

1 **Morpho-tectonics of transpressional systems: insights from analog**
2 **modeling**

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14 **Key Points:**

- 15 • Feedback between fault and drainage network development regulates the deformation,
16 exhumation, and morphology of transpressional systems
17 • The progression from distributed deformation to full strike-slip strain partitioning is
18 accelerated in more erosive systems
19 • Exhumation in a transpressional wedge is maximized along the master fault and axial
20 valley due to heightened rock uplift and incision

21 **Abstract**

22 Transpressional margins are widespread, and their dynamics are relevant for plate boundary
23 evolution globally. Though transpressional orogen evolution involves a topographic response to
24 deformation, many studies focus only on the structural development of the system ignoring surface
25 processes. Here, we present a new set of analog models constructed to investigate how tectonic
26 and surface processes interact at transpressive plate boundaries and shape topography.
27 Experiments are conducted by deforming a previously benchmarked crustal analog material in a
28 meter-scale plexiglass box while controlling erosion through misting nozzles mounted along the
29 transpressional wedge. We used a laser scanner to generate digital elevation models throughout
30 the model evolution and photos for particle image velocimetry analysis. We focus on three
31 experiments that cover a range of rainfall and convergence settings, with two end-member erosion
32 settings and a dry reference. In all experiments, a bivergent wedge forms, and strain partitioning
33 broadly evolves according to previously established models. Regarding drainage networks, we
34 find that the streams in our models develop differently through feedback between fault
35 development and drainage rearrangement processes. Differences between end-member erosional
36 models can be explained by the varying response of streams to structure modulated by rainfall.
37 Additionally, erosion may influence the structural evolution of transpressional topography, leading
38 to accelerated strike-slip partitioning. From these results, we create a model for developing
39 structures, streams, and topography where incision and valley formation along main structures
40 localize exhumation. We apply insights from the models to natural transpressional systems,
41 including the Transverse Ranges, CA., and the Venezuelan Andes.
42

43 1. Introduction

44 Coupling between tectonics and surface processes may affect the localization of deformation
45 and morphological evolution of orogenic systems (e.g., Burbank & Anderson, 2011; Graveleau et
46 al., 2015; Koons, 1995; Molnar & England, 1990; Willett, 1999). When orogenesis is accompanied
47 by a degree of obliquity, the resultant deformation is termed transpression, describing the pairing
48 of wrenching and thrusting structures to accommodate strain (Sanderson & Marchini, 1984). In
49 natural transpressional systems, tectonic strain may be partitioned so that a single vertical strike-
50 slip fault or pairs of strike-slip faults oriented sub-parallel to the zone boundary accommodate the
51 wrench component of oblique convergence (Teyssier et al., 1995). Since most plate boundaries are
52 oblique ($> 10^\circ$ obliquity; Philippon & Corti, 2016), understanding the erosion-tectonic feedback
53 and its relationship with strain partitioning in such settings is essential to accurately constrain,
54 interpret, and model the evolution of the crust and surface.

55 Recent field observations from transpressional settings suggest that climatic variability may
56 affect deformation patterns, exhumation, and topographic change around major faults (Cochran et
57 al., 2017; Cruz et al., 2007). The stream network response to such change may also vary depending
58 on precipitation and bedrock erodibility (Reitman et al., 2022). Generally, faults control drainage
59 geometries through entrainment (Chorley et al., 1984; Koons, 1994, 1995) and preferential incision
60 by mechanical weakening (Koons, 1994, 1995). These mechanisms are important in orogenic
61 systems since fluvial incision is a primary driver of mass transfer. However, a general
62 understanding of how stream networks and fault structure modify the morphotectonic evolution of
63 a transpressional wedge remains to be established.

64 “Erosion-tectonic” sandbox models provide valuable insight into transpressional systems by
65 combining tectonic deformation and surface mass transport using appropriate analog materials and
66 misting systems that realistically simulate the erosional processes acting on a deforming wedge
67 (Guerit et al., 2016; Guerit et al., 2018). Previous erosion-tectonic models have been used to study
68 the passive rotational response of drainages to oblique convergence (Guerit et al., 2016) and the
69 transient nature of landscapes under transpression (Guerit et al., 2018). Observations from these
70 studies show that streams have a predictable response to deformation in the absence of
71 confounding variables and can be used to characterize deformation in an oblique wedge.
72 Furthermore, analog models by Malavieille et al. (2021) showed that mass transfer by erosional
73 processes could influence the location of major faults, the topographic response to internal
74 deformation partitioning, and, therefore, the long-term evolution of the wedge.

75 Here, we present erosion tectonic sandbox experiments that investigate the relationships
76 between fault structure, stream networks, and the strain field in transpressional systems. We
77 attempt to identify the potential feedback between these components to explain morphological and
78 deformational differences between experimental wedges for high- and low-erosion endmembers.
79 Through analyses of digital elevation models and velocity fields from particle tracking, we address
80 1) how stream networks evolve in transpressional systems under variable erosional conditions, 2)
81 if and how erosion influences the structural and morphological evolution of transpressional
82 mountain belts, and 3), how strain partitioning evolves and is affected by structural and stream
83 network development. The components of wedge evolution related to these questions are
84 highlighted in Figure 1. We extend our results and analyses to natural transpressional prototypes,

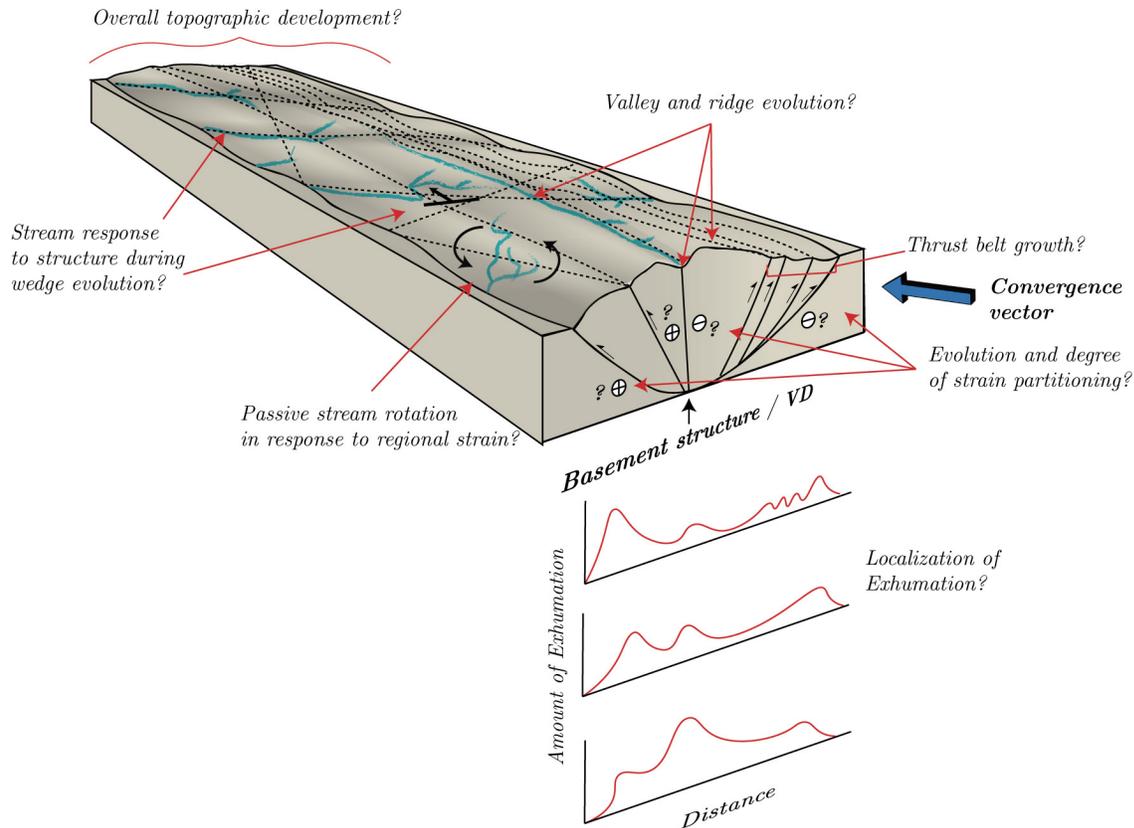


Figure 1. Illustration highlighting the unresolved components of the growth of a transpressional (left-lateral) wedge addressed in this paper. The velocity discontinuity (VD) and convergence vector are in bold font, dashed lines on topography show fault traces. The x-y plots show hypothetical exhumation patterns (red lines) across the wedge. Examples of stream responses to structure shown include drainage deflection (black arrow showing offset), headward erosion, and entrainment along faults. Black rotation arrows indicate the direction of rotation in a left-lateral transpressional orogen.

85 mainly focusing on the Merida Andes of Venezuela and the central Transverse Ranges along the
86 San Andreas fault system in California, U.S.

87 2. Analog model: Erosion–tectonics sandbox

88 Previous analog studies of the evolution of transpressional mountain belts focus on the
89 structural development of the model without including surface processes (e.g., Barcos et al., 2016;
90 Lallemand et al., 1994; Leever et al., 2011a; Leever et al., 2011b; Pinet & Cobbold, 1992). Some
91 workers have conducted laboratory studies that included the effects of erosion and sedimentation
92 by removing and applying material by hand (e.g., Bonnet et al., 2007, 2008; Konstantinovskaia &
93 Malavieille, 2005; J. Malavieille et al., 1993; Perrin et al., 2013). However, this approach limits
94 the internal control of the system. Only a few models combine tectonic stresses and surface
95 processes using misting systems that more realistically simulate the erosional processes acting on
96 a deforming wedge (Graveleau et al., 2015; Graveleau & Dominguez, 2008; Guerit et al., 2016,
97 2018; Lague et al., 2003; Mao et al., 2021; Reitano et al., 2022; Viaplana-Muzas et al., 2015,
98 2019). These “erosion–tectonic” laboratory studies are often limited to purely compressional or
99 extensional settings with few strike-slip (e.g., Graveleau et al., 2015) or transpressional (e.g.,
100 Guerit et al., 2016; Guerit et al., 2018) investigations. In the presented experiments, we add to the

101 current collection of erosion-tectonic studies of transpression and aim to understand how
102 deformation and wedge morphology evolve under the influence of structural and fluvial
103 mechanisms in different erosional regimes.

104 *2.1 Experimental material*

105 Analog materials used in erosion-tectonic experiments should account for the first-order
106 deformational and erosional behavior of the lithosphere (e.g., Graveleau et al., 2011). In addition,
107 the material should scale appropriately, demonstrating geometric, kinematic, and dynamic
108 similarity (Hubbert, 1951). Many granular single component (e.g., crushed quartz, silica powder)
109 and composite materials (e.g., Mat IV or CM2) have been tested and shown to behave similarly to
110 natural cases in a variety of geodynamic experiments (Graveleau et al., 2011). Our material, CM2,
111 is a combination of 40 wt. % glass microspheres, 40 wt. % silica powder, and 20 wt. % PVC
112 powder (Reitano et al., 2020). Reitano et al. (2020) characterized CM2 following the work of
113 Graveleau et al. (2011), who developed a similar material, Mat IV. Mat IV has the same
114 composition as CM2, yet 2 wt. % graphite powder and 18 wt. % PVC. These authors show that
115 CM2 and Mat IV deform following the Mohr-Coulomb failure criterion, exhibit natural basin and
116 channel characteristics, and appropriately balance hillslope diffusion and channel incision. They
117 are also velocity-weakening materials leading to stick-slip. Since Reitano et al. (2020) only report
118 the hydrated frictional behavior (20 wt. % H₂O) of CM2, we conduct rotary shear tests to
119 characterize its dry frictional behavior. These tests were conducted using an energy controlled
120 rotary shear apparatus (see Conrad et al., 2023) fitted with a granular sample holder (Conrad et al.,
121 2020).

122 *2.2 Experimental setup*

123 We conduct experiments in a 2 m × 1 m × 0.5 m plexiglass box, with ends left open for
124 drainage (Figure 2a). The basal slope is fixed at 1° to ensure water exits the system. We set a mylar
125 sheet inside the box and fix a plexiglass board cut to the desired obliquity to the sidewall. By
126 pulling the mylar sheet beneath the board, we simulate oblique convergence (Figure 2b). We load
127 the board-sheet set up with a ~5 cm thick package of the experimental material hydrated to ~20
128 wt. % water (see section 2.1). The length and width of the material package are controlled to ensure
129 that the edges do not influence the wedge's evolution or reach the sidewalls of the box. Free
130 boundaries are particularly important on the fixed side of the model as it allows the wedge to form
131 independent of the geometry of a rigid backstop (e.g., Guerit et al., 2016). This independence arises
132 because the material properties, rather than the backstop dip, control the geometry of the wedge.
133 Additionally, the thrust can propagate beyond the location of a would-be backstop.

