



**HAL**  
open science

## Drainage network evolution and patterns of sedimentation in an experimental wedge

Marc Viaplana-Muzas, Julien Babault, Stéphane Dominguez, Jean van den Driessche, Xavier Legrand

► **To cite this version:**

Marc Viaplana-Muzas, Julien Babault, Stéphane Dominguez, Jean van den Driessche, Xavier Legrand. Drainage network evolution and patterns of sedimentation in an experimental wedge. *Tectonophysics*, 2015, 664, pp.109-124. 10.1016/j.tecto.2015.09.007 . insu-01252482

**HAL Id: insu-01252482**

**<https://insu.hal.science/insu-01252482>**

Submitted on 8 Feb 2024

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

# 1 Drainage network evolution and patterns of sedimentation 2 in an experimental wedge

3  
4 Marc Viaplana-Muzas<sup>a</sup>, Julien Babault<sup>a</sup>, Stéphane Dominguez<sup>b</sup>, Jean Van Den Driessche<sup>c</sup>,  
5 Xavier Legrand<sup>d</sup>

6 <sup>a</sup> Departament de Geologia, Universitat Autònoma de Barcelona, 08193 Bellaterra (Barcelona),  
7 Spain.

8 [marc.via.mu@gmail.com](mailto:marc.via.mu@gmail.com), [julien.babault@uab.es](mailto:julien.babault@uab.es)

9 <sup>b</sup> Géosciences Montpellier, Université Montpellier II, F-34095, France.

10 [dominguez@gm.univ-montp2.fr](mailto:dominguez@gm.univ-montp2.fr)

11 <sup>c</sup> Géosciences Rennes, Université de Rennes 1, Campus de Beaulieu, Rennes, France.

12 [jean.van-den-driessche@univ-rennes1.fr](mailto:jean.van-den-driessche@univ-rennes1.fr)

13 <sup>d</sup> Petronas CariGali, Twin Tower KLCC, 50088, Kuala Lumpur, Malaysia.

14 [legrand.xavier@petronas.com.my](mailto:legrand.xavier@petronas.com.my)

## 15 Abstract

16 *In fold and thrust belts drainage organization and patterns of sedimentation depend conceptually on the ability*  
17 *or not for preexisting reaches to incise uplifting thrust sheets. In this study we investigate experimentally the*  
18 *dynamics of drainage network in a wedge submitted to shortening and erosion. It allows us to reproduce and to*  
19 *monitor the interactions between tectonics, erosion and sedimentation during the development of up to five successive*  
20 *thrust sheets. In the experiments channels adjust to uplift rate by both increasing their slope and narrowing their*  
21 *channels as it is observed in nature. The series of experiments shows that the proportion of persistent preexisting*  
22 *transverse channels increases with the ratio of rainfall over shortening rates. The experiments confirm the view that*  
23 *the competition between discharge and tectonic uplift controls along-strike variations in sediment flux in sedimentary*  
24 *basins by controlling the drainage organization. If the transverse channels draining a wedge are not diverted, a line-*  
25 *source dispersal system develops in front of the active structure. If channels are diverted in the backlimb of the*  
26 *frontal structure it results in point-sourced depositional systems separated by areas fed only by small channels*  
27 *developing in the front of the wedge. Fans accumulated in front of the active structures reveal two stages of*  
28 *sedimentation, one of progradation, while the frontal structure is active and a second one of valley backfilling and*  
29 *thrust sealing during internal deformation of the wedge. The experiments also suggest that spatial variations in rock*  
30 *uplift rate along a thrust front may be evidenced by minimum-discharge variations of persistent transverse channels.*

31 Keywords: experimental modeling, accretionary wedge, drainage network organization, river diversion,  
32 sedimentation patterns, source-to-sink.

## 33 **1 Introduction**

34 The drainage network at the Earth's surface exerts a first order control on the relief  
35 dynamics and the erosion of mountain belts. Beyond its role in shaping the topography, the  
36 drainage network is a main controlling factor of the coupling between surface processes and  
37 deep crustal deformation, as well as of the relations between tectonics and climate variations,  
38 and the sedimentary record in basins. During mountain building, the drainage network  
39 development involves river diversions and captures at different scales (e.g. Babault et al., 2012;  
40 Castelltort et al., 2012; Giletycz et al., 2015) resulting in rapid changes in sediment routing and  
41 spatial distribution of erosion. Indeed, in response to folding or thrusting, transverse rivers are  
42 commonly diverted into longitudinal reaches, parallel to the structures, that gather into greater  
43 rivers, which maintain gorges through growing structures (e.g. Burbank et al., 1999; Burbank  
44 and Vergés, 1994; Jackson et al., 1996; Jolley et al., 1990; Oberlander, 1985; Talling et al.,  
45 1995). By controlling the spacing between outlets, drainage diversion may eventually control  
46 the loci of sediment supply and the stratigraphic architecture in foreland-basins (Gupta, 1997;  
47 Horton and DeCelles, 2001; Tucker and Slingerland, 1996). Conceptually, the ability or not for  
48 preexisting reaches to incise uplifting structures controls the number of diversions, and by  
49 extension drainage organization, a view confirmed by numerical models (Champel et al., 2002;  
50 Humphrey and Konrad, 2000; Koons, 1994; Koons, 1995; Sobel et al., 2003; Tomkin and  
51 Braun, 1999; van der Beek et al., 2002). In particular, numerical models, where deformation is  
52 reduced to uplift and advection over a thrust, show that the proportion of persistent preexisting  
53 transverse channels scales with the ratio of precipitation over shortening rate (Champel et al.,  
54 2002; Tomkin and Braun, 1999). Alternatively, it has been proposed that aggradation in the  
55 backlimb of emergent thrusts also helps transverse rivers to balance uplift rates allowing them  
56 to maintain their course instead of being diverted (Humphrey and Konrad, 2000), or that axial  
57 slopes, controlled by the dip of the décollement layer where the thrusts are rooted, may divert  
58 preexisting transverse channels before they reach the uplifts (Champel et al., 2002; van der Beek  
59 et al., 2002).

60 In this study we investigate experimentally how the interaction between drainage network  
61 and deformation controls along-strike variations in sediment flux in wedges submitted to  
62 shortening and erosion. We first investigate the similarities of behavior (geometries and  
63 kinematics) between experimental and natural channels evolving under uplifting conditions. We  
64 examine the differences in both the drainage organization and along-strike variations in  
65 sediment accumulations as a function of the ratio of rainfall rate over shortening rate. We  
66 determine the factors that control (1) the capacity for channels to incise uplifting structures and  
67 (2) the patterns of sedimentation that resulted from the drainage organization.

68

## 69 **2 Method**

### 70 **2.1 Setup**

71 The experimental set-up used in this study is adapted from the setup used by Graveleau and  
72 Dominguez, (2008) and Graveleau et al., (2011). The deformation device dimensions are 80 x  
73 150 cm and are constituted by a basal film pulled beneath a static buttress. The film is overlaid  
74 by the analogue material that models the upper part of the crust. Shortening induces material  
75 deformation and generates an accretionary wedge composed of imbricated thrusts. The rainfall  
76 system is composed by sprinklers that deliver water micro-droplets over the model. Sprinklers  
77 deliver water micro-droplets in sequences of 10 seconds with rain and 3 seconds without rain in  
78 order to improve channel incision processes. During the 3 seconds without rain, slope erosion  
79 processes are mostly inhibited whereas in the river network, water, collected by the channel  
80 catchments during the 10 seconds rain phase, continues to flow for a while (a few seconds up to  
81 tens of seconds depending on the length of the considered channel). During the dry time period,  
82 channel incision is then enhanced generating a more incised topography and favoring fluvial  
83 and alluvial terrace formation. This protocol was implemented to limit channel widening  
84 induced by the high river flow dynamics. Droplet size is small enough (diameter  $\leq 100 \mu\text{m}$ ) to  
85 reduce rain-splash effect and potential surface craterization. Rainfall in the experiment allows  
86 water runoff to generate both diffusive erosion processes on hillslopes and incision/lateral  
87 erosion in channels but it does not intend to simulate real water droplets (Graveleau et al., 2015).  
88 Spatial variation of rainfall rates due to air convection induced by water supply from sprinklers  
89 have been measured to be minor ( $<10\%$  on average).

90 The analogue material is composed of three different materials: glass microbeads, silica  
91 powder and plastic powder (PVC). To obtain experiments with thrusts spaced enough to avoid  
92 the burying of thrust backlimbs by fans, we adjust the composition of material mixtures, the  
93 thickness of the material and layering of different materials. As mentioned above, aggradation  
94 in the backlimb of an uplifting thrust helps transverse rivers to balance uplift rates. Two different  
95 analogue material mixtures were derived and deposited in two layers in the sand-box. The upper  
96 layer is the material submitted to erosion and it is made up with 46% of glass microbeads, 30%  
97 of silica powder and 24% of plastic powder (PVC) plus some graphite ( $<1\%$ ) necessary for  
98 photo-correlations (material IV). This mixture is slightly different to the material IV used in  
99 Graveleau et al. (2008), Graveleau et al. (2011) and Strack et al., (2011). The lower layer is made  
100 of 50% of glass microbeads and 50% of PVC. We performed several tests and we found  
101 empirically that a total thickness of 55 mm made by a basal layer of 5 mm of glass microbeads  
102 (d collement layer), overlaid by 45 mm of the analog materials plus a thin layer (5 mm) of glass  
103 microbeads within the upper mixture results in 14-cm-spaced thrust sheets (Fig. 1). The thin  
104 layer of glass microbeads within the upper mixture allows slip to occur within the material IV

105 resulting in folding at the surface above the ramps, as in nature. The deformation style  
106 reproduces well an accretionary wedge pattern made of individualized in-sequence thrust faults  
107 dipping toward the buttress. The material IV submitted to erosion has enough cohesion for  
108 valleys and crests to develop but not too high for basal shear stress applied by the fluid on the  
109 riverbed to exceed the thresholds for detachment and transport.

110 Digital Elevation Models (DEM) are acquired by an optical measurement bench composed  
111 by high resolution cameras coupled to a laser interferometer. This device acquires digital  
112 topographies at a 3D resolution close to 0.2 mm. It requires stops in both shortening and rainfall  
113 systems during 30-45 minutes to dry the uppermost 1-2 mm of the model surface and avoid  
114 bright laser points that could affect DEM resolution. Photograph cameras allow us to document  
115 the dynamical evolution of the relief by video movies and image correlation analyses (Graveleau  
116 and Dominguez, 2008). Finally, the model is cut in serial cross-sections in order to study the 3D  
117 geometry of thrusts and syntectonic deposits.

118 In this work we present eight experiments run under shortening rates ranging from 4 cm/h  
119 to 100 cm/h, of which six were submitted to a precipitation rate of 9 mm/h (named A1 to A6)  
120 and the other two to a precipitation rate of 18 mm/h (named B1 and B2).

121  
122  
123  
124

## 125 **2.2 Analysis of the accretionary wedges**

126 Most of the deformation in the experimental wedges occurs by frontal accretion due to  
127 forward propagation of the thrust sequence as observed classically in experiments (e.g. Davis et  
128 al., 1983). Two domains can be differentiated in the experimental wedges. The prowedge,  
129 located in the frontal part of the wedge, and formed by forethrusts and the retrowedge located in  
130 the rear of the wedge and formed by backthrusts. Prowedge slope is ruled by the Coulomb  
131 wedge theory (Davis et al., 1983) and depends on the internal friction of the material, the basal  
132 friction and the dip angle of the décollement level (flat in our experiments). The spacing  
133 between thrusts and the dip of the thrusts also depends on the basal friction and on the total  
134 thickness of the layer but also on the mechanical resistance (cohesion) of the layer. We  
135 measured prowedge slopes, spacing between thrusts in the external part of the prowedge, when  
136 a new thrust appears, and we measured the dip of the thrusts to check if the rheology of the bulk  
137 material was equivalent in all the experiments.

138

## 139 2.3 Scaling

140 Experimental modeling is ruled by a set of geometric, kinematic and dynamic similarity  
141 criteria that imply, respectively, proportionality of length, time and forces between nature and  
142 model(Hubbert, 1951; Paola et al., 2009). Applying these similarity criteria to the Mohr-  
143 Coulomb equation and neglecting inertial forces, leads to the expression:

$$144 C_o^* = \rho^* \cdot L^* \quad (1)$$

145 where  $C_o^*$ ,  $\rho^*$  and  $L^*$  are the ratios between model and nature for cohesion, density and length,  
146 respectively. We calculated the cohesion of the erodible material using a modified Hubbert-type  
147 set-up, following the methodology described in Graveleau et al. (2011). This device measures  
148 frictional properties of water-saturated granular materials. It consists of two aluminum rings, 8  
149 cm in internal diameter and 50 cm<sup>2</sup> in horizontal cross-sectional area, which contain the  
150 granular material sample. The lower ring, 3 cm deep, is fixed on a mobile plateau controlled by  
151 a stepping motor. The upper ring is fixed on a rigid pendulum that rests on a force gauge. Its  
152 internal surface is covered with a low friction material (Teflon) to reduce sidewall friction.  
153 Between both rings there is a gap of 0.1 mm filled with low viscosity grease to prevent water  
154 and material from escaping. When the lower ring moves, material inside the ring is sheared and  
155 the upper ring presses the force gauge. We performed several tests with increasing normal loads  
156 to measure normal and tangential stress couples at failure and to deduce a yield locus.  
157 Displacement and force gauge data are recorded and allow for the generation of a Mohr-  
158 Coulomb failure envelope. Best fit to the data gives a slope equal to the coefficient of internal  
159 friction ( $\mu$ ) and an extrapolated cohesion ( $C_o$ ). Results for our mixture, material IV (matIV), for  
160 stable friction, are  $\mu = 0.67$  ( $\phi_{\text{matIV}} = 35^\circ$ ) and cohesion of  $C_o = 610$  Pa for stabilized water  
161 saturation of 21-22 % (Fig.2).Cohesion in natural rocks ranges between  $10^7$  Pa for sandstone or  
162 granite and  $10^8$  Pa for marble and limestone(Schellart, 2000), hence the cohesion ratio  $C_o^*$   
163 ranges between  $6.1 \cdot 10^{-5}$ -  $6.1 \cdot 10^{-6}$ . Considering the bulk density of matIV ( $\rho_{\text{matIV}}=1600$  kg.m<sup>-3</sup>)  
164 and natural rocks (between 2000 kg.m<sup>-3</sup> for sandstone and 2800 kg.m<sup>-3</sup> for granite), the density  
165 ratio  $\rho^*$  ranges between 0.57 and 0.80. According to equation (1) $L^*$  is  $1 \cdot 10^{-5}$ -  $8 \cdot 10^{-6}$ , that is, 1  
166 cm in the experiments equals ~0.1 km to ~1km in nature.

167

## 168 2.4 Analysis of the digital topographies

### 169 2.4.1 Quantification of the stream power erosion law in the experiments

170 We searched for relationships between metrics of the topography in the framework of the  
171 stream power theory for understanding and quantifying the landscape evolution and the drainage  
172 evolution in response to external perturbations.We analyzed the relationship between the  
173 steepness indices (a metric based on slope-area plots), the relative uplift rates as defined in

174 Babault et al. (2005), the upstream velocity of propagation of knickpoints, the channel widths  
 175 and the drainage areas of steady-state reaches identified within the channels that cross uplifting  
 176 thrust sheets. The stream power theory assumes that fluvial erosion is a power law function of  
 177 the drainage area,  $A$ , and the channel slope,  $S$ , (Howard and Kerby, 1983). Stream power  
 178 erosion laws have the form:

$$179 \frac{dz}{dt} = U - f(Q_s) \cdot K \cdot \frac{A^m}{W} \cdot S^n \quad (2)$$

180 where the rate of change of elevation,  $z$ , with time,  $t$ , depends on the imposed uplift rate,  $U$ , and  
 181 on a power law function of drainage area,  $A$  (as a proxy for catchment discharge), channel  
 182 width,  $W$ , and downstream channel slope,  $S$ . The exponents  $m$  and  $n$  are positive,  $f(Q_s)$  describes  
 183 the tool and cover effects of sediment supply and it is often taken to be equal to 1. In the  
 184 experiments, uplifts induce waves of incision and knickpoints that propagate upstream  
 185 indicating conditions close to the detachment-limited model of incision (Cowie et al., 2008;  
 186 Whipple and Tucker, 2002). We consider that there is no effect of  $Q_s$  on erosion efficiency in the  
 187 experiments and  $f(Q_s)$  equals 1.  $K$  is a parameter describing erosional efficiency. Channel  
 188 width ( $W$ ), may be described as a power law function of  $A$  (and is often taken to scale with  
 189  $\sqrt{A}$ ) (Montgomery and Gran, 2001; Schellart, 2000; Whittaker et al., 2007a), in which case its  
 190 effect can be subsumed into exponent  $m$  and coefficient  $K$  (Attal et al., 2008; Whipple and  
 191 Tucker, 1999). However, it has been shown in a few settings that in response to an increment of  
 192 the uplift rate, incision rate can be increased by the narrowing of channels in order to increase  
 193 the stream power (e.g. Amos and Burbank, 2007; Finnegan et al., 2005; Whittaker and Boulton,  
 194 2012; Whittaker et al., 2007a, 2007b).

195 At steady-state, by definition the channel erosion rate is equal to the uplift rate of rock ( $E =$   
 196  $U$ ), and the steady-state channel gradient ( $S_e$ ) can be found by solving equation (2) for slope  
 197 under this condition (Howard, 1980):

$$198 S_e = \left(\frac{WU}{K}\right)^{\frac{1}{n}} \cdot A^{-\frac{m}{n}} \quad (3)$$

199 The  $m/n$  ratio is predicted to depend only on the rate of increase of water discharge and channel  
 200 width with drainage area (Whipple and Tucker, 1999), with a typical value of  $\sim 1/2$  in nature. The  
 201 exponents  $m$  and  $n$  reflect the mechanics of the dominant incision process, or combination of  
 202 processes (Whipple, 2004). Equation (3) has the same form of Flint's law that relates the channel  
 203 slope,  $S$ , to drainage area,  $A$ , with the normalized channel steepness index,  $k_{sn}$ , and concavity,  
 204  $\theta$  (we used a reference concavity equal to 0.5 in this study):

$$205 S = k_{sn} \cdot A^{-\theta}, \quad (4)$$

206 Steady-state channels adjust to increasing uplift rate by increasing their slope (normalized  
 207 channel steepness index,  $k_{sn}$ ) (Bookhagen and Strecker, 2012; DiBiase et al., 2010;  
 208 Gudmundsdottir et al., 2013; Kirby and Ouimet, 2011; Kirby and Whipple, 2012; Ouimet et al.,  
 209 2009; Perron and Royden, 2013; Wobus et al., 2005) and/or by narrowing their width, which

210 concentrates runoff and increases the stream power(Amos and Burbank, 2007; Finnegan et al.,  
 211 2005; Whittaker et al., 2007a). The stream power erosion law predicts at steady-state that the  
 212 relation between the erosion/uplift rate,the normalized steepness index and channel width  
 213 depends on parameter  $n$ :

$$214 \quad U = \frac{K}{W} \cdot k_{sn}^n \quad (5)$$

215 If channel width scales with drainage areas then equation (5) can be rewritten as:

$$216 \quad U = K \cdot k_{sn}^n \quad (6)$$

217 We calculated the steepness indices and we measured the channel width of the steady-state  
 218 reaches to understandthe way channels adjust to uplift rates. $K$  is constant inthe series of  
 219 experiments run under the same rainfall rate. We estimated the parameter  $n$  using equation (5).

220 The family of stream power models also predicts equation (2) takes the form of a nonlinear  
 221 kinematic wave of erosion in response to external perturbation(Rosenbloom and Anderson,  
 222 1994; Royden et al., 2000; Whipple, 2001; Whipple and Tucker, 1999) and can be rewritten as:

$$223 \quad \frac{\partial z}{\partial t} = -\frac{K}{W} \cdot A^m \cdot S^{n-1} \left| \frac{\partial h}{\partial x} \right|, \quad \frac{\partial h}{\partial x} < 0 \quad (7)$$

224

225 where

$$226 \quad C = \Psi_K \cdot A^m \cdot S^{n-1}, \text{ with } \Psi_K = \frac{K}{W} \quad (8)$$

227  $C$  is the wave celerity which has been used to represent knickpoint retreat rate(e.g. Berlin and  
 228 Anderson, 2007; Bishop et al., 2005; Crosby and Whipple, 2006; Loget and Van Den Driessche,  
 229 2009; Parker and Anderson, 1977; Tucker and Whipple, 2002; Whipple and Tucker, 2002;  
 230 Whittaker et al., 2008; Whittaker and Boulton, 2012; Wobus et al., 2006a; Wobus et al.,  
 231 2006b)and $\Psi_K$  embeds lithological and width effects. Field evidence shows that knickpoint  
 232 migration rates scalewith drainage area(e.g. Berlin and Anderson, 2007; Bishop et al., 2005;  
 233 Crosby and Whipple, 2006; Loget and Van Den Driessche, 2009; Parker and Anderson,  
 234 1977)and equation (8) reduces to:

$$235 \quad C = \Psi_A \cdot A^m \quad (9)$$

236  $m$  is shown to be equal to 0.5 and  $\Psi_A$  is a coefficient of retreat efficiency ( $L^{(1-2h)}T^{-1}$ ). This  
 237 relationship implies that the upstream drainage area is the predominant parameter on knickpoint  
 238 migration, with regard to lithology(e.g. Berlin and Anderson, 2007; Bishop and Cowell, 1997;  
 239 Bishop et al., 2005; Loget and Van Den Driessche, 2009). We first estimated the parameter  $m$   
 240 using equation (9), we also studied the width effect on parameter  $m$ using a simplified version of  
 241 equation (8) relating thevariables  $C$ ,  $A$  and  $W$ :

$$242 \quad C = \Psi_{AW} \cdot \frac{A^m}{W} \quad (10)$$

243 where  $\Psi_{AW}$  is the product  $K \cdot S^{n-1}$  with a dimension depending on  $m$ .

244 The width ( $W$ ) and the distance of knickpoint propagation used to calculate  $C$  have been  
245 measured using the GridVisual program (P. Davy, University of Rennes 1,  
246 <http://terrasse.geosciences.univ-rennes1.fr/?p=1>). In all the experiments we observed that the  
247 ramp dips vary at depth, with a constant dip from the trace of the thrust up to a distance of  $\leq 8$ -  
248 10 cm. Beyond this distance the ramp flattens before rooting to the décollement layer of glass  
249 microbeads. In order to study the controls on  $W$ ,  $k_{sn}$  and knickpoints propagation velocities ( $C$ ) at  
250 a given uplift rate, we restricted the study of knickpoints to the first 8-10 cm from the trace of  
251 the thrust where uplift rates can be considered spatially uniform. Mean channel widths,  $W$ , in  
252 uplifting thrust sheets have been calculated from three measures spaced by  $\sim 4$ -5 cm and  
253 located in the hanging wall of the thrust between fan apexes and knickpoints.  $k_{sn}$  of each pixel in  
254 the channels located in the uplifting thrust sheets have been calculated using GridVisual and  
255 they have been averaged for comparison to uplift rates. Knickpoint location corresponds to the  
256 point where channel local slopes increase downstream. The distance over which a knickpoint  
257 has propagated has been measured between the knickpoints and the point where a thrust starts to  
258 emerge. Repeated measures indicate typical error values of 1 mm. This error has been used to  
259 derive an error of the values of the knickpoint propagation velocities ( $C$ ).

