knickpoints can be initiated even in the absence of base level fall, due to the
193 interaction between the channel bed and supercritical flow conditions. The passage of
194 supercritical knickpoints provide an efficient process for the abandonment of strath terraces,
195 and should be considered when using the presence of strath terraces within landscapes to
196 infer past fluvial incision rates or climatic changes, which may be more spatially and
197 temporally variable than previously thought.

198 **ACKNOWLEDGMENTS**

199 We thank F. Métivier for the loan of the sediment feeder, P. Davy for providing the Floodos
200 code and T. Croissant for his help with the model runs. We thank J. Scheingross, A. Mather,
201 A. Limaye and an anonymous reviewer for constructive comments that helped improve the
202 final version of this manuscript. The research was funded by Marie Skłodowska-Curie
203 Actions Individual Fellowship No. 703230 (to E.B).

204 **REFERENCES**

205 Baynes, E.R.C., Attal, M., Niedermann, S., Kirstein, L.A., Dugmore, A.J., and Naylor, M.,
206 2015, Erosion during extreme flood events dominates Holocene canyon evolution in

207 northeast Iceland: Proceedings of the National Academy of Sciences of the United States of
208 America, v. 112, p. 2355–2360, <https://doi.org/10.1073/pnas.1415443112>.

209 Baynes E.R.C., Lague D., Attal M., Gangloff A., Kirstein L.A., Dugmore A.J., 2018, River self-
210 organisation inhibits discharge control on waterfall migration: Scientific Reports 8, 2444,
211 <https://doi.org/10.1038/s41598-018-20767-6>

212 Bull, W.B., 1990, Stream-terrace genesis: Implications for soil development: Geomorphology,
213 v. 3, p. 351–367, [https://doi.org/10.1016/0169-555X\(90\)90011-E](https://doi.org/10.1016/0169-555X(90)90011-E).

214 Crosby, B.T., and Whipple, K.X., 2006, Knickpoint initiation and distribution within fluvial
215 networks: 236 waterfalls in the Waipaoa River, North Island, New Zealand: Geomorphology,
216 v. 82, p. 16–38, <https://doi.org/10.1016/j.geomorph.2005.08.023>.

217 Davy, P., Croissant, T., and Lague, D., 2017, A precipitation method to calculate river
218 hydrodynamics, with applications to flood prediction, landscape evolution models and
219 braiding instabilities: Journal of Geophysical Research. Earth Surface, v. 122, p. 1491–1512,
220 <https://doi.org/10.1002/2016JF004156>.

221 Dubinski, I.M., and Wohl, E., 2013, Relationships between block quarrying, bed shear stress,
222 and stream power: A physical model of block quarrying of a jointed bedrock channel:
223 Geomorphology, v. 180–181, p. 66–81, <https://doi.org/10.1016/j.geomorph.2012.09.007>.

224 Finnegan, N.J., 2013, Interpretation and downstream correlation of bedrock river terrace
225 treads created from propagating knickpoints: Journal of Geophysical Research. Earth
226 Surface, v. 118, p. 54–64, <https://doi.org/10.1029/2012JF002534>.

227 Finnegan, N.J., and Dietrich, W.E., 2011, Episodic bedrock strath terrace formation due to
228 meander migration and cutoff: Geology, v. 39, p. 143–146, <https://doi.org/10.1130/G31716.1>.

229 Finnegan, N.J., and Balco, G., 2013, Sediment supply, base level, braiding, and bedrock
230 river terrace formation: Arroyo Seco, California, USA: Geological Society of America Bulletin,
231 v. 125, p. 1114–1124, <https://doi.org/10.1130/B30727.1>.

232 Finnegan, N.J., Schumer, R., Finnegan, S., 2014, A signature of transience in bedrock river
233 incision rates over timescales of 104–107 years: Nature, v. 505, p. 391–394,
234 <https://doi.org/10.1038/nature12913>

235 Fuller, T.K., Perg, L.A., Willenbring, J.K., and Lepper, K., 2009, Field evidence for climate-
236 driven changes in sediment supply leading to strath terrace formation: Geology, v. 37, p.
237 467–470, <https://doi.org/10.1130/G25487A.1>.

238 Hancock, G.S., and Anderson, R.S., 2002, Numerical modelling of fluvial strath-terrace
239 formation in response to oscillating climate: *Geological Society of America Bulletin*, v. 114, p.
240 1131–1142.

241 Hasbargen, L.E., and Paola, C., 2000, Landscape instability in an experimental drainage
242 basin: *Geology*, v. 28, p. 1067-1070, [https://doi.org/10.1130/0091-
243 7613\(2000\)28<1067:LIIAED>2.0.CO;2](https://doi.org/10.1130/0091-7613(2000)28<1067:LIIAED>2.0.CO;2)

244 Haviv, I., Enzel, Y., Whipple, K.X., Zilberman, E., Matmon, A., Stone, J., and Fifield, K.L.,
245 2010, Evolution of vertical knickpoints (waterfalls) with resistant caprock: Insights from
246 numerical modelling: *Journal of Geophysical Research*, v. 115, p. F03028,
247 <https://doi.org/10.1029/2008JF001187>.

248 Hooke, R. 1968, Model Geology: Prototype and Laboratory Streams: Discussion: *Geological*
249 *Society of America Bulletin*, v. 79, p. 391–94, [https://doi.org/10.1130/0016-
250 7606\(1968\)79\[391:MGPALS\]2.0.CO;2](https://doi.org/10.1130/0016-7606(1968)79[391:MGPALS]2.0.CO;2)

251 Izumi, N., Yokokawa, M., and Parker, G., 2017, Incisional cyclic steps of permanent form in
252 mixed bedrock-alluvial rivers: *Journal of Geophysical Research. Earth Surface*, v. 122, p.
253 130–152, <https://doi.org/10.1002/2016JF003847>.

254 Johnson, J.P.L., and Whipple, K.X., 2007, Feedbacks between erosion and sediment
255 transport in experimental bedrock channels: *Earth Surface Processes and Landforms*, v. 32,
256 p. 1048–1062, <https://doi.org/10.1002/esp.1471>.

257 Lavé, J., and Avouac, J.P., 2001, Fluvial incision and tectonic uplift across the Himalayas of
258 central Nepal: *Journal of Geophysical Research*, v. 106, p. 26561–26591,
259 <https://doi.org/10.1029/2001JB000359>.

260 Limaye, A.B.S., and Lamb, M.P., 2016, Numerical model predictions of autogenic fluvial
261 terraces and comparison to climate change expectations: *Journal of Geophysical Research.*
262 *Earth Surface*, v. 121, p. 512–544, <https://doi.org/10.1002/2014JF003392>.

263 Mackey, B.H., Scheingross, J.S., Lamb, M.P., and Farley, K.A., 2014, Knickpoint formation,
264 rapid propagation, and landscape response following coastal cliff retreat at the last
265 interglacial sea-level highstand: Kaua'i, Hawai'i: *Geological Society of America Bulletin*, v.
266 126, p. 925–942, <https://doi.org/10.1130/B30930.1>.

267 Merritts, D.J., Vincent, K.R., and Wohl, E.E., 1994, Long river profiles, tectonism, and
268 eustasy: A guide to interpreting fluvial terraces: *Journal of Geophysical Research*, v. 99, p.
269 14031–14050, <https://doi.org/10.1029/94JB00857>.

270 Parker, G., and Izumi, N., 2000, Purely erosional cyclic and solitary steps created by flow
271 over a cohesive bed: *Journal of Fluid Mechanics*, v. 419, p. 203–238,
272 <https://doi.org/10.1017/S0022112000001403>.

273 Pazzaglia, F.J., Gardner, T.W., and Merritts, D., 1998, Bedrock fluvial incision and
274 longitudinal profile development over geologic time scales determined by fluvial terraces, in
275 Wohl, E., and Tinkler, K., eds., *Bedrock channels: American Geophysical Union Geophysical*
276 *Monograph Series 107*, p. 207–235, <https://doi.org/10.1029/GM107p0207>.

277 Scheingross, J.S., and Lamb M.P., 2017, A mechanistic model of waterfall plunge-pool
278 erosion into bedrock, *Journal of Geophysical Research. Earth Surface*, v. 122, p. 2079-2104,
279 <https://doi:10.1002/2017JF004195>.

280 Schumm, S.A., and Parker, R.S., 1973, Implications of Complex Response of Drainage
281 Systems for Quaternary Alluvial Stratigraphy: *Nature*, v. 243, p. 99–100.

282 Stokes, M., Mather, A.E., and Harvey, A.M., 2002, Quantification of river-capture-induced
283 base-level changes and landscape development, Sorbas Basin, SE Spain: *Geological*
284 *Society of London, Special Publications*, v. 191, p. 23–35,
285 <https://doi.org/10.1144/GSL.SP.2002.191.01.03>.