134 We use the velocity discontinuity (VD) between the fixed board and moving sheet to
135 localize deformation, forming a bivergent wedge in the material package. This approach is similar
136 to Leever et al. (2011a,b) and the classic wrench experiments of (Riedel, 1929), where the VD
137 simulates a basement fault beneath a homogenous sediment cover. We initiate surface processes
138 using misting nozzles mounted on an aluminum crossbar aligned with the wedge trend. These
139 nozzles maintain a droplet size of fewer than 100 μm to avoid rain splash erosion (Bonnet et al.,
140 2007; Graveleau et al., 2012; Lague et al., 2003; Reitano et al., 2022; Viaplana-Muzas et al., 2015,
141 2019).

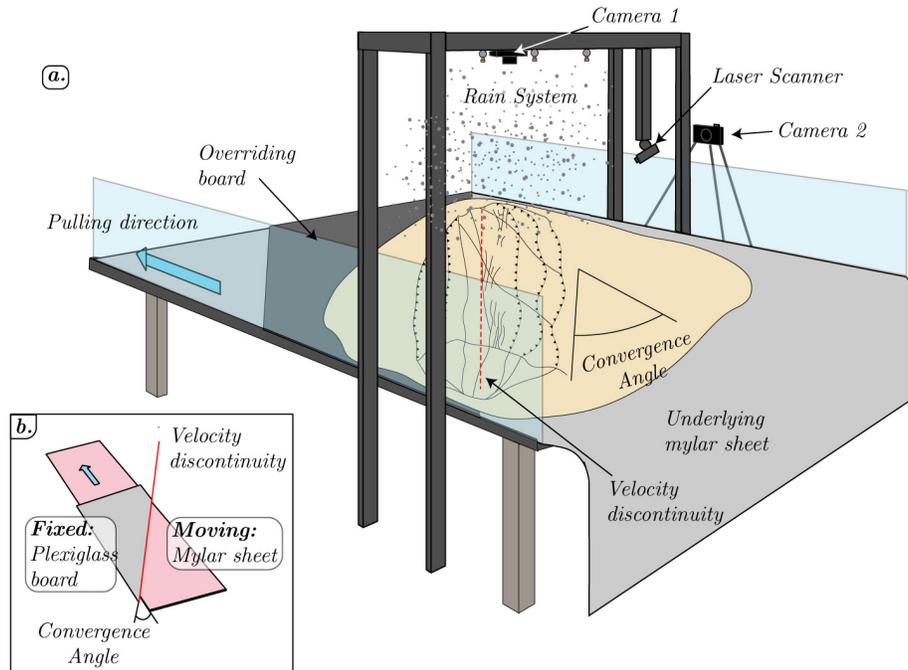


Figure 2. (a) Cartoon showing the experimental set-up. The experiment is comprised of a $2 \times 1 \times 1$ m plexiglass box set on a tabletop. The beige feature represents the material loaded on top. The red line shows the velocity discontinuity and the blue arrow on the left is the pulling direction. The motor that pulls the mylar sheet is not pictured. (b) The velocity discontinuity set-up that was put into the box to generate a deformational wedge. The gray color represents the plexiglass board cut to the desired convergence angle. The pink represents the mylar sheet that was pulled under the board. The velocity discontinuity (red line) is the interface between the fixed and moving materials.

142 2.3 Parameters varied

143 We selected three representative experiments (bold font in Table 1) out of ten that explored a
 144 more extensive range of rainfall and convergence settings. These additional experiments test the
 145 model sensitivity, ensure reproducibility, and explore the parameter space. Most experiments were
 146 performed with a convergence angle of 20° to investigate wrench-dominated transpression (see
 147 Teyssier et al., 1995). For comparison, we conducted one model at 40° . Convergence rates ranged
 148 from 65 to 325 mm hr^{-1} , and rainfall rates from 0 to 26 mm hr^{-1} . The three presented experiments
 149 provide the most robust and comparable datasets considering the scope of this paper: “dry”
 150 (D_62422), “low erosion” (W_71322), and “high erosion” (W_62722). The prefixes D and W are
 151 for dry and wet experiments and the suffix is the date experiments were conducted. In the table,
 152 the CR refers to a dimensionless quantity defined as the convergence over rainfall rate (Reitano et
 153 al., 2022). An infinite CR means the system is completely dry, whereas a CR of 0 indicates the
 154 system is tectonically quiescent. We present these tests as representative “wet” (“low” and “high
 155 erosion”) and “dry” CR scenarios at a convergence angle of 20° . Due to the highly differing
 156 boundary conditions between a completely dry experiment and one subject to consistent misting,
 157 the wet experiments are more directly comparable.

158 We consider results robust if fault geometries and the dimensions and general shape of the
 159 wedge are reproducible between models. Furthermore, we ensured that throughout the main stages,
 160 there was no connectivity between drainages propagating from the edge of the model and drainages
 161 within the wedge. The latter is necessary because the edge drainages provide a lower base level.

Table 1. All experiments conducted using the sandbox set-up. Prefixes *D* and *W* denote wet and dry experiments, respectively. The *CR* is the ratio between convergence and rainfall rates.

Experiment #	Convergence Angle (°)	Convergence rate (mm/hr)	1 σ	Rainfall rate (mm/hr)	1 σ	CR	Referred to in text as:
D_62422	20	320	40			Inf.	"Dry"
W_62321	20	240	9	20	6	12	
W_62421	20	80	6	15	4	5	
W_71321	20	230	11	31	13	7	"Low erosion"
W_70221	20	80	14	26	11	3	
W_71521	40	70	11	26	11	3	
W_62722	20	70	24	34	11	2	"High erosion"

162 Therefore, once connected with the wedge network, these outlets would localize and increase
 163 the erosion mass flux out of the wedge, dominating the topography and drainage network
 164 morphology (Leopold & Bull, 1979).

165 2.4 Scaling

166 To dynamically scale experiments to natural systems, we should follow the principles outlined by
 167 Hubbert (1937) and Ramberg (1981). However, given uncertainties about the physics of surface
 168 transport, it is not entirely clear how to upscale surface processes. We follow previous work by
 169 considering kinematic similarity only rather than dynamic scaling. This approach might be
 170 appropriate if erosional processes are scale-invariant (Sapozhnikov & Fofoula-Georgiou, 1996),
 171 leading to potential scale independence (see Paola et al., 2009, for discussion). In terms of
 172 geometric comparison with nature, we define a length scaling factor, $l^* = l_{model}/l_{nature}$. Given the
 173 approximate dimensions of transpressional mountain belts ($l = \sim 10^5$ m, $w = \sim 10^4$ m) and that of
 174 the wedges generated in the experiment ($l = 1$ m, $w = 10^{-1}$ m), $l^* = 10^{-5}$, meaning 1 cm in the model
 175 represents 1 km in nature. To derive the time scaling factor, t^* , we use the erosion number scaling
 176 approach of Reitano et al. (2022) to estimate $t^* = 4 \times 10^{-11} - 4 \times 10^{-10}$. These values suggest that 1
 177 hour of model time corresponds to 300 kyr – 3 Ma, as in prior work (Graveleau et al., 2011; Mao
 178 et al., 2021; Reitano et al., 2022). This scaling approach assumes that material accreted to the
 179 wedge is balanced by material removed by erosion. In this case, the difference in material
 180 erodibility, k , between model and nature can be evaluated as $k^* = v^*/4l^*$, where v^* is the
 181 convergence rate scaling factor ($v^* = 10^4$ - 10^5). Using this v^* and the l^* found above, k^* is between
 182 10^9 - 10^{10} .

183 2.5 Analysis

184 We monitor the structural and surficial evolution by scanning the model incrementally with
 185 a laser scanner to create digital elevation models (DEMs) and conducting particle image
 186 velocimetry (PIV) from photos taken every minute. Vertical and horizontal resolutions for the laser
 187 are 0.07 mm and 0.05 mm, respectively. Scans are taken at 10 cm, 15 cm, 25 cm, and 35 cm of
 188 convergence. We chose the first increment based on our preliminary experiments, where we
 189 determined that this much convergence creates sufficient relief (~ 1 cm) for realistic drainages to
 190 develop. The subsequent increases provide 10 cm increments (+30% shortening) of experiment
 191 evolution up to a maximum of 35 cm (considered 100% shortening). These stages are appropriate

192 given spatiotemporal constraints, including the influence of edge effects, which grow with rain
 193 time and total displacement. After this amount of shortening, especially for high rainfall
 194 experiments, the influence of drainages and faults propagating from the boundary cannot be
 195 neglected. We use the MATLAB software TopoToolbox (Schwanghart & Scherler, 2014) to
 196 analyze DEMs and the corresponding stream networks across the experimental stages. Structural
 197 interpretations are made of each stage by pairing DEMs with photographs, which more clearly
 198 display structures lacking sufficient vertical offset to be resolved by the laser scanner.

199 We derive the evolution of the velocity field using a 2-D cross-correlation technique,
 200 Particle Image Velocimetry (PIV, see Raffel et al., 2007), with the MATLAB toolbox PIVlab
 201 (Thielicke & Sonntag, 2021). However, there are some unavoidable limitations and high amounts
 202 of noise associated with using this technique in the presence of a rain system, as surface transport
 203 is also partially tracked, and mist affects the quality of the images. Yet, with a 1-minute capture
 204 rate, image pre-processing, and velocity filtering, PIV can provide insight into the differences
 205 between end member erosional cases. We preprocess images using a contrast-limited, adaptive-
 206 histogram equalization filter and auto-contrast stretch. To generate velocity fields, we use a Fast
 207 Fourier Transform PIV algorithm across a region of interest of 2253×784 px with an initial
 208 interrogation area of 116 px, and three passes down to 32 px, and a gauss 2×3 point sub-pixel
 209 estimator (see Thielicke & Sonntag, 2021). Resultant velocity fields were calibrated using a photo
 210 reference and analyzed and plotted using Generic Mapping Tools (Wessel et al., 2019). We
 211 calculate the horizontal component of the velocity, u , the maximum horizontal shear strain rates, $\dot{\epsilon}_s$,
 212 and the dilatational strain rate, $\dot{\epsilon}_m$. The strain components $\dot{\epsilon}_s$ and $\dot{\epsilon}_m$ allow us to analyze the
 213 localization of strike-slip and compressional/extensional deformation, respectively. For ease of
 214 comparison between frames of each experiment, we normalize the values by the standard deviation
 215 in each frame and denote the normalized u , $\dot{\epsilon}_s$, and $\dot{\epsilon}_m$ as \hat{u} , $\hat{\epsilon}_s$, and $\hat{\epsilon}_m$, respectively.

216 **3 Results**

217 *3.1 Structural evolution*

218 Figure 3a shows the DEM results of the three presented experiments (dry, low erosion,
 219 high erosion), and Figure 3b shows the interpreted structural evolution of all models. For reference
 220 we show contrast-enhanced images of the final model stages in Figure 3c-d. To describe our
 221 models, we use the wrench fault terminology of Naylor et al. (1986). Fault progression begins with
 222 the appearance of *en-echelon* synthetic shears (R , $15^\circ - 30^\circ$) that initiate sub-parallel with the
 223 convergence direction and delineate rhomboidal packages within the wedge. Within the viewing
 224 frames, these features form at least three clear packages, which can be directly compared between
 225 models. R -shears are accompanied by antithetic shears (R' , $65^\circ - 90^\circ$) and connecting splays (S ,
 226 $>17^\circ$). The left-lateral displacement of material packages initiates the main inboard (on the fixed
 227 side of the model) thrust (bottommost thrust in Figure 3b), followed by the formation of an
 228 outboard (on the moving side of the model) back-thrust. Together these features form “*pop-up*”
 229 structures, which accommodate the uplift of the blocks. While somewhat obscure in erosional
 230 models due to erosion/sedimentation, a notable low-angle shear striking in the opposite direction
 231 (P , $180^\circ - 165^\circ$) forms in all models outlining the bottom-left portion of an elongate diamond-
 232 shaped or “*pug-nosed*” landform (Figure 3a).