260

#### 261 **2.4.2 Quantification of erosion and uplift**

262 To estimate the amounts of erosion and uplift rates in the experiments we subtracted the  
263 eroded topography from reconstructed initial topographies. The initial topographies are derived  
264 by the highest values surrounding a given pixel and measured within a rectangular-moving  
265 window. The width of the window, parallel to the transverse-channel width, is determined by  
266 the widest valleys and it has been set to 80 mm from the experiment A1. The length, parallel to  
267 the transverse channels, has been set to 1 mm to take into account the folding of the initial  
268 topography. The uplift rates we calculated are relative uplift rates as defined in Babault et al.  
269 (2005). We divided the change of elevation above the apex of the fans deposited in front of the  
270 structures by the time span of uplift.

271

#### 272 **2.4.3 Quantification of the proportion of deviations**

273 We calculate the proportion of persistent antecedent channels,  $R$ , defined as the ratio  
274 between the number of transverse streams cutting through the uplifted ridge and the initial  
275 number of streams (Champel et al., 2002):

$$276 \quad R = \frac{N_p}{N_{ch}} \quad (11)$$

277 The variable  $R$  varies between 0 (all rivers deflected) and 1 (all rivers cut through the growing  
278 ridge). The initial number of channels has been obtained by counting the number of incisions on  
279 the surface of a thrust sheet, when it appears. This number may be affected by processes that

280 merge initial channels during the growth of a structure, as valley widening, and by deviations  
281 not controlled by the growing frontal structure but by backthrusts in the backlimb. Fan  
282 aggradation in the backlimb also helps transverse streams to maintain their course instead of  
283 being diverted (Humphrey and Konrad, 2000). The variable  $R$  measured in this study strictly  
284 reflects the processes of deviation by uplift.

285

#### 286 **2.4.4 Quantification of sedimentation rates**

287 By sprinkling tinted analogue material we added time markers on the surface of the fans at  
288 the foot of the thrust sheets. Because these color layers would alter the acquisition of digital  
289 elevation models from the surface of the experiments, they were sprinkled after the  
290 digitalization of the surface. After drying, the experiments are cut and these timelines are used  
291 to calculate the sedimentation rates from the cross-sections of the fans.

292

### 293 **3 Results**

#### 294 **3.1 Style of deformation**

295 We do not observe structural differences between the models. In the series, the mean slopes  
296 of the prowedge ( $\alpha$ ) range between  $8^\circ$  and  $11^\circ$ , thrust dips ( $\phi$ ) range between  $20^\circ$  and  $30^\circ$  with a  
297 mean dip of  $24 \pm 2^\circ$  and the spacing between thrust sheets is  $\sim 14 \pm 2$  cm (Figs. 3C, 3D and 3E).  
298 These results indicate that the rheology of the material (internal friction and basal friction) is  
299 almost constant in all the models and that the pattern of deformation in the series is not  
300 influenced by the two orders of magnitude of shortening rates we applied. Experiment results  
301 also reveal that erosion has no effect on the pattern of deformation under the rates of shortening  
302 and precipitation we applied.

303 When a new thrust appears, it is accompanied by a small amplitude fold located above the  
304 ramp (Fig. 4A). The maximum structural offset on the thrusts is  $\sim 10$  cm, most of it being  
305 completed in the initial stage of activity of a structure in the external part of the wedge. Once a  
306 new structure appears the older ones remain only weakly active. Since thrust dips are almost  
307 constant in the series of models, uplift rate is ruled by shortening rate. The faster the shortening  
308 rates, the faster the uplift rates. In most experiments we observed small thrusts rooted in the  
309 intermediate layer of glass microbeads that break the surface (offset  $\sim 1$  cm). Less often we  
310 observed back thrusts with displacement  $< 1$  cm. These secondary structures have little influence  
311 on the uplift rates of the external structures.

312

313 **3.2 Surface processes**

314 In the experiments, we observe channel incision by runoff, sapping and landslide. As  
315 described in previous experiments (Graveleau et al., 2011), the high rate of precipitation (9 and  
316 18 mm/h) we apply leads to the formation of a permanent sheet flow in the external and flat areas  
317 of the model, the thickness of which ranges between 0.5 mm and 1 mm. When a new structure  
318 emerges in the external part of the wedge the resulting slope makes the water to converge, and  
319 channels to form leading to the onset of incision of the erodible material (Figs. 4A and 4B). The  
320 small channels rapidly grow and they attain 0.5-1 mm in depth and 2-5 mm in width (Fig. 4C).  
321 Ongoing shortening and uplift trigger retreating incision on the structure and a wave of incision  
322 migrates to the back limb of the fold-thrust, i.e. toward the inner part of the wedge (Figs. 4D, 4E  
323 and 4F). Not all of the initial channels (antecedent) succeed to cut in a fold-thrust (see animation  
324 1). We discuss in detail in the next section the controls of such behavior. Qualitatively, we  
325 observe that transverse channels with faster velocities of knickpoint retreat achieve to cross-cut  
326 the uplifting thrust sheets and we refer to them as persistent transverse channels. The number of  
327 these persistent transverse channels decreases when shortening rates increase resulting in  
328 drainage networks dominated by longitudinal reaches (Fig. 5). Lateral erosion in the channels  
329 generates terraces, which is common in such kind of experiments (Graveleau and Dominguez,  
330 2008), and widens the valleys. The widening of these valleys may eventually merge adjacent  
331 channels. The products of erosion are transported over a small distance that is a function of the  
332 nature of the grain. Most of the composite material (Silica Powder, PVC, Glass microbeads)  
333 accumulates at the foot of the thrust sheets, forming centimeter-scale fans by avulsion (Fig. 4D).  
334 A small fraction of the eroded products, the smaller and lighter grains of the mixture (mostly  
335 PVC), covers the flat undeformed part of the models up to 15 cm to 20 cm away from the frontal  
336 structure (Fig. 4A), and another small fraction in suspension leaves the model (<1%).

337 In the headwaters of the channels we observe that drainage expansion is achieved by  
338 sapping of the upper millimeter of the erodible material (Figs. 6A and 6B). These results imply  
339 that erosion by groundwater flow is very limited. Landslides may also occur in the external limb  
340 of the compressive structures, usually in areas preserved from channel incision. A landslide can  
341 start when the slope of the external limb becomes steeper than 20°. Normal fault scarps first  
342 appear, immediately followed by mass movement involving areas of 2×2 cm, as observed in  
343 experiment A4 (Figs. 4E and 4F) and up to 20×5 cm as in experiment A6 (Figs. 6C and 6D). As  
344 expected landslides are more common at high shortening rates when the ratio between uplift and  
345 surface erosion rates reach a limit allowing the topography to attain and exceed its equilibrium  
346 slope.

347

348 The accretion of new structures in the wedge tilts the above thrust sheet toward the inner  
349 part (Fig. 3A), decreasing the local slopes of the topography thus inhibiting erosion. Below a  
350 threshold slope of c.a.  $3^\circ$ , the cohesion of the erodible material strongly decreases the erosion rate.  
351 In all the experiments we observe preserved, non-eroded surfaces. The extent of these surfaces  
352 is higher in the experiments run at a high shortening rate (Figs. 6C and 6D). We also observe that  
353 drainage areas in the prism shrink through time, because shortening is not only accommodated  
354 by the external thrusts, but also by minor thrust reactivations within the wedge that  
355 induce underthrusting. In experiment A6, the distance between thrusts 2 and 3 decreases from 13  
356 cm to 10 cm after 12 cm of shortening (Figs. 4B and 4F). At the scale of the whole prism, the  
357 underthrusting results in a ~20% decrease in drainage areas.

358

### 359 **3.3 Amounts of erosion and erosion rates**

360 For a given precipitation rate (9 mm/h), the mean volumes of channel erosion, achieved  
361 during the activity of the external thrust sheets, show that the amount of erosion in valleys is five  
362 to twelve fold higher in the experiments running under shortening rates between 4 cm/h and 20  
363 cm/h, compared to experiments at higher shortening rates (Fig. 7A). However, volumes  
364 mobilized by landslides are three to four times higher at shortening rates higher than 10  
365 cm/h. These results indicate that in all these experiments, surface mass balance is dominated by  
366 channel erosion and transport at shortening rates equal to or lower than 9 cm/h (Fig. 7A), and it is  
367 dominated by landslides at high shortening rates (A4 and A5).

368 Steady-state is never achieved at the scale of the active thrust sheets and erosion rates are  
369 roughly one eighth of relative uplift rate (Fig. 7B), far from equilibrium as evidenced from the  
370 flat surfaces preserved from erosion and from the continuing increase in elevation of the active  
371 thrust sheets. However, the channels crossing an active thrust sheet as in figure 4D do adapt to  
372 the relative uplift rates. We can see that the wave of incision propagates upstream toward the  
373 inner parts of the uplifting topography, being limited in its upstream part by a knickpoint (Fig.  
374 8). Downstream of a knickpoint the incision rate balances the relative uplift rate, indicating that  
375 this portion of a channel is at steady-state.

376

### 377 **3.4 Scaling erosion law with steady-state reaches**

378 We measured the width of the steady-state channel reaches once the knickpoints have  
379 propagated in the hanging wall of the thrusts, as in figure 8C, in the series of experiments run  
380 under a precipitation rate of 9 mm/h. In this series, relative uplift rates vary from ~0.1 mm/min in  
381 experiment A1 up to ~2.6 mm/min in experiment A6. Uplift rates are balanced by incision rates

382 in the persistent transverse channels. In all the experiments we observe that the relationship  
 383 between  $W$  and  $A$  follows a power law:

$$384 \quad W = W_n A^b, \quad (12)$$

385 where  $W_n$  is a reference width measured as width coefficient following Whittaker and Boulton  
 386 (2012). Best fits are obtained for  $b=0.2$ . Channel widths narrow in response to uplift rates  
 387 predicted by the stream power theory (Equation 5). The channel width coefficient decreases by a  
 388 factor of three over one order of magnitude of uplift rate (Fig. 9).

389 Equation (3) also predicts that channels at steady-state are expected to steepen with  
 390 increasing uplift rate. In particular, the stream power theory predicts that the exponent on the  
 391 slope, parameter  $n$ , controls the nonlinearity between uplift rates,  $U$ , and normalized steepness  
 392 indices,  $k_{sn}$  (Equation 6). Even if there is some dispersion in the data at high uplift rates, we  
 393 observe a positive relation between  $k_{sn}$  and uplift rates, which is non-linear (Fig. 10A) in the  
 394 series of experiments run under a precipitation rate of 9 mm/h:

$$395 \quad U \propto k_{sn}^{p'} \quad (13)$$

396 where best fit of exponent  $p'$  is 2. However, we know that channels also adjust to uplift rate by  
 397 narrowing. Combining the adjustment of channel steepness and width to uplift rate we obtain:

$$398 \quad U \propto \frac{1}{W} \cdot k_{sn}^p \quad (14)$$

399 with exponent  $p=1.5\pm 0.2$  (Fig. 10B). Using both approaches we observe a non-linear relationship  
 400 between channel steepness indices and uplift rates. The combination of a threshold for incision  
 401 and a non-uniform distribution of discharge events could explain a non-linear relationship  
 402 between channel steepness and erosion rate at steady-state (DiBiase and Whipple, 2011; Lague  
 403 et al., 2005; Snyder et al., 2003b; Tucker, 2004). If true in the experiments, it would imply  
 404 that the exponent  $p$  may not be equal to the parameter  $n$  in the erosion law (Equation 2).  
 405 However, in the experiments rainfall rate is constant and runoff distribution events are rather  
 406 uniform, implying that they should not affect erosion rates. On the other hand, downstream of  
 407 the retreating knickpoints, the transverse reaches are at steady-state (Fig. 8C), indicating that the  
 408 threshold for incision is overcome. This suggests that the slope exponent  $n$  is higher than 1 and  
 409 taking into account the width effect we infer  $n = 1.5\pm 0.2$ .

410 As predicted by the stream power theory (Equation 9) we obtain, in the series of  
 411 experiments run under precipitation rates of 9 mm/h and 18 mm/h, a power law relation between  
 412 the velocity of knickpoint retreat,  $C$ , and the drainage area,  $A$  with  $m = 0.4$ . We observe that the  
 413 knickpoint retreat rates are higher at higher uplift rates (Fig. 11A) as observed in nature (e.g.  
 414 Loget and Van Den Driessche, 2009; Weissel and Seidl, 1998; Whittaker and Boulton, 2012)  
 415 and, as predicted by the stream power theory for values of  $n > 1$  (Whipple and Tucker, 1999).  
 416 This finding is consistent with the value of  $n$  recovered from the uplift rate, steepness index and  
 417 width relationship. Taking into account the width effect (Equation 10), we infer  $m = 0.8\pm 0.2$ , a

418 slightly highvalue, in the series A of experiments run under a precipitation rate of 9 mm/h(Fig.  
419 11B).In the experimentB1run under a rainfall rate two times higher (18 mm/h), rates of  
420 knickpoint retreats are higherfor a given drainage area when compared to the experiment A1 run  
421 under 9 mm/h of rainfall. We also observe that smaller catchments managed to incise the uplifts  
422 (Fig. 11A).

423

### 424 **3.5 Patterns of drainage network and sedimentation**

425 During the emergence of a fold-thrust not all the preexisting-transverse channels  
426 (antecedent) (Fig.12A) succeed to cut through an uplifting thrust sheet, some antecedent  
427 channels are defeated (Fig.12B). In the model A1 at low shortening rate, most of the preexisting  
428 channels maintained their course through the emerged thrust sheet and these channels are  
429 referred to as persistent antecedent channels. In the model A2, some antecedent-transverse  
430 channels are defeated, and the discharge they lose is diverted into longitudinal channels located  
431 in the backlimb of the thrust sheets. As a consequence, the downstream part of a defeated  
432 channel that has lostits upstream area eventually drainonly the forelimbs of the active  
433 thrusts(Fig. 12).The channels crossing an active thrust sheet adapt to the uplift rates, and if  
434 knickpoints retreat fast enough through the uplifting structure the antecedent transverse  
435 channels persist.We have shown that knickpoint retreat depends on drainage area or discharge  
436 (Fig.11A). The drainage area of a persistent channel is either initially large enough or it  
437 increases when adjacent diverted channels merge with it.

438 In the experiment A1, that is, at low uplift rates,75% of the antecedent channels manage to  
439 incise the uplifting thrust sheets and the 25% remaining are diverted into longitudinal reaches  
440 developed in the backlimb of the fold-thrusts (Fig. 5). At higherrates of uplift and under the  
441 same precipitation rate,<35% of the antecedent channels are persistent.As expected in the frame  
442 of the above-outlined conceptual model, the proportion of persistent antecedent channels  
443 decreases with shortening rate, i.e., with uplift rate (Fig.13A). We also observe that the  
444 proportion of persistent preexisting transverse channels, R, does not scale linearly with the ratio  
445 of precipitation over shortening rate:

$$446 \quad R \propto \frac{\text{Rainfall rate}}{\text{shortening rate}}^{0.65 \pm 0.2} \quad (15)$$

447 At high R values diversions are sparse and the spacing between outlets of transverse  
448 persistent channels is low. Lower values of R mean that the spacing is higher. By controlling the  
449 spacing between the outlets of the transverse channels, the magnitude of the uplift rates also  
450 control the dispersal of the products of erosion in front of the active thrust. Low spacing  
451 between outlets leads to a line-source dispersal system where fans grow by lateral avulsion and  
452 finally coalesce in a bajada-like piedmont (Fig. 14A).Some fans are fed by larger drainage areas  
453 while most of them are small-scale fans fed by small channels that drain the external limb of the

454 next to last thrust sheet and the active frontal thrust sheet (Fig. 14B). At higher values of  
455 spacing between outlets, a point-sourced depositional system develops and the reduced number  
456 of fans may not merge. Large catchments feed large prograding fans that may locally bury the  
457 backlimb of the active thrust sheet (Figs. 14C and 14D). We also see in figure 14 that the  
458 old sedimentary zones have been incorporated into the prism as wedge-top basins (highlighted in  
459 red in Fig. 14). Although the process of wedge widening implies the incorporation of channels  
460 into larger integrated networks, we do not observe an increase of sediment outflux, which results  
461 from a limitation of the experimental setup. The bulk of erosion in all the experiments comes  
462 mostly from the active external thrust sheet and in minor proportion from the next to last thrust  
463 sheets. Thrust sheets are back tilted when incorporated into the wedge, resulting in the decrease  
464 of channel steepness below the critical slope where erosion is strongly reduced (threshold slope  
465 of c.a. 3°).

466 In the experiment A1, the thickness of sediments accumulated in front of the thrust 5 at  
467  $t=160$  min has reached ~1.5 cm and in cross-section the area covered by sediments is 480  
468  $\text{mm}^2$  (Fig. 15A). In the experiment A4, the area of sedimentation in cross-section reached 460  
469  $\text{mm}^2$  (at the same stage as Fig. 15B), i.e., the volume of sediment accumulated in front of an  
470 active thrust is more or less the same whatever the shortening rate. This is explained by the fact  
471 that sediments shed by persistent channels come mainly from the erosion of the external thrust  
472 sheets during thrust activity and that the geometries of the valleys of the persistent channels are  
473 very similar in all the series of experiments. The volumes of sediment in experiments A1 and  
474 A4 have been accumulated during ~160 min and ~55 min, respectively, leading to  
475 sedimentation rates much higher in the experiment evolving under the higher shortening rate  
476 (Fig. 16).

477 We observe in all the experiments two sequences of aggradation. In the first one the growth  
478 of the prograding fans is evidenced by downlaps while in the second stage a thin layer of  
479 sediments seals the thrust that has become inactive (Fig. 15B). The second sequence is almost  
480 not visible in the cross-section of the experiment A1 (Fig. 15A) because the trace of the cross-  
481 section is slightly oblique to the bottom of the valley where sediments were accumulated.  
482 Backtilting, in the backlimb of the youngest thrust sheets, triggers erosion of the distal parts of  
483 the wedge-top basins, resulting in an unconformity between the first and the second sequence.  
484 Once a basin is incorporated within the wedge deformation eventually folds both sequences and  
485 the thin second sequence may be totally eroded.

486

## 487 4 Discussion

### 488 4.1 Dynamics of the erosion in the experimental models and comparison to 489 numerical models

490 We infer from uplift rates, incision rates and metrics of the topography that erosion in the  
491 persistent transverse channels scales nonlinearly to slope and drainage area (discharge):

$$492 E = K \cdot \frac{A^{0.8 \pm 0.2}}{W} \cdot S^{1.5 \pm 0.2} \quad (16)$$

493 The inferred exponents  $m$  and  $n$  are different to findings from other experiments undertaken at  
494 Geosciences Montpellier where  $m = n = 1$  (Graveleau et al., 2011; Strak et al., 2011). These  
495 differences likely reflect different material proportions in the mixture, since the basic materials  
496 used in the above cited experiments and in this study are the same. An exponent on the slope  $n$   
497 higher than one is consistent with the high degree of knickpoint preservation.  
498 Knickpoints should gradually become smoother with time following the stream power theory if  $n$   
499 would be less than 1 (Tucker and Whipple, 2002). The relationship between faster knickpoint  
500 celerity and large uplift rates (Fig. 11) is also explainable if  $n > 1$  (Eq. 10), taking into account  
501 that the channels steepen under higher uplift rates (Fig. 10). We obtain  $\Psi_{AW} \propto U^{0.5 \pm 0.1}$  supporting  
502  $n = 1.5$  inferred from the scaling of  $k_{sn}$  with  $W$  and  $U$ . It is a higher value than inferred in studies  
503 from rates of erosion or uplift that shows that  $n$  vary between  $n = 1$  in the Siwalik Hills,  
504 Nepal (Wobus et al., 2006a) and in Hawaii (Ferrier et al., 2013), and  $n < 1$  for the same Siwalik  
505 Hills (Kirby and Whipple, 2001). Recent theoretical and field studies demonstrated a nonlinear  
506 dependence between steepness indices and erosion rates with an exponent  $n > 1$  that can be  
507 explained by thresholds in the process of erosion combined to rainfall and runoff  
508 variability (DiBiase and Whipple, 2011; Finnegan, 2013; Lague, 2014). In our experiments the  
509 discharge in the persistent-transverse channels is by definition largely higher than the threshold  
510 for erosion and we do not observe temporal variations in flow velocity in these channels  
511 indicating that a threshold cannot account for the high value of  $n$ . The inferred high value is  
512 however consistent with some studies that inferred  $n > 1.3$  in Turkey and Italy (Whittaker and  
513 Boulton, 2012) and with theoretical estimations inferring  $n$  to vary from 1.05 to 1.45 (Attal et al.,  
514 2008). Although  $m$  is generally believed to be close to 0.5 (e.g. Berlin and Anderson, 2007;  
515 Bishop and Cowell, 1997; Bishop et al., 2005; Loget and Van Den Driessche, 2009) we found  $m$   
516  $= 0.8 \pm 0.2$ . The experiments also show a dependence of channel width  $W$  to drainage area and  
517 uplift rate that implies that a simple scaling of  $W$  with drainage area  $A$  is not correct and that the  
518 experimental channels adjust to uplift rate by both increasing their slope and narrowing their  
519 width as it is observed in nature (Duvall et al., 2004; Lavé and Avouac, 2001; Snyder et al.,  
520 2003a; Turowski et al., 2009; Whittaker and Boulton, 2012; Whittaker et al., 2007a; Yanites and  
521 Tucker, 2010). The propagation of a knickpoint during the uplift of a thrust suggests that the

522 erosion model in the experiments is not a transport-limited model (e.g., Whipple and Tucker,  
523 2002) except maybe in the areas of sedimentation. The model that best characterizes the areas  
524 under erosion in the experiments is a detachment-limited model or a hybrid model including  
525 transport capacity of channels (Whipple and Tucker, 2002). Our results indicate that the  
526 experimental models have geometric and kinematic similarities to what is observed in natural  
527 landscapes even if rigorous scaling is not feasible for these types of experiments (see discussions  
528 in Bonnet, 2009; Lague et al., 2003; Paola et al., 2009).