286 Stokes, M., Mather, A.E., Belfoul, M., Faik, F., Bouzid, S., Geach, M.R., Cunha, P.P.,
287 Boulton, S.J., and Thiel, C., 2017, Controls on dryland mountain landscape development
288 along the NW Saharan desert margin: Insights from Quaternary river terrace
289 sequences(Dades River, south-central High Atlas, Morocco): *Quaternary Science Reviews*,
290 v. 166, p. 363–379, <https://doi.org/10.1016/j.quascirev.2017.04.017>.

291 Turowski J.M., 2007, Controls on bedrock channel morphology: experimental and theoretical
292 investigations and comparison with natural channels, University of Cambridge PhD Thesis
293 (<http://ethos.bl.uk/OrderDetails.do?uin=uk.bl.ethos.613289>).

294 Turowski J.M., Lague D., Hovius N., Crave A., 2005, Dynamics and steady-state geometry of
295 an experimental channel incising cohesive material; *Geophysical Research Abstracts. EGU*
296 *General Assembly 2005, Apr 2005, Vienna, Austria. European Geosciences Union,*
297 pp. EGU05-A-05294, 2005, Vol 7

298 Turowski, J.M., Lague, D., Crave, A., and Hovius, N., 2006, Experimental channel response
299 to tectonic uplift: *Journal of Geophysical Research*, v. 111, F03008,
300 <https://doi.org/10.1029/2005JF000306>.

301 Wegmann, K.W., and Pazzaglia, F.J., 2002, Holocene strath terraces, climate change, and
302 active tectonics: The Clearwater River basin, Olympic Peninsula, Washington State:

303 Geological Society of America Bulletin, v. 114, p. 731–744, [https://doi.org/10.1130/0016-](https://doi.org/10.1130/0016-7606(2002)114<0731:HSTCCA>2.0.CO;2)
304 7606(2002)114<0731:HSTCCA>2.0.CO;2.

305 Wobus, C.W., Tucker, G.E., and Anderson, R.S., 2010, Does climate change create
306 distinctive patterns of landscape incision?: Journal of Geophysical Research, v. 115, p.
307 F04008, <https://doi.org/10.1029/2009JF001562>.

308 Wohl, E., 2008, The effect of bedrock jointing on the formation of straths in the Cache la
309 Poudre River drainage, Colorado Front Range: Journal of Geophysical Research, v. 113, p.
310 F01007, <https://doi.org/10.1029/2007JF000817>.

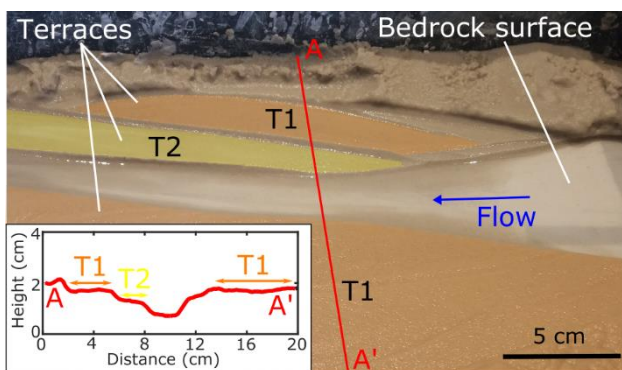
311 Yokokwa, M., Kotera, A., Kyogoku, A., and Izumi, N., 2013, Cyclic steps by bedrock incision,
312 in Fukuoka, S., et al., eds., Advances in River Sediment Research: The Netherlands, CRC
313 Press, p. 629–633.

314 Zaprowski, B.J., Evenson, E.B., Pazzaglia, F.J., and Epstein J.B., 2002, Knickzone
315 propagation in the Black Hills and northern High Plains: A different perspective on the late
316 Cenozoic exhumation of the Laramide Rocky Mountains, Geology, v. 29, p. 547-550,
317 [https://doi:10.1130/0091-7613\(2001\)029<0547:kpitbh>2.0.co;2](https://doi:10.1130/0091-7613(2001)029<0547:kpitbh>2.0.co;2).

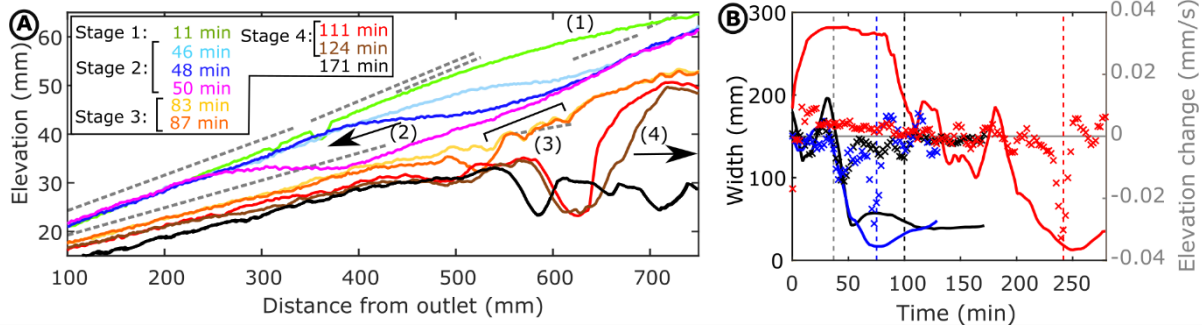
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319 FIGURES

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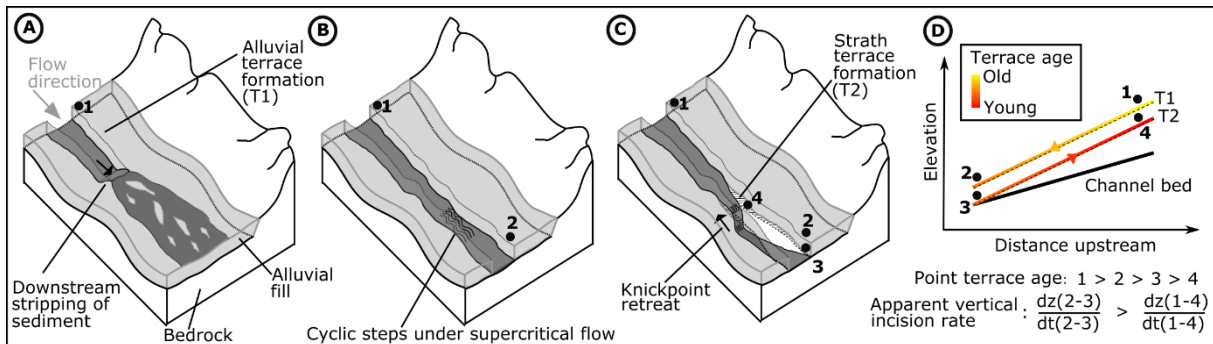


321 *Figure 1. : Example of terraces generated in the experiments. The white surface is the silica*
322 *paste acting as cohesive bedrock and the brown material is sand, acting as coarse material*
323 *transported as bedload. Progressive channel incision and channel narrowing leads to the*
324 *abandonment of the previously active river bed, preserved as strath terraces. Inset shows*
325 *topographic cross-section from A to A', with terraces labeled in yellow and orange.*



326

327 Figure 2. (A). Extracted long profiles during experiment 1. Green line ($t = 11$ min) shows the
 328 profile of the alluvial river, with the subsequent stripping of the sediment cover progressively
 329 exposing the bedrock surface within the channel. Black solid line ($t = 171$ min) shows the
 330 final long profile. Grey dashed lines indicate the location of preserved terraces at the end of
 331 the experiment. Instabilities form mid-channel, developing into a single knickpoint step that
 332 propagates upstream. (1): Initial alluvial channel. (2): Downstream stripping of alluvial cover.
 333 (3): Cyclic step development. (4): Knickpoint retreat in headward direction. (B): Evolution of
 334 channel width (lines) and rate of vertical elevation change (crosses) through time for the
 335 experiments (expt 1 = black, expt 2 = blue, expt 3 = red) at a cross-section 25% distance
 336 from the channel inlet. Width is plotted using a 10 point moving average and elevation
 337 change is plotted using a 5 point moving average. Dashed vertical lines indicates the timing
 338 of the perturbation in experiments 1 and 2 at 37 mins (gray), and the timing of the presence
 339 of a knickpoint (colored lines). After the perturbation, there was a reduction in channel width
 340 during both experiments 1 and 2. Focusing of the flow into a single thread occurred due to
 341 the gradual tilting in experiment 3. There was a short period of channel widening between
 342 150 and 175 min associated with the erosion of the sediment deposited at the inlet during the
 343 initial stages, unrestricting the flow and inundating a larger area and an apparent increase in
 344 the channel width. The highest incision rates are associated with (i) the rapid downstream
 345 stripping of sediment and (ii) the headward migration of a knickpoint through bedrock.