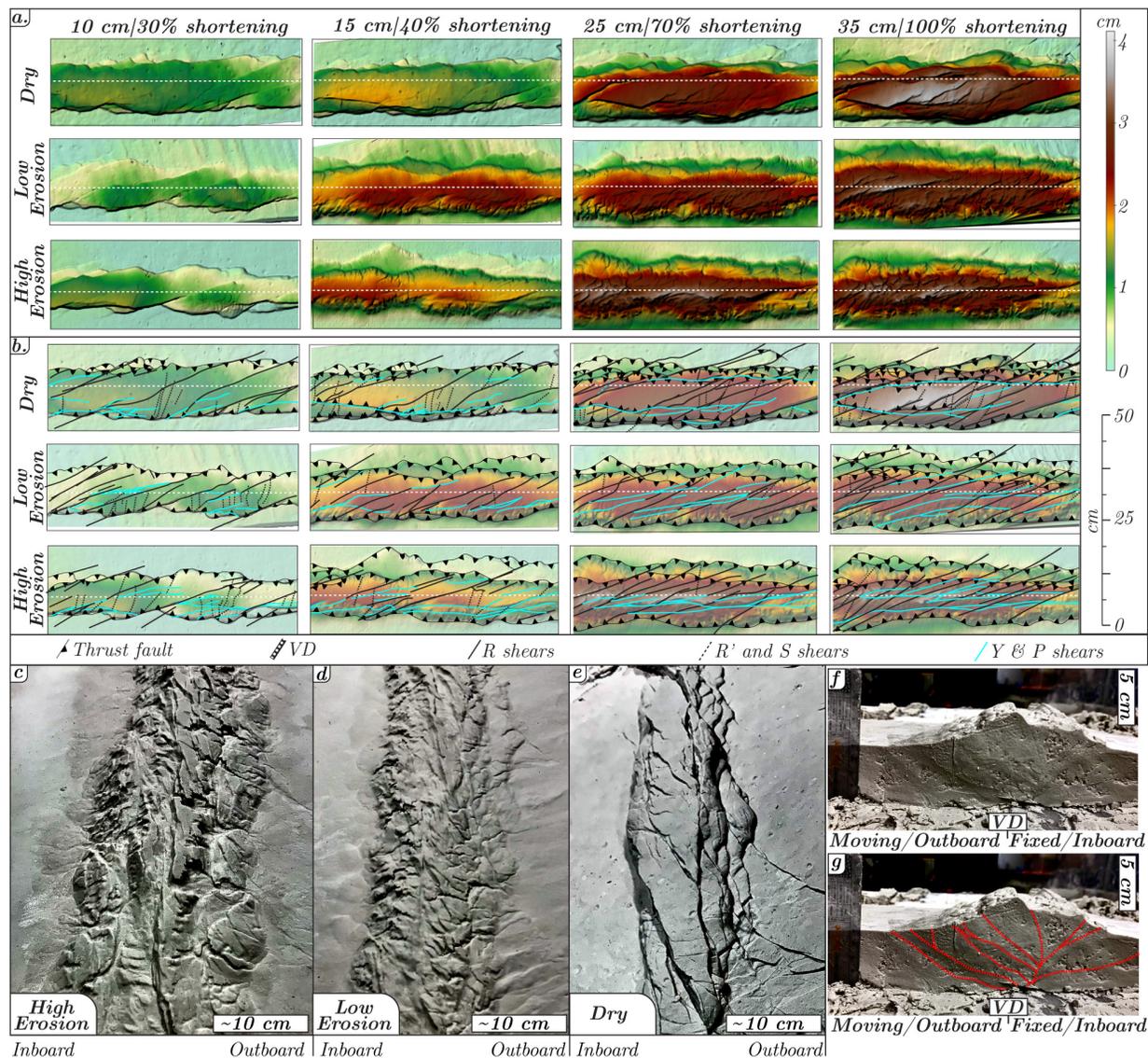


Figure 3: Overview of the results of selected experiments. **(a)** Digital elevation models (DEMs) across shortening stages. Rows show increasing erosion and columns show increasing convergence/shortening. White lines represent the location of the velocity discontinuity (VD). **(b)** Interpreted structures showing thrusts (decorated lines), the VD (white dashed line), riedel (R) shears and connecting splay (S) faults (black lines), anti-riedel (R') shears (dotted black lines), and low-angle faults (Y- and P-shears, see text for description, cyan lines). **(c-e)** Contrast enhanced oblique images at the final stage of each model. Scale bars are only accurate at the bottom of the images due to perspective distortion. Note the alluvial fans in the erosion models **(c,d)**. **(f-g)** Contrast enhanced images showing the uninterpreted **(f)** and **(g)**

233 The main inboard thrust feature moves only a few centimeters during the evolution of each
 234 model (2.3 cm, 3.4 cm, and 3.3 cm for dry, low erosion, and high erosion models, respectively).
 235 The outboard thrust belt propagates throughout the model at distances depending on the
 236 presence/amount of erosion. Thrust sheets nucleate at the tips of the R-shears that extend into the
 237 undeformed inboard and outboard sections of the model. With further convergence, lower angle
 238 R-shears (Y, 0° - 15°) form and coalesce with P-shears into an anastomosing VD-parallel master

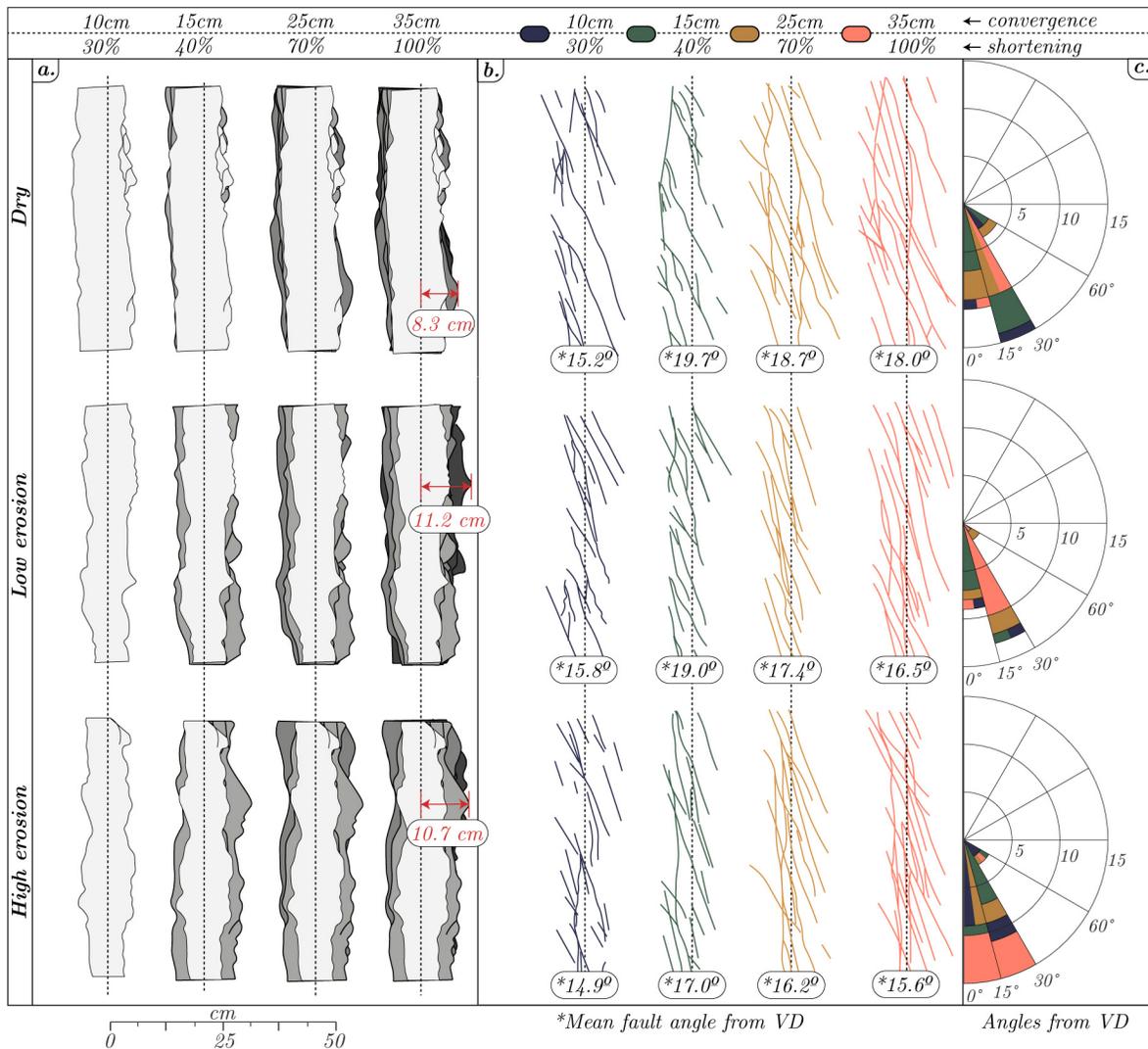


Figure 4: The evolution of thrust sheets and internal strike-slip faults with increasing shortening. The upper, middle, and lower rows show the dry, low erosion, and high erosion models, respectively. **(a)** Change in thrust belts with increasing convergence. Each grey shade represents a horizontal slice through the wedge. These slices are stacked at each convergence step. The horizontal black dashed lines are the velocity discontinuity. **(b)** The evolving internal strike-slip structure of the models. **(c)** Polar histograms showing the number of structures within 15° bins from 0 to 60°. Colors in the histograms correspond to the convergence stages as shown at the top of the figure. Split bins represent subsequent stages with no change in count. Faults and fractures > 60° and < 0° were included in those nearest bins, because they are uncommon and deviate only small amounts from those angles.

239 fault zone. In some cases, R' oriented fractures accommodate an extensional component. In either
 240 case, these fractures have an apparent clockwise rotation as they are offset by the left lateral R -
 241 shear system. R -shears may have an extensional component after they no longer accommodate
 242 strike-slip motion, especially when they optimally intersect with the evolving master fault and
 243 develop a releasing bend. This occurrence is prevalent in the later stages of the model. R -shears
 244 also tend to form arcs concaving into the VD and sometimes form sharp cusps at the VD-fault
 245 intersections (Figure 3b). Overall, the structural evolution of the experiments agrees with what is
 246 described by prior analog studies of wrench-dominated fault zones (e.g., Casas et al., 2001; Cloos,

247 1928; Leever et al., 2011a,b; Naylor et al., 1986; Pinet & Cobbold, 1992; Riedel, 1929; Schreurs
 248 & Colletta, 1998, 2002; Tchalenko, 1968; Wilcox et al., 1973). Though challenging to interpret,
 249 as the material is monochromatic, high-contrast photos of cross sections cut through the high
 250 erosion model (Figure 3f,g) show the complex internal deformation within the wedge, interpreted
 251 as a thrust-bounded, upward tulip shaped structure (Figure 3g).

252 To compare the evolution of thrust faults across the models, we show superimposed
 253 horizontal topographic slices for each model and stage (Figure 4a). From dry to high erosion, the
 254 number of thrust sheets increases at each time stage yet are narrower, forming distinct half-moon-
 255 shaped salients. Contrary to the other models, the first thrust sheet that initiates in the high erosion
 256 model remains dominant for most of the experiment. This sheet is nearly as wide as the maximum
 257 extent of the thrust sheet in the low erosion model and exceeds that of the dry model (3.3 cm).
 258 Subsequent thrust initiation in the low erosion model eventually overtakes that of the high erosion
 259 model. The maximum thrust toe distance from the VD across all stages is 8.3 cm, 11.2 cm, and
 260 10.7 cm for the dry, low, and high erosion experiments, respectively. Thus, the drier systems
 261 initiate more yet thinner thrust sheets but still achieve nearly equal or greater cumulative widths at
 262 17 cm, 21 cm, and 16 cm, respectively.

263 There are evident changes in the evolution of the intrawedge strike-slip faults between
 264 experiments. We illustrate this in Figures 4b and 4c by extracting the surface traces of these faults,
 265 calculating their orientation (Figure 4b), and binning them into rose diagrams (Figure 4c). Dry
 266 models show a slight change in the geometry of shears through each stage, with a subtle indication
 267 of the eventual through-going, master wrench-fault formation. This final-stage fault seems to
 268 reactivate the initial outboard thrust fault plane. Conversely, erosion models show a more
 269 significant change in intrawedge shear geometry and an earlier coalescence of shears into a clear,
 270 through-going wrench fault. From inspection of such traces, low-angle faults that begin to merge
 271 into the master fault dominate the later stages of the high erosion model. However, faults are more
 272 distributed with more *R*-shears for the low erosion and dry experiments. Visualizing this in the
 273 rose diagrams (Figure 4c), the high erosion model has more fault traces ($n = 15$) within 15° of the
 274 velocity discontinuity.