529 In the models experiencing the higher shortening rates (A5 and A6), the time for runoff-  
530 driven erosion to incise the forelimb of the uplifting thrust sheets is too short, leaving the slopes  
531 preserved from incision and submitted to generalized mass wasting. Even in dry land context,  
532 extensive non-eroded surfaces are usually not so preserved indicating that these high shortening  
533 rate experiments represent an end-member. The preservation of flat surfaces from erosion  
534 is inherent to this kind of models, which are composed of cohesive materials that satisfy the  
535 mechanical properties for large fold-thrust to develop, in return inhibiting erosion below slopes  
536 of  $\sim 3^\circ$ . However, in the models run under the slower rates of convergence, most of the erosion is  
537 achieved by runoff (Fig. 6A) and the topographies dissected by channels show valleys, channels,  
538 terraces, and other features found in natural landscapes.

539 Until now, the influence of uplift on drainage diversion has been studied by the use of  
540 kinematic numerical models, where deformation is reduced to uplift and advection over a thrust  
541 (e.g. Champel et al., 2002; Humphrey and Konrad, 2000; Tomkin and Braun, 1999; van der  
542 Beek et al., 2002). In this study, we consider an accretionary wedge system that allows us to  
543 reproduce and to monitor the dynamics of drainage network, erosion and sedimentation  
544 patterns, above and during the development of up to five successive thrust sheets. Onset of  
545 thrusting and spacing between thrust sheets emerge naturally from the experiment. The  
546 deformation is controlled by the mechanical properties of the experimental material and the  
547 setup (layering and thicknesses). Except at very high rates of shortening, surface processes in the  
548 experimental models are dominated by runoff-erosion, transport (advection) and sedimentation  
549 (see also Graveleau et al., 2011 and Strak et al., 2011). In particular, this experimental setup  
550 allows investigating the effects of shortening and rainfall rates on drainage organization and  
551 sediment dispersal unlike 3D numerical studies of thrust wedges implementing 2D diffusive  
552 surface process (Ruh et al., 2013). Recently a study coupling 3D mechanical and surface process  
553 model (Collignon et al., 2014) has shown that where folds grow (thrust are not modeled) they all  
554 deviate the main channels resulting in a longitudinal-dominated network. To date no specific  
555 treatment of the controlling factors on drainage network evolution and sedimentation pattern in  
556 wedges has been done using a coupled 3D mechanical/surface process model.

557

## 558 4.2 Controls on spacing of persistent transverse channels

559 In fold-and-thrust belts drainage organization depends conceptually on the capability for  
560 preexisting reaches to incise or not uplifting thrust sheets (Jackson et al., 1996). Progradation of  
561 fans in wedge top-like basins potentially makes aggradation in the backlimb of emergent thrusts  
562 to balance uplift rates allowing transverse rivers to maintain their course instead of being  
563 diverted (Humphrey and Konrad, 2000). We designed our sand-box experiments in order for the  
564 spacing between thrusts to be large enough (14 cm) to avoid the burying of the backlimbs  
565 (Fig. 4). When a fold-thrust emerges, it propagates laterally and an axial slope parallel to the  
566 structure and perpendicular to the regional slope may be created in the backlimb. Like frontal  
567 uplift, this axial slope has, if large enough, the potential to divert preexisting transverse  
568 channels. Axial slopes are controlled by the ratio of the lateral to frontal propagation rate and by  
569 the dip of the décollement layer where the thrusts are rooted (Champel et al., 2002; van der Beek  
570 et al., 2002). In our experiments the basal décollement is flat, and when lateral growth occurs,  
571 the lateral propagation rates are 10 to 100 times higher than the convergence rates, both  
572 elements preventing the development of channel diversion by the growth of axial slopes  
573 (see figure 4B). The setup we used allows studying the influence of the rate of uplifting thrust  
574 sheet alone on channel diversion. The series of experiments shows that the proportion of  
575 persistent preexisting transverse channels,  $R$ , does not scale linearly with the ratio of  
576 precipitation over shortening rate (Fig. 13B). It has been shown numerically that this ratio scales  
577 linearly with the ratio of precipitation over shortening rate if the law of erosion is linearly  
578 proportional to the drainage area and to the slope (Champel et al., 2002; Tomkin and Braun,  
579 1999; van der Beek et al., 2002). The scaling of the erosion law suggests erosion is not linearly  
580 dependent on drainage area and slope, which seems to be confirmed by the non-linear behavior  
581 of the number of diversions in response to uplift. We observe that the minimum drainage area of  
582 the persistent transverse channels increases with the rate of shortening and that it decreases  
583 under wetter conditions (Fig. 11A). It means that the minimum discharge needed to overcome a  
584 new uplifting thrust sheet increases with uplift rate. Figure 17 shows that the minimum  
585 discharge increases slightly exponentially with uplift rate. Theoretically, at steady-state,  
586 discharge is proportional to uplift rate raised to the power of  $1/m$ . Our data follow this relation  
587 using  $m=0.8$  ( $1/m=1.25$ ). This relation ultimately controls the spacing of persistent transverse  
588 channels.

589  
590 As mentioned above, it is generally accepted that the spacing of persistent antecedent rivers  
591 depends on the competition between stream power and uplift (Gupta, 1997; Jackson et al., 1996;  
592 Tomkin and Braun, 1999), although aggradation in the backlimb and the décollement dip of a  
593 thrust may complicate this simple relation (Champel et al., 2002; Humphrey and Konrad, 2000;

594 van der Beek et al., 2002). At the scale of an orogen, lithology does not appear to control the  
595 diversion of rivers (Gupta, 1997). Consequently, if lateral variations in drainage deflection are  
596 neither controlled by thrust geometry (décollement dip) nor by substantial changes in  
597 lithology, spatial variations in rock uplift rate along a thrust front may be evidenced by  
598 minimum-discharge variations of persistent transverse channels.

599

### 600 **4.3 Controls on sediment dispersal**

601 The drainage organization is an additional control on sedimentary flux to sedimentary  
602 basins. If the transverse channels draining the wedge are not diverted, a line-source dispersal  
603 system develops in front of the active structure. The merger of drainage basins results in point-  
604 sourced depositional systems separated by areas fed only by small channels developing in the  
605 external limb of the active thrust sheet. We reproduce similar sediment patterns that occur at the  
606 Himalayan and Andean fronts (Gupta, 1997; Horton and DeCelles, 2001). The experiments  
607 confirm the view that the competition between discharge and tectonic uplift controls along-  
608 strike variations in sediment flux in sedimentary basins by controlling drainage  
609 organization (Tucker and Slingerland, 1994). The magnitude of drainage reorganization should  
610 modulate the magnitude of stratigraphic architecture variations along strike. However, we do  
611 not observe significant larger volumes of sediments in front of larger drainage basins with  
612 respect to smaller ones of the same experiments as it is observed in nature. This is an  
613 experimental limit due to the inhibition of erosion in the interior of the prism. It also prevents the  
614 lateral expansion of drainage network in the interior of the prism, which is believed to maintain  
615 the spacing ratio of outlets (half mountain width over outlet spacing) during orogen  
616 widening (Horton and DeCelles, 2001; Hovius, 1996). In the experiments this ratio is not  
617 honored because it can only increase during widening of the prism. Once a thrust becomes  
618 inactive, it is incorporated, with its associated basin, into the prism, and minor erosion and  
619 bypass dominate. Because the spacing between thrusts is the same at different convergence  
620 rates, the time of activity of the external thrusts is lower at higher rates of shortening. This  
621 explains why the total amount of sediments delivered for different convergence rates (two  
622 orders of magnitude) does not vary although we show higher rates of sedimentation and erosion  
623 under higher rates of shortening (Fig. 16).

624

### 625 **4.4 Prism widening**

626 In all models the spatial extent of the wedge-top basins, as well as the distance between the  
627 traces of the thrusts, both shrink through time indicating that the internal structural units are  
628 slightly underthrust with ongoing shortening. As in previous studies we observe that the

629 growth of a prism fluctuates between periods of frontal accretion and internal thickening(e.g.  
630 Gutscher et al., 1996; Hoth et al., 2006; Konstantinovskaia and Malavieille, 2005; Naylor and  
631 Sinclair, 2007; Naylor et al., 2005; Simpson, 2006). Detail cross-sections of fans accumulated in  
632 front of a thrust sheet not yet active reveal two stages of sedimentation, one of progradation and  
633 a second one of backfilling and sealing of the thrust (Fig. 15B). We suggest that this two-phase  
634 sedimentary infill may be used as a marker for the transition between frontal accretion and  
635 internal growth in orogenic wedges. A proposed timescale for the activity of external thrust is  
636 0.1-5 Myr(Naylor and Sinclair, 2007). In between periods of frontal accretion, the ratio of  
637 drainage area lost during internal growth represents ~20% of a thrust sheet initial area since the  
638 thrust emerged. The fluctuation between periods of frontal accretion and internal thickening, as  
639 well as drainage area shrinking, may modulate the sedimentary flux at these timescales in  
640 foreland basins.

641 It is interesting to note that local slopes created by folding and faulting develop since the  
642 initial stages of tectonic growth of a wedge, and that this structurally-controlled relief is  
643 incorporated within a wedge during widening. Local slopes may divert runoff in directions  
644 perpendicular to the regional slope and create longitudinal-dominated drainage networks.  
645 Therefore tectonic-related relief may influence the long-term drainage network organization not  
646 only in the external part of an orogen but also in its interior. This can occur during the inversion  
647 of a rift (Babault et al., 2012) or during widening of a wedge by forward propagation and  
648 incorporation of thrust sheets as in the experiments.

649

## 650 **5 Conclusions**

651 Our results suggest that the rule of erosion in the experimental models has many  
652 similarities to what is observed in natural landscape. We infer from uplift rates, incision rates  
653 and metrics of the topography that erosion scales nonlinearly to slope and drainage area. We  
654 found for the exponents on the slope ( $n$ ) and on the drainage ( $m$ ) the respective values of  
655  $1.5 \pm 0.2$  and  $0.8 \pm 0.2$ . The experimental channels adjust to uplift rate by both increasing their  
656 slope and narrowing their channels as it is observed in nature above active faults.

657 Drainage pattern in fold-and-thrust belts depends on the capability for antecedent reaches  
658 to incise or not the uplifting thrust sheets. We observe that the minimum drainage area for  
659 transverse channels to incise the uplifting structure increases with the rate of shortening and  
660 decreases under wetter conditions, suggesting that discharge is the controlling factor. Indeed, a  
661 minimum discharge exists for a new uplifting thrust sheet to be overcome and it increases  
662 exponentially with uplift rate. This minimum discharge ultimately controls the spacing of  
663 persistent transverse channels. Consequently, if lateral variations in drainage deflection are  
664 neither controlled by thrust geometry (décollement dip) nor by substantial changes in lithology

665 as shown in previous studies, then the spatial variations in rock uplift rate along a thrust front  
666 could be evidenced by studying minimum-discharge variations of persistent transverse channels.

667 The experiments confirm the view that competing discharge and tectonic uplift control  
668 drainage pattern resulting in along-strike variations of sediment flux in sedimentary basins.  
669 When the transverse channels that drain a wedge are not diverted, a line-source dispersal system  
670 develops in front of the active thrust sheet. The merger of drainage basins in the backlimb of the  
671 frontal thrust sheet results in point-sourced depositional systems separated by areas fed by small  
672 channels that develop in the external limb.

673 Detail cross-sections of fans accumulated in front of a thrust sheet that became inactive  
674 reveal two stages of sedimentation, one of progradation and a second one of backfilling and  
675 sealing of the thrust. We suggest that this two-phase sedimentary infill may be used as a marker  
676 for the transition between frontal accretion and internal growth in orogenic wedges.

677

## 678 **Acknowledgments**

679 Financial support was provided to J. Babault by REPSOL, for M. Viaplana-Muzas PhD  
680 grant, and they also benefited from Spanish national projects (CGL2010-1516 and  
681 CONSOLIDER-Ingenio CSD2006-00041). We thank C. Romano from Géosciences  
682 Montpellier for technical assistance and we thank J. Malavieille for helpful discussions. Thanks  
683 to M. González, C. Díaz and E. Álvarez de Buergo for discussions on preliminary results of this  
684 work. We thank the editor Rob Govers, Kosuke Ueda and one anonymous reviewer for their  
685 constructive comments and suggestions that have contributed to the improvement of the  
686 manuscript.

687

688

## 689 **References**

690 Amos, C.B., Burbank, D.W., 2007. Channel width response to differential uplift. *Journal of*  
691 *Geophysical Research: Earth Surface* 112, F02010.

692 Attal, M., Tucker, G.E., Whittaker, A.C., Cowie, P.A., Roberts, G.P., 2008. Modeling fluvial  
693 incision and transient landscape evolution: Influence of dynamic channel adjustment. *J.*  
694 *Geophys. Res.-Earth Surf.* 113, 16.

695 Babault, J., Bonnet, S., Crave, A., Van Den Driessche, J., 2005. Influence of piedmont  
696 sedimentation on erosion dynamics of an uplifting landscape: An experimental approach.  
697 *Geology* 33, 301-304, doi: 310.1130/G21095.21091.

698 Babault, J., Van Den Driessche, J., Teixell, A., 2012. Longitudinal to transverse drainage  
699 network evolution in the High Atlas (Morocco): The role of tectonics. *Tectonics* 31, TC4020.

700 Berlin, M.M., Anderson, R.S., 2007. Modeling of knickpoint retreat on the Roan Plateau,  
701 western Colorado. *Journal of Geophysical Research: Earth Surface* 112, F03S06.

702 Bishop, P., Cowell, P., 1997. Lithological and Drainage Network Determinants of the Character  
703 of Drowned, Embayed Coastlines. *The Journal of Geology* 105, 685-700.

704 Bishop, P., Hoey, T.B., Jansen, J.D., Artza, I.L., 2005. Knickpoint recession rate and catchment  
705 area: the case of uplifted rivers in Eastern Scotland. *Earth Surface Processes and Landforms* 30,  
706 767-778.

707 Bonnet, S., 2009. Shrinking and splitting of drainage basins in orogenic landscapes from the  
708 migration of the main drainage divide. *Nature Geoscience* 2, 897-897.

709 Bookhagen, B., Strecker, M.R., 2012. Spatiotemporal trends in erosion rates across a  
710 pronounced rainfall gradient: Examples from the southern Central Andes. *Earth and Planetary  
711 Science Letters* 327–328, 97-110.

712 Burbank, McLean, Bullen, Abdrakhmatov, Miller, 1999. Partitioning of intermontane basins by  
713 thrust-related folding, Tien Shan, Kyrgyzstan. *Basin Research* 11, 75-92.

714 Burbank, D.W., Vergés, J., 1994. Reconstruction of topography and related depositional  
715 systems during active thrusting. *Journal of Geophysical Research: Solid Earth* 99, 20281-20297.

716 Castelltort, S., Goren, L., Willett, S.D., Champagnac, J.-D., Herman, F., Braun, J., 2012. River  
717 drainage patterns in the New Zealand Alps primarily controlled by plate tectonic strain. *Nature  
718 Geosci* 5, 744-748.

719 Champel, B., van der Beek, P., Mugnier, J.-L., Leturmy, P., 2002. Growth and lateral  
720 propagation of fault-related folds in the Siwaliks of western Nepal; rates, mechanisms, and  
721 geomorphic signature. *Journal of Geophysical Research* 107.

722 Collignon, M., Kaus, B., May, D., Fernandez, N., 2014. Influences of surface processes on fold  
723 growth during 3-D detachment folding. *Geochemistry, Geophysics, Geosystems* 15, 3281-3303.

724 Cowie, P.A., Whittaker, A.C., Attal, M., Roberts, G., Tucker, G.E., Ganas, A., 2008. New  
725 constraints on sediment-flux-dependent river incision: Implications for extracting tectonic  
726 signals from river profiles. *Geology* 36, 535-538.

727 Crosby, B.T., Whipple, K.X., 2006. Knickpoint initiation and distribution within fluvial  
728 networks: 236 waterfalls in the Waipaoa River, North Island, New Zealand. *Geomorphology* 82,  
729 16-38.

730 Davis, D., Suppe, J., Dahlen, F.A., 1983. Mechanics of fold-and-thrust belts and accretionary  
731 wedges. *Journal of Geophysical Research: Solid Earth* 88, 1153-1172.

732 DiBiase, R.A., Whipple, K.X., 2011. The influence of erosion thresholds and runoff variability  
733 on the relationships among topography, climate, and erosion rate. *Journal of Geophysical  
734 Research: Earth Surface* 116, F04036.

735 DiBiase, R.A., Whipple, K.X., Heimsath, A.M., Ouimet, W.B., 2010. Landscape form and  
736 millennial erosion rates in the San Gabriel Mountains, CA. *Earth and Planetary Science Letters*  
737 289, 134-144.

738 Duvall, A., Kirby, E., Burbank, D., 2004. Tectonic and lithologic controls on bedrock channel  
739 profiles and processes in coastal California. *Journal of Geophysical Research: Earth Surface*  
740 109, F03002.

741 Ferrier, K.L., Huppert, K.L., Perron, J.T., 2013. Climatic control of bedrock river incision.  
742 *Nature* 496, 206-209.

743 Finnegan, N., 2013. Interpretation and downstream correlation of bedrock river terrace treads  
744 created from propagating knickpoints. *Journal of Geophysical Research: Earth Surface* 118, 54-  
745 64.

746 Finnegan, N.J., Roe, G., Montgomery, D.R., Hallet, B., 2005. Controls on the channel width of  
747 rivers: Implications for modeling fluvial incision of bedrock. *Geology* 33, 229-232.

748 Giletycz, S., Loget, N., Chang, C.P., Mouthereau, F., 2015. Transient fluvial landscape and  
749 preservation of low-relief terrains in an emerging orogen: Example from Hengchun Peninsula,  
750 Taiwan. *Geomorphology* 231, 169-181.

751 Graveleau, F., Dominguez, S., 2008. Analogue modelling of the interaction between tectonics,  
752 erosion and sedimentation in foreland thrust belts. *Comptes Rendus. Geoscience* 340, 324-333.

753 Graveleau, F., Hurtrez, J.E., Dominguez, S., Malavieille, J., 2011. A new experimental material  
754 for modeling relief dynamics and interactions between tectonics and surface processes.  
755 *Tectonophysics* 513, 68-87.

756 Graveleau, F., Strak, V., Dominguez, S., Malavieille, J., Chatton, M., Manighetti, I., Petit, C.,  
757 2015. Experimental modelling of tectonics–erosion–sedimentation interactions in  
758 compressional, extensional, and strike–slip settings. *Geomorphology*.

759 Gudmundsdottir, M.H., Blisniuk, K., Ebert, Y., Levine, N.M., Rood, D.H., Wilson, A., Hilley,  
760 G.E., 2013. Restraining bend tectonics in the Santa Cruz Mountains, California, imaged using  
761 <sup>10</sup>Be concentrations in river sands. *Geology* 41, 843-846.

762 Gupta, S., 1997. Himalayan drainage patterns and the origin of fluvial megafans in the Ganges  
763 foreland basin. *Geology* 25, 11-14.

764 Gutscher, M.-A., Kukowski, N., Malavieille, J., Lallemand, S., 1996. Cyclical behavior of thrust  
765 wedges: Insights from high basal friction sandbox experiments. *Geology* 24, 135-138.

766 Horton, B.K., DeCelles, P.G., 2001. Modern and ancient fluvial megafans in the foreland basin  
767 system of the central Andes, southern Bolivia: implications for drainage network evolution in  
768 fold-thrust belts. *Basin Research* 13, 43-63.

769 Hoth, S., Adam, J., Kukowski, N., Oncken, O., 2006. Influence of erosion on the kinematics of  
770 bivergent orogens: Results from scaled sandbox simulations. *Geological Society of America*  
771 *Special Papers* 398, 201-225.

772 Hovius, N., 1996. Regular spacing of drainage outlets from linear mountain belts. *Basin*  
773 *Research* 8, 29-44.

774 Howard, A.D., 1980. Thresholds in river regimes. *Thresholds in geomorphology*, 227-258.

775 Howard, A.D., Kerby, G., 1983. Channel changes in badlands. *GSA Bull.* 94, 739-752.

776 Hubbert, M.K., 1951. Mechanical basis for certain familiar geologic structures. *Geological*  
777 *Society of America Bulletin* 62, 355-372.

778 Humphrey, N.F., Konrad, S.K., 2000. River incision or diversion in response to bedrock uplift.  
779 *Geology* 28, 43-46.

780 Jackson, J., Norris, R., Youngson, J., Wojtal, S.F., 1996. The structural evolution of active fault  
781 and fold systems in central Otago, New Zealand; evidence revealed by drainage patterns.  
782 *Journal of Structural Geology* 18, 217-234.

783 Jolley, E., Turner, P., Williams, G., Hartley, A., Flint, S., 1990. Sedimentological response of an  
784 alluvial system to Neogene thrust tectonics, Atacama Desert, northern Chile. *Journal of the*  
785 *Geological Society* 147, 769-784.

786 Kirby, E., Ouimet, W., 2011. Tectonic geomorphology along the eastern margin of Tibet:  
787 insights into the pattern and processes of active deformation adjacent to the Sichuan Basin.  
788 *Geological Society, London, Special Publications* 353, 165-188.

789 Kirby, E., Whipple, K., 2001. Quantifying differential rock-uplift rates via stream profile  
790 analysis. *Geology* 29, 415-418.

791 Kirby, E., Whipple, K.X., 2012. Expression of active tectonics in erosional landscapes. *Journal*  
792 *of Structural Geology* 44, 54-75.

793 Konstantinovskaia, E., Malavieille, J., 2005. Erosion and exhumation in accretionary orogens:  
794 Experimental and geological approaches. *Geochem. Geophys. Geosyst.* 6, Q02006.

795 Koons, P.O., 1994. Three-dimensional critical wedges: Tectonics and topography in oblique  
796 collisional orogens. *Journal of Geophysical Research: Solid Earth* 99, 12301-12315.

797 Koons, P.O., 1995. Modeling the topographic evolution of collisional belts. *Annual Review of*  
798 *Earth and Planetary Sciences* 23, 375-408.

799 Lague, D., 2014. The stream power river incision model: evidence, theory and beyond. *Earth*  
800 *Surface Processes and Landforms* 39, 38-61.

801 Lague, D., Crave, A., Davy, P., 2003. Laboratory experiments simulating the geomorphic  
802 response to tectonic uplift. *Journal of geophysical research* 108, 2008,  
803 doi:2010.1029/2002JB001785.

804 Lague, D., Hovius, N., Davy, P., 2005. Discharge, discharge variability, and the bedrock  
805 channel profile. *Journal of Geophysical Research: Earth Surface* 110, F04006.

806 Lavé, J., Avouac, J.P., 2001. Fluvial incision and tectonic uplift across the Himalayas of central  
807 Nepal. *Journal of Geophysical Research: Solid Earth* 106, 26561-26591.

808 Loget, N., Van Den Driessche, J., 2009. Wave train model for knickpoint migration.  
809 *Geomorphology* 106, 376-382.

810 Montgomery, D.R., Gran, K.B., 2001. Downstream variations in the width of bedrock channels.  
811 *Water Resour. Res.* 37, 1841-1846.

