Morpho-tectonics of transpressional systems: insights from analog modeling

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Key Points:

- Feedback between fault and drainage network development regulates the deformation, exhumation, and morphology of transpressional systems
- Increased erosion accelerates the progression from distributed deformation to complete strike-slip strain partitioning
- Due to heightened rock uplift and incision, the maximum exhumation in a transpressional wedge is along the master fault and axial valley

Abstract

Transpressional margins are widespread, and their dynamics are relevant for plate boundary evolution globally. Though transpressional orogen evolution involves a topographic response to deformation, many studies focus only on the structural development of the system ignoring surface processes. Here, we present a new set of analog models constructed to investigate how tectonic and surface processes interact at transpressive plate boundaries and shape topography. Experiments are conducted by deforming a previously benchmarked crustal analog material in a meter-scale plexiglass box while controlling erosion through misting nozzles mounted along the transpressional wedge. To analyze the experiments, we generate digital elevation models from laser scans and conduct image correlation analysis on photos taken during experiments. We focus on three experiments that cover a range of erosional conditions and shortening stages (two end-member erosion models and a dry reference). In all experiments, a bivergent wedge forms, and strain partitioning broadly evolves according to previously established models. Regarding drainage networks, we find that the streams in our models develop differently through feedback between fault development and drainage rearrangement processes. Differences between end-member erosional models can be explained by the varying response of streams to structure modulated by rainfall. Additionally, erosion may influence the structural evolution of transpressional topography, leading to accelerated strike-slip partitioning. From these results, we create a model for developing structures, streams, and topography where incision and valley formation along main structures localize exhumation. We apply insights from the models to natural transpressional systems, including the Transverse Ranges, CA, and the Venezuelan Andes.

1. Introduction

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

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

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

Here, we present erosion tectonic sandbox experiments that investigate the relationships between fault structure, stream networks, and the strain field in transpressional systems. We attempt to identify the potential feedback between these components to explain morphological and deformational differences between experimental wedges for high- and low-erosion endmembers. Through analyses of digital elevation models and velocity fields from particle tracking, we address 1) how stream networks evolve in transpressional systems under variable erosional conditions, 2) if and how erosion influences the structural and morphological evolution of transpressional mountain belts, and 3) how strain partitioning evolves and is affected by structural and stream network development. The components of wedge evolution related to these questions are highlighted in Figure 1. We extend our results and analyses to natural transpressional prototypes, mainly focusing on the Merida Andes of Venezuela and the central Transverse Ranges along the San Andreas fault system in California, U.S.

2. Analog model: Erosion–tectonics sandbox

Previous analog studies of the evolution of transpressional mountain belts focus on the structural development of the model without including surface processes (Barcos et al., 2016; Cooke et al., 2020; Lallemand et al., 1994; Leever et al., 2011a,b; Pinet & Cobbold, 1992). Some workers have conducted laboratory studies that included the effects of erosion and sedimentation by removing and applying material by hand (e.g., Bonnet et al., 2007, 2008; Konstantinovskaia & Malavieille, 2005; Malavieille et al., 1993; Perrin et al., 2013). However, this approach limits the internal control of the system. Only a few models combine tectonic stresses and surface processes using misting systems that more realistically simulate the erosional processes acting on a deforming wedge (Graveleau et al., 2015; Graveleau & Dominguez, 2008; Guerit et al., 2016, 2018; Lague et al., 2003; Mao et al., 2021; Reitano et al., 2022; Viaplana-Muzas et al., 2015, 2019). These "erosion–tectonic" laboratory studies are often limited to purely compressional or extensional settings with few strike-slip (e.g., Graveleau et al., 2015) or transpressional (e.g., Guerit et al., 2016; Guerit et al., 2018) investigations. In the presented experiments, we add to the current collection of erosion-tectonic studies of transpression and aim to understand how deformation and wedge morphology evolve under the influence of structural and fluvial mechanisms in different erosional regimes.