275 3.2 Stream Evolution

276 Figure 5 shows snapshots of the drainage evolution of the high and low erosion models.
 277 After we engaged the rain/mist system, streams nucleated orthogonal to the trace of the thrust
 278 sheets (transverse orientation). As convergence continued, streams evolved following various
 279 well-described mechanisms: headward erosion, drainage deflection by strike-slip motion, drainage
 280 capture, and drainage beheading (see Bishop et al., 1995, for review). From observations of
 281 pictures and DEMs, it is apparent that faults strongly control the initiation of streams and pathways
 282 of headward erosion. As a result, we generally observe asymmetric forked to rectangular drainage
 283 patterns with consistent spacing and sharp angles defining tributary junctions.

284 Streams that initiate in the *R-shear* direction erode headward throughout the evolution of
 285 the model and follow the reorientation of *R-shears* described above. With the left-lateral deflection
 286 of transverse drainages along *R-shears*, these ‘*R*-streams’ are captured, resulting in sharp cusps in
 287 the drainage topology. This pattern is apparent on both the main thrust and thrust belt sides of the
 288 wedge. However, on the main thrust portion, transverse streams dominate the evolving networks.
 289 In both models, major valleys capture the flow of several transverse streams following the *R*-shear

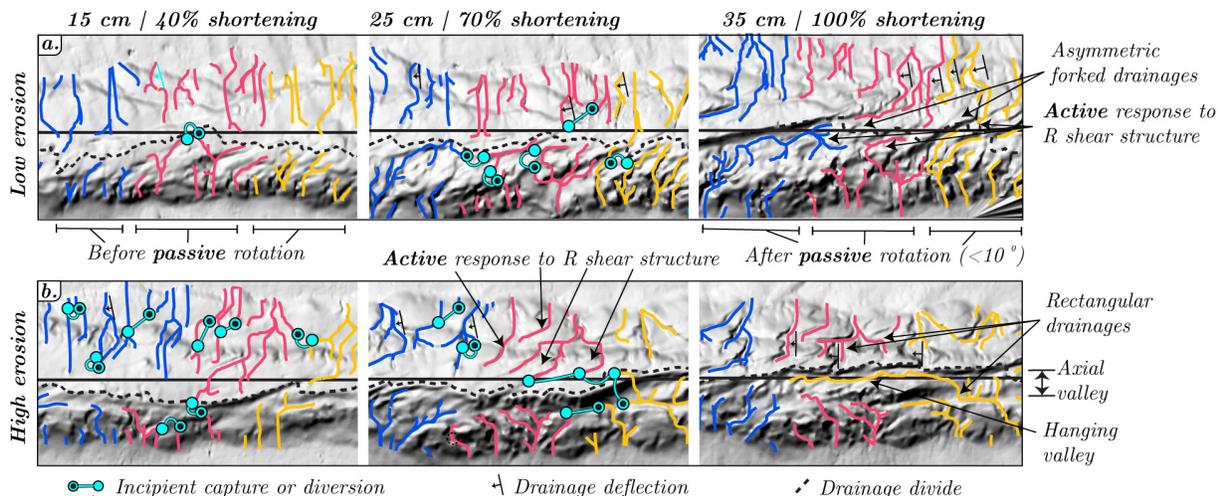


Figure 5: Snapshots of stream evolution between erosion models. Streams are cut to the lowest thrust sheet and colored as a visual aid to tracking reorganization mechanisms between frames. **(a)** Low erosion model. The active drainage response to structure is more delayed, while the passive rotation is better expressed (preserved). Drainages networks exhibit asymmetric forked patterns in the final stage. **(b)** CR2 model. The active drainage response to structure occurs earlier, yet there is a less apparent passive rotation response. Final drainage patterns are rectangular in form.

290 structures in the low erosion model and the VD-parallel master fault in the high erosion model.
 291 The capture of this valley by transverse tributary drainages causes punctuated erosion events.

292 In the high erosion model (Figure 5b), the primary drainage system shows less branching
 293 and is more aligned with the velocity discontinuity. Furthermore, *R*-streams initiate early and are
 294 more rapidly captured by transverse streams, resulting in more rectangular drainage networks with
 295 sharper junctions. Similarly, captures are less prevalent in the low erosion model - with one capture
 296 pair (linked blue dots) compared to eight in the high erosion model at 40% shortening (Figure 5a).
 297 As a result, faults more consistently entrain streams in the direction of structures forming forked
 298 asymmetric drainages. Lastly, as described in section 4.2, the intersection of *R*-shears with the
 299 master fault may form a releasing bend, expressed geomorphologically as a partially restricted
 300 lofted valley in our high erosion system.

301 3.3 PIV analysis of velocities and strain-rates

302 From the normalized velocities, \hat{u} (Figure 6a) and strain-rates, $\hat{\epsilon}_s$ (Figure 6b), and $\hat{\epsilon}_m$
 303 (Figure 6c) derived through our PIV analysis, we recognize three main phases of strain-rate field
 304 evolution common for all models. In Figure 6d, we extract the structures corresponding to sharp
 305 gradients in \hat{u} (Figure 6a) and related bands of high $\hat{\epsilon}_s$ (Figure 6a). Velocities are particularly useful
 306 in verifying shear zones and differentiating them from noise imposed by landsliding or mist
 307 interference (red blobs in Figure 6a).

308 Some early organization phases occur before the first panel in Figure 6, beginning with
 309 distributed deformation followed by shear strain localization along *R* faults at 6-9% shortening
 310 (not pictured). Before the first frame in Figure 6, there is also a stage of distributed deformation
 311 and shear strain localization on *R*-faults (stage 1 in Figure 6). Shortly after, elevated $\hat{\epsilon}_s$ values are
 312 focused on the sides of the wedge (10 cm convergence / 30 % shortening), marking a phase of
 313 incomplete partitioning and oblique faulting on the major thrust structures (stage 2 in Figure 6).

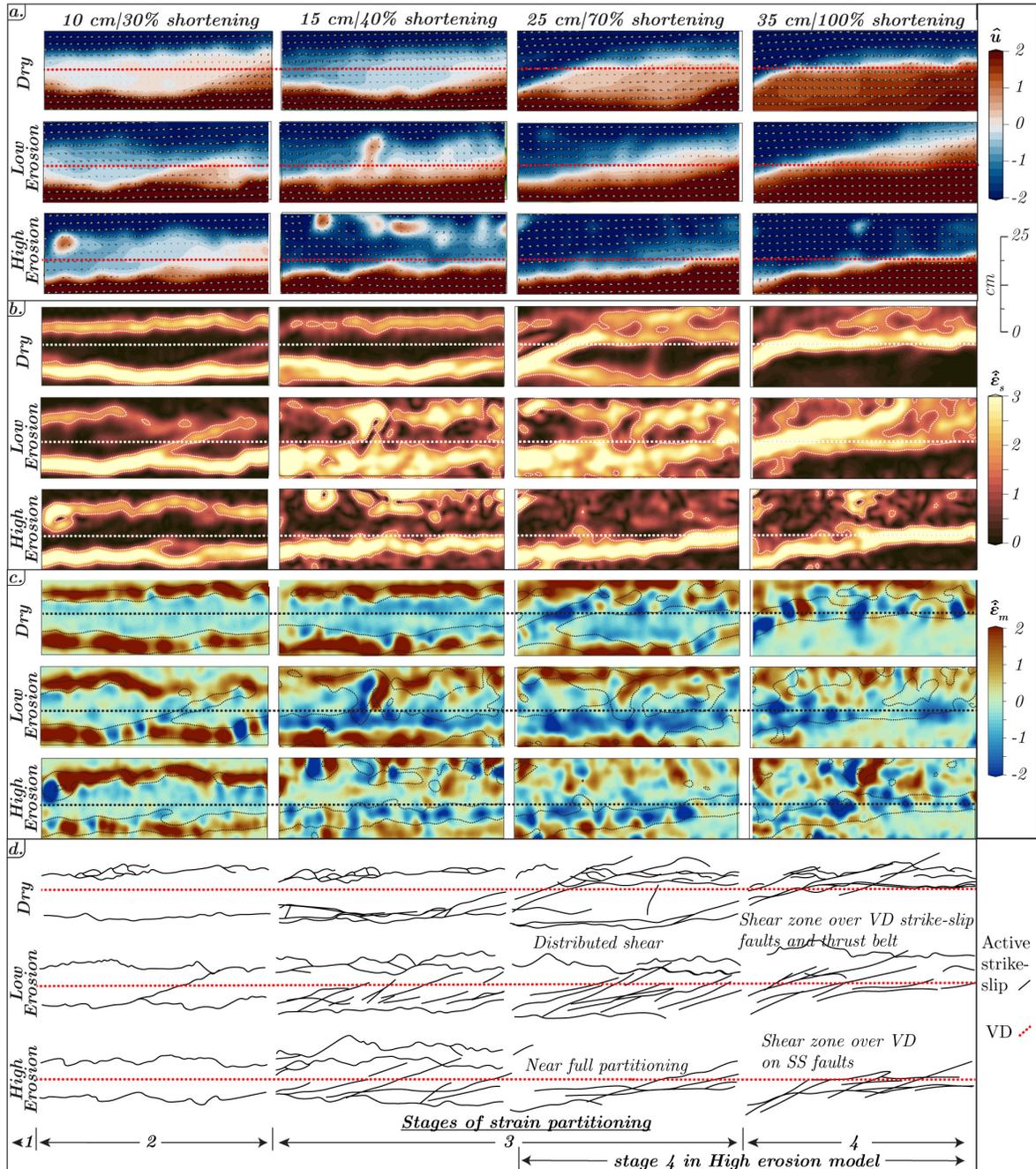


Figure 6: Evolution of strain partitioning from PIV at the same stages shown in Figures 2, 3, and 4. **(a)** The normalized horizontal velocity components (\hat{u}). Sharp color contrast indicates locations of strike-slip deformation. Irregular red blobs show locations of land sliding or noise caused by mist interference. The dashed red line indicates the velocity discontinuity (VD). **(b)** The normalized maximum horizontal shear strain-rate at each stage ($\hat{\epsilon}_s$). White dotted lines are superimposed $\hat{\epsilon}_s = 1.5$ contours. The dashed white line indicates the VD. **(c)** The normalized dilatational strain-rate, ($\hat{\epsilon}_m$). Positive values (red) indicate compression, negative values (blue) indicate extension. Black dotted lines are superimposed $\hat{\epsilon}_s = 1.5$ contours. The dashed black line indicates the VD. **(d)** Active faults determined by interpreting fault locations with respect to \hat{u} and $\hat{\epsilon}_s$. The dashed red line indicates the VD. The evolution is broken into four stages of strain partitioning: 1) distributed strain and en-echelon R-shear formation, 2) oblique slip on bivergent thrusts, 3) transitional strain partitioning, 4) full partitioning to throughgoing structure (s).

315 The experiments then enter a transitional stage (stage 3 in Figure 6). Synchronous with the
 316 structural evolution, strike-slip motion becomes increasingly localized on a narrow band of
 317 anastomosing strike-slip faults (Figure 6c). The high erosion experiment achieves near-complete
 318 strike-slip strain partitioning at 70% shortening (stage 4 in Figure 6). For the other experiments,
 319 complete strike-slip partitioning does not occur until the final frame (35 cm | 100% shortening) at
 320 the earliest (i.e., if the model continues, the evolution likely progresses). This difference suggests
 321 that the strain partitioning evolution is accelerated in the high erosion model. The more prevalent
 322 noise from land sliding attests to more vigorous sediment routing out of the wedge. Balanced by
 323 compression at the boundaries, the interior of the wedge is under extension in each model and
 324 stage (Figure 6c). This band becomes localized to the master-fault zone as the models progress
 325 and compression becomes less organized. Additionally, more restraining bends (red anomalies in
 326 blue extensional bands) occur along the master-fault zone in the dry model compared to wet
 327 models.

328 3.4 Topographic evolution

329 Initial topography forms along pop-up structures as rhomboidal slices (Figure 3). With the
 330 onset of thrust belt propagation, the topography develops transverse asymmetry, with one steep
 331 side corresponding to the main thrust and a broader side corresponding to the thrust belt. As can
 332 be seen from the differences between the dry and erosion models, hillslope diffusion and stream
 333 erosion drastically modify the topography by incising valleys and causing fault scarps to retreat
 334 inward toward the velocity discontinuity. Alluvial fans fill the recessed portions of ridges. As
 335 expected, there are broader and higher volume alluvial fans and more deeply incised channels in
 336 the high erosion model (Figure 3c-e). In the dry model, the final topography resembles an uplifted
 337 and broadly concave plateau. With increased erosion, the topography is more rugged and
 338 characterized by steeper peaks and more incised valleys.