812 Naylor, M., Sinclair, H.D., 2007. Punctuated thrust deformation in the context of doubly vergent  
813 thrust wedges: Implications for the localization of uplift and exhumation. *Geology* 35, 559-562.

814 Naylor, M., Sinclair, H.D., Willett, S., Cowie, P.A., 2005. A discrete element model for  
815 orogenesis and accretionary wedge growth. *Journal of Geophysical Research: Solid Earth* 110,  
816 B12403.

817 Oberlander, T.M., 1985. Origin of drainage transverse to structures in orogens. *Binghamton*  
818 *Symposia in Geomorphology: International Series* 15, 155-182.

819 Ouimet, W.B., Whipple, K.X., Granger, D.E., 2009. Beyond threshold hillslopes: Channel  
820 adjustment to base-level fall in tectonically active mountain ranges. *Geology* 37, 579-582.

821 Paola, C., Straub, K., Mohrig, D., Reinhardt, L., 2009. The "unreasonable effectiveness" of  
822 stratigraphic and geomorphic experiments. *Earth-Science Reviews* 97, 1-43.

823 Parker, G., Anderson, A.G., 1977. Basic principles of river hydraulics. *Journal of the Hydraulics*  
824 *Division* 103.

825 Perron, J.T., Royden, L., 2013. An integral approach to bedrock river profile analysis. *Earth*  
826 *Surface Processes and Landforms* 38, 570-576.

827 Rosenbloom, N.A., Anderson, R.S., 1994. Hillslope and channel evolution in a marine terraced  
828 landscape, Santa Cruz, California. *Journal of Geophysical Research: Solid Earth* 99, 14013-  
829 14029.

830 Royden, L., Clark, M., Whipple, K., 2000. Evolution of river elevation profiles by bedrock  
831 incision: Analytical solutions for transient river profiles related to changing uplift and  
832 precipitation rates. *Eos Trans. AGU* 81, 48.

833 Ruh, J.B., Gerya, T., Burg, J.-P., 2013. High-resolution 3D numerical modeling of thrust  
834 wedges: Influence of décollement strength on transfer zones. *Geochemistry, Geophysics,*  
835 *Geosystems*, n/a-n/a.

836 Schellart, W.P., 2000. Shear test results for cohesion and friction coefficients for different  
837 granular materials: scaling implications for their usage in analogue modelling. *Tectonophysics*  
838 324, 1-16.

839 Simpson, G.D.H., 2006. Modelling interactions between fold–thrust belt deformation, foreland  
840 flexure and surface mass transport. *Basin Research* 18, 125-143.

841 Snyder, N.P., Whipple, K.X., Tucker, G.E., Merritts, D.J., 2003a. Importance of a stochastic  
842 distribution of floods and erosion thresholds in the bedrock river incision problem. *Journal of*  
843 *Geophysical Research: Solid Earth* 108, 2117.

844 Snyder, N.P., Whipple, K.X., Tucker, G.E., Merritts, D.J., 2003b. Channel response to tectonic  
845 forcing: field analysis of stream morphology and hydrology in the Mendocino triple junction  
846 region, northern California. *Geomorphology* 53, 97-127.

847 Sobel, E.R., Hilley, G.E., Strecker, M.R., 2003. Formation of internally drained contractional  
848 basins by aridity-limited bedrock incision. *Journal of Geophysical Research* 108, 2344.

849 Strak, V., Dominguez, S., Petit, C., Meyer, B., Loget, N., 2011. Interaction between normal  
850 fault slip and erosion on relief evolution: Insights from experimental modelling. *Tectonophysics*  
851 513, 1-19.

852 Talling, P.J., Lawton, T.F., Burbank, D.W., Hobbs, R.S., 1995. Evolution of latest Cretaceous–  
853 Eocene nonmarine deposystems in the Axhandle piggyback basin of central Utah. *Geological*  
854 *Society of America Bulletin* 107, 297-315.

855 Tomkin, J.H., Braun, J., 1999. Simple models of drainage reorganisation on a tectonically active  
856 ridge system. *New Zealand Journal of Geology and Geophysics* 42, 1-10.

857 Tucker, G.E., 2004. Drainage basin sensitivity to tectonic and climatic forcing; implications of a  
858 stochastic model for the role of entrainment and erosion thresholds. *Earth Surface Processes and*  
859 *Landforms* 29, 185-205.

860 Tucker, G.E., Slingerland, R., 1994. Erosional dynamics, flexural isostasy, and long-lived  
861 escarpments: a numerical modeling study. *Journal of Geophysical Research* 99, 12,229-212,243.

862 Tucker, G.E., Slingerland, R., 1996. Predicting sediment flux from fold and thrust belts. *Basin*  
863 *Research* 8, 329-349.

864 Tucker, G.E., Whipple, K.X., 2002. Topographic outcomes predicted by stream erosion models:  
865 Sensitivity analysis and intermodel comparison. *Journal of Geophysical Research: Solid Earth*  
866 107, 2179.

867 Turowski, J.M., Lague, D., Hovius, N., 2009. Response of bedrock channel width to tectonic  
868 forcing: Insights from a numerical model, theoretical considerations, and comparison with field  
869 data. *Journal of Geophysical Research: Earth Surface* 114, F03016.

870 van der Beek, P., Champel, B., Mugnier, J.-L., 2002. Control of detachment dip on drainage  
871 development in regions of active fault-propagation folding. *Geology* 30, 471-474.

872 Weissel, J.K., Seidl, M.A., 1998. Inland Propagation of Erosional Escarpments and River  
873 Profile Evolution Across the Southeast Australian Passive Continental Margin, *Rivers Over*  
874 *Rock: Fluvial Processes in Bedrock Channels*. American Geophysical Union, pp. 189-206.

875 Whipple, K.X., 2001. Fluvial Landscape Response Time: How Plausible Is Steady-State  
876 Denudation? *Am J Sci* 301, 313-325.

877 Whipple, K.X., 2004. Bedrock rivers and the geomorphology of active orogens. *Annual Review*  
878 *of Earth and Planetary Sciences* 32, 151-185.

879 Whipple, K.X., Tucker, G.E., 1999. Dynamics of the stream-power river incision model:  
880 Implications for height limits of mountain ranges, landscape response timescales, and research  
881 needs. *Journal of Geophysical Research* 104, 17,661-617,674.

882 Whipple, K.X., Tucker, G.E., 2002. Implications of sediment-flux-dependent river incision  
883 models for landscape evolution. *Journal-of-geophysical-research* 107, ETG3.1-ETG3.20.

884 Whittaker, A.C., Attal, M., Cowie, P.A., Tucker, G.E., Roberts, G., 2008. Decoding temporal  
885 and spatial patterns of fault uplift using transient river long profiles. *Geomorphology* 100, 506-  
886 526.

887 Whittaker, A.C., Boulton, S.J., 2012. Tectonic and climatic controls on knickpoint retreat rates  
888 and landscape response times. *Journal of Geophysical Research: Earth Surface* 117, F02024.

889 Whittaker, A.C., Cowie, P.A., Attal, M., Tucker, G.E., Roberts, G.P., 2007a. Bedrock channel  
890 adjustment to tectonic forcing: Implications for predicting river incision rates. *Geology* 35, 103-  
891 106.

892 Whittaker, A.C., Cowie, P.A., Attal, M., Tucker, G.E., Roberts, G.P., 2007b. Contrasting  
893 transient and steady-state rivers crossing active normal faults: new field observations from the  
894 Central Apennines, Italy. *Basin Research* 19, 529-556.

895 Wobus, C., Heimsath, A., Whipple, K., Hodges, K., 2005. Active out-of-sequence thrust  
896 faulting in the central Nepalese Himalaya. *Nature* 434, 1008-1011.

897 Wobus, C., Whipple, K.X., Kirby, E., Snyder, N., Johnson, J., Spyropolou, K., Crosby, B.,  
898 Sheehan, D., 2006a. Tectonics from topography; procedures, promise, and pitfalls. *Special*  
899 *Paper - Geological Society of America* 398, 55-74.

900 Wobus, C.W., Crosby, B.T., Whipple, K.X., 2006b. Hanging valleys in fluvial systems; controls  
901 on occurrence and implications for landscape evolution. *Journal of Geophysical Research* 111.

902 Yanites, B.J., Tucker, G.E., 2010. Controls and limits on bedrock channel geometry. *Journal of*  
903 *Geophysical Research: Earth Surface* 115, F04019.

904

905

906

907

908

909

910

911

912

913

914

915

916

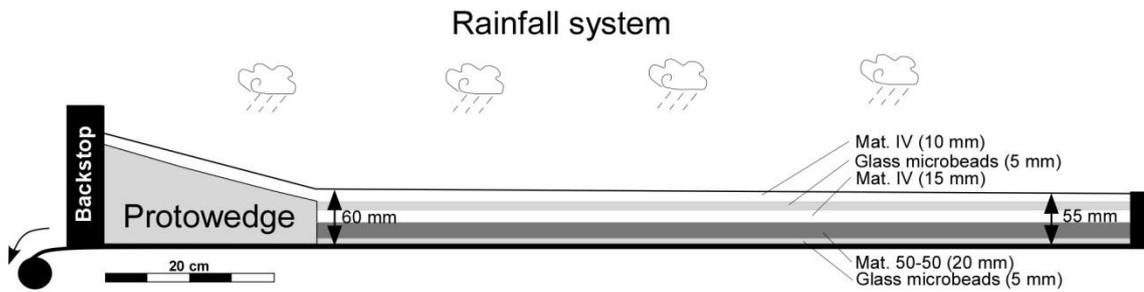
917

918

919

**Figures and captions:**

921



922

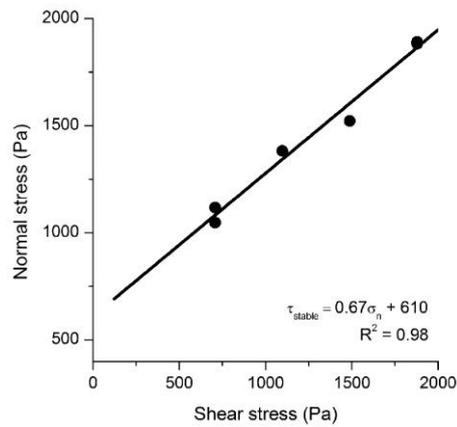
923

924

925

926

**Figure 1.** Scheme of the experimental setup. We imposed a slope of 0.5° at the surface of the models in order to enhance runoff connectivity. The initial thickness of the layered material is 60 mm close to the protowedge and 55 mm in the distal part.

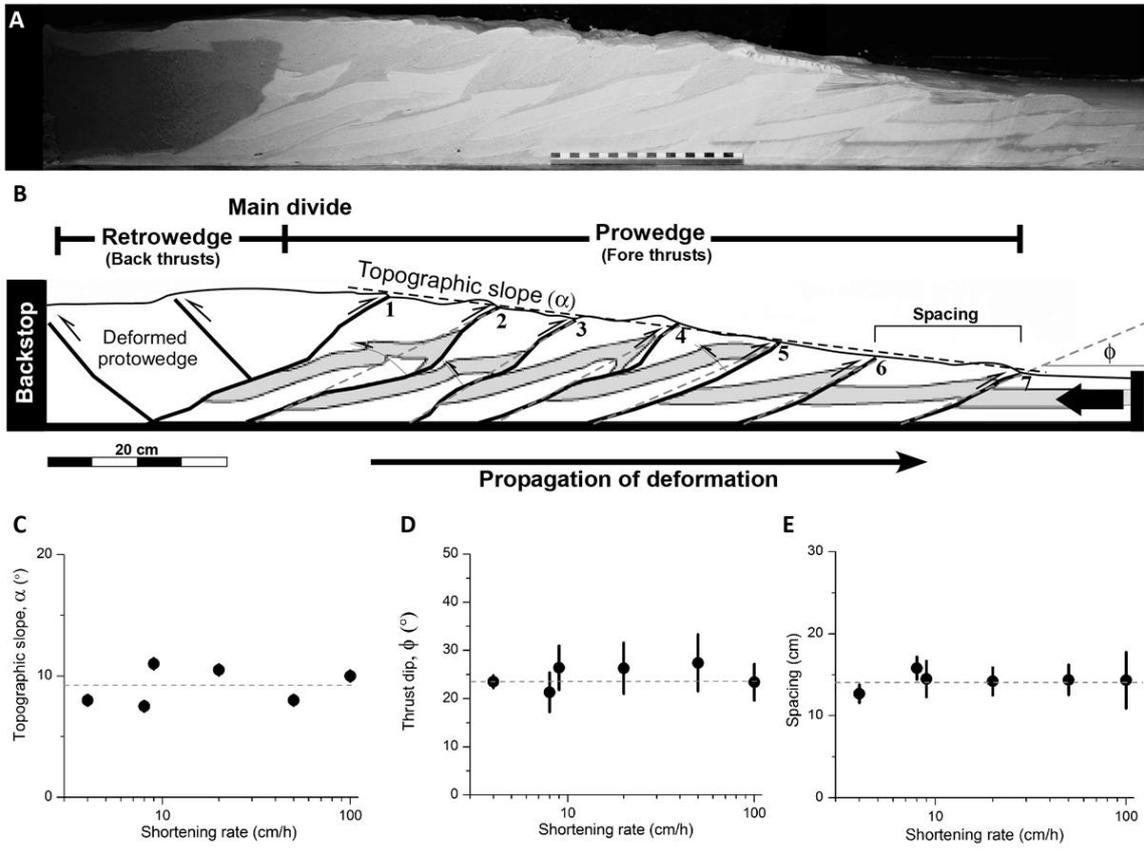


927

928

929

**Figure 2.** Mohr–Coulomb envelope (stable friction measurements) for failure of the erodible mixture (material IV) used in our experiments.



930

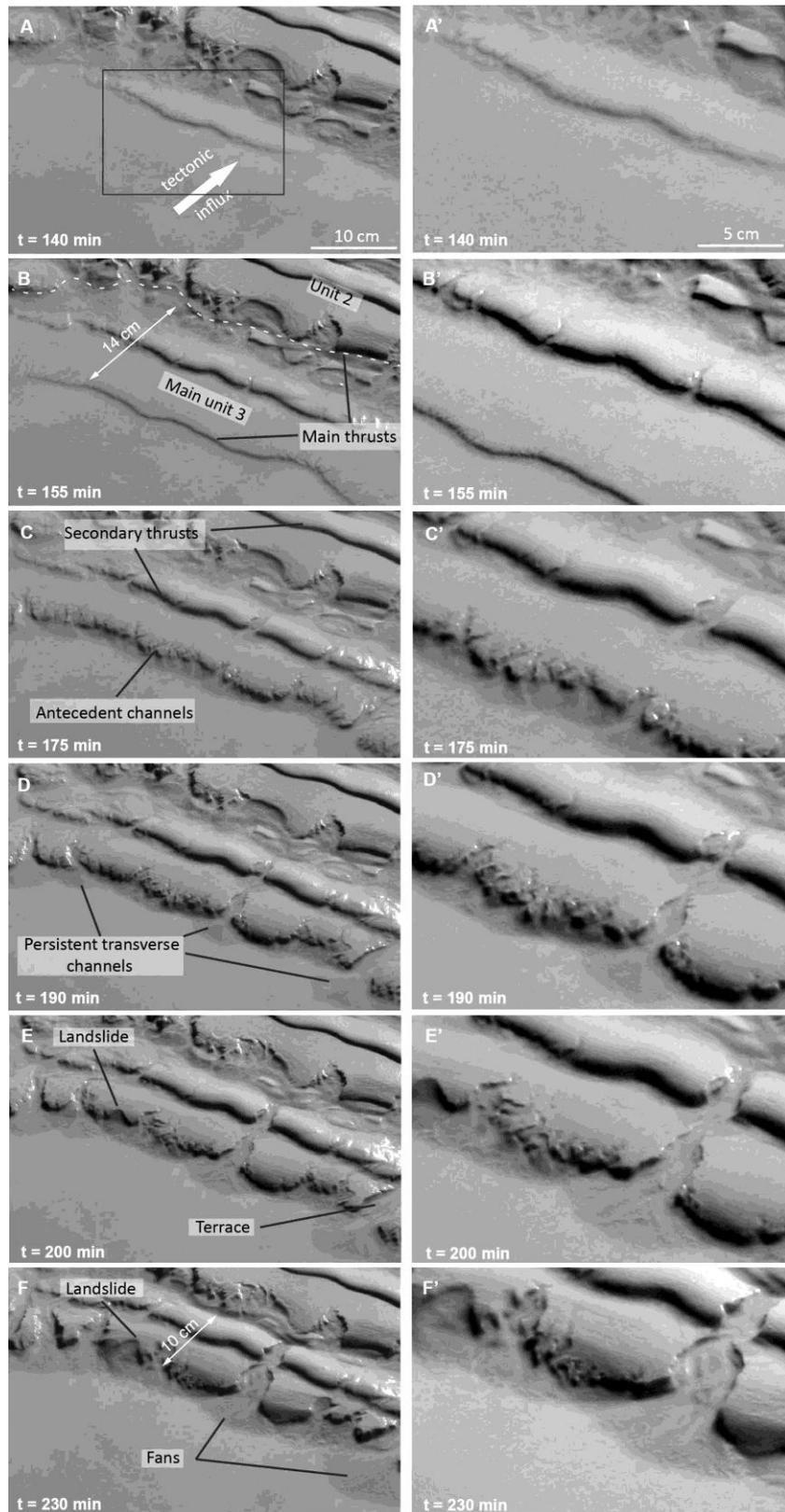
931

932

933

934

**Figure 3.** (A) Cross-section of the model A4 after 1 m of shortening. (B) Interpretation showing the main thrusts. For all the experiments we calculated the topographic mean slopes ( $\alpha$ ) of the prowedges (C), the thrust dips ( $\phi$ ) measured between the root and the tip of the thrust (dashed grey lines in B) (D) and the thrust spacing measured in the external part when a new thrust emerges (E).



935

936

937

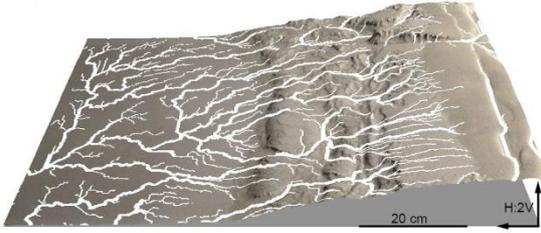
938

939

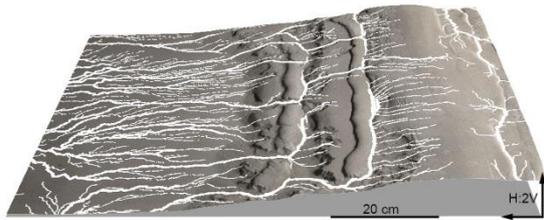
940

**Figure 4.** Photos of the experiment, A2, from  $t=140$  min, (A), before thrust 3 emerges, to  $t=230$  min (F) when the third thrust sheet has totally emerged (animation 1 add in supplementary data). The amount of shortening between (A) and (F) is 12 cm.

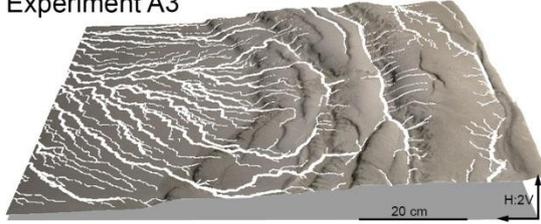
Experiment A1



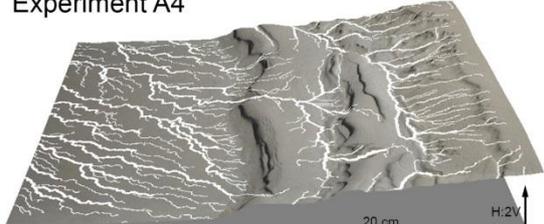
Experiment A2



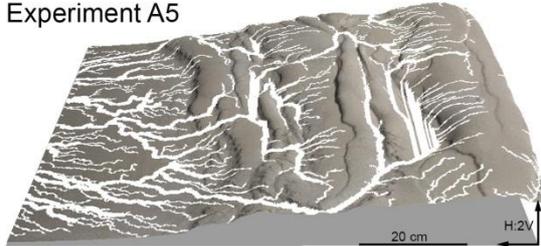
Experiment A3



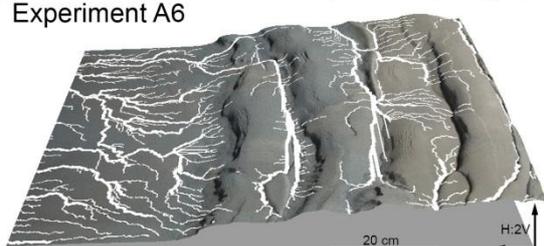
Experiment A4



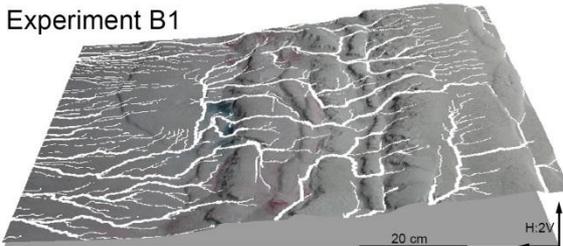
Experiment A5



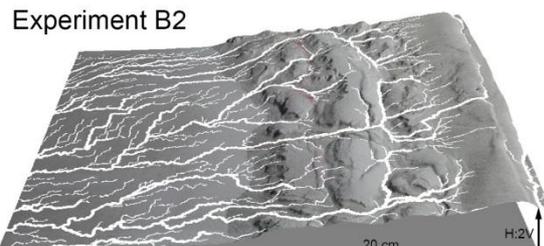
Experiment A6



Experiment B1

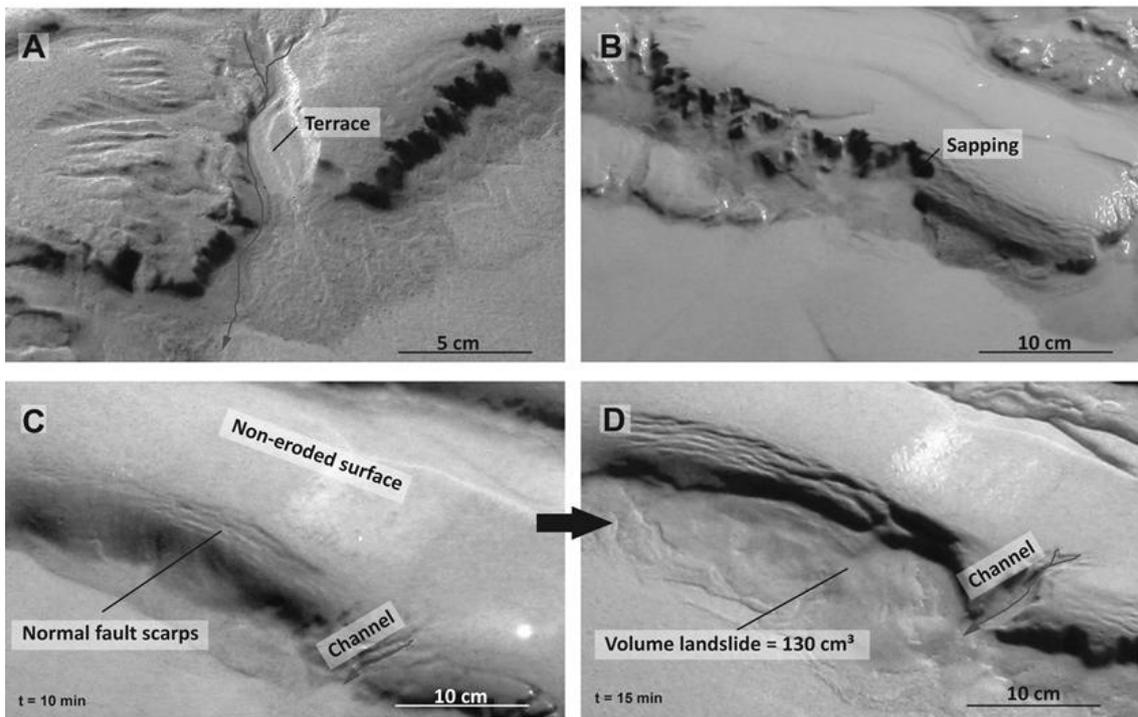


Experiment B2



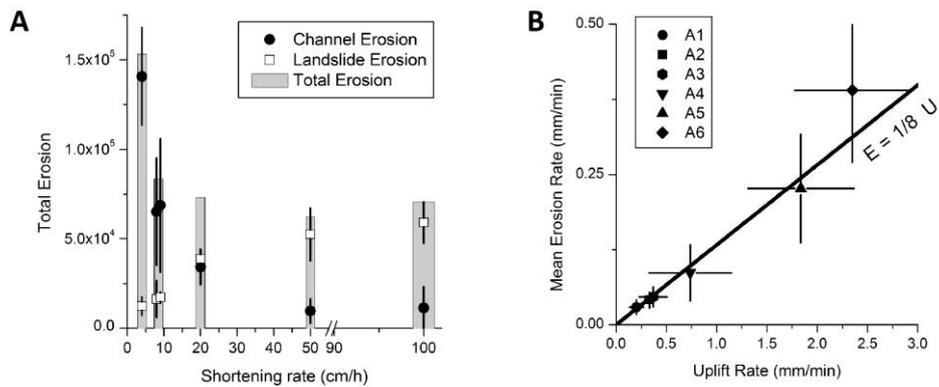
941  
 942  
 943  
 944  
 945  
 946

**Figure 5.** 3D views of all the models after the third thrust activity( the photographs are overlapped on the DEM). The drainage network is superimposed in white. The models run under 9 mm/h of rainfall rate in the A series with the shortening rate ranging between 4-100 cm/h (A1: 4 cm/h; A2: 8 cm/h; A3: 9 cm/h; A4: 20 cm/h; A5: 50 cm/h; A6: 100 cm/h). In the models run under 18 mm/h of rainfall rate, the B series, the shortening rate varied between 8 cm/h (B1) and 18 cm/h (B2).