346

347 Figure 3. Conceptual model showing the stages of supercritical terrace formation. (A)
 348 Downstream stripping of sediment cover following perturbation to Q_s/Q_{sc} . (B) Under

349 *supercritical flow conditions, the channel bed destabilizes into series of cyclic steps. (C)*
350 *Cyclic steps develop into a 'supercritical knickpoint', which migrates in an upstream direction*
351 *abandoning a 'supercritical river terrace'. Channel narrows at knickpoint, and widens*
352 *downstream. (D) Conceptual representation of the elevation of the channel bed and*
353 *corresponding terraces following response shown in (A-C). Arrows on terraces indicate the*
354 *direction of the incision wave that abandoned them. The upstream part of T1 is abandoned*
355 *first by the downstream migrating incision wave, and the downstream part of T2 is*
356 *abandoned first by the upstream migrating knickpoint. Vertical incision rates calculated from*
357 *point terraces ages (1–4 and 2–2) provide misleading spatial variation in vertical incision, as*
358 *the majority of the incision actually occurs through the passage of the supercritical*
359 *knickpoint.*

360

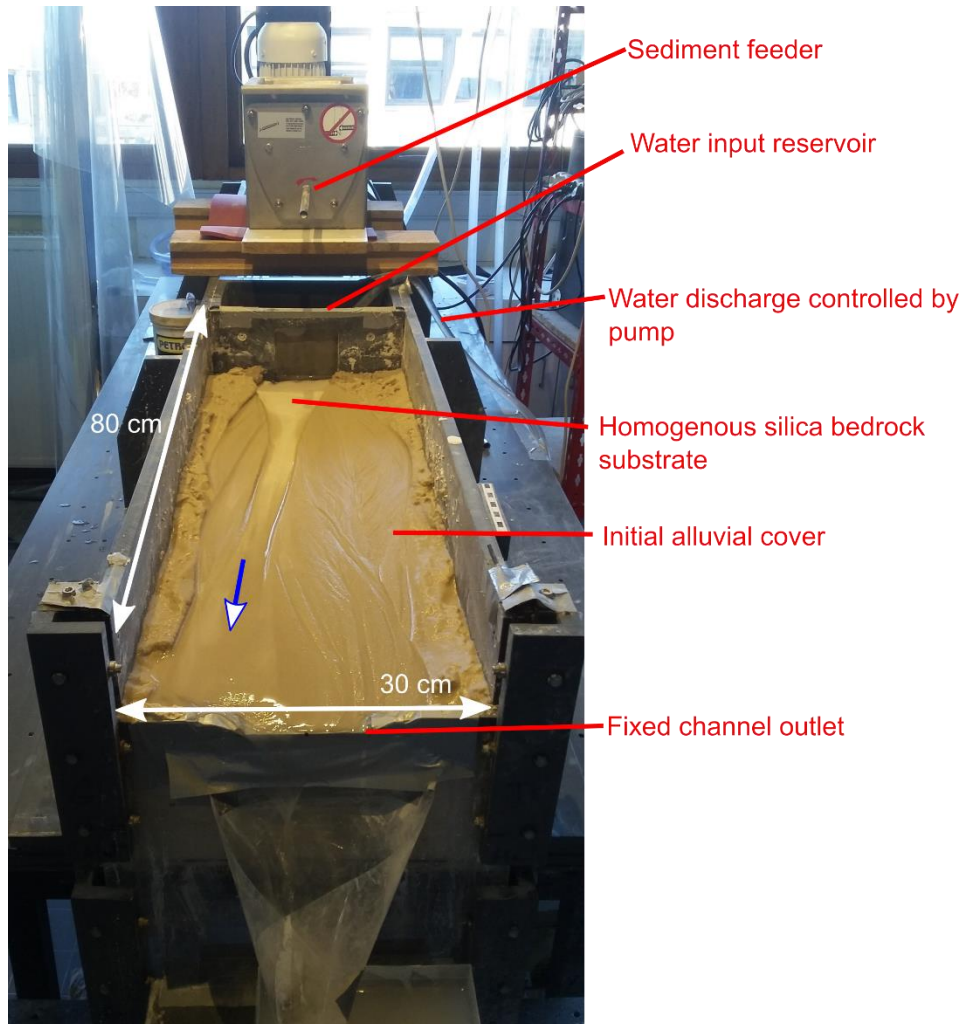
361 **Supplementary Information**

362 **SI Section 1: Relevance of experiments for natural channels and experimental setup.**

363 Experimental modelling studies of landscape and fluvial dynamics are typically performed
364 using one of two approaches. The first directly scales the experimental prototype to the
365 natural system by ensuring that non-dimensional hydraulic parameters such as the Froude
366 number are scaled strictly and the physical processes (e.g., sediment transport) are the
367 same in both the experimental and natural systems. The second approach relaxes this strict
368 scaling, and seeks a 'similarity of process' between the experiment and the natural channel
369 (Hooke, 1968) whereby the behaviour of the processes within the experiment system are
370 qualitatively similar and 'reasonably effective' at replicating the behaviour of natural systems
371 despite large differences in the governing dimensionless numbers (Paola et al., 2009). This
372 second approach has the benefit of allowing temporal and spatial timescales to be relaxed,
373 as materials such as silica cement that is eroded by clear water flow can be used to replicate
374 bedrock, and experiments exploring bedrock erosion processes can be performed relatively
375 quickly. It should be noted that the results from these experiments cannot be directly scaled
376 up to any particular natural system. Additionally, the experimental channel is always 'active',
377 and the experiments do not include the periods of time between flood events when the
378 channel is not being eroded, creating a difficulty when exploring relative timescales of
379 geomorphic response between the experiments and natural rivers. However, processes such
380 as flow acceleration above the knickpoint lip, cyclic step development, plunge pool erosion,
381 undercutting of the knickpoint face and channel banks, cantilever failure, erosion and
382 transport of cohesive material by hydraulic shear are present in the experiments. The typical
383 width-depth ratio of the experimental channels at the start of the experiments (alluvial
384 channel conditions) is 50-80, similar to that of natural gravel-bedded rivers (Finnegan et al.,
385 2005), and the scaling of the channel width with discharge observed in natural rivers is also
386 reproduced in the experiment. Therefore, we can be confident that the experiments are
387 'reasonably effective' in replicating natural processes, and the response of the channels to
388 the perturbations (sediment supply, discharge or slope) in the experiments can help inform
389 the response of natural channels to similar forcing.

390 Figure S1 shows the setup used during our experiments in the Bedrock River Experimental
391 Incision Tank at the Université de Rennes. The initial alluvial cover was composed of well-
392 sorted sand (grain size = 250 μm), and the homogenous cohesive silica paste used as a
393 bedrock simulant was composed of 61.5% angular silica, 20.5% spherical beads and 18%
394 water. The silica paste was mixed in a cement mixer before being transferred into the box
395 flume and homogenised using a high frequency vibrating needle to re-liquefy the paste (see

396 Baynes et al., 2018 for further details related to the properties of the silica paste and the
397 preparation procedure). Flow in the channel is typically laminar with low Reynolds numbers
398 (< 1000), and is input into the channel through a reservoir and the discharge magnitude
399 controlled using a pump.

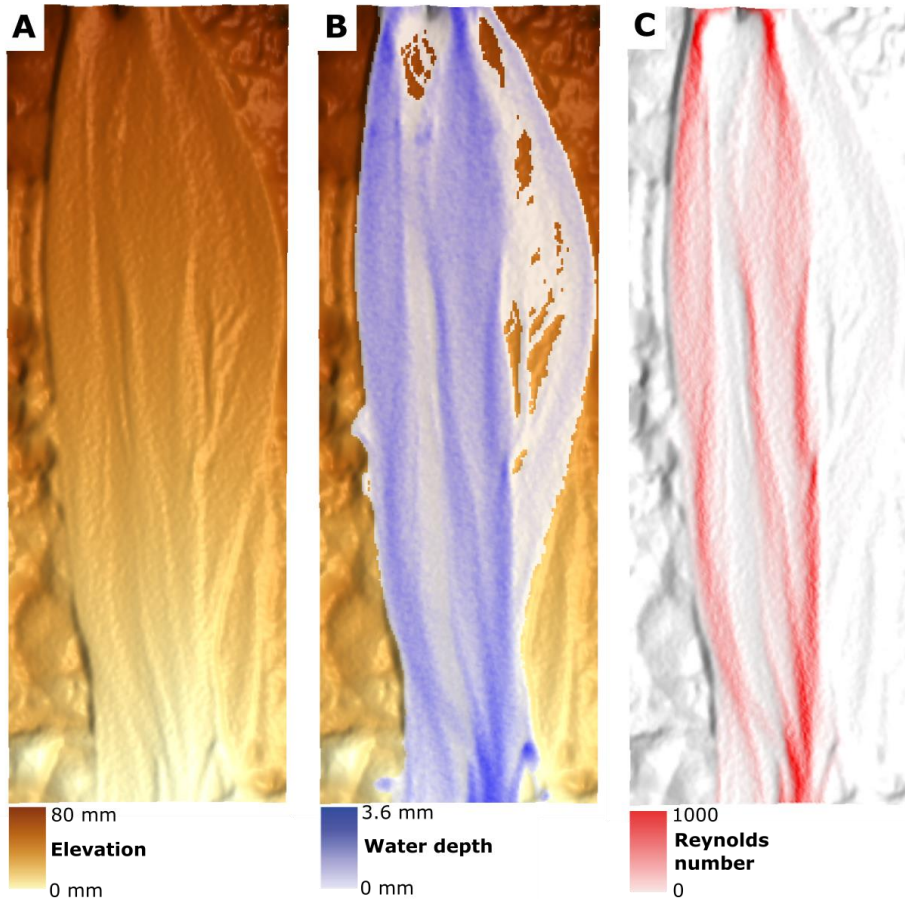


400

401 **Fig. S1:** Photograph of the Bedrock River Experimental Incision Tank at Université de
402 Rennes. The blue arrow shows the flow direction within the channel.