339 Figure 7a shows the changing width of the wedges. The high erosion model shows an increase
 340 in width of >5 cm after the first erosional stage, then a plateau with continued convergence. On
 341 the other hand, the low erosion model width shows a slightly smaller increase, approximately 4
 342 cm, yet continues to grow as the experiment continues. The dry model has a broader initial
 343 topography and shows slow and steady growth in the wedge width from 14 to 15.8 cm. As the
 344 experiment evolves, it is marked by a higher curvature thrust belt and more salients and recesses.
 345 Furthermore, there is no channel incision, resulting in a broad wedge dissected only by strike-slip
 346 structures. The characteristic diamond or “pug-nose” shape of wrench-dominated systems is most
 347 evident in the dry model due to the intersection of *P* and *R-shears* (Figure 3a). With increased
 348 erosion, structures that do not accommodate significant displacement become less apparent. For
 349 instance, in the erosion models, the scarp of the uplifted *P* and *R-shears* that delineate the diamond
 350 structure is eroded in the outboard direction, nearly hiding the feature altogether. There are also
 351 differences in relief across strike-slip faults between models. For the high erosion model, *Y*
 352 structures have more relief. In the dry and low erosion model, relief is higher on *R* structures.

353 The surface uplift seems to progress similarly in all models, with only subtle differences in the
 354 rate and magnitude. Notably, the high erosion model reaches 3.5 cm earlier at 70% shortening and
 355 remains constant until 100% shortening. From 0 to 25 centimeters, each experiment shows an
 356 initial phase of more rapid uplift (3 – 3.5 cm at 25 cm of convergence) followed by a slow rise,
 357 perhaps approaching a limit of around 4 cm (Figure 7b). Figure 7c shows the elevation of the
 358 wedge divided by the distance of the thrust toe from the *VD*. The inverse tangent of the plotted

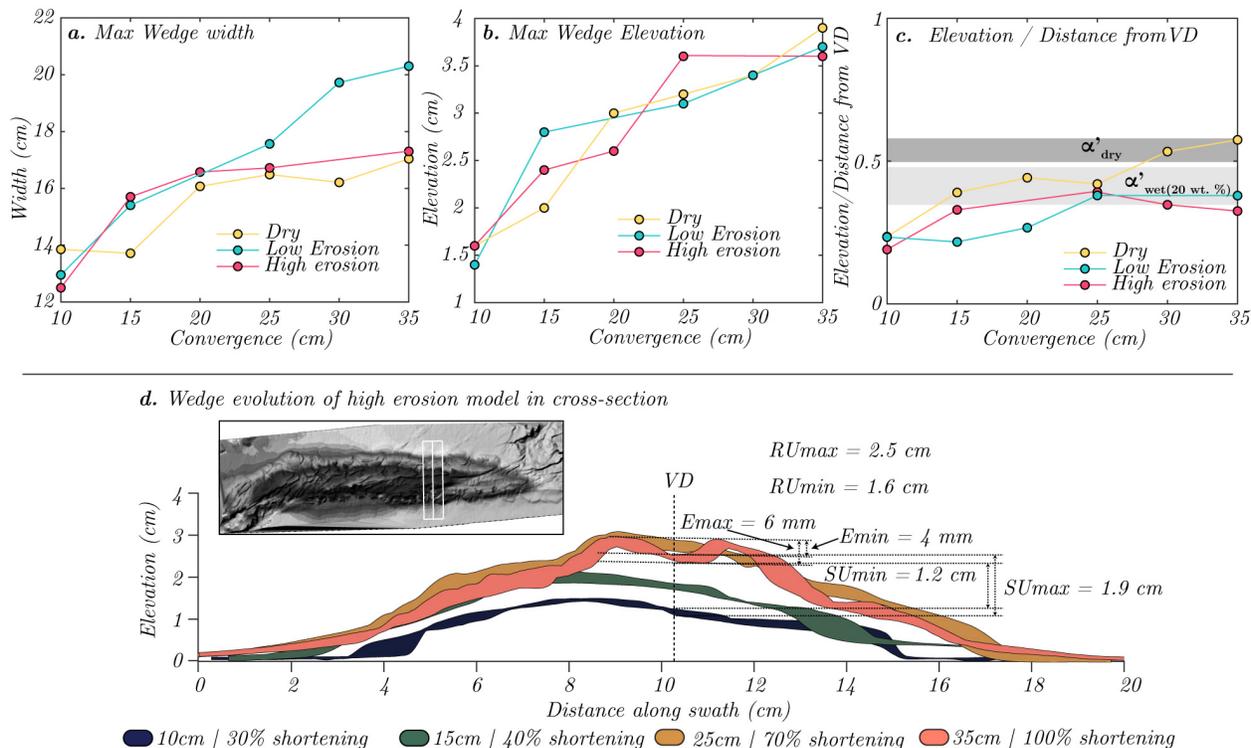


Figure 7. The topographic evolution of the presented models. **(a)** Maximum wedge width, **(b)** elevation, and **(c)** elevation divided by distance from the velocity discontinuity (VD). The gray bars are the error windows for the theoretical wedge slope calculated from critical taper theory (see Dahlen, 1990) using the material parameters for wet (20 wt. % H₂O) CMII from Reitano et al. (2020), α'_{wet} , and dry CMII (measured), α'_{dry} . **(d)** Wedge evolution of the high erosion model in cross section with estimates of erosion (E), rock uplift (RU), and surface uplift (SU). The inset shows the location of the swath at the final stage (convergence = 35 cm / shortening = 100%).

359 values is effectively the slope of the thrust belt. Using the peak angle of internal friction, ϕ , of wet
 360 CMII ($\phi = 25$ - 36 , reported in Reitano et al. 2020) and the measured ϕ of dry CMII ($\phi = 44$ - 48) the
 361 error window for the inverse tangent of the theoretical dry and wet wedge slope angle, α , from
 362 critical taper theory (see Dahlen, 1990) is $\alpha_{dry} = 27^\circ - 30^\circ$ and $\alpha_{wet} = 19^\circ - 26^\circ$. The tangents of
 363 these alpha windows, α' , are plotted in Figure 7c ($\alpha'_{dry} = 0.53 - 0.57$, $\alpha'_{wet} = 0.35 - 0.48$). Initially,
 364 the thrust belt slope in the dry and low erosion models increases into the α'_{wet} window, reaching a
 365 value of approximately 0.4 at 70% shortening. The wedge slope in the high erosion model slightly
 366 decreases at first, corresponding to the propagation of a wide thrust sheet, then similarly increases
 367 to 0.4 at 70% shortening. Besides a slight decrease of ~ 0.05 in the low erosion model, from 70%
 368 to 100% shortening, the wedge slope reaches a steady state in the wet models. In contrast, in the
 369 dry model, it continues to steepen, nearly exceeding the maximum α_{dry} .

370 A cross-section of wedge evolution is shown in Figure 7d with corresponding estimates of rock
 371 and surface uplift and erosion. The highest rock uplift is located at the VD, which in the high
 372 erosion case corresponds with the trend of the axial valley where surface uplift is locally low.
 373 Therefore, exhumation is at a maximum at this location.

374 3. Discussion

375 4.1 Drainage evolution in response to transpressional tectonics

376 Considering the experimental results (Figure 5) and the modes of drainage reorganization
 377 described in the literature (e.g., Bishop, 1995; Bloom, 1998; Castellort et al., 2012; Hallet &
 378 Molnar, 2001; Koons, 1994, 1995; Ramsey et al., 2007), we group stream response mechanisms
 379 to tectonic deformation into two categories:

- 380 1. a dynamic reorganization response influenced by the structural evolution of the orogen.
 381 2. a passive response to local strain.

382 The primary drainage rearrangement mechanisms that enable the dynamic reorganization
 383 response to structure are entrainment by fault block growth, diversion and beheading by lateral
 384 displacement, and lengthening and capture by headward erosion and preferential fault plane
 385 incision (Bloom, 1991; Koons, 1994, 1995; Bishop, 1995). Figure 8a-j shows several examples of
 386 these dynamic responses. The evolution of each drainage network can be considered the result of
 387 the linear combination of these mechanisms.

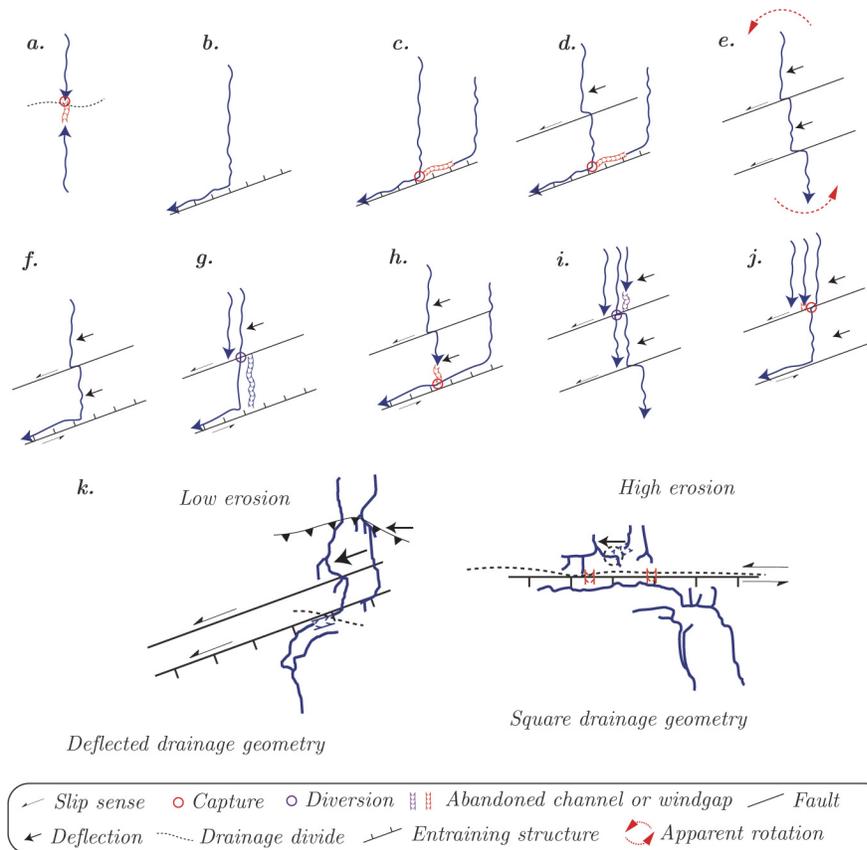


Figure 8: Examples of drainage reorganization mechanisms leading to the drainage patterns in the high and low erosion models. **(a)** across divide capture, **(b)** structural entrainment, **(c)** structural entrainment and capture upstream along R-shear, **(d)** deflection, entrainment, and capture upstream along R-shear, **(e)** deflection, **(f)** deflection and entrainment, **(g)** deflection, beheading, diversion, and entrainment, **(h)** deflection and capture of transverse stream by stream along R-shear, **(i)** deflection, diversion, and beheading, and **(j)** deflection, diversion, beheading, and entrainment. **(k)** shows representative streams extracted from high and low erosion DEMs and thus, how these mechanisms may combine to dictate differences in drainage patterns between models. Clearly, the structural evolution has an impact on the geometry forming more rectangular drainages in the high erosion model.

388 Differences in stream network geometry between erosion models suggest that the rate at which
 389 the drainage system responds to structure controls the potential feedback with structural evolution
 390 (Figure 5). More erosive conditions (headward erosion, capture) favor some dynamic stream
 391 response mechanisms and, thus, a shorter response time to structural change. However, other
 392 mechanisms, such as deflection and preferential incision, rely on more structurally dominated
 393 stream paths (Koons et al., 1994). In Figure 5, drainages in the high erosion model respond more
 394 quickly to the structural evolution of the model and deflection in the direction of *R*-shears, with a
 395 response evident at 70% shortening. There are also more capture pairs at 40% and 70% shortening,
 396 with 12 in the high erosion model compared to 6 in the low erosion model. Due to this heightened
 397 response, high erosion stream networks are more rectangular, and a clear axial valley forms with
 398 4-6 mm of incision. Alternatively, streams in the low erosion model have a more delayed response
 399 forming asymmetric forked drainage networks in the final stage. To highlight these differences, in
 400 Figure 8k, we show single characteristic drainage basin networks from both the high and low
 401 erosion models.