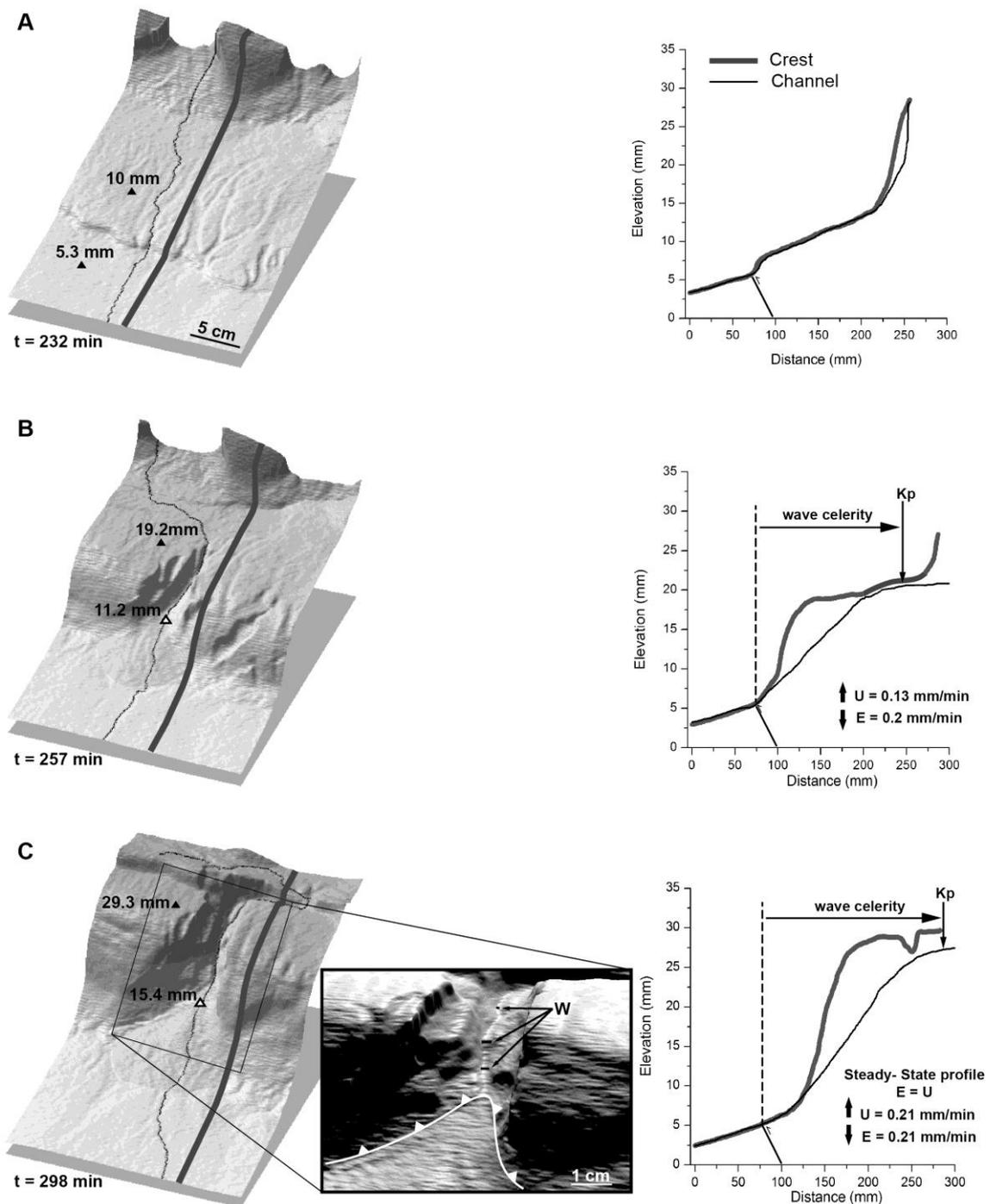
947  
948  
949  
950  
951  
952  
953

**Figure 6.** Detail pictures of the principal surface processes active in experiments (A). Detail of a channel crossing the fourth thrust sheet in the experiment A4 at  $t = 350$  min. Note the formation of a terrace in the right part of the channel and the alluvial fan located in the channel outlet. (B) Detail of sapping in the external part of the second thrust sheet of the experiment B2 at  $t = 70$  min. (C) and (D) Evolution of a landslide in the experiment A6, from  $t = 10$  min to  $t = 15$  min, situated in the external part of the third thrust sheet.



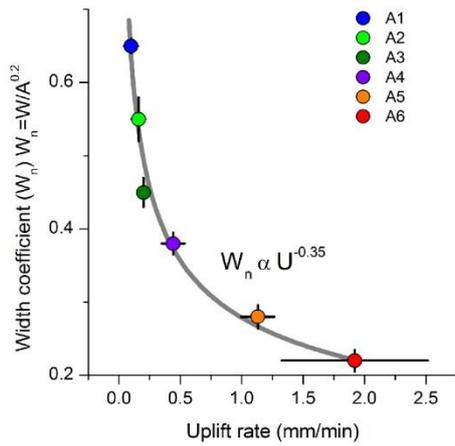
954  
955  
956  
957  
958  
959

**Figure 7.** (A) Mean amounts and standard deviations of erosion of the frontal thrust sheets, during their activity, calculated for experiments run under  $9$  mm/h of precipitation rates. (Dots) Erosion done by persistent channels and small channels developed on the forelimb of the fold-thrust. (White squares) Erosion done by landslides those develop in the forelimbs of the thrusts. (B) Mean erosion rates and standard deviations calculated from channel-driven amounts of erosion (dots in A) against uplift rate (landslide-derived erosion not included):  $E \cong \frac{1}{8} U$ .

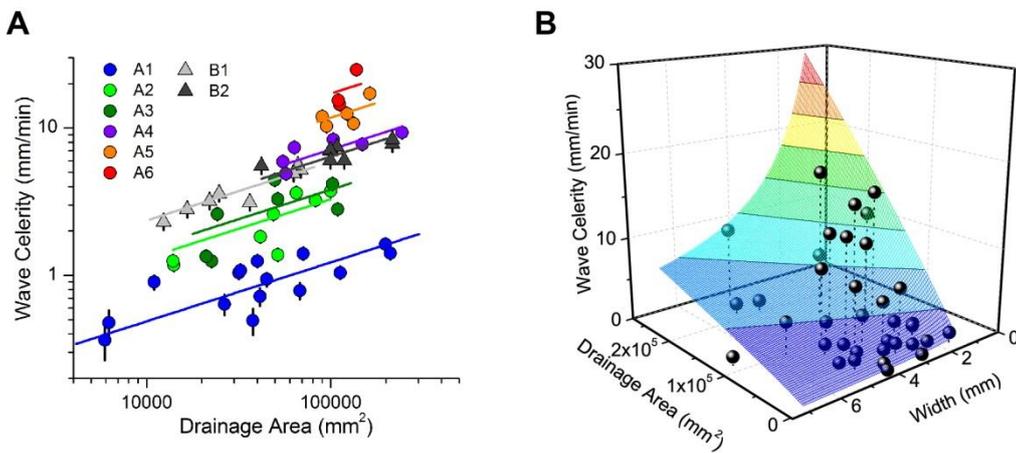


960  
 961  
 962  
 963  
 964  
 965  
 966  
 967  
 968  
 969

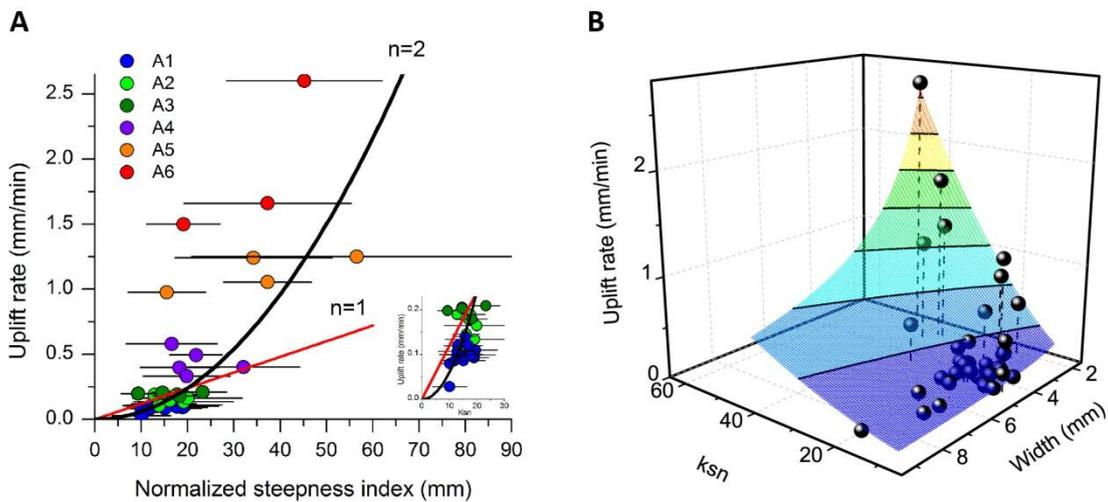
**Figure 8.** (Left) 3D detail views of a persistent-channel incision during thrust activity (experiment A2). (Right) Crest and channel topographic profiles (traces superimposed on 3D views) showing the formation of a knickpoint (Kp), when the third thrust sheet starts to emerge at  $t=232$  min (A). The knickpoint migrates upstream (B) until the fourth thrust sheet starts to emerge (C). At this moment channel erosion rate (E) balances relative uplift rate (U), 0.21 mm/min, indicating that this portion of the channel is at steady-state. The inset illustrates where the width of channels, W, has been measured along the channel. The length of the channel in this figure is 6 cm long.



970  
 971 **Figure 9:** Channel width coefficient,  $W_n$  (channel width normalized for drainage area), of persistent transverse  
 972 channels against mean relative uplift rate (error bars are standard deviation,  $1\sigma$ ), in the series of experiments run  
 973 under a precipitation rate of 9 mm/h. (Black line) Best fit using a non-linear regression:  $W_n = 0.3 \cdot U^{-0.35}$ .  
 974

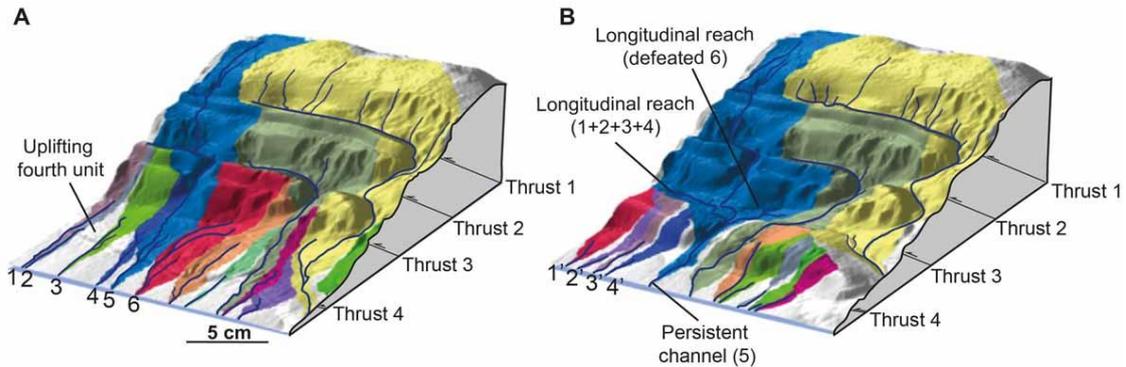


975  
 976 **Figure 10:** (A) Normalized steepness index ( $k_{sn}$ ) of the persistent antecedent channels against uplift rates, (Series A).  
 977 The equation that best fits is  $U = k_{sn}^2$ . (B) Graphic combining the adjustment of channel steepness and width to uplift  
 978 rate.  
 979

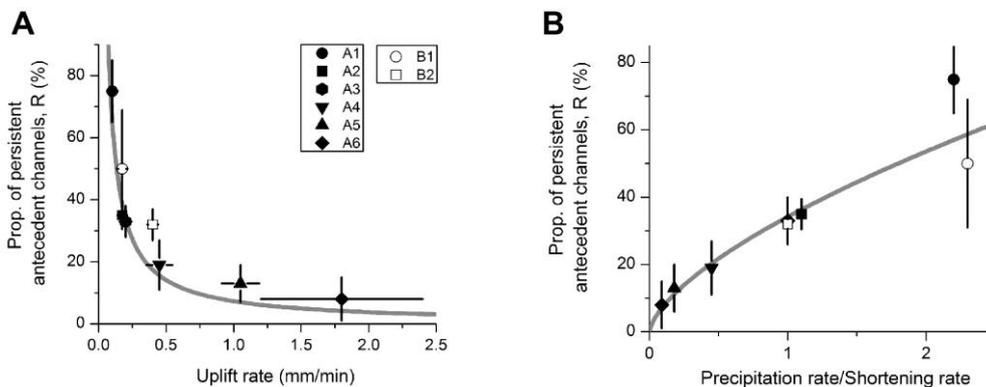


980  
 981 **Figure 11:**(A) Knickpoint wave celerity of the antecedent persistent channels against drainage area best fit using eq.  
 982 (9):  $C = \Psi_A \cdot A^{0.4}$ . Error bars come from the error in measuring the amount of knickpoint propagation, typically  $< 1$

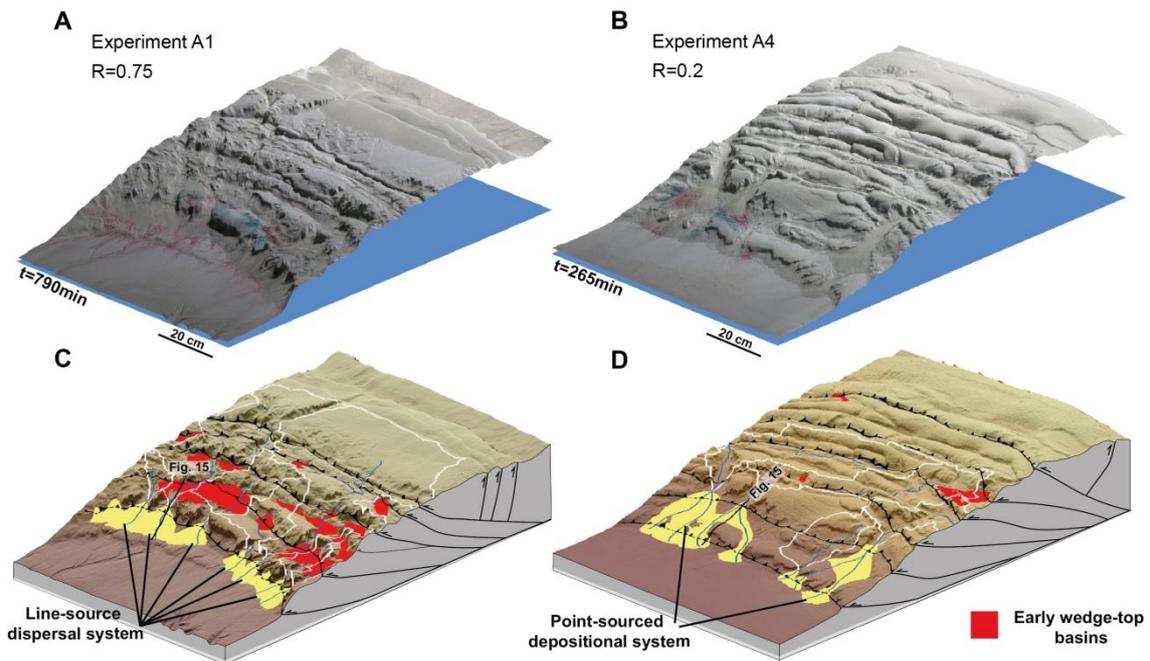
983 mm. (B) Knickpoint wave celerity as a function of drainage area and channels width, in the series of experiments ran  
 984 under a precipitation rate of 9 mm/h, best fit using eq. (10):  $C = \Psi_{AW} \frac{A^{0.8}}{W}$   
 985  
 986



987  
 988 **Figure 12:** Drainage areas of the catchments draining the emerging thrust sheet 4 of experiment A2 (rainfall 9  
 989 mm/h) at  $t=232$  min (A) and at  $t=298$  min (B). The channels 1 to 6 are pre-existing transverse channels. The  
 990 antecedent channel 5 managed to cut through the uplifting thrust sheet 4 (persistent channel). Channels 1', 2', 3' and  
 991 4' are the downstream parts of the small catchments 1, 2, 3 and 4 that lost their upstream drainage area by diversion  
 992 above the thrust sheet the 4 (defeated channels). Their upstream drainage areas are gathered together in the  
 993 longitudinal reach of channel 5 in the backlimb of the active thrust 4.  
 994  
 995

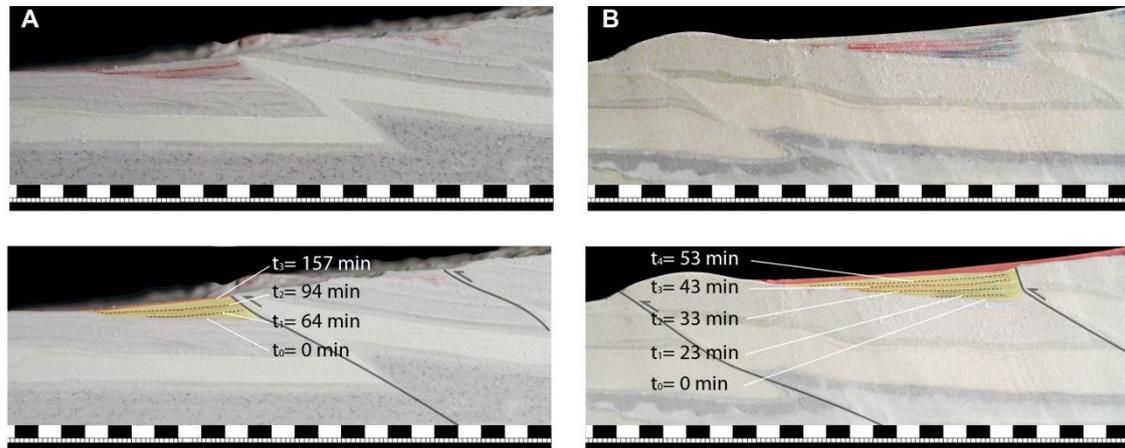


996  
 997 **Figure 13:** (A) Proportion of persistent transverse channels against uplift rates. Note the decrease of the proportion  
 998 of persistent transverse channels with the increase of the uplift rate in both series of experiments run under different  
 999 rainfall rates. We obtained a good fit ( $R^2=98\%$ ) using a power law regression  $R=7.3 \cdot U^{-0.95}$ . (B) Proportion of  
 1000 persistent antecedent channels against precipitation over shortening rate (labels are the same as in (A)). We obtained  
 1001 a good fit ( $R^2=98\%$ ) using a power law regression  $R=35 \cdot (\text{Prec./Short.})^{0.65}$ .  
 1002



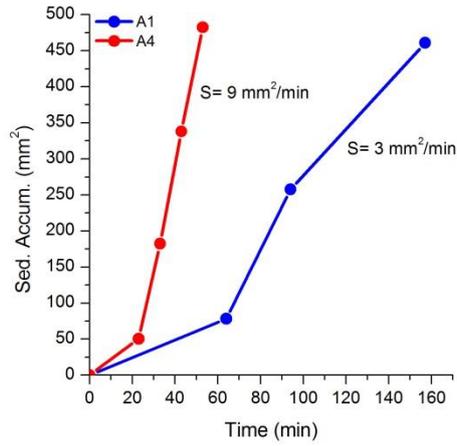
1003  
1004  
1005  
1006  
1007  
1008

**Figure 14:** Patterns of sedimentation in two experiments run under low (A) and high (B) shortening rates (rainfall is 9 mm/h). (C) Experiment A1, the high proportion of persistent transverse channels ( $R=75\%$ ) results in a line-source dispersal system. (D) Experiment A4, the low proportion of persistent transverse channels ( $R=20\%$ ) results in 2 large point-sourced depositional systems plus some small fan systems.