403 **SI Section 2: Floodos parameters**

404 *Floodos* is a numerical precipitation-based hydrodynamic model (see Davy et al., 2017 for full
405 technical description of the model), and was used here to assess the hydraulic conditions
406 during the experiments. The input topography for *Floodos* was produced directly by gridding
407 the point clouds, collected from the terrestrial laser scanner at regular time intervals during
408 the experiments, with a pixel size of 2 mm. *Floodos* was run using the laminar flow option
409 consistent with the relatively low Reynolds number of the experiment (< 1000) across most of
410 the channel under alluvial conditions (Fig. S2).



411

412 **Fig. S2:** (A) Digital Elevation Model from Experiment 1, when the channel is under fully
 413 alluvial conditions ($t = 33$ minutes). (B) Modelled water depth mask using Floodos
 414 hydrodynamic model (Davy et al., 2017). (C) Calculated Reynolds number for each pixel of
 415 the flow, using the Floodos model output.

416

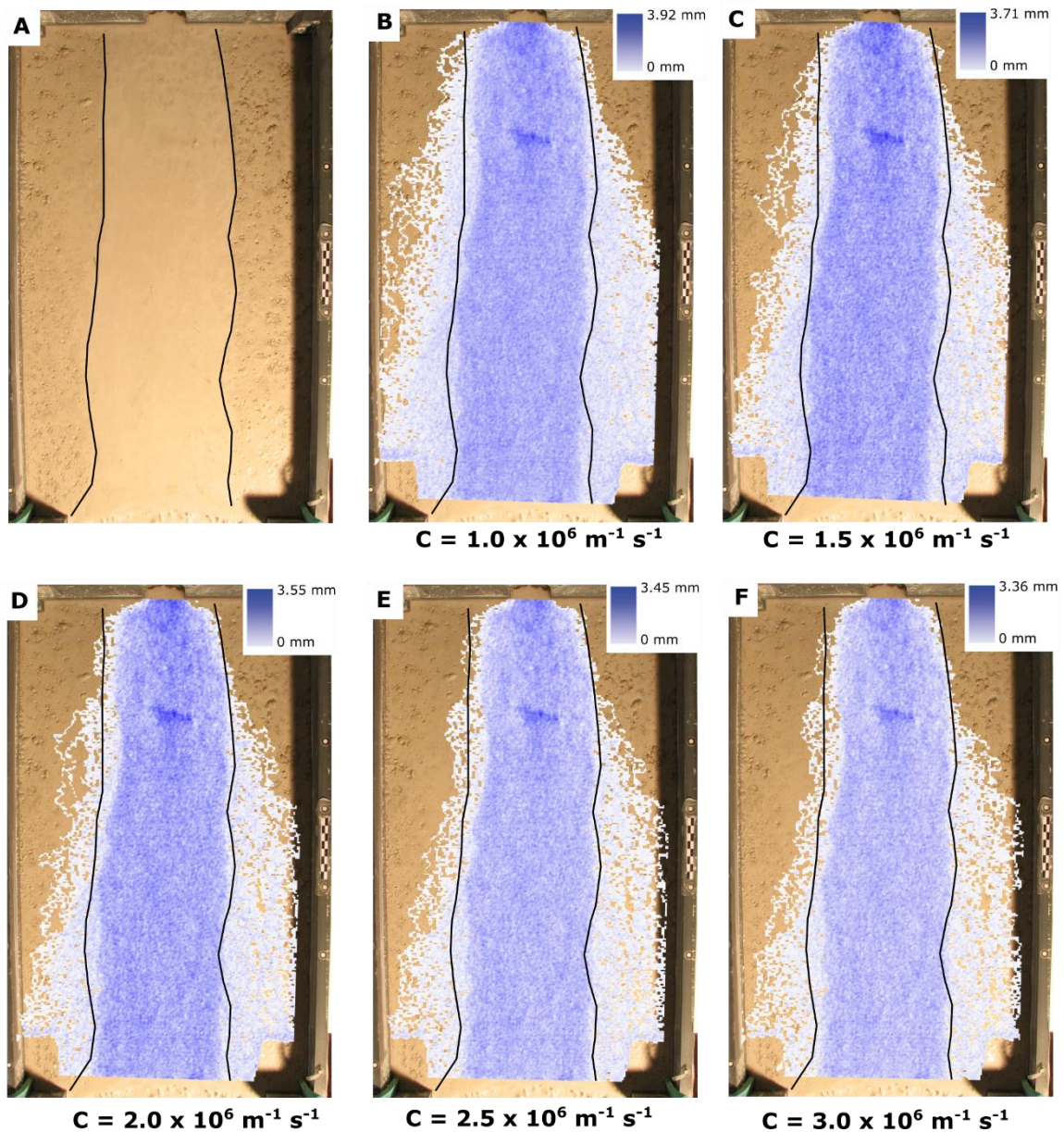
417 Under the laminar flow conditions, the local vertically averaged velocity U is approximated
 418 by:

419
$$U = CH^2S, \quad (S1)$$

420 where C is a friction coefficient in $[L^{-1}.T^{-1}]$, H is local water depth and S is the local channel
 421 slope. Equation S1 is obtained by integrating vertically a laminar flow velocity profile
 422 assuming steady-uniform flow. In pure laminar flow, C is entirely set by the water viscosity μ
 423 and should be approximately $C \sim \rho g / 3\mu$ where ρ is water density and g is gravitational
 424 acceleration. This predicts that at 10°C , $C \sim 2.5 \times 10^6 \text{ m}^{-1}.\text{s}^{-1}$. Fig. S2 shows the impact of
 425 different values of C on the predicted extent of water flowing over a flat surface compared to
 426 manual measurements of the wetted area obtained from a photograph taken from above the
 427 experiment (Fig. S3A). The simulations of flow extent are relatively similar for all modelled

428 values of C , although lower values of C predict a greater level of overflow on the edges in the
 429 downstream part. We therefore used the value of C ($2.5 \times 10^6 \text{ m}^{-1} \cdot \text{s}^{-1}$) for the simulations
 430 presented here, due to good consistency with the manual measurements and the theoretical
 431 prediction of the friction coefficient based on the water viscosity for flow over a flat surface
 432 (Fig. S3) and for channelised flow (Fig. S4).

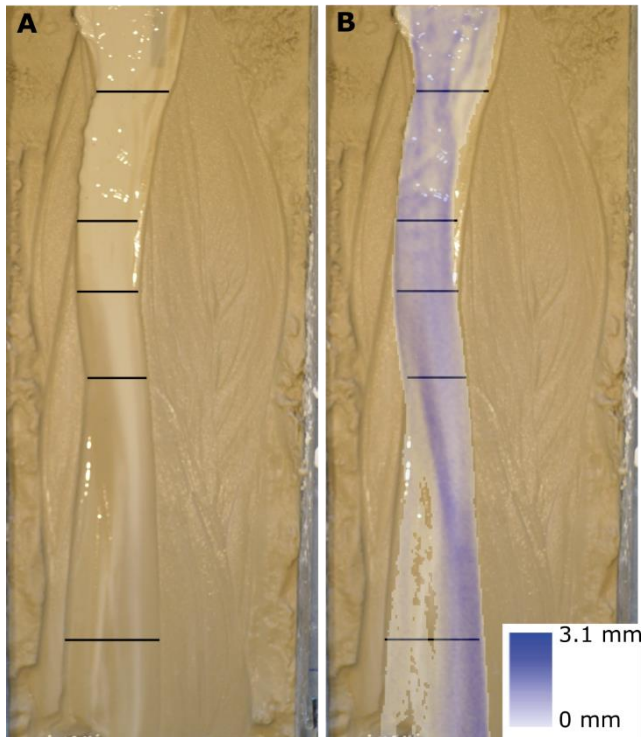
433



434

435 **Fig. S3:** (A) Photograph of water flowing over a flat surface (slope = 3.3° , $Q = 1 \text{ l/min}$), with
 436 black lines indicating edge of water extent measured manually from the image. Flow is from
 437 top to bottom in the image. (B-F) Floodos hydrodynamic model output using laminar flow
 438 option and different values of the friction coefficient. Black lines are the same as (A) and
 439 shown for comparison of modelled against observed flow extent. The predicted theoretical

440 value of the friction coefficient ($2.5 \times 10^6 \text{ m}^{-1} \text{ s}^{-1}$) has a good match with the observed flow
441 extent. For this value, the lateral areas where overflow occurs have flow depth that are
442 extremely low ($< 0.1 \text{ mm}$) and represent a negligible fraction of the total discharge.