402 The processes described above may explain the formation of a more incised axial valley by
 403 drainage redirection in the high erosion model (Figure 5) due to higher strain localization on the
 404 master fault (Figure 6b) and pervasive along-strike extension (Figure 6c). The material in these
 405 shear zones is weakened by the concentration of mechanical strain and erosional energy along fault
 406 damage zones as a function of the strain-weakening behavior of the material (Vermeer & De Borst,
 407 1984). Such strain-weakening behavior was described in the material characterization of Reitano
 408 et al. (2020). In both cases, vertical offsets along main structures (1 mm – 10 mm) entrain streams
 409 so that they reflect the orientation of the active fault system, especially at later stages once
 410 extension becomes concentrated on the master fault, and the stream-structure feedbacks are well-
 411 developed.

412 Once deeply incised, streams may also rotate with the local strain field, described here as
 413 the passive response to local strain (Castelltort et al., 2012; Goren et al., 2015; Guerit et al., 2018;
 414 Hallet & Molnar, 2001; Ramsey et al., 2007; Zeitler et al., 2001). In both the high and low erosion
 415 models, the passive response is less commonly observed but nevertheless tracks the anticlockwise
 416 rotation of some blocks up to a few degrees (Figure 5). The stepwise left-lateral deflection of
 417 stream segments (Figure 8e) further assists the apparent rotation.

418 For a stream to be a passive strain marker, the initial orientation of streams should be nearly
 419 perpendicular to the trend of the wedge, so they are ideal for rotation with the strain-rate field and
 420 can be reliably measured. The initial orientation of such streams seems to be controlled by the *R'*
 421 fracture structure. These streams follow the nucleation and rotation of *R'* fractures ($< 10^\circ$) with
 422 continued convergence (Figure 5a). Passive streams must also be in a place where shear strain is
 423 distributed equally across their length because shear strain is localized differently depending on
 424 the stage of the experiment. Therefore, even with poor fault exposure, streams can provide insight
 425 into where shear strain is localized in a wedge and how mature the orogen is regarding the
 426 evolution of strike-slip partitioning.

427 *4.2 Links between fault structure, erosion, and the evolution of strain partitioning*

428 In general, the structural and strain partitioning results of our experiments agree with the
 429 experiments of Leever et al. (2011a) using a dry quartz sand pack laid upon a velocity
 430 discontinuity. Building on the work of Pinet & Cobbold (1992), Leever et al. (2011a) described a
 431 3-stage evolution of the strain field during transpression from distributed strain to full partitioning.

432 Expanding the work of these authors, we describe the progression observed in our models (Figure
 433 6) by combining wrench tectonics within the wedge (e.g., Naylor et al., 1986; Tchalenko, 1970;
 434 Wilcox et al., 1973) with the evolution described by Leever et al. (2011a).

435 Beginning with stage 1, following the period of distributed strain, strike-slip deformation is
 436 first accommodated along *R*-shear structures, as the principal infinitesimal strain axes are
 437 horizontal in wrench-dominated transpression (Tikoff & Teyssier, 1994). In stage 2, a slow-
 438 growing thrust forms on both sides of the VD eventually resulting in a bivergent wedge (30%
 439 shortening in presented models). The complete formation of thrust structures bounding the material
 440 packages provides pervasive discontinuities in the system where oblique motion preferentially
 441 concentrates. The system then enters stage 3, a transitional stage (40% shortening in presented
 442 models), where low-angle structures ($<17^\circ$ to the VD) and splay faults form, grow, and eventually
 443 link (see Naylor et al., 1986 for discussion). Stage 4 begins when a VD-parallel anastomosing
 444 “master fault” zone becomes apparent. Synchronously the zones of extension and principal shear
 445 narrow over the VD (70% and 100% shortening in high and low erosion models, respectively).
 446 Subsequent deformations are mostly independent, and bivergent thrusts now have a purely VD
 447 perpendicular dip-slip component while the master fault system fully accommodates the strike-
 448 slip component of bulk strain.

449 We observe the above stages of strain evolution across all the presented models (Figure 6).
 450 However, the difference in shortening between when the high and low erosion models enter stage 4
 451 suggests that strain partitioning is also dependent on the erosion/rainfall rate relative to the
 452 convergence rate. By 71% shortening in the high erosion model, a VD-parallel master fault system
 453 is evident (Figure 4b, c) with high strike-slip partitioning (Figure 6). In the low erosion model,
 454 while there is an indication that these paired features are developing, the structure is geometrically
 455 and kinematically immature – in the context of a fully connected anastomosing master fault zone
 456 with localized strike-slip deformation. Considering a fully partitioned fault a continuous band of
 457 $> 1.5 \hat{\epsilon}_s$, for the dry model, the VD parallel strike-slip system is well-formed by 100% shortening
 458 with $\hat{\epsilon}_s > 1.5$, yet there is also $> 1.5 \hat{\epsilon}_s$ on the back thrust.

459 The development of shear zones in the high erosion model coincides with the development of
 460 an axial valley. This observation suggests that the accelerated progression of the model through
 461 the stages of strain partitioning is linked to the erosion of fault scarps and incision by structurally
 462 controlled drainages. Therefore, feedback between the evolving stream and fault networks may
 463 accelerate strain partitioning in more erosive systems. The entrainment of streams by major faults
 464 leads to preferential incision along these structures and a positive interference with *Y*-shear
 465 formation through drainage capture, ultimately leading to the earlier appearance of a fully
 466 partitioned master wrench fault. This series is shown in Figure 4c by the increase in the 0° – 15° bin
 467 and in Figure 5b by the incision of a VD parallel drainage along the trace of the master fault and
 468 capture of the headwaters of adjacent streams.

469 We identify two potential explanations for the earlier formation of a VD-parallel valley and
 470 earlier strike-slip partitioning: 1) focused mass removal by incision changing the stress balance in
 471 the material, thus exposing and localizing deformation earlier along actively developing *Y*- and *P*-
 472 shears, and 2) weakening of the fault by infiltration and water-induced friction reduction.
 473 However, since faults are unconfined, there should be no substantial increase in fluid pressure that
 474 would promote failure.

475 Considering the first mechanism, a simple analysis of the stress on the wedge-bounding thrusts
 476 and strike-slip faults shows how reducing the overburden on the wedge decreases the strength of
 477 each fault. Given a simple wedge geometry, the force normal to the thrust, F_{thrust} , is diminished by
 478 $mg\cos(\theta)$, where m is the mass of the removed material, g is the acceleration due to gravity, and
 479 θ is the dip of the thrust. Correspondingly, the force normal to a VD-parallel strike-slip fault, $F_{strike-}$
 480 $slip$, is reduced by $mg\cos(\theta)\sin(\theta)$. The normal stress change, $\Delta\sigma_n$, is each force value over the
 481 fault area.

482 To estimate the unclamping effect caused by negative $\Delta\sigma_n$ for the presented models, we must
 483 first calculate the volume of eroded material. To do so and capture variable incision patterns in
 484 each wedge we create 5 cm wide swath profiles for the wet models at 70% shortening, with a
 485 centerline across the midsection of the innermost rhomboidal package in both models. We then
 486 integrate the difference in maximum and minimum elevations across the length of the profile to
 487 calculate the volume of eroded material. Using the density of CMII from Reitano et al. (2020), we
 488 calculate its mass. For the strike-slip fault, we estimate A given a 5 cm long fault with the height
 489 of the maximum average swath elevation. Alternatively, we estimate the A of the thrust fault as a
 490 5 cm long patch dipping 35°. For the friction coefficient, μ , we use the wet (20 wt. % H₂O) values
 491 for CMII from Reitano et al. (2020). With these values, we estimate $\Delta\sigma_n$ on the thrust faults as
 492 ~141 Pa and ~105 Pa and the $\Delta\sigma_n$ on the strike-slip faults as 69 Pa and 51 Pa for the high and low
 493 erosion cases, respectively.

494 It is helpful to put $\Delta\sigma_n$ estimates into the context of the range of stresses in the wedge. For this
 495 effort, we assume a Mohr-Coulomb failure criterion (Coulomb, 1776; Mohr, 1900) and
 496 Andersonian principal stress orientations for thrust and strike-slip faults (Anderson, 1905). Given
 497 these assumptions, we estimate the normal stress on the thrust fault imposed by the mass of an
 498 equally long wedge-shaped package as ~1.9 KPa and ~0.9 KPa for a strike-slip fault striking 20°
 499 from the greatest principal stress direction. Therefore, compared to a dry scenario, erosion results
 500 in an overall “unclamping” effect on both faults by 7%-8%. There is a difference of ~2% between
 501 high and low erosion cases. Since the initial slip style of the wedge-bounding thrusts are initially
 502 oblique, the reduction in vertical stress by erosion also assists in the rotation of the stress field so
 503 that the least principal stress is vertical, and the fault behaves as purely dip slip. The different stress
 504 states in the dry versus wet cases are exemplified by the notion that the models with erosion reach
 505 a steady state by 70% shortening, yet the dry case does not until perhaps after 100% shortening
 506 (Figure 7).

507 *4.3 Coupling between fault and stream networks to shape topography*

508 In an oblique collision zone, the topography of the resulting mountain belt is that of a thrust
 509 bounded wedge. Relief is generally subdued but rises abruptly into a steep backslope to the main
 510 divide, which falls steeply to the indenter forming the inboard slope (Koons et al., 1994). We
 511 observe the same general morphology in our models (Figure 3). Yet, at shorter length scales (< 5
 512 cm), there are apparent differences between the dry, high, and low erosion models. Here we argue
 513 that these topographic variations between our models depend on this faulting-surface process
 514 feedback and its impact on strain partitioning.

515 The interplay between tectonic and erosional factors manifests in part as the trends we see in
 516 the lateral growth of the wedge between different models (Figures 3, 4, and 7; cf. Dahlen & Suppe,
 517 1988; Steer et al., 2014). The high erosion model has fewer, yet wider, thrust sheets. The width of
 518 these sheets is a function of gradual thickening by syntectonic deposition of alluvium and more

519 rapid and widespread erosion, leading to the preferential propagation of the basal thrust further
520 away from the wedge and reduction of surface slope (Bonnet et al., 2007; Fillon et al., 2013;
521 Malavieille, 2010; Reitano et al., 2022; Simpson, 2006; Stockmal et al., 2007). The formation of
522 additional thrusts is also delayed due to the crustal thickening, and thus, further shortening required
523 to propagate deformation into the foreland. In contrast, the low erosion model continues to expand
524 with the formation of thrust sheets and outpaces the growth of the high erosion model.

525 In Figure 7, for the models including erosion, the relationship between surface uplift and wedge
526 propagation is well explained by the critical taper model (Dahlen, 1990). In these experiments the
527 wedge reaches a critical state by 70% shortening. At this point, material accretion is fully balanced
528 by erosion out of the wedge and the wedge ceases to grow (Hilley & Strecker, 2004; Willett, 1999).
529 The onset of a steady state condition is coincident with the appearance of a fully strike-slip
530 partitioned master fault with a component of dilation, and the establishment of a subparallel main
531 axial or *R*-shear drainage. Conversely, in the dry model, wedge growth begins to stabilize by 100%
532 shortening, yet a critical state is never fully attained. This observation suggests that the stream-
533 fault feedback is fundamental in attaining an erosional steady state condition.

534 The location and amplitude of salients and recesses in the thrust belt are also controlled by the
535 along-trend distribution of erosion (Graveleau & Dominguez, 2008; Liu et al., 2020; Marshak,
536 2004). With localized erosion, as expected where the tips of synthetic faults intersect thrust toes,
537 the wedge will be driven back to a supercritical state. At an intensified condition, the propagation
538 of the thrust belt will be locally limited at these intersections. Such interactions also help to explain
539 the relatively slow wedge growth in the high erosion model. As structurally controlled drainages
540 develop more rapidly, they concentrate sediment discharge and hamper wedge growth (Liu et al.,
541 2020). The width of a wedge in an oblique system is further affected by lateral block motion along
542 *R*- and *Y*-shears. The bookshelf-style faulting along these features generally reduces the width of
543 the wedge, linking the width of the wedge to the degree of strain partitioning.