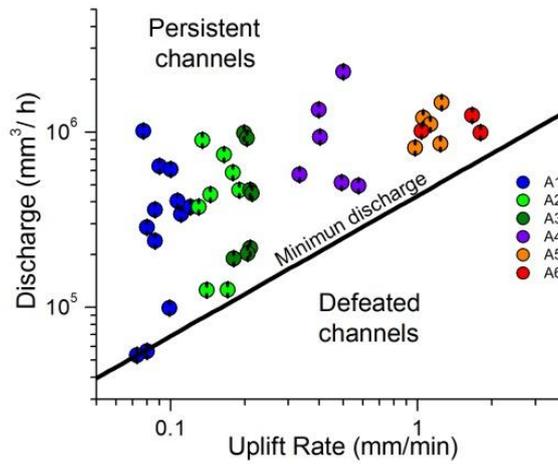
1009  
1010  
1011  
1012  
1013  
1014

**Figure 15:** (A) Detail cross-section of the sedimentation zone in front of thrust 4 in the model A1. (B) Detail cross section of the sedimentation zone in front of thrust 5 in the model A4. Photos are interpreted below and the traces of the cross sections are located in figure 14. The areas of the sedimentation zones in yellow accumulated during thrust activity and the deposits in red accumulated after thrust activity. Note that the second sequence of sediments (in red) seals the inverse faults.



1015  
1016  
1017  
1018

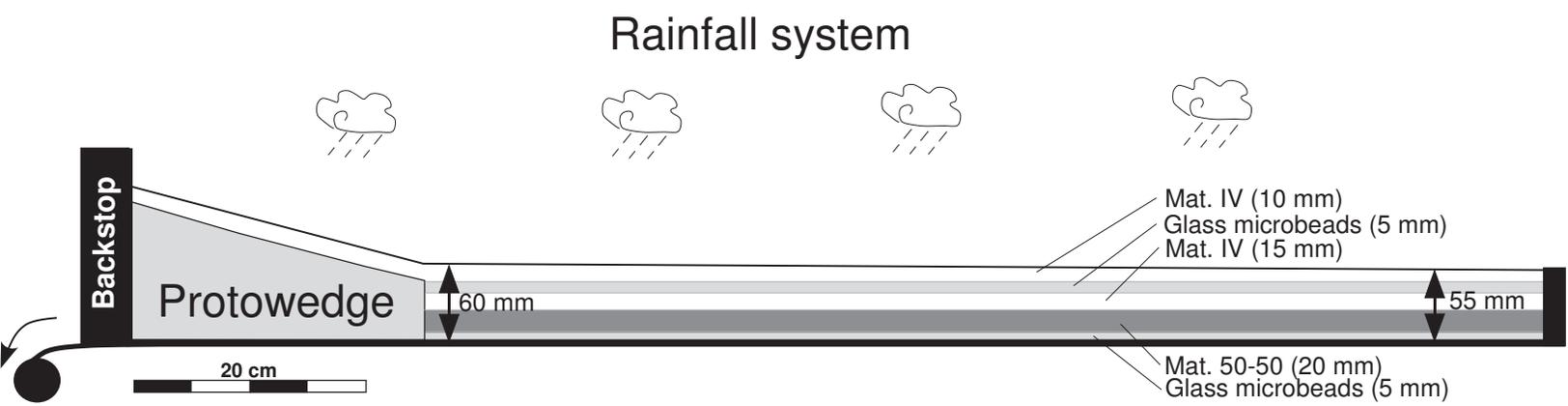
**Figure 16:** Mean rates of aggradation ( $S$ ) calculated from the area of sedimentation measured in the cross-sections presented in figure 15.



1019  
1020  
1021  
1022  
1023  
1024

**Figure 17:** Discharge of persistent transverse channels against uplift rate. Note that the minimum discharge needed to overcome a new uplifting thrust sheet increases with uplift rate. This minimum discharge increases slightly exponentially with uplift rate:  $Q_w \propto U^{1/m}$  with  $m=0.8$ .

Figure



Figure

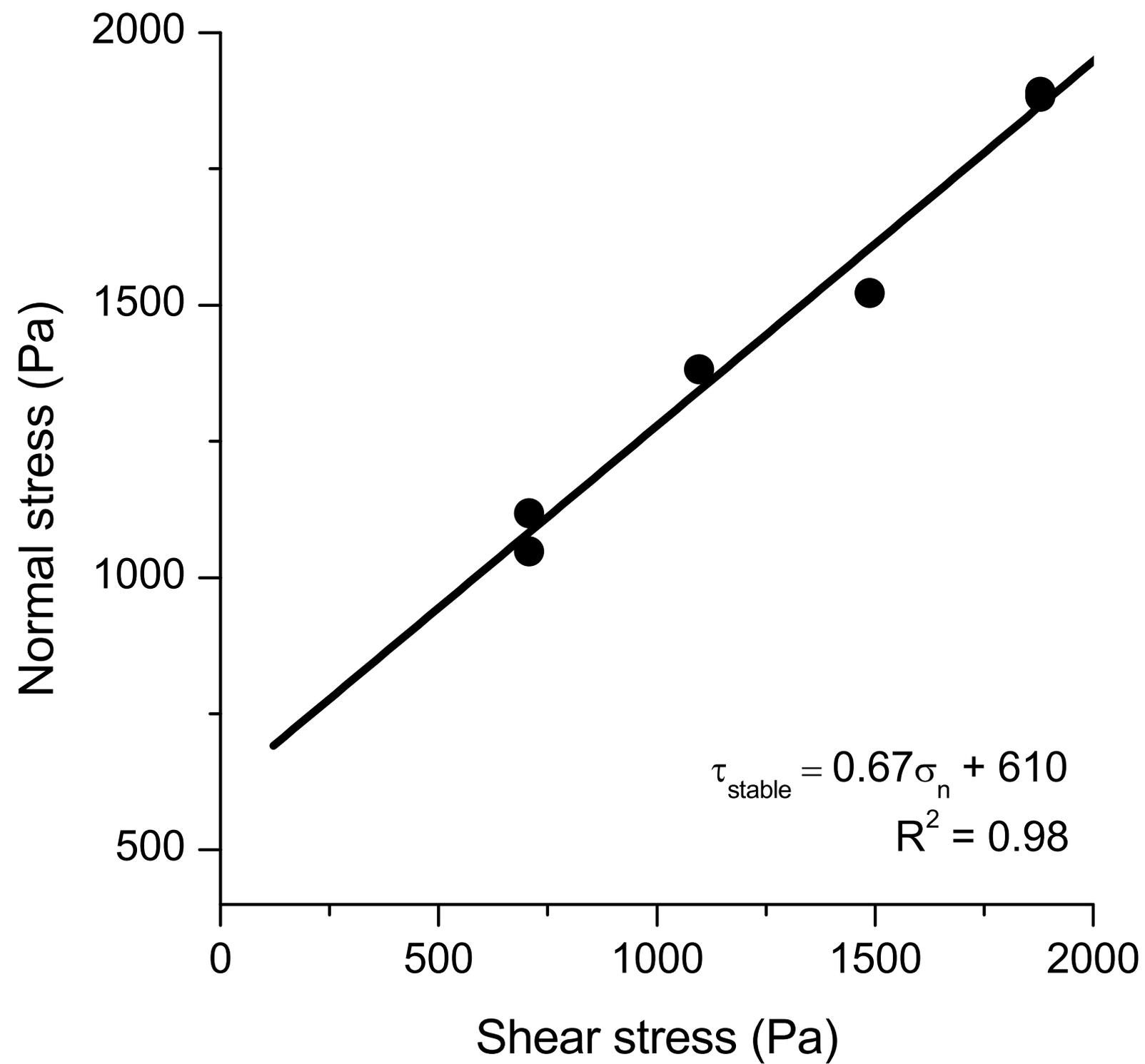
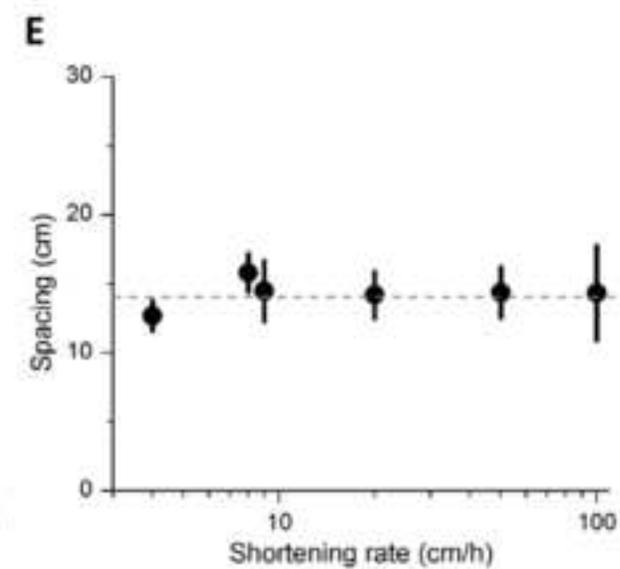
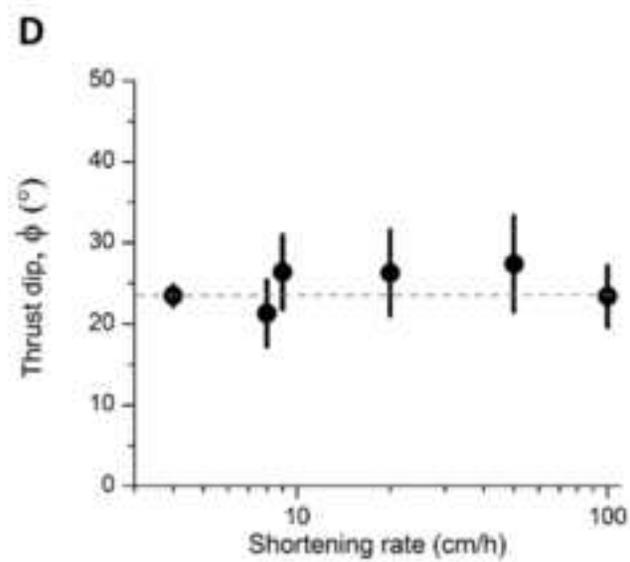
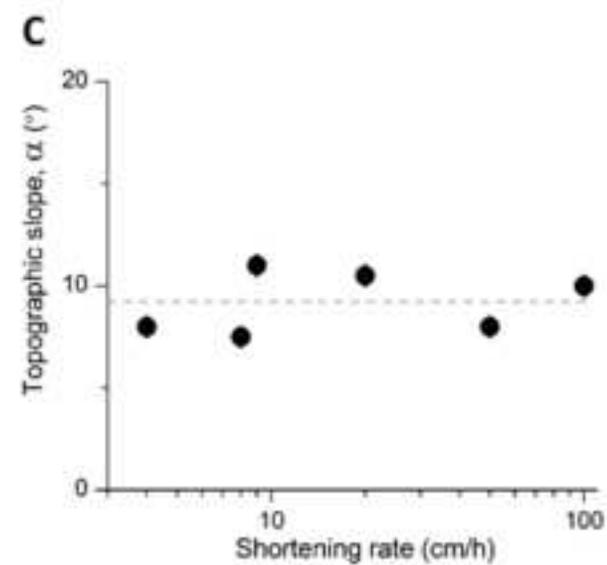
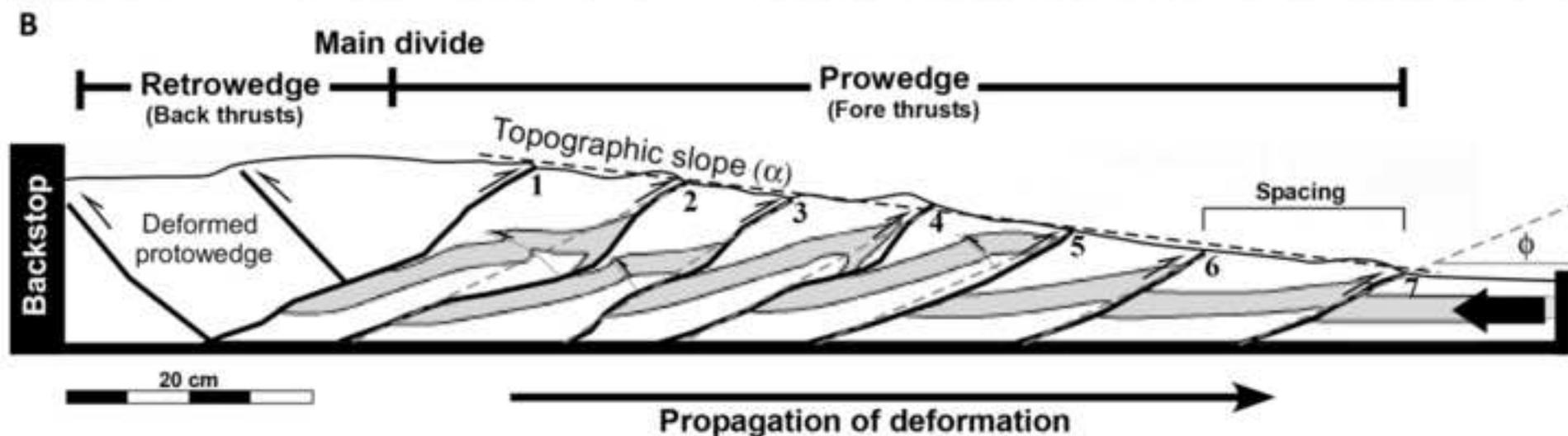
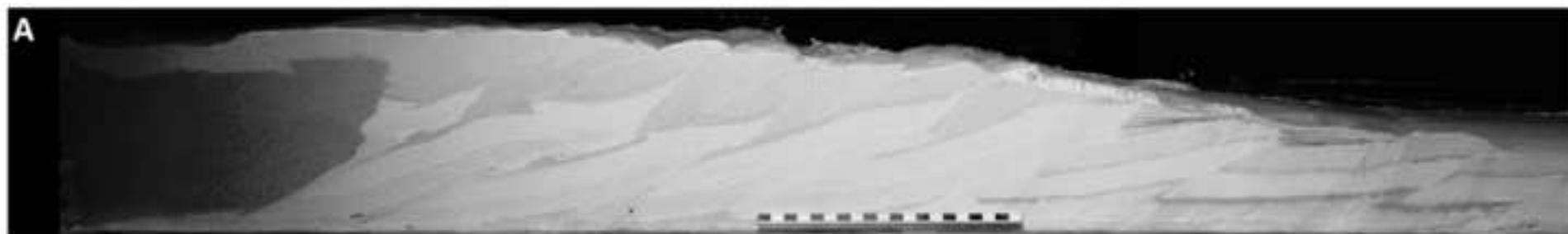
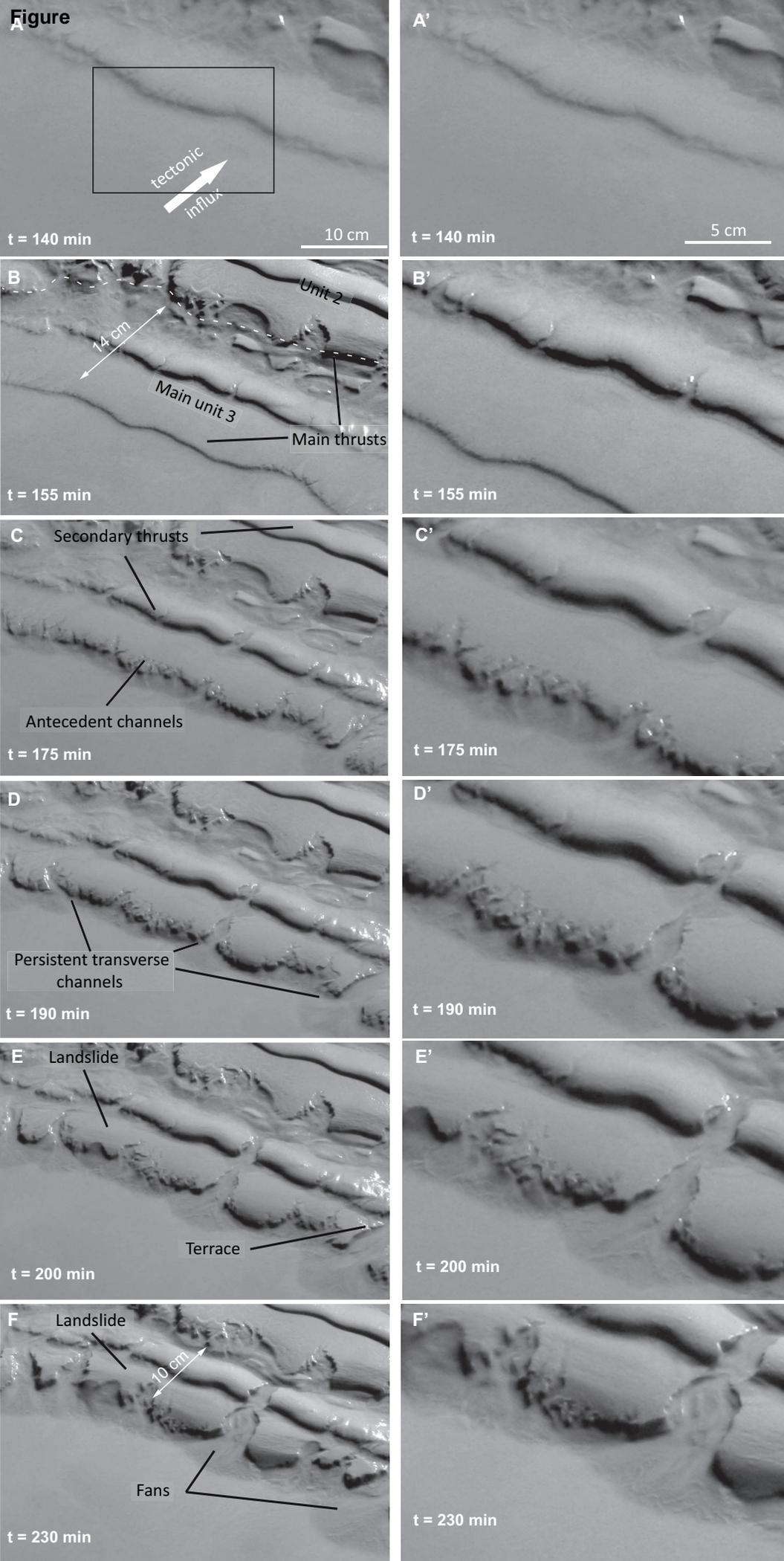


Figure  
[Click here to download high resolution image](#)