443

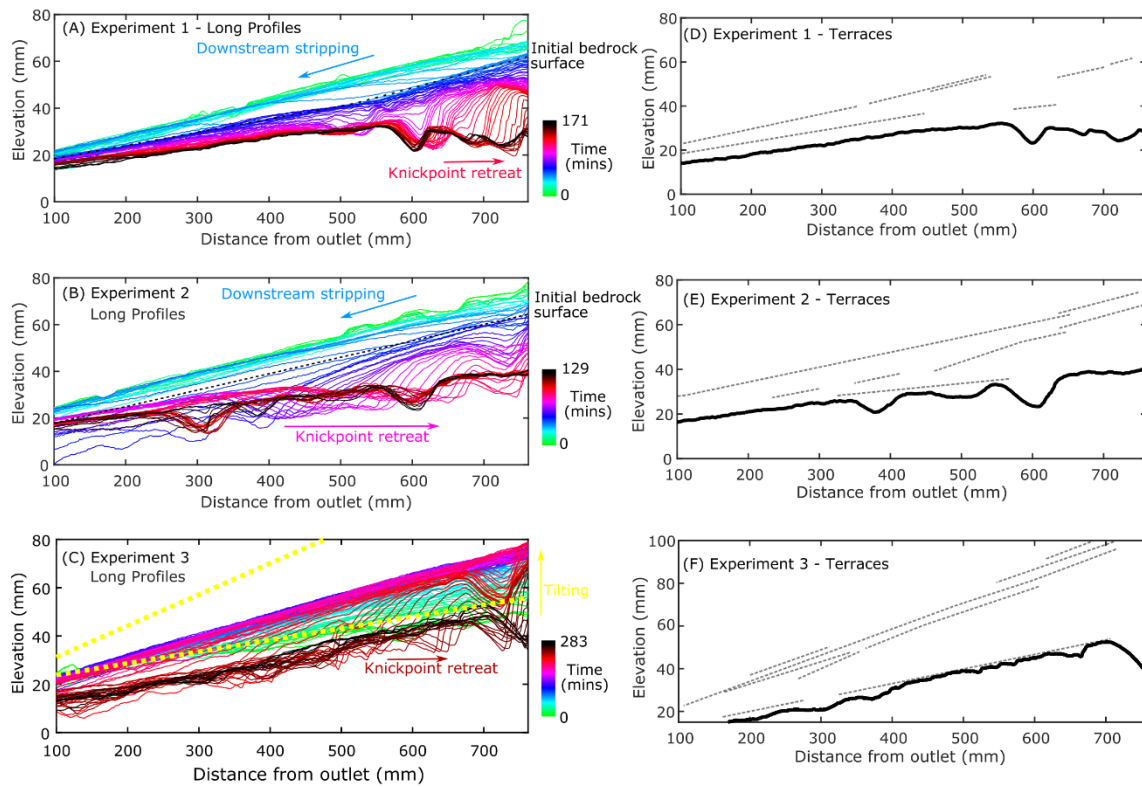
444 **Fig. S4:** (A) Photograph of experiment 1 ($t = 66 \text{ mins}$), with wetted area indicated by the
445 black bars. Flow is from top to bottom in the image. (B) Floodos model water mask coloured
446 by water depth, using friction coefficient of $2.5 \times 10^6 \text{ m}^{-1} \text{ s}^{-1}$, for the same time period overlain
447 over the photograph, with the same black bars from (A). The Floodos water mask closely
448 matches the manual measurements of channel width from the photograph throughout the
449 length of the channel. At the left hand side in the downstream reach, a terrace is starting to
450 be abandoned, hence the very low flow depths in this region.

451

452 **SI Section 3: Evolution of channel and terrace geometry during experiments**

453 This SI section provides the complete experimental data for the evolution of the channel
454 geometry for each of the three experiments. The long profiles were extracted by finding the
455 point of minimum elevation for each row of the grid (Fig. S5). Terraces were defined from the
456 DEM as flat surfaces (elevation within 1mm) and at least 2 cm wide. The channel cross-
457 sections (Fig. S6) were extracted from the data at the location of the grid that was 25% of the
458 distance from the inlet to the outlet of the channel.

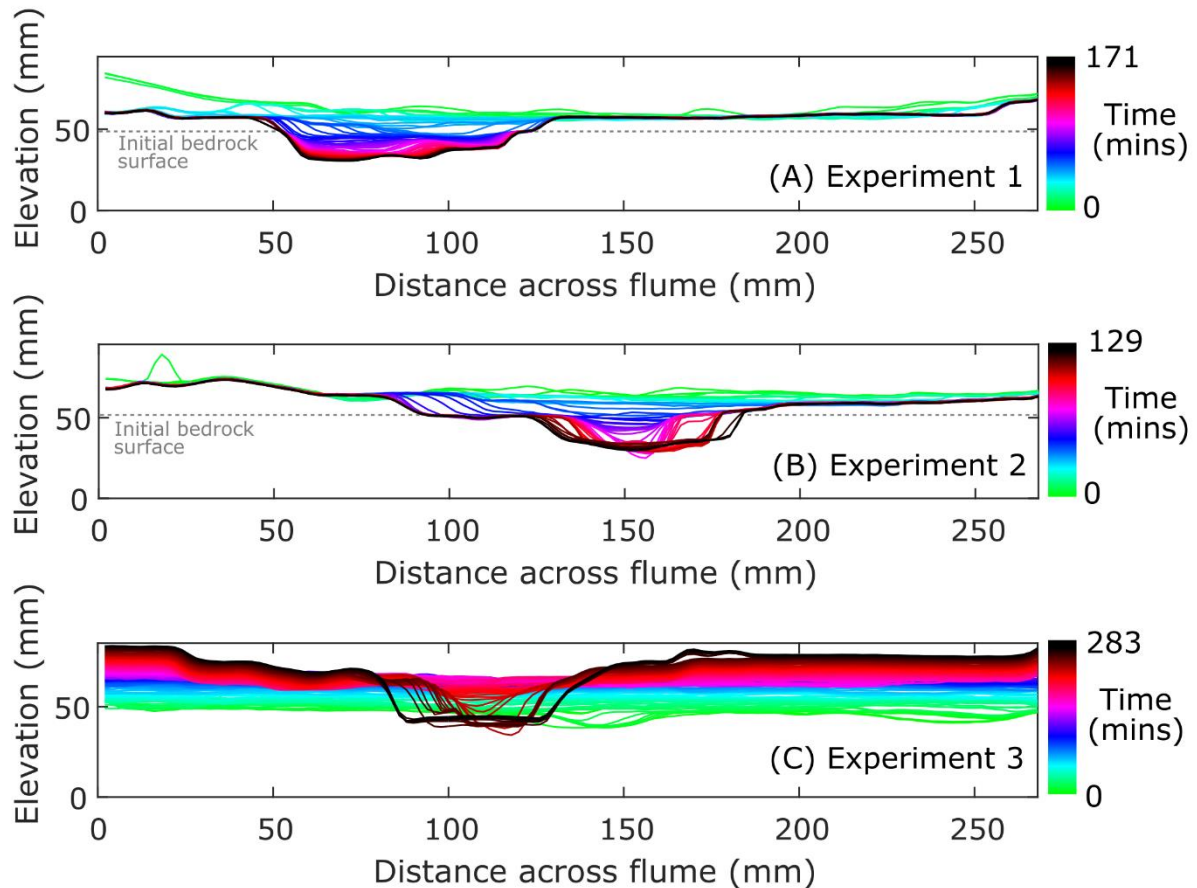
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460

461 **Fig. S5:** (A-C) Extracted long profiles from the three experiments. (A) Experiment 1, where
 462 the input sediment flux was stopped after 37 minutes (80 profiles in total). (B) Experiment 2,
 463 where the discharge was doubled to 3 l/min after 37 minutes (61 profiles in total). (C)
 464 Experiment 3, where the slope was increased at 1° per 60 minutes from an initial slope of 2°
 465 for 280 minutes (132 profiles in total). The colour of the lines indicates the evolution during
 466 the experiments. The dashed yellow lines in (C) show the initial slope of the experiment
 467 (2.5°), and the projected slope of the channel at the end of the tilting if there had been no
 468 erosion of the alluvial cover or bedrock (7.2°). The maximum slope achieved in the channel
 469 before the onset of the incision of bedrock was $\sim 4.5^\circ$. Grey dashed lines in (A) and (B)
 470 indicate the elevation of the bedrock surface at the start of each experiment, to show the
 471 transition from alluvial stripping and bedrock incision. The same line is not shown in (C) as
 472 the elevation of the bedrock surface changed during the course of the experiment due to the
 473 tilting. (D-F) Preserved terrace surfaces at the end of each experiment. Solid lines indicate
 474 the long profile of the active channel, and dashed lines indicate the location and elevation of
 475 terraces. The slope of the upper terraces is greater than the incised active channel.