544 The swath profiles from the high erosion model shown in Figure 7d highlight how erosion,
545 incision, and deformation modify the elevation and relief of the wedge at different shortening
546 stages. The corresponding surface uplift, rock uplift, and exhumation measurements from the
547 initial to the final stage are maximum estimates for the erosion-dominated end member. We find
548 that along the master fault-axial valley, there is 1.6-2.5 cm of rock uplift and 4-6 mm of
549 exhumation, corresponding to 1.2 – 1.9 cm of surface uplift. These estimates are reasonable when
550 transferred to natural systems. With our l^* of 1×10^{-5} , these measurements translate to 1.6-2.5 km,
551 400-600 m, and 1.2-1.9 km of rock uplift, exhumation, and surface uplift, respectively.
552 Considering these figures, exhumation in a transpressional setting is greatest where the major trunk
553 streams intersect the velocity discontinuity, which from Figure 5, is approximated by the drainage
554 divide. Thus, in thermochronometric studies, we should expect the youngest dates in these zones.

555 Our results indicate that the morphology of a transpressional wedge is linked to the systematics
556 of the potential feedback between faulting and incision. Specifically, valley orientation and shape
557 vary based on the amount of precipitation/erosion, the geometry of drainage networks, and the
558 degree and duration of strike-slip partitioning. Figure 9 synthesizes the evolution of stream and
559 fault networks, strain partitioning, and topography between high and low erosion systems. The
560 synthesized stages shown here follow an initial stage of distributed strain and correspond to the
561 strain evolution stages in section 5.2.

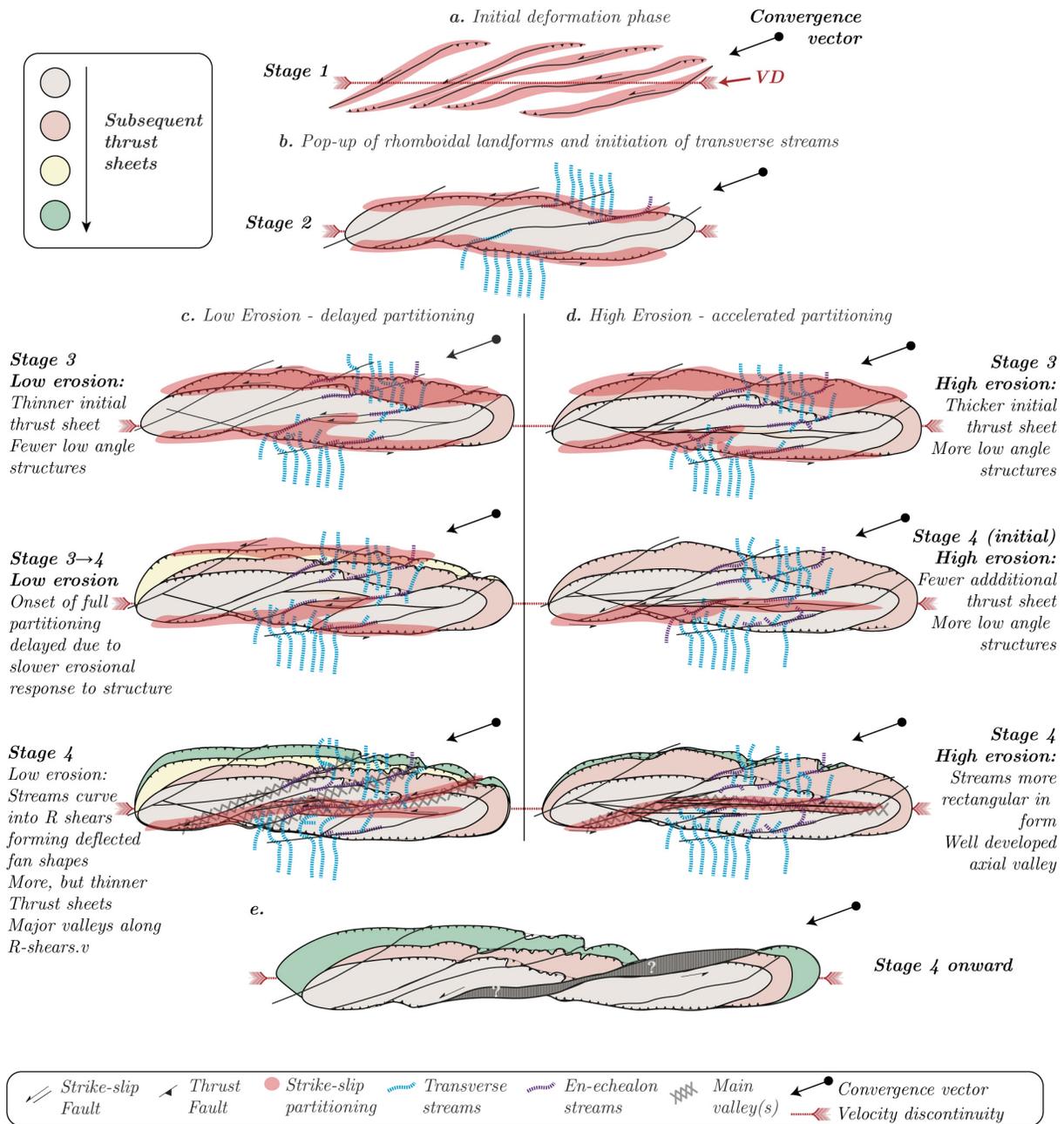


Figure 9: Illustration highlighting the differences between the evolution of fault and stream networks, strain partitioning, and topography. **(a)** The initial deformation phase. **(b)** Beginning stages of wedge development. **(c)** Stages 3 (transitional) – 4 (full partitioning) for low erosion systems. **(d)** Stages 3 (transitional) – 4 (full partitioning) for high erosion systems. **(e)** Wedge development after full partitioning.

562 **Stage 1 (pre-erosion):** R-shear structures accommodate strike-slip deformation and link
 563 laterally along thrust structures.

564 **Stage 2:** With the onset of erosion, transverse streams form along uplifted rhomboidal
 565 packages. Since wedge-bounding thrusts accommodate both the strike-slip and VD-perpendicular
 566 components of deformation, streams are mainly offset along these features.

567 **Stage 3:** A transitional phase when drainages actively respond to the progressive evolution of
 568 faults toward parallelism with underlying VD or passively rotate with simple shearing. Ridges and
 569 valleys transiently develop following the fault-stream feedback and progression toward complete
 570 partitioning. As controlled by the critical taper of the wedge, thrust sheets begin to form and
 571 propagate as a function of erosion rate and strike-slip displacement on *R-shears*. The low erosion
 572 model has more but thinner thrust sheets. In comparison, in the high erosion scenario, a wide thrust
 573 sheet forms early on and accommodates most convergence throughout the model evolution. The
 574 high erosion model also has more low-angle strike-slip structures (*Y* and *P* shears). Within a radius
 575 equal to the wavelength of the dominant valleys (~5 cm), the local topography along these
 576 structures exceeds 5 mm, the highest for incised valleys in the models. These deep valleys indicate
 577 that the collocated trunk streams and shears are significant incision points. Sediment routing out
 578 of these incised valleys locally induces a supercritical state in the wedge, limiting its propagation
 579 in the near field.

580 **Stage 4:** The wedge is fully partitioned with a well-developed master fault system. The main
 581 drainage is created by stream entrainment along the master fault forming a distinct axial valley.
 582 High volumes of sediment are routed out of orogen along this valley, and exhumation is localized.
 583 This phase is reached at lower shortening in the high erosion model due to the accelerated erosion-
 584 strain partitioning feedback. The rapid evolution of strain partitioning is facilitated by heightened
 585 incision and headward erosion, more vigorous stream reorganization, and mass removal along
 586 stream networks. For streams in the low erosion model, deflection, diversion, and erosion along *R*
 587 structures lead to asymmetric forked stream networks that curve into the wedge in the direction
 588 parallel to the convergence vector. In contrast, streams in the high erosion model are more
 589 rectangular, reflecting the more prevalent capture mechanisms and the change in prevailing
 590 structures from *R* to the more VD-parallel *Y*-shears and master fault.

591 **Post strain partitioning and wedge development:** Kinematic separation of rhomboidal
 592 landforms along the master fault–main valley feature with continuous strike-slip deformation and
 593 exhumation along the master fault system.

594 *5.1 Comparison with natural systems*

595 Our models have numerous simplifications, including the absence of a more ductile lower
 596 crust, which we know affects strain localization (e.g., Roy & Royden, 2000a, 2000b). Moreover,
 597 because our experiments couple nonlinear deformational and topography forming processes, it is
 598 challenging to extrapolate observations made within the time frame of experimental systems or
 599 over the seismic cycle to the deformation patterns observed in large and long-lived collisional
 600 zones. While attempts have been made, there is still work to be done to fully characterize the
 601 scaling of the material transport processes and the material's deformational behavior (see section
 602 2.1).

603 Concerning the boundary conditions, at depth, the VD set-up and activation of slip along the
 604 basement fault idealizes the propagation of a fault from a basement structure to an undeformed
 605 homogenous cover. Furthermore, there may be edge effects on the side of the shear zone, affecting
 606 the drainage patterns at later stages. These effects result from offsetting one side of the material
 607 package and exposing void space against the evolving orogen. Lastly, we ignore the impact of
 608 some erosional processes and modifiers, with examples including glaciation and vegetation.

609 Given those assumptions, the patterns in the models presented here still provide some insights
610 into fault development and propagation, strain partitioning, dynamic river network processes, and
611 topographic formation in transpressional orogens. There are several active or recently
612 transpressive systems around the globe where the results of this study are relevant, some of many
613 include the Central-Western Colombian Andes (Figure 10a; Cortés et al., 2005; Suter et al., 2008),
614 the Merida Andes in Venezuela (Figure 10b; Audemard & Audemard, 2002; Erikson et al., 2012),
615 and the central Transverse Ranges along the San Andres fault system (Figure 10c; Binnie et al.,
616 2008; Blythe et al., 2002; Matti & Morton, 1993). The annual precipitation in each range varies
617 between these orogens ranging from 200-300 cm yr^{-1} in the Central–Western Colombian Andes,
618 50–250 cm yr^{-1} in the Merida Andes, and 25 – 100 cm yr^{-1} in the Central Transverse Ranges
619 (estimated from WorldClim2, Fick & Hijmans, 2017). The key morphostructural similarities
620 between each natural prototype and presented models are exhibited in Figure 10.

621 Curiously, the early-stage macroscopic topographic features we observe only exist in some of
622 these orogens. Notably, orogen-scale repeated *R*-shear structures occur in few locations of wrench-
623 dominated deformation (e.g., Tchalenko & Ambraseys, 1970). While these strike-slip structures
624 are commonly observed at the micro or outcrop scale (e.g., Katz et al., 2004; Tchalenko, 1968) or
625 in analog experiments (Burbidge & Braun, 1998; Casas et al., 2001; Naylor et al., 1986; Schreurs
626 & Colletta, 1998; Tchalenko, 1970; Wilcox et al., 1973), the appearance of orogen-scale repeated
627 *R*-shear structures seems to only occur in some locations of wrench dominated deformation (e.g.,
628 Tchalenko & Ambraseys, 1970). Based on experiments and natural observations, Keller et al.
629 (1997) posit that *P* shears dominate over *R*-shears in zones of oblique convergent deformation.
630 The rarity of *R*-shear structures is likely linked to the varying kinematic modes of bulk strain
631 accommodation in zones of transpressional deformation due to lithological complexity, pre-
632 existing structural anisotropy, convergence angle, poor relief on faults or nearfield sedimentation,
633 and climatic control.

634 Furthermore, as shown in experiments from the literature en-echelon, *R* features are short-lived
635 structures considering the long-term evolution of an orogen (Wilcox et al., 1973), and thus, so is
636 the time scale to complete strain partitioning. As follows, en-echelon strike-slip structures should
637 be observed only in tectonically young orogeny (< 10 Myrs) with relatively consistent rheology
638 and constant convergence angles across the zone. In other words, the global examples of
639 transpressional tectonics only provide a snapshot of the overall evolution of a transpressional
640 wedge. Thus, the rapid progression through the initial stages biases observations toward the final
641 *R*-shear-absent configurations.

642 For the evolution and later topographic expression of transpression in nature to strongly
643 resemble our experiments, the region must also have a nearly single-phase tectonic history with a
644 limited amount of inherited structural anisotropy. The convergence angle also plays a significant
645 role. At low obliquities, there is little to no strike-slip partitioning. Whereas at high obliquities,
646 such as along the San Andreas fault system (Figure 9b), there are high degrees of strike-slip
647 partitioning (Teyssier et al. 1995). Furthermore, salt tectonics can modify the partitioning state and
648 lead to highly complex structures and landforms (e.g., Archer et al., 2012; Lohr et al., 2007).