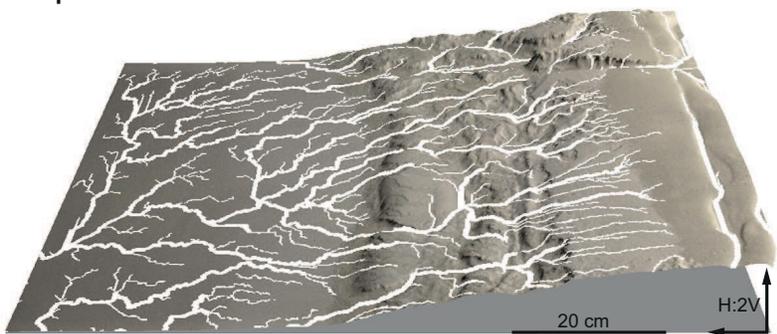


Figure

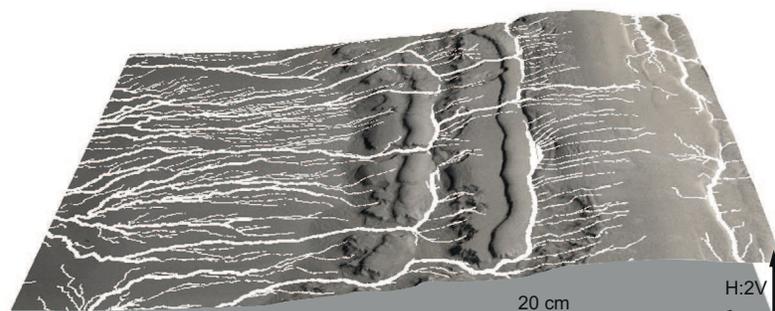


Figure

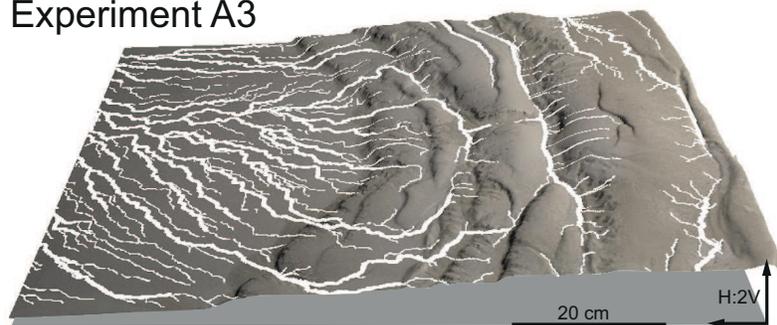
Experiment A1



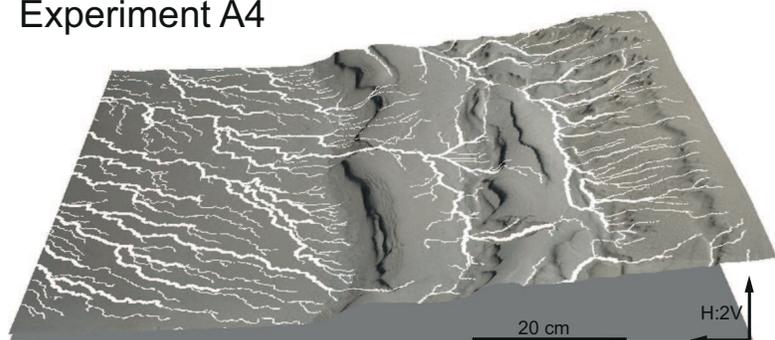
Experiment A2



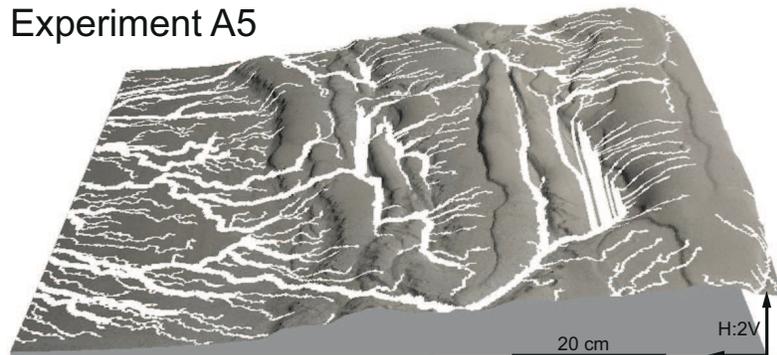
Experiment A3



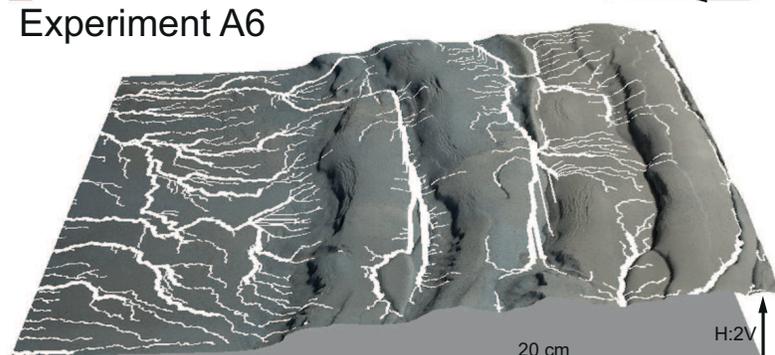
Experiment A4



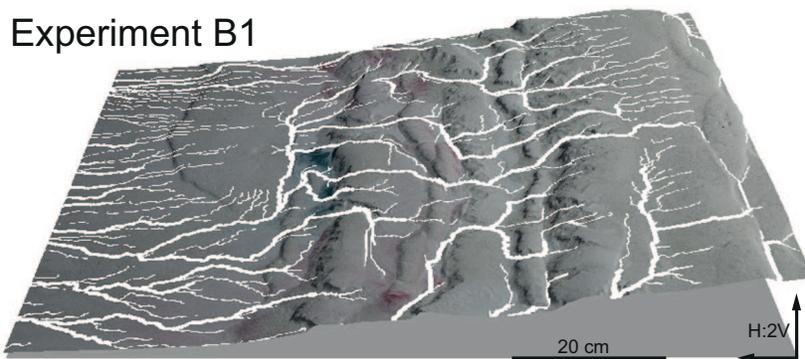
Experiment A5



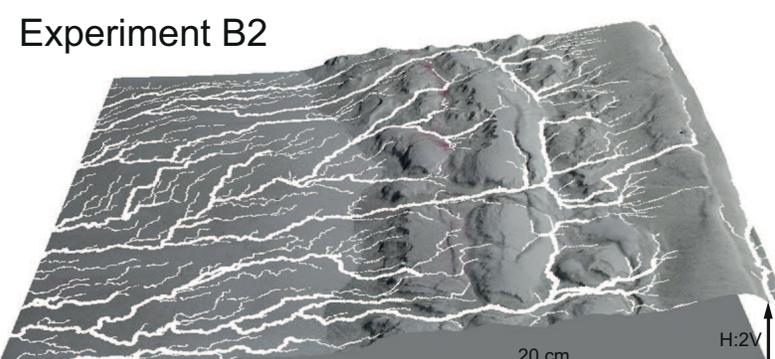
Experiment A6



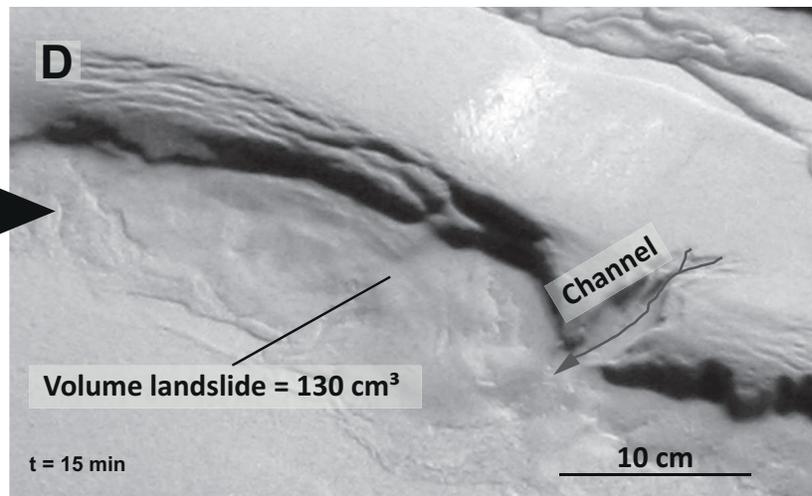
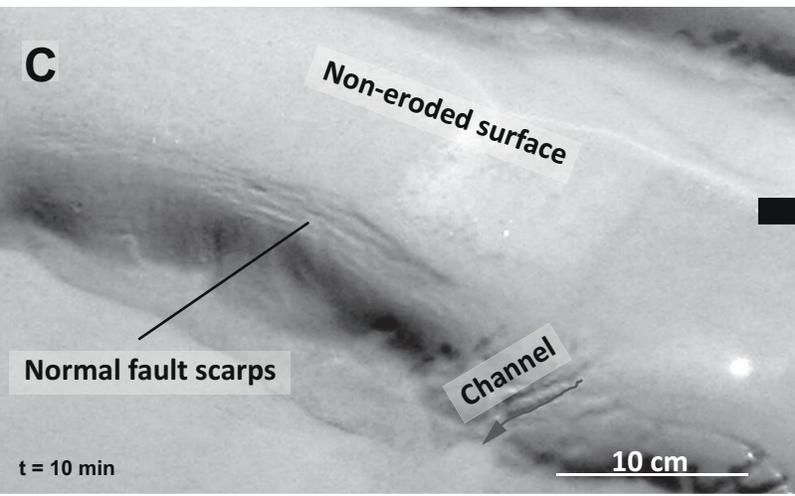
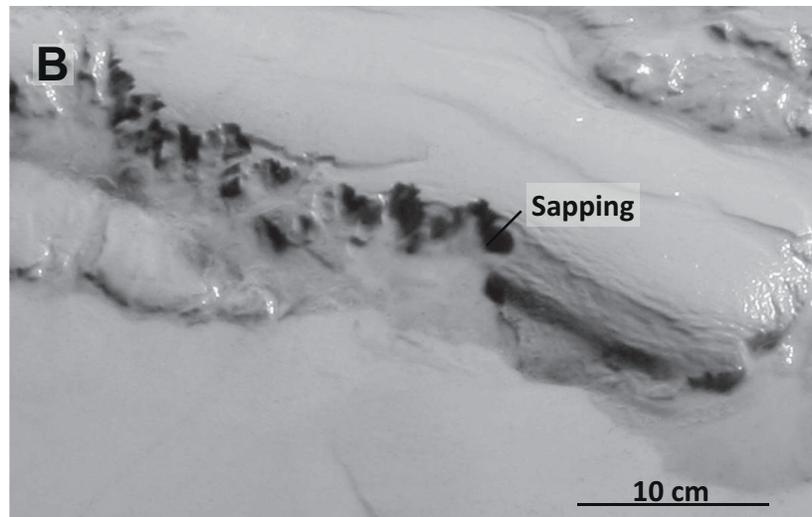
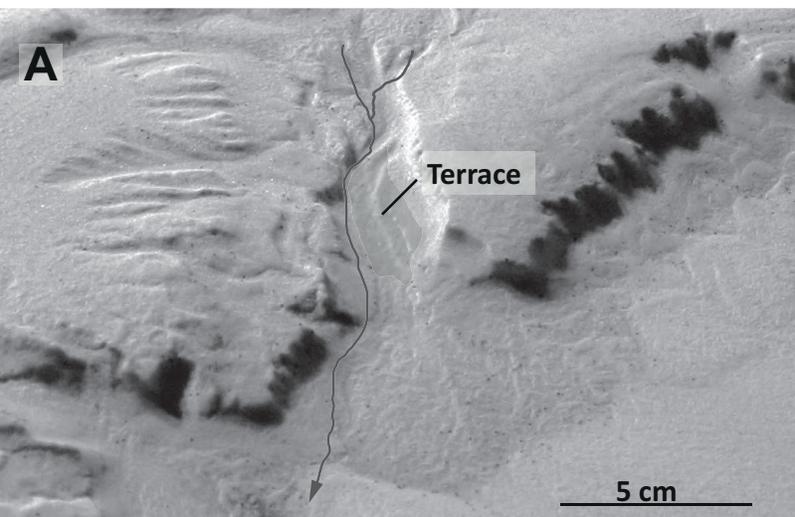
Experiment B1



Experiment B2



Figure



Figure

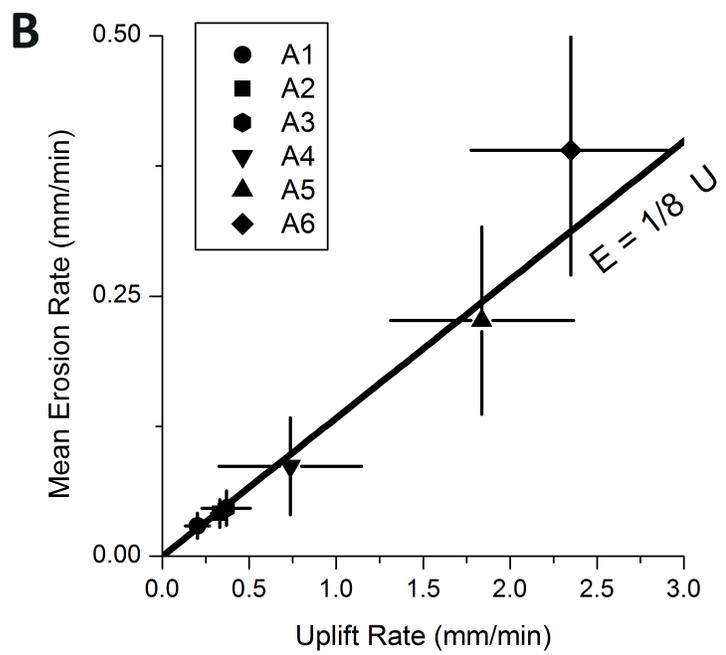
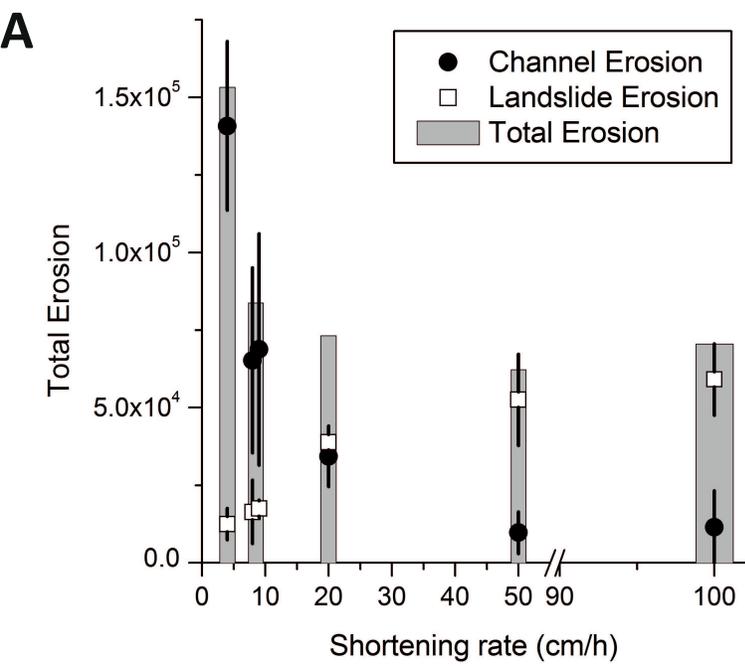
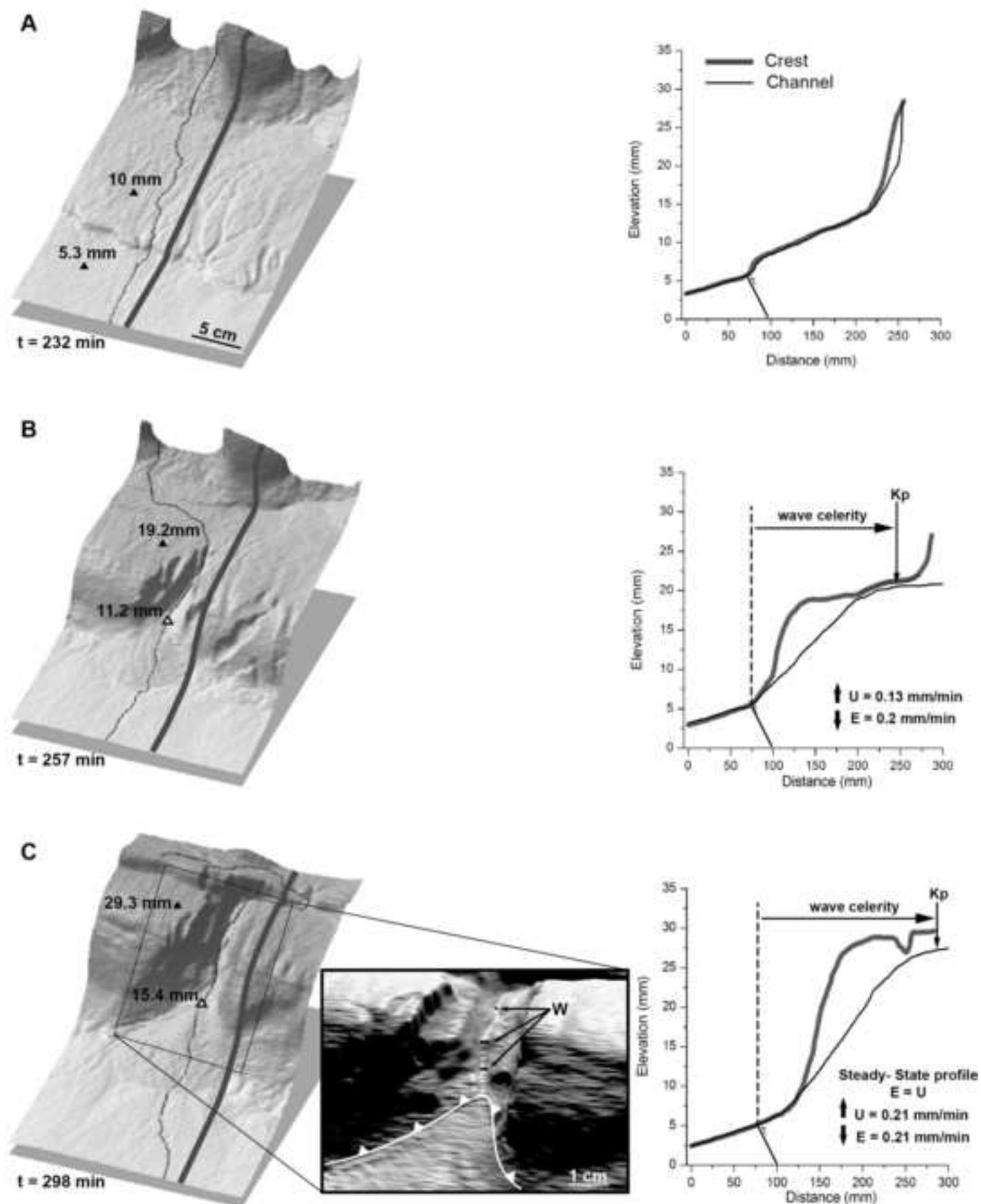
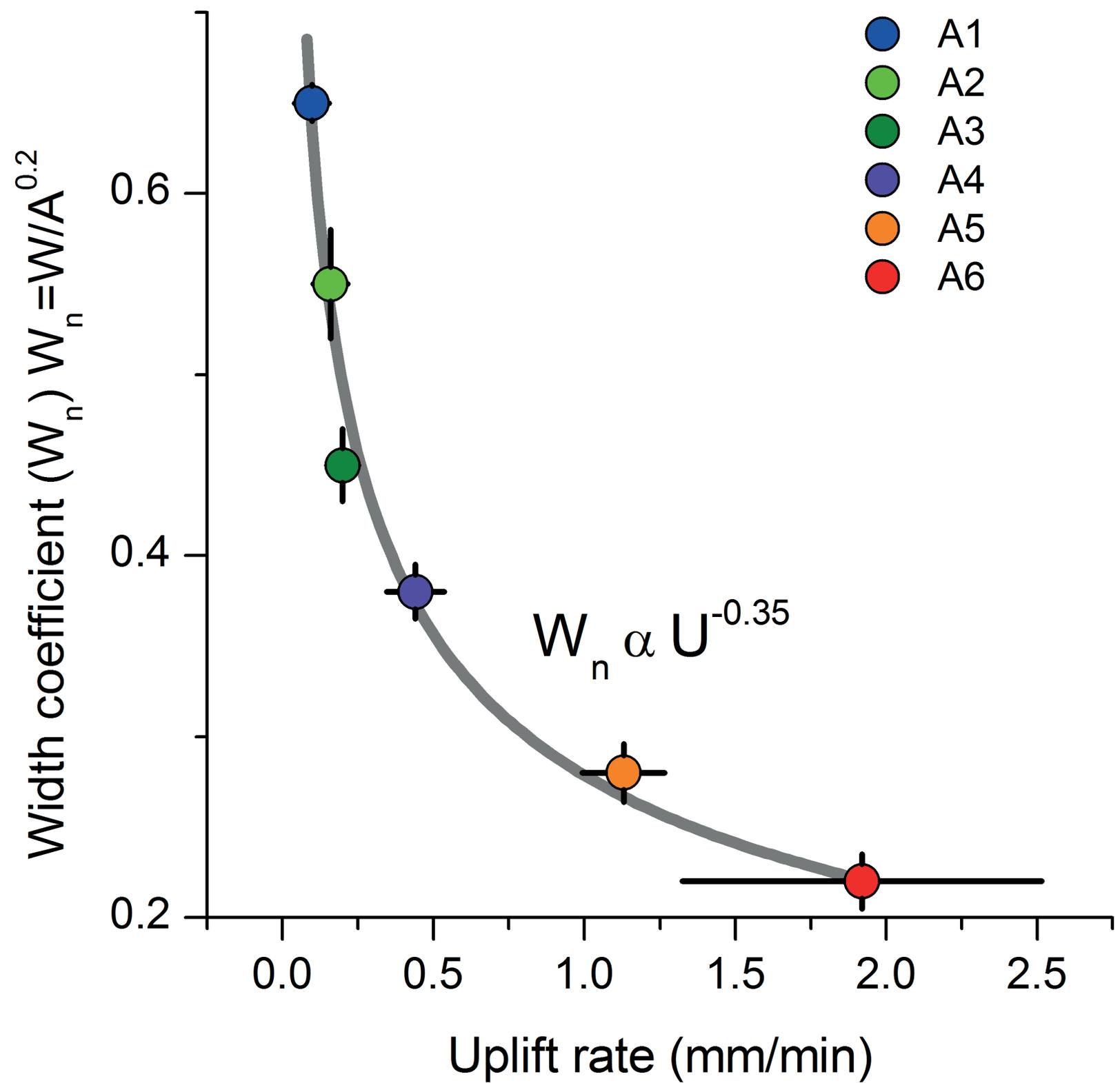


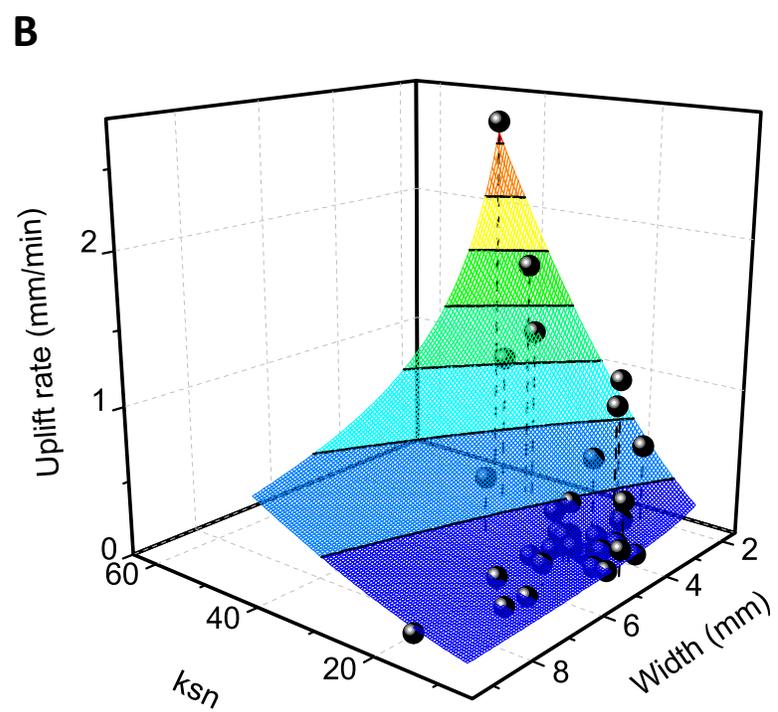
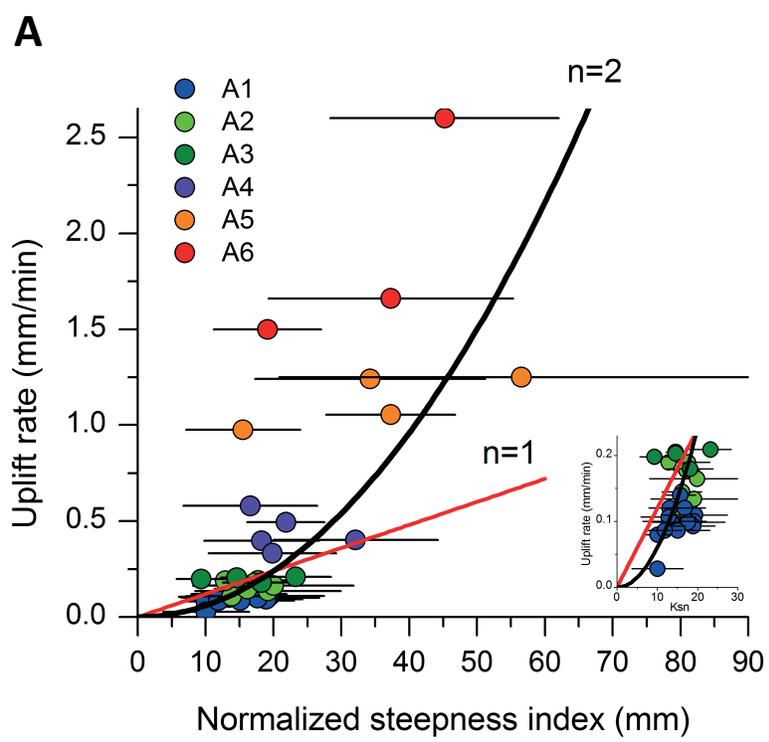
Figure  
[Click here to download high resolution image](#)