476



477

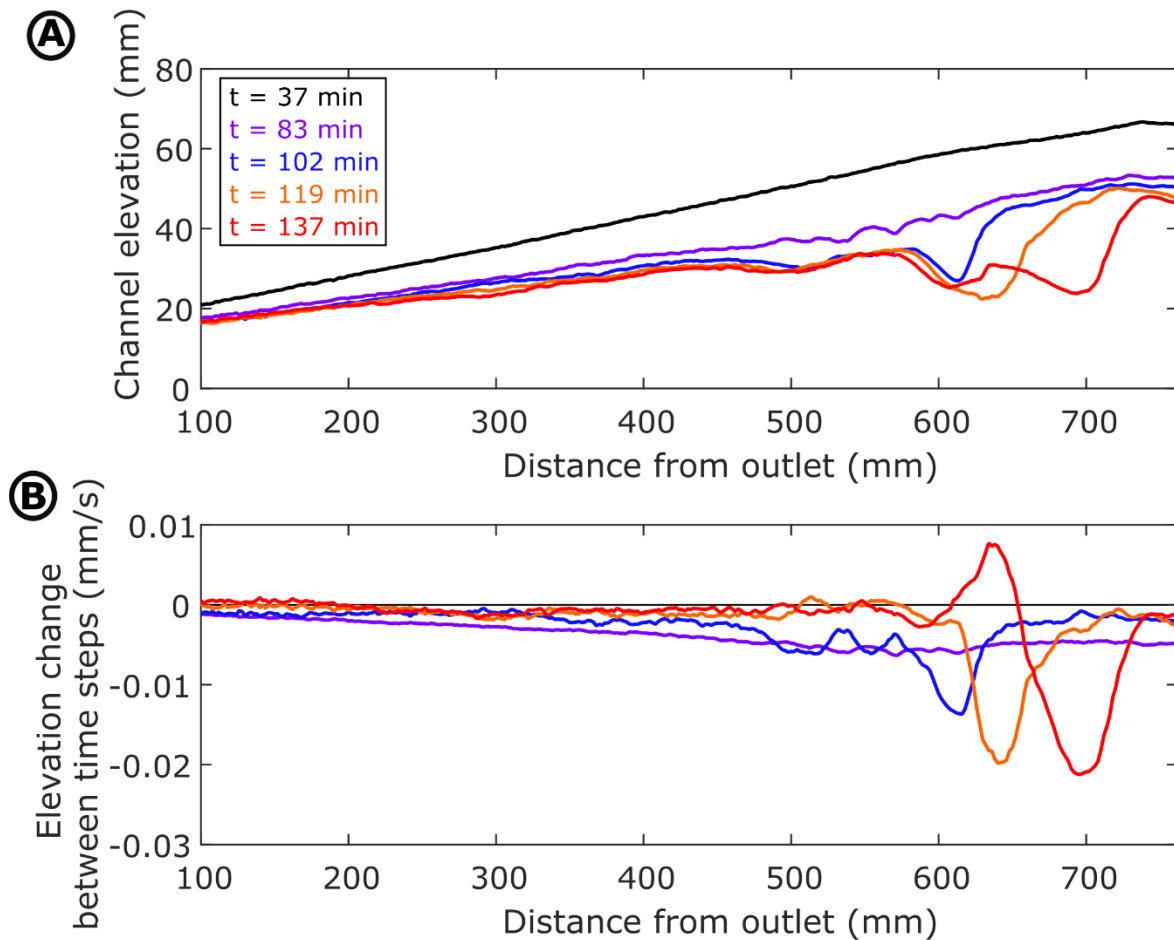
478 **Fig. S6:** Evolution of channel cross-sections at the 25% position from the water inlet. Cross-
 479 sections were extracted from this location to ensure it was located over an area that
 480 experienced incision and the development and propagation of a knickpoint. Colours
 481 represent the same time points as Fig. S1. It can be seen that channel incision occurred
 482 under narrow channel width conditions, followed by lateral erosion of the banks leading to
 483 channel widening and minor channel aggradation in experiments 2 and 3. Grey dashed lines
 484 in (A) and (B) indicate the elevation of the bedrock surface at the start of each experiment, to
 485 show the transition from alluvial stripping and bedrock incision. The same line is not shown in
 486 (C) as the elevation of the bedrock surface changed during the course of the experiment due
 487 to the tilting.

488

489 **SI Section 4: Demonstration of the incisional efficiency of the headward erosion**
 490 **process and width-incision coupling**

491 Here, we provide further analysis to demonstrate the bedrock erosional efficiency of the
 492 headward migration of knickpoints compared to the vertical erosion by hydraulic shear
 493 upstream of the knickpoint lip (Fig. S7). Fig. S7A shows selected long profiles from
 494 experiment 1 and Fig. S8B shows the rate of elevation change along the profiles between the

495 time steps. The largest rates of bedrock elevation change correspond to the locations and
496 time periods when the knickpoint is migrating upstream.



497

498 **Fig. S7:** (A) Selected long profiles showing alluvial channel just before sediment supply
499 stopped ($t = 37$ min; black), channel containing cyclic steps, after sediment has been stripped
500 away from the bedrock profile ($t = 83$ min; purple) and then three long profiles from different
501 stages of knickpoint retreat ($t = 102$, 119, 137 min; blue, orange and red, respectively). (B)
502 The variation in rate of elevation change with distance along the channel between the
503 different time steps. The consistently high rate of elevation change from $t = 37$ minutes to $t =$
504 83 minutes (purple line) corresponds to the time period when the alluvial cover was being
505 stripped downstream.