649 The two natural prototypes that most resemble the models presented here are the Merida Andes
650 of Venezuela and the central Transverse Ranges of the San Andreas fault system (Figure 10b, c).
651 Though these systems are presently exposed to different climatic regimes (San Andreas – semi-
652 Arid; Merida Andes – tropical), we focus mainly on the general results of our experiments because
653 of other confounding variables present in natural systems (lithological heterogeneity, climatic

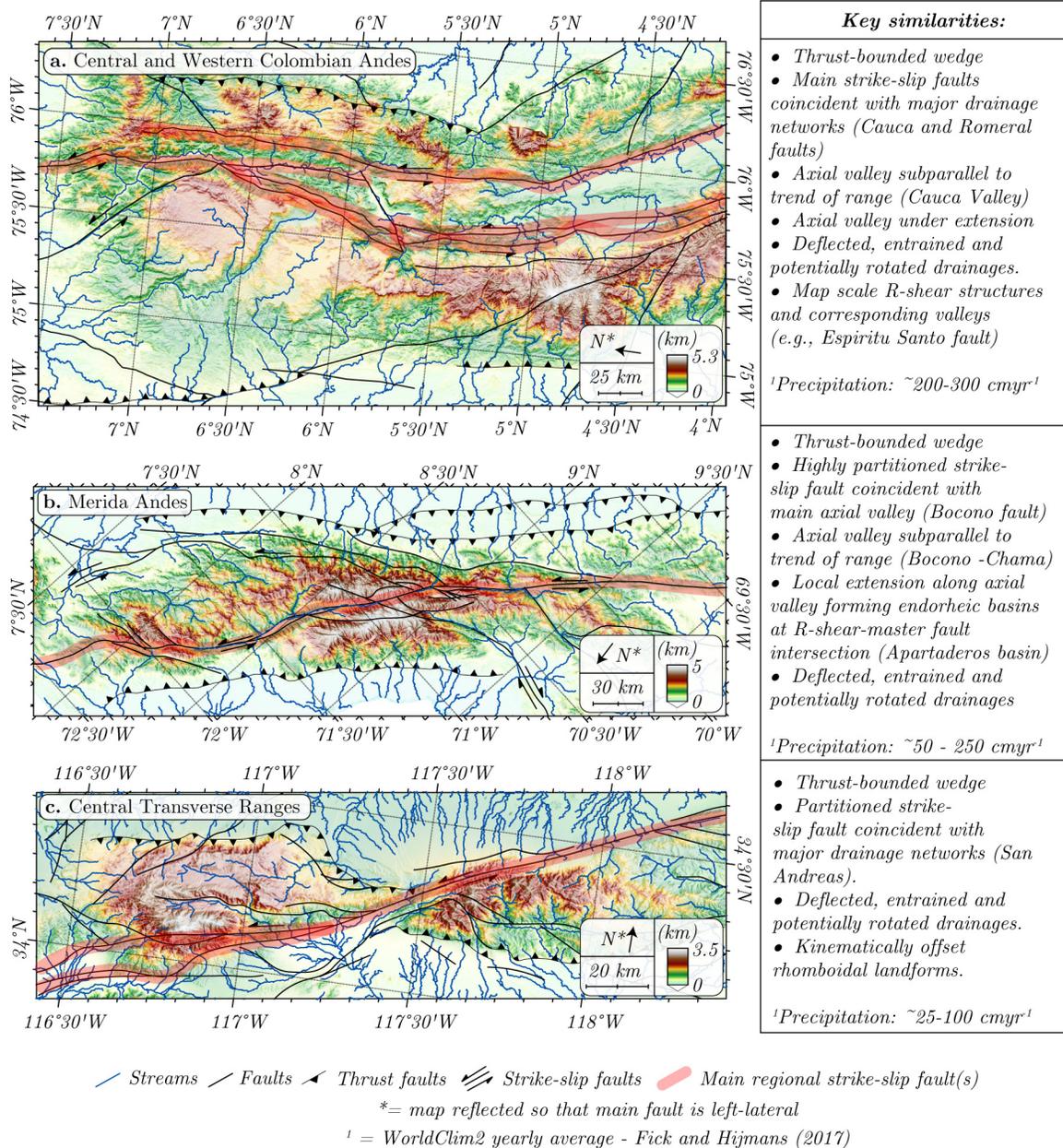


Figure 10: Natural prototypes with similar morphostructural characteristics as our models. **(a)** The Merida Andes in Venezuela. **(b)** The Central Transverse Ranges along the San Andreas Fault north of Los Angeles, California. **(c)** The Central and Western Colombian Andes. The similarities between each setting and the models are listed in the boxes on the right of the figure. Precipitation data is derived from WorldClim2 (Fick and Hijmans, 2017)

654 gradients, preexisting structures). These transpressional ranges predominantly exhibit the
 655 bivergent wedge structure bounding an uplifted zone of internally deformed topography that we
 656 observe in our models.

657 The Merida Andes (Figure 10b) is a roughly 350 km long × 100 km wide dextral
 658 transpressional mountain range that is thought to have begun significant deformation in the Late
 659 Miocene (F. Audemard, 1992; Colletta et al., 1997; Stephan, 1982). Strike-slip deformation is
 660 highly partitioned to the Bocono fault, a 500 km dextral strike-slip system, since 15 ± 2 Ma with
 661 slip rates of 7.3 – 10.7 mm/a (Audemard, 2003). River systems in the Merida Andes exhibit similar

662 patterns as those described in the laboratory models, including irregular or rectangular drainages,
663 prevalent wind gaps, beheaded or diverted rivers, and densely dissected fault scarps (Audemard,
664 1999). River channels exhibit deep incision with valley walls as high as 200 – 300 m (Audemard
665 & Audemard, 2002). The high incision and active drainage reorganization result from the highly
666 erosive setting with around 200 cm of yearly rainfall (Martin et al., 2020). The main valley cutting
667 through the Merida Andes follows the trace of the Bocono fault, as we observe in the high erosion
668 model. Considering the rock uplift estimates derived from our models, the trace of the Bocono
669 fault, especially in the center of the orogen, should also correspond to the highest amounts of rock
670 uplift and exhumation. Furthermore, the jog in the Bocono fault is appropriately aligned with the
671 ideal *R*-shear orientation and is likely influenced by pre-existing *R*-shear structures. At the tips of
672 the Bocono fault, large alluvial fans reflect its role in sediment routing from the internal portion of
673 the orogen. The ~160 km long by ~40 km wide triangular feature visible in the northwestern part
674 of the Merida Andes is likely formed by the intersection of the master fault with *R* and *P* shear
675 structures. These similarities with model results suggest that the Merida Andes is in stage 4 (Figure
676 9) of the development of the transpressional system. At this point, deformation and exhumation
677 may be localized to the Bocono fault system by the stream fault feedback. Additionally, pull apart
678 basins present along this structure (Audemard and Audemard, 2002) express the concentrated
679 extension we note in erosion model $\hat{\epsilon}_m$ maps. Projecting into the future, we expect the rhomboidal
680 landforms cut by the Bocono fault to offset left-laterally, as in Figure 9e.

681 The central Transverse Ranges of the San Andreas fault system (Figure 10b) are composed of
682 two distinct lenticular mountain ranges, the San Gabriel and San Bernadino mountains, separated
683 by the main strand of the San Andreas fault. Both mountain ranges are roughly 35 km wide and
684 100 km long. Beginning with the activation of the San Gabriel fault around 12 Ma, which presently
685 bounds the San Gabriel mountains to the south, the transverse ranges were uplifted to elevations
686 > 2000 m and vastly reconfigured. The San Bernadino block was translated as much as 200 km to
687 the southwest by motion along the main San Andreas fault strand starting as early as 5 Ma (Blythe
688 et al., 2002; Matti & Morton, 1993). The evolution of the Transverse ranges strongly resembles
689 the presented model for the topographic and deformational development of transpressional wedges
690 with the present configuration beyond stage 4 (Figure 9e).

691 Considering model observations, valleys corresponding to the master fault system should be
692 the locus of exhumation in transpressional ranges (Figure 8). Where erosion/denudation estimates
693 are available, the presented natural prototypes support this claim. In the central Transverse ranges
694 (Figure 10b), low-temperature thermochronometric ages (Buscher & Spotila, 2007) and
695 denudation rates from radionuclide dating (Binnie et al., 2008) support the hypothesized trends
696 with erosion/denudation rates increasing toward the main San Andreas fault strand. The same
697 trends are apparent further to the northwest in the San Emigdio and Mount Pinos regions, where
698 the western Transverse ranges accommodate most transpressional deformation. There, a low-
699 temperature north-south thermochronometric transect shows a substantial decrease in
700 thermochronometric dates from 19.4 ± 2.4 Ma to 4.4 ± 0.7 Ma across ~10 km. The youngest dates
701 are at higher elevations than the older dates within a hanging stream valley, similar to that observed
702 in our models (Niemi et al., 2013). In the Merida Andes (Figure 10a), though only higher
703 temperature $^{40}\text{Ar}/^{39}\text{Ar}$ thermochronometric data are available, the youngest Muscovite dates
704 (approximately 135 – 200 Ma) lie along the Bocono fault on the edges of the Chama valley near
705 the city of Mérida. Muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ dates are older outside this valley at around 250 – 425 Ma
706 (van der Lelij et al., 2016). Cosmogenic radionuclide dating (Ott et al., 2023) and low-temperature
707 thermochronology (Pérez-Consuegra et al., 2022) from the central and western Colombian Andes

708 (Figure 10c) show the highest erosion/denudation rates in the lofted Cauca valley (1 km elevation)
709 along the Romeral-Cauca fault systems. Though perhaps coincidentally, we note that the trend of
710 the main valley in each natural prototype supports the observed differences between the final
711 configurations of the presented high and low erosion models. In the more arid central Transverse
712 Ranges, the wedge trend and main valley trend are dissimilar. Meanwhile, in the wetter Central-
713 Wester Colombian Andes and Merida Andes the main valley – strike-slip feature is subparallel to
714 the wedge trend.

715 **5 Conclusions**

716 Erosion plays a significant role in the morphostructural evolution of transpressional systems.
717 High erosion models are characterized by more rectangular drainages and the earlier appearance
718 of low-angle (*Y*- and *P*-shear) structures. In the final stage, a highly partitioned master fault and
719 velocity discontinuity parallel (VD) axial valley form. Conversely, low erosion models have
720 drainage networks in the form of deflected fans. Their structural evolution progresses more slowly
721 with the protracted formation of a fully partitioned shear zone. Morphologically, major valleys in
722 the wedge instead follow the traces of synthetic *R*-shears. We propose that these differences are
723 the result of a feedback between stream and fault network development. With more erosion, this
724 feedback is augmented as drainages rearrange more vigorously and incise incipient and actively
725 evolving structures. Mass removal by incision leads to an adjustment in wedge stresses and
726 accelerated structural reconfigurations which accommodate greater portions of the wrench
727 component of deformation.

728 The results of our experiments assist in understanding patterns of rock and surface uplift and
729 exhumation in natural transpressional systems. The proposed feedbacks between incision and
730 strike-slip strain localization suggest that, in nature, deeply incised valleys should form along the
731 master fault. The location of this valley is influenced by the concentration of erosion energy due
732 to crustal weakening along fault strike. Maximum rock uplift occurs along the wedge axis, which
733 roughly aligns with the VD. Therefore, neglecting confounding variables, including lithology, and
734 pre-existing structure, the intersection of the VD with the throughgoing master fault–main valley
735 feature should be the location of the maximum amount of exhumation throughout the wedge. The
736 Merida Andes, Transverse Ranges, and central-western Colombian Andes each show patterns
737 demonstrating this trend.

738 We demonstrate that fault and drainage network development are linked to deformation and
739 exhumation patterns in a transpressional system. However, work must be done to fully understand
740 the complexities of the stream-fault feedback, including additional analog, numerical, and field
741 studies. Numerical models that pair the thermomechanical evolution of the wedge with surface
742 processes would be beneficial to clarify the physics of the system and more deeply explore the
743 parameter space. Additionally, continued tectonic-geomorphic field studies focused on continental
744 transpressional would provide the data necessary to interpret model results more rigorously in the
745 context of natural systems.

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757 **Open Research**

758 Digital elevation models, images used for particle image velocimetry analysis, and grid files of
759 velocity field are available for download from the Texas Data Repository (Conrad, 2023)

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