Figure

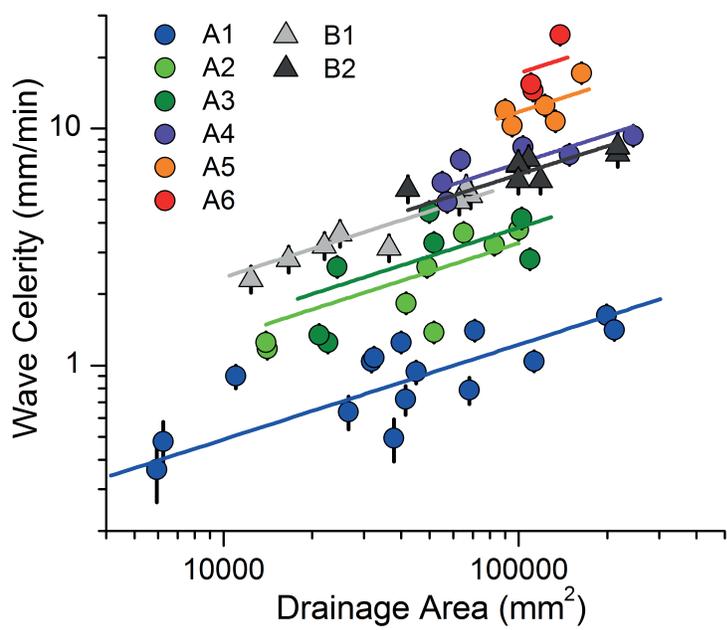


Figure

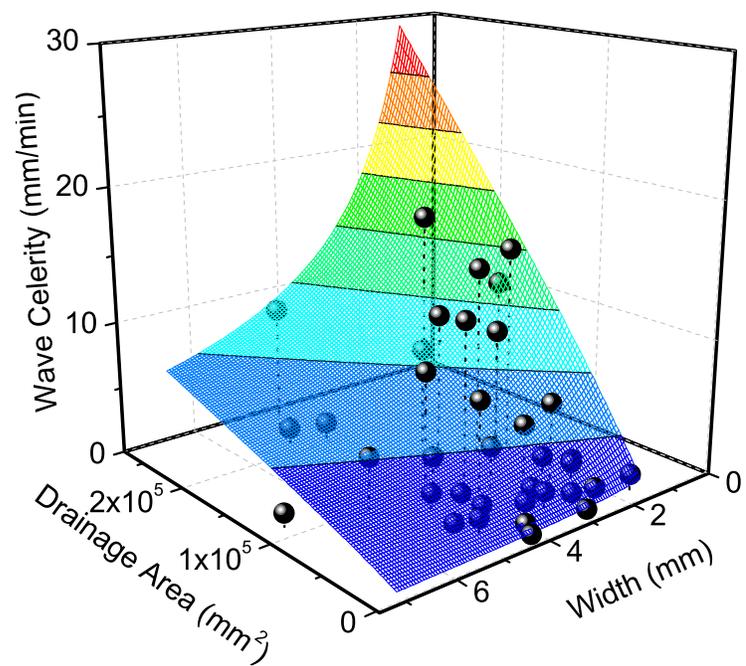


Figure

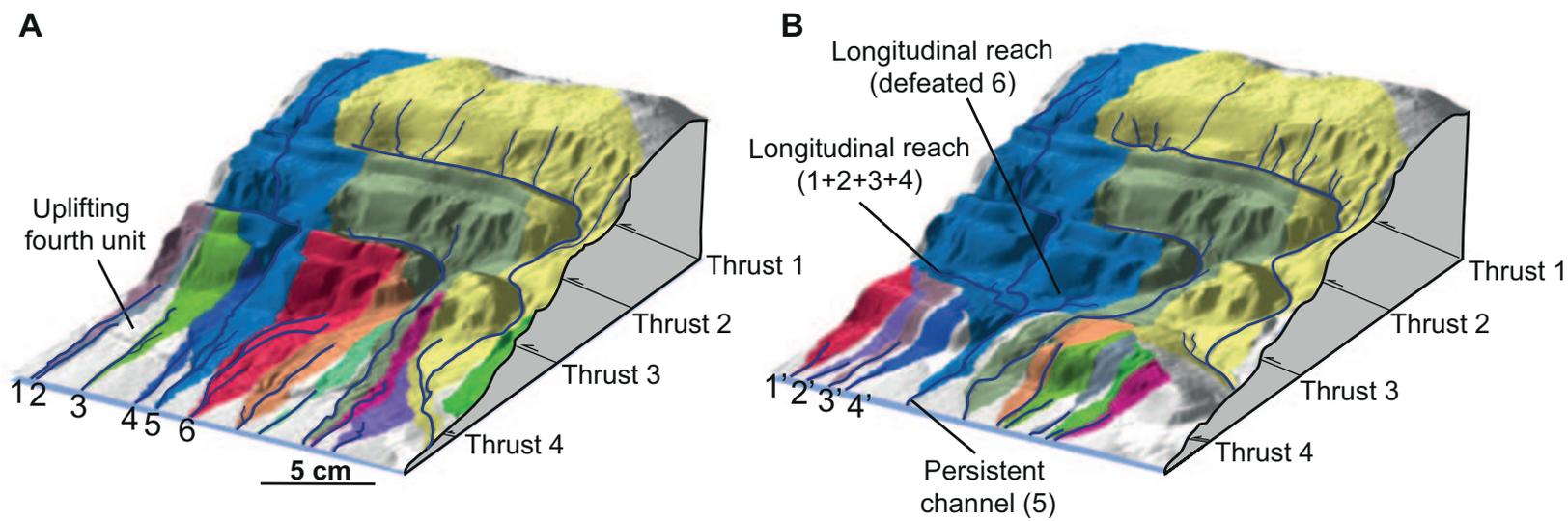
**A**



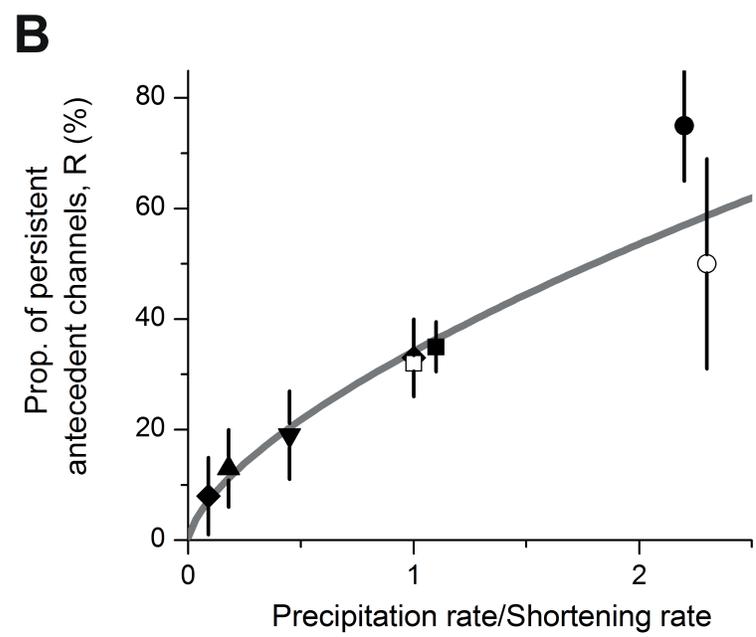
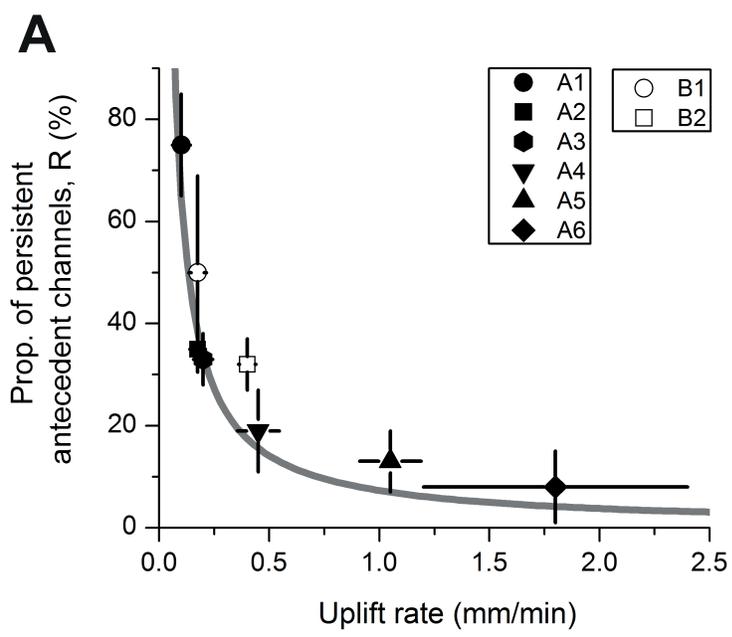
**B**



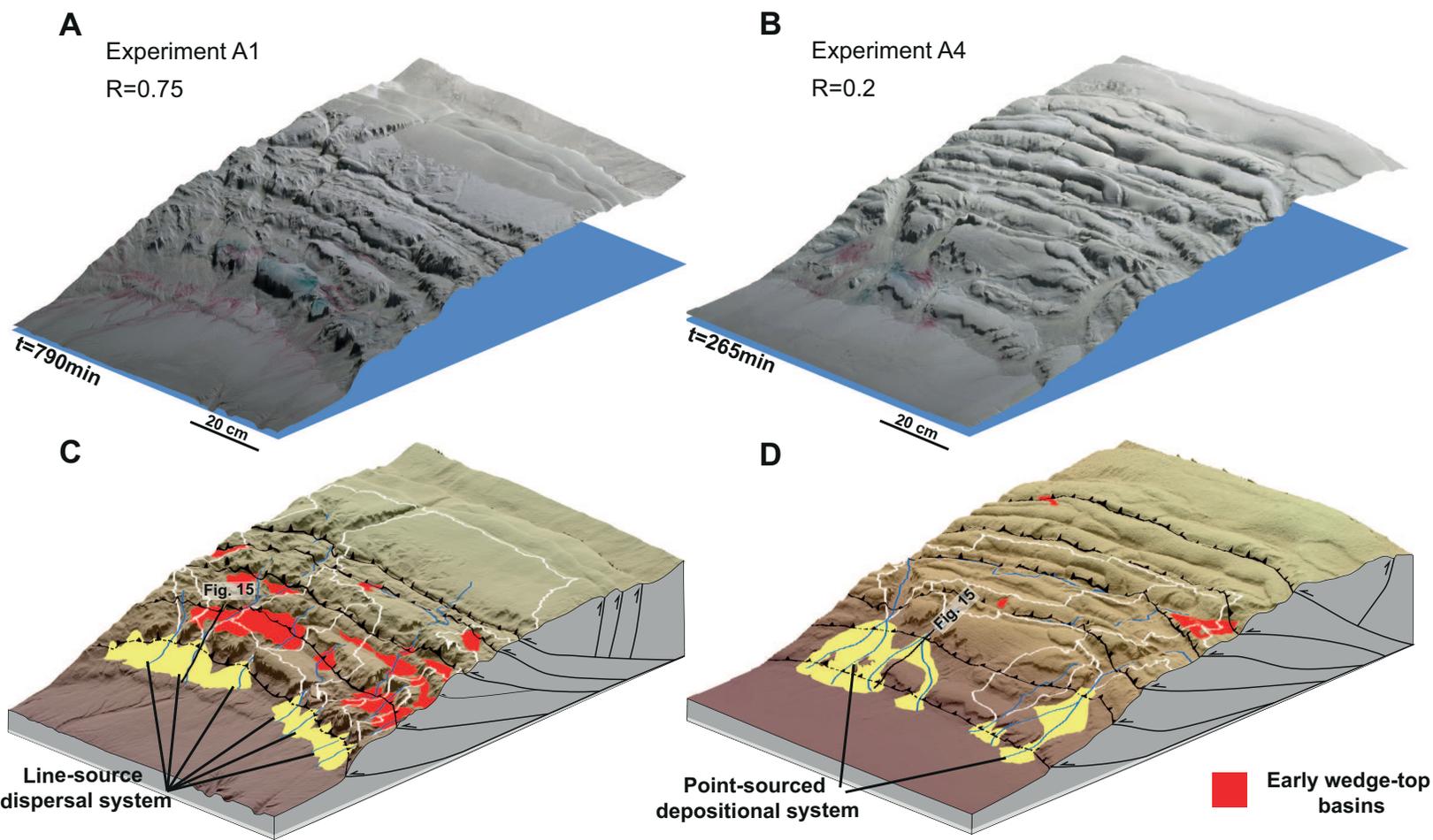
Figure



Figure

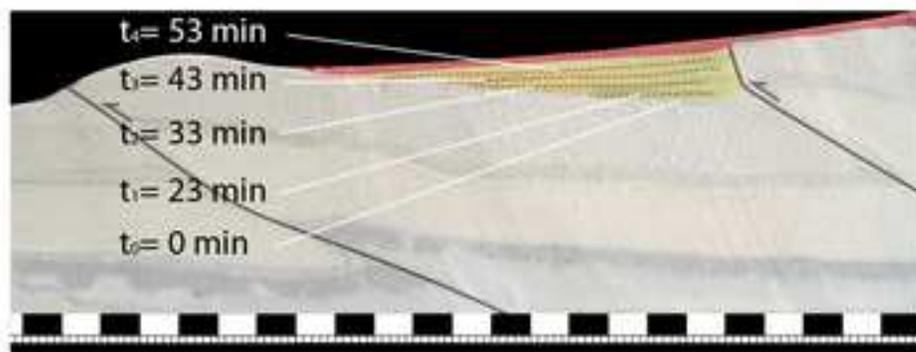
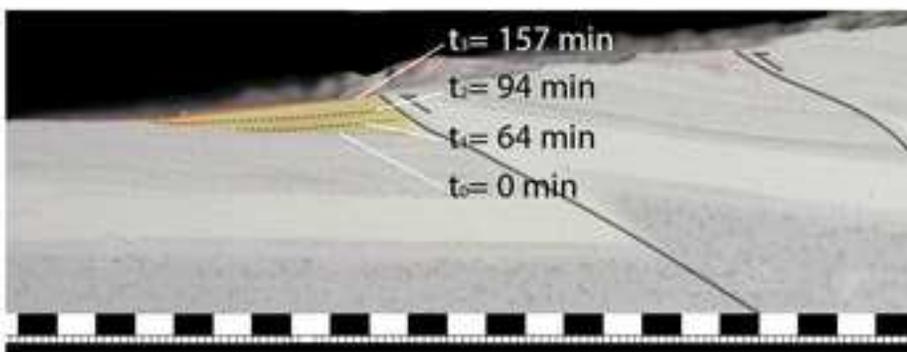
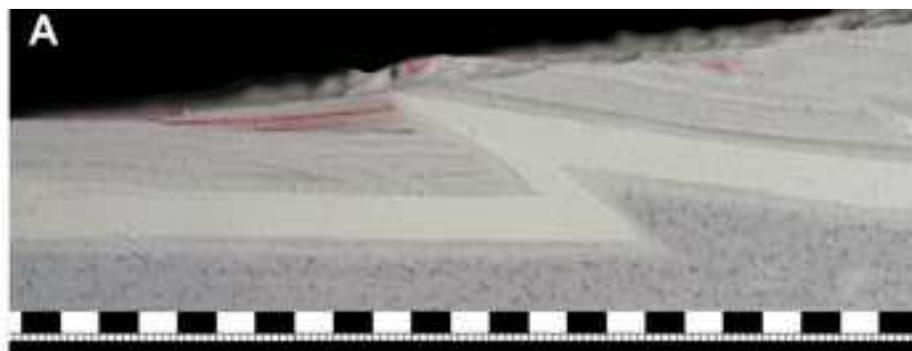


Figure

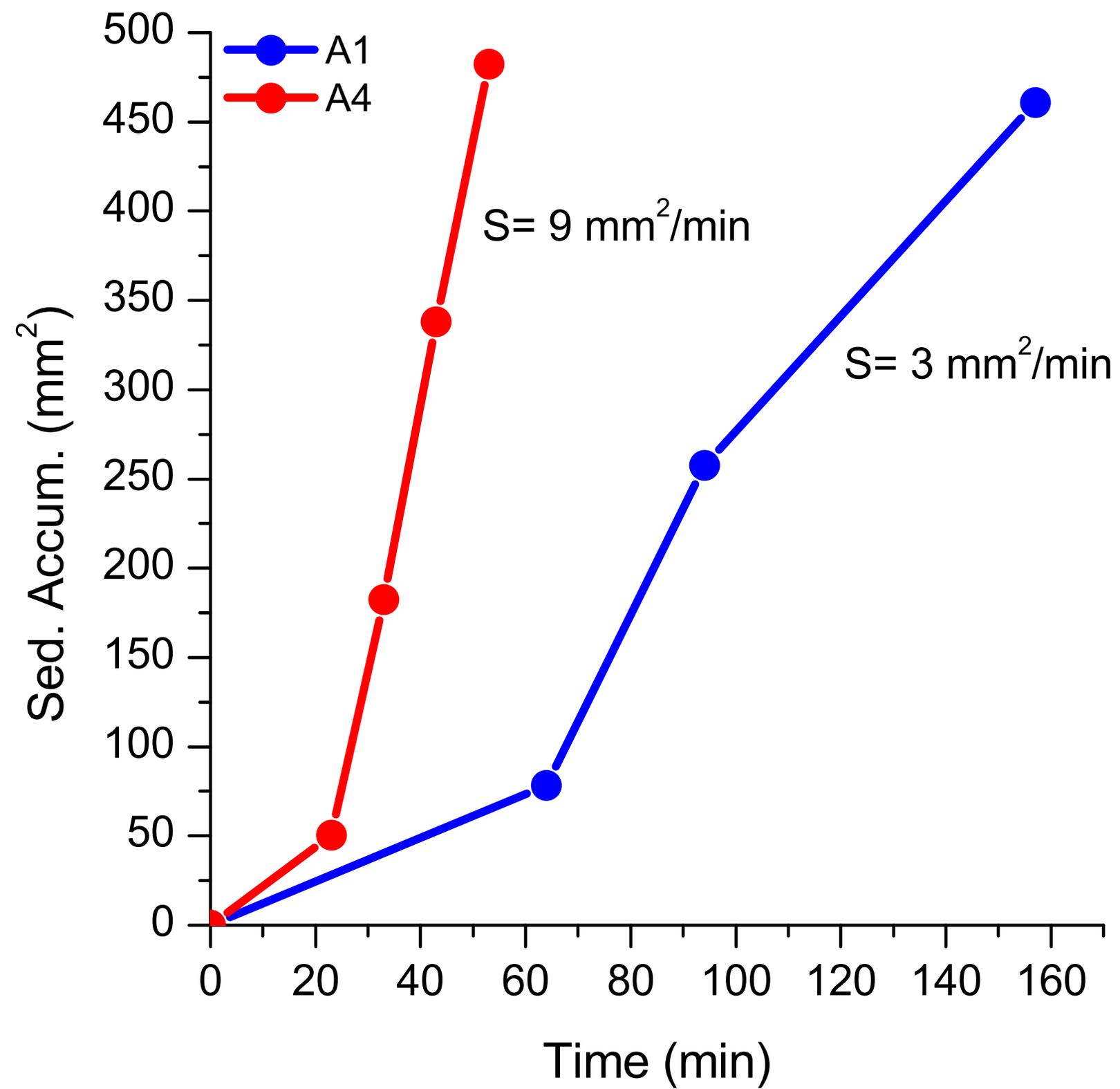


Figure

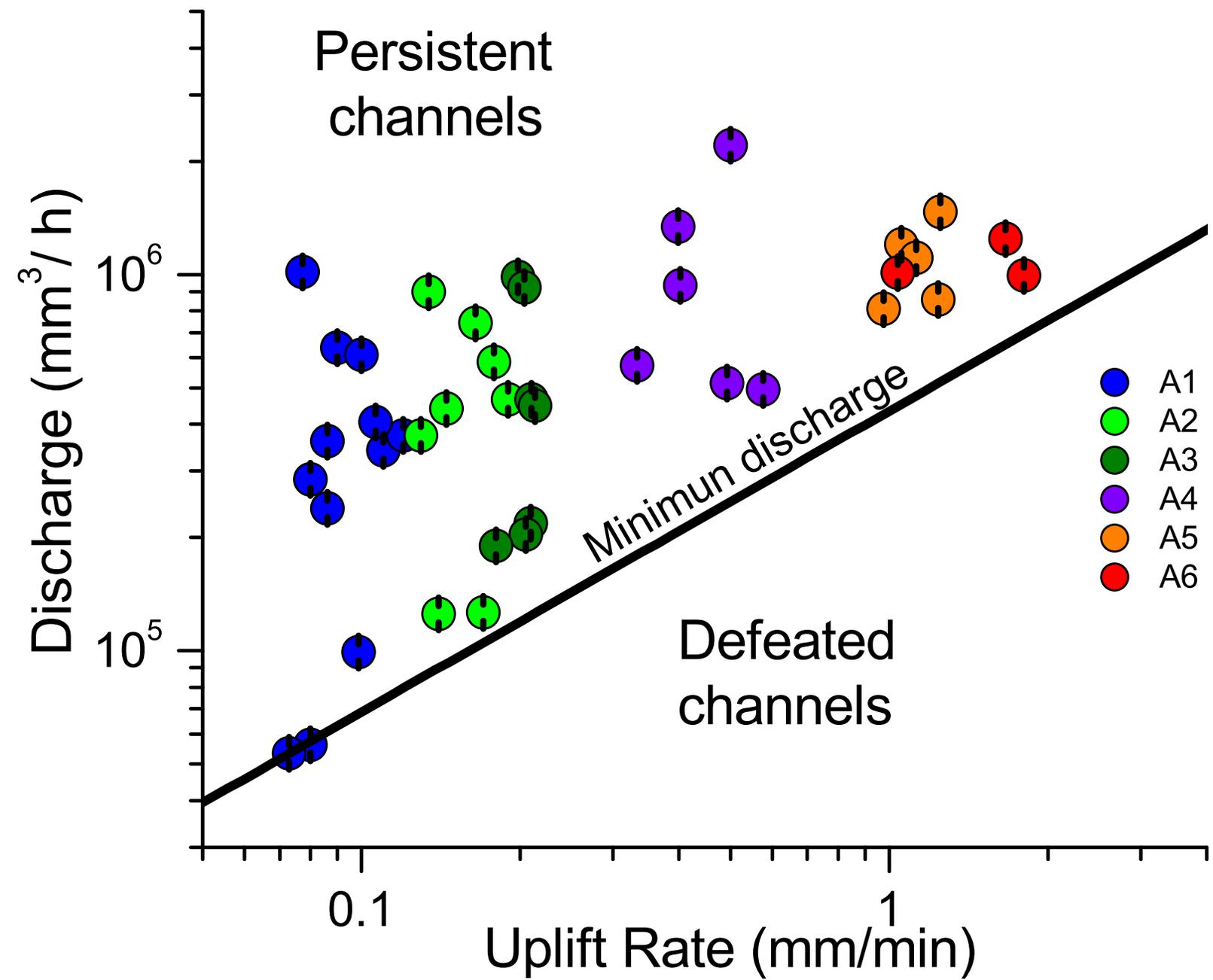
[Click here to download high resolution image](#)



Figure



Figure



**Supplementary material for online publication only**

[Click here to download Supplementary material for online publication only: Animation1.mov](#)

**Supplementary material for online publication only**

[Click here to download Supplementary material for online publication only: Supplementary\\_Table.xlsx](#)