506

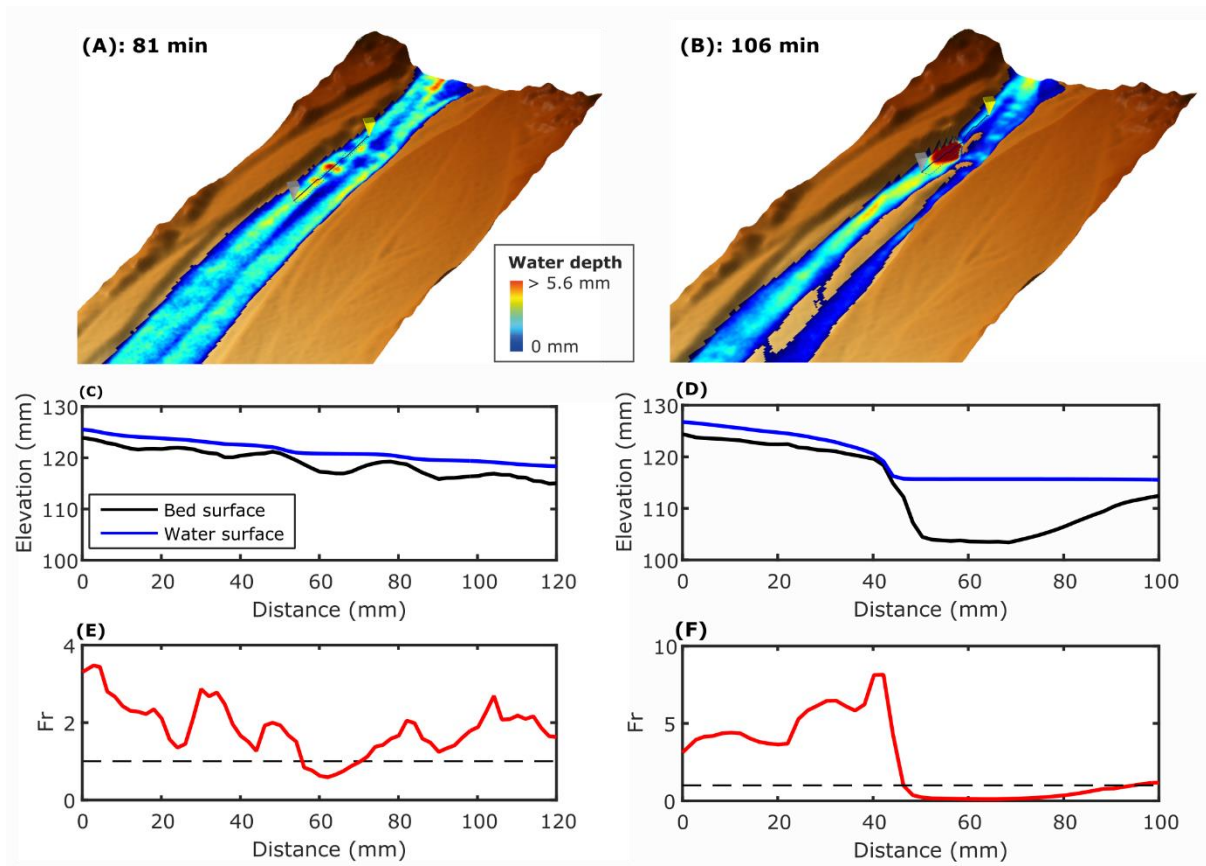
507 **SI Section 5:**

508 **Conditions for cyclic step and knickpoint development**

509 'Supercritical knickpoints' were generated during the experiments in the absence of any base
510 level fall, after the development of instabilities in the channel bed possibly associated with
511 supercritical flow conditions, when the Froude number ($Fr = U/(gH)^{0.5}$), is > 1 (Fig. S9; Parker
512 and Izumi, 2000). In this section, we used the Floodos hydrodynamic model output to
513 calculate the Froude number for all locations within the channel (Fig. S8) for selected times
514 of experiment 1 ($t = 81$ min, when bed instabilities were developing; Fig. S8A, and $t = 106$
515 min, when a single knickpoint had formed; Fig. S8B).

516 Cyclic steps form under supercritical flow conditions due to an interaction between the
517 channel bed and the flow above it (Izumi et al., 2017), developing a series of steps with
518 transcritical flow that alternates between supercritical and subcritical flow with hydraulic
519 jumps (Fig. S8A). After a single knickpoint formed, a pronounced change in the Froude
520 number and a hydraulic jump is present at the knickpoint lip, with subcritical flow conditions
521 present in the plunge pool downstream (Fig. S8B). The absolute values of Fr are possibly
522 slightly overestimated due to the use of the laminar flow option in *Floodos*, and some
523 uncertainty related to whether the flow is purely laminar or transitional (Fig. S2). However,
524 the pattern of Fr variation is consistent with cyclic step development as observed and
525 predicted theoretically under supercritical flow conditions (Parker and Izumi, 2000; Izumi et
526 al., 2017).

527



528

529 **Fig. S8:** Exploration of flow conditions during the development of the self-formed knickpoints.

530 (A-B) Topography during the experiments at selected time steps, with the colours indicating

531 the water depth. (C-D) Profiles for the bed and water surfaces for the reach along the black

532 lines in (A) and (B). (E-F) Extracted Froude number along the same reach as the (C) and (D).

533

534 Supercritical flow in natural rivers

535 Here, we performed a simple test to consider whether supercritical flow conditions can be
 536 sustained in natural rivers under steady, uniform flow, by exploring the impact of varying
 537 channel width and discharge on the Froude number, assuming a rectangular channel with a
 538 constant slope.

539 The flow velocity can be estimated using the Manning's equation ($U = n^{-1} \cdot H^{2/3} \cdot S^{0.5}$), where n
 540 is the Manning's roughness coefficient, H is the flow depth, and S is the channel slope. The
 541 flow velocity is also given by $U = Q / WH$, where Q is the discharge and W is the channel
 542 width. Substituting these two equations into each other gives the following equation for the
 543 flow depth:

$$544 \quad H = \left(n \frac{Q}{W} \right)^{\frac{3}{5}} S^{-\frac{3}{10}} \quad (S2)$$

545 The Froude number is calculated from the flow velocity, the acceleration due to gravity (g)
 546 and the flow depth: ($Fr = U/(gH)^{0.5}$). Substituting equation S2 into this equation gives the
 547 following relationship between the Froude number and the discharge, slope, width and
 548 Manning's roughness coefficient:

$$549 \quad Fr = Q^{\frac{1}{10}} S^{\frac{9}{20}} W^{-\frac{1}{10}} n^{-\frac{9}{10}} g^{-\frac{1}{2}} \quad (S3)$$

550 Eq (S3) shows that in steady uniform flow conditions, the Froude number is mostly sensitive
 551 to slope and roughness, with a weak dependency on width and discharge.

552 According to Parker et al., (2007), the channel width and channel slope of a gravel bed river
 553 at bankfull discharge conditions can be calculated using the following equations:

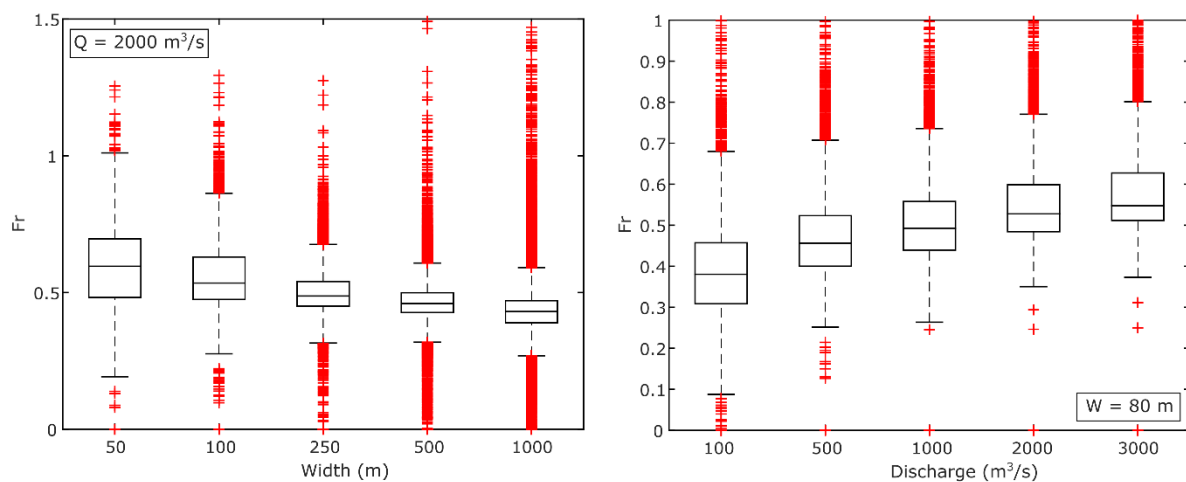
$$554 \quad W_{bf} = \frac{4.63}{g^{\frac{1}{5}}} Q_{bf}^{0.4} \left(\frac{Q_{bf}}{\sqrt{gD_{50} D_{50}^2}} \right)^{0.0667} \quad (S4)$$

$$555 \quad S_{bf} = 0.101 \left(\frac{Q_{bf}}{\sqrt{gD_{50} D_{50}^2}} \right)^{-0.344} \quad (S5)$$

556 Where Q_{bf} is the bankfull discharge, W_{bf} is the channel width at bankfull conditions, S_{bf} is the
 557 channel slope at bankfull conditions and D_{50} is the median grain size. Using these
 558 relationships, we calculated the Froude number for bankfull conditions at a range of
 559 discharges in an idealized gravel bed river, assuming a median grain size of 10 cm and two
 560 values of the Manning's roughness coefficient. The Froude number at equilibrium conditions
 561 using the relationships proposed by Parker et al., (2007) are subcritical, typically in the range
 562 0.2-0.3, whatever the value of the bankfull discharge. This shows that in steady uniform flow
 563 conditions, single thread gravel bed rivers are not expected to be close to supercriticality,
 564 and would not do so even after significant narrowing due to the weak dependency of Fr on
 565 channel width (eq. S3). The Parker et al., (2007) relationships (equations S4, S5) are
 566 empirical based on a dataset of 72 gravel bed rivers in the UK, Canada, and the USA, which
 567 exhibit a degree of universality between them, but do not pertain to braided rivers which are
 568 very often the state of rivers during terrace aggradation phases (e.g., Hanson et al., 2006).
 569 Using high resolution lidar topography and the *Floodos* hydrodynamic model, we explored
 570 the characteristics of the flow conditions at bankfull conditions for a braided river, the Rakaia
 571 river in New Zealand. The present day conditions of the Rakaia are a good analogue both for
 572 the initial conditions of the experiments and for the former conditions of many present day
 573 rivers that are incised with flights of terraces abandoned on the valley sides (e.g., Charwell
 574 River, New Zealand; Bull, 1990).

575 The supercritical flow conditions that initiated the development of cyclic steps in the
 576 experiments (Fig. S9) occurred after channels had narrowed following the initial stripping of

577 the sediment cover. Therefore, we performed some simple tests using the DEM of the
 578 Rakaia river to explore the impact of different constant channel widths with a fixed discharge
 579 on the Froude number (Fig. S9A), and the impact of different discharges in a constrained
 580 fixed width channel (Fig. S9B). For a fixed discharge, a narrower channel increases the
 581 average value of the Froude number and for a fixed width, increased discharge also
 582 increases the average values of the Froude number.



583

584 **Fig. S9:** (A) Impact of channel width on Froude number for a fixed discharge ($Q = 2000 \text{ m}^3 \text{ s}^{-1}$). (B) Impact of discharge on Froude number for a fixed width ($W = 80 \text{ m}$). As predicted by
 585 equation S3, both width and discharge have a small impact on the average Froude number
 586 in the channel, but the flow can become supercritical in a few locations.
 587

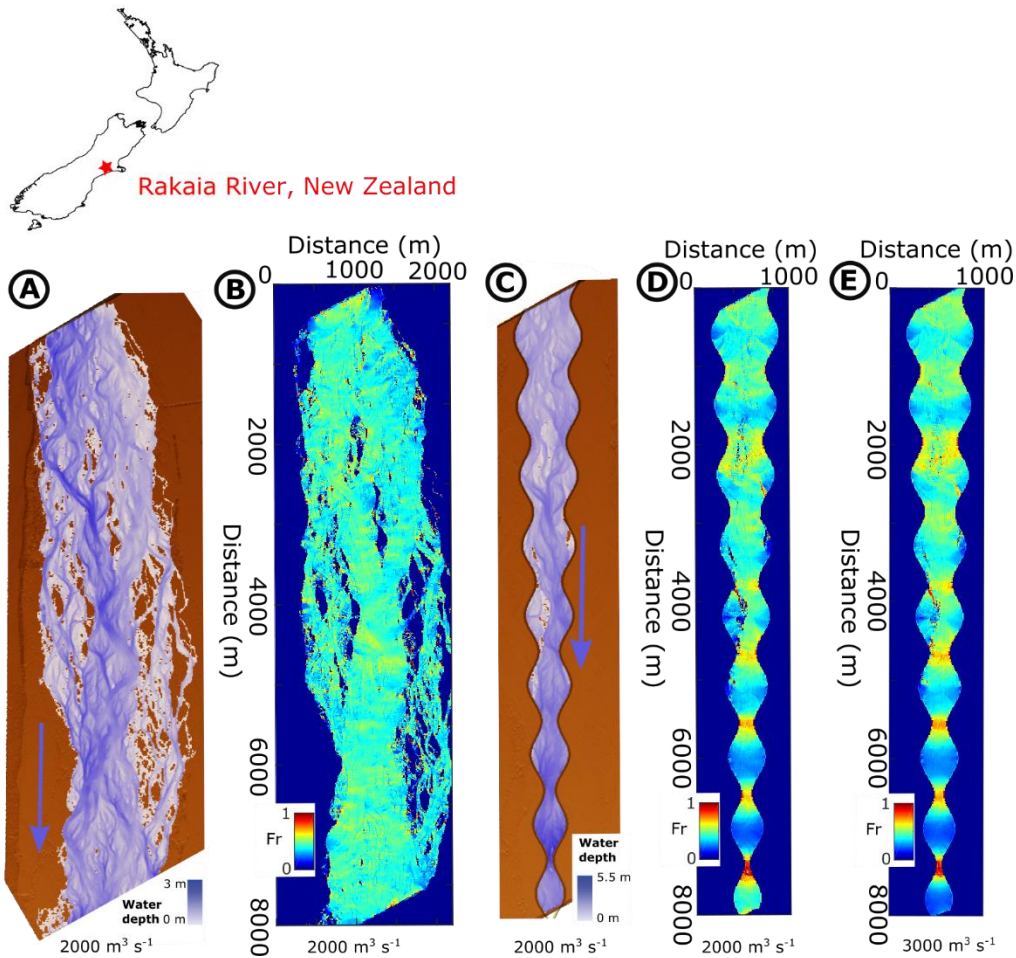
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589 The average values of the Froude number remain sub-critical for the scenarios presented in
 590 Fig. S9, but the peak values of Fr approach supercriticality due to the local spatial variations
 591 in the flow conditions. The numerical simulations in Fig. S11 were carried out using a straight
 592 channel, which is overly simplistic for a natural river system. The maps of Froude number for
 593 the Rakaia (Fig. S10B) river indicate that flow conditions are not spatially uniform at bankfull
 594 discharge and are subcritical ($Fr < 1$) but the values of the Froude number can be higher
 595 ($\sim 0.4-0.6$) than the values predicted by the relationships of Parker et al., (2007). However, it
 596 is possible to have supercritical flow conditions during frequent floods when the flow is
 597 focussed into a narrower channel with a spatially non-uniform channel width (Fig. S10C-E).
 598 Such conditions may be present in natural settings following a perturbation to the sediment
 599 supply or transport capacity and the subsequent stripping of sediment cover and focussing of
 600 the flow into a significantly narrower channel with spatially variable width (Fig. 2A). The
 601 simulations in Fig. S10 were carried out with no alteration to the channel slope. If the slope of
 602 the bedrock surface is steeper than the slope of the alluvial channel flowing over the

603 sedimentary deposits (e.g. Finnegan and Balco, 2013), the likelihood of supercritical flow and
 604 the possible development of bed instabilities that evolve into knickpoints is increased. We
 605 therefore suggest that the processes leading to rapid strath terrace generation through self-
 606 formed knickpoint retreat, in the absence of any vertical perturbation to the bed elevation,
 607 identified in the experiments are possible in natural environments.

608

609



610

611 **Fig. S10.** (A) Map of water depth for a reach of the Rakaia river, New Zealand, under
 612 bankfull conditions ($Q = 2000 \text{ m}^3 \text{ s}^{-1}$), modelled using Floodos on a Lidar generated DEM
 613 (Released under Creative Commons By Creative Commons Attribution 3.0 New Zealand
 614 Link: <http://data.linz.govt.nz/license/attribution-3-0-new-zealand/>; downloaded from
 615 www.opentopography.org; DOI: 10.5069/G9JQ0XZV). (B) Map of Froude number, typically
 616 ~ 0.4 , with some local areas slightly higher at ~ 0.6 . Channel width variability and degree of
 617 channel constriction has impact on the water depth (C) and Froude number at $Q = 2000 \text{ m}^3 \text{ s}^{-1}$
 618 (D) and $Q = 3000 \text{ m}^3 \text{ s}^{-1}$ (E), with increasing degree of supercriticality with increased
 619 channel constriction and increased discharge.

620 **SI References**

621 Baynes E.R.C., Lague D., Attal M., Gangloff A., Kirstein L.A., Dugmore A.J., 2018, River self-
622 organisation inhibits discharge control on waterfall migration: Scientific Reports 8, 2444,
623 <https://doi:10.1038/s41598-018-20767-6>

624 Davy P, Croissant T, Lague D (2017) A precipitation method to calculate river hydrodynamics,
625 with applications to flood prediction, landscape evolution models and braiding
626 instabilities. *Journal of Geophysical Research. Earth Surface*, v. 122, p. 1491–1512,
627 <https://doi.org/10.1002/2016JF004156>

628 Finnegan N.J., Roe G., Montgomery D.R., Hallet B., 2005, Controls on the channel width of
629 rivers: Implications for modelling fluvial incision of bedrock: *Geology*, v. 33 (3), p. 229-
630 232

631 Hanson PR, Mason JA, Goble RJ (2006) Fluvial terrace formation along Wyoming's Laramie
632 Range as a response to increased late Pleistocene flood magnitudes. *Geomorphology*
633 76, 12-25

634 Hooke, R. 1968, Model Geology: Prototype and Laboratory Streams: Discussion: *Geological*
635 *Society of America Bulletin*, v. 79, p. 391–94, [https://doi.org/10.1130/0016-](https://doi.org/10.1130/0016-7606(1968)79[391:MGPALS]2.0.CO;2)
636 [7606\(1968\)79\[391:MGPALS\]2.0.CO;2](https://doi.org/10.1130/0016-7606(1968)79[391:MGPALS]2.0.CO;2)

637 Izumi N, Yokokawa M, Parker G (2017) Incisional cyclic steps of permanent form in mixed
638 bedrock-alluvial rivers. *Journal of Geophysical Research. Earth Surface*, v. 122, p. 130–
639 152, <https://doi.org/10.1002/2016JF003847>.

640 Paola, C, K M Straub, D C Mohrig, and L Reinhardt. 2009. “The “unreasonable
641 Effectiveness” of Stratigraphic and Geomorphic Experiments.” *Earth-Science Reviews* v.
642 97, p. 1–43.

643 Parker G, Izumi N (2000) Purely erosional cyclic and solitary steps created by flow over a
644 cohesive bed. *Journal of Fluid Mechanics*, v. 419, p. 203–238,
645 <https://doi.org/10.1017/S0022112000001403>.

646 Parker G, Wilcock PR, Paola C, Dietrich WE, Pitlick J (2007) Physical basis for quasi-
647 universal relations describing bankfull hydraulic geometry of single-thread gravel bed
648 rivers. *Journal of Geophysical Research*, v. 112, F04005.

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