2.1 Experimental material

Analog materials used in erosion-tectonic experiments should account for the first-order deformational and erosional behavior of the lithosphere (e.g., Graveleau et al., 2011). In addition, the material should scale appropriately, demonstrating geometric, kinematic, and dynamic similarity (Hubbert, 1951). Many granular single component (e.g., crushed quartz, silica powder) and composite materials (e.g., Mat IV or CM2) have been tested and shown to behave similarly to natural cases in a variety of geodynamic experiments (Graveleau et al., 2011). Our material, CM2, is a combination of 40 wt. % glass microspheres, 40 wt. % silica powder, and 20 wt. % PVC powder hydrated to 20 wt. % H_2O relative to the bulk mixture (Reitano et al., 2020). The water content of the material was measured by mass and volume, knowing the dry density of the material. Reitano et al. (2020) characterized CM2 following the work of Graveleau et al. (2011), who developed a similar material, Mat IV. Mat IV has the same composition as CM2, yet with 18 wt. % PVC and the addition of 2 wt. % graphite powder. These authors show that CM2 and Mat IV deform following the Mohr-Coulomb failure criterion and exhibit basin and channel characteristics akin to nature, including a balance between hillslope diffusion and channel incision. They are also velocity-weakening materials leading to stick-slip. We chose to use 20 wt.% H_2O because at this level, the material is close to saturation; with additional water, the material becomes fluidized and is no longer supported by the grain structure. At saturation, the infiltration capacity (the maximum rainfall rate that the material can absorb) is constant. If the rainfall rate exceeds the infiltration capacity, surface processes must move water through the system (Horton et al., 1933). The infiltration capacity of the material was designed to be minimal by optimizing its porosity, permeability, and grain size distribution (Reitano et al., 2020; Graveleau et al., 2011). The upper layer of the material is rather soft during the experiments but, like the crust, strengthens downward due to compaction and lithostatic pressure. Graveleau et al. (2011) posit that after roughly 1 cm, its mechanical behavior is consistent with frictional values measured at 20 wt. % H_2O. Thus, it is reasonable to assume that once the material package is set, its mechanical behavior is independent of differences in the amount of water delivered to the surface by variable rainfall rates.

2.2 Experimental setup

We conduct experiments in a 2 m × 1 m × 0.5 m plexiglass box, with ends left open for drainage (Figure 2a). The basal slope is fixed at 1º to ensure water exits the system. We set a mylar sheet inside the box and fix a 2 mm thick right-trapezoidal (25 × 100 × 60 × 96 cm) shaped plexiglass board to the sidewall. By pulling the mylar sheet beneath the board, we simulate oblique convergence (Figure 2b). The mylar sheet is pulled by attaching its ends to a wooden plank mounted to a *MecVel screwjack* and Electric Motors *Vela stm* electric motor. We load the board–sheet set up with a ~5 cm thick package of the experimental material hydrated to ~20 wt. % water (see section 2.1). The length and width of the material package are controlled to ensure that the edges do not influence the wedge's evolution or reach the sidewalls of the box. Free boundaries are particularly important on the fixed side of the model as it allows the wedge to form independent of the geometry of a rigid backstop (e.g., Guerit et al., 2016). This independence arises because the material properties, rather than the backstop dip, control the geometry of the wedge. Additionally, the thrust can propagate beyond the location of a would-be backstop.

We use the velocity discontinuity between the fixed board and moving sheet to localize deformation, forming a bivergent wedge in the material package. This approach is similar to Leever et al. (2011a,b) and the classic wrench experiments of Riedel (1929), where the velocity discontinuity simulates a basement fault beneath a homogenous sediment cover. We initiate surface processes using misting nozzles mounted on an aluminum crossbar aligned with the wedge trend (Figure 2a). These nozzles maintain a droplet size of fewer than 100 µm to avoid rain splash erosion (Bonnet et al., 2007; Graveleau et al., 2012; Lague et al., 2003; Reitano et al., 2022; Viaplana-Muzas et al., 2015, 2019).

2.3 Parameters varied

We selected three representative experiments (bold font in Table 1) out of six total that explored a more extensive range of rainfall and convergence settings. These additional experiments test the model sensitivity, ensure reproducibility, and explore the parameter space. The experiments were performed with a convergence angle of 20º to investigate wrench-dominated transpression (see Teyssier et al., 1995). Convergence rates ranged from 70 to 320 mm/hr, and rainfall rates from 20 to 34 mm/hr. The three presented experiments provide the most robust and comparable datasets considering the scope of this paper: "dry" (D_62422), "low erosion" (W_71322), and "high erosion" (W_62722). The prefixes D and W are for dry and wet experiments, and the suffix is the date experiments were conducted. In the table, the CR refers to a dimensionless quantity defined as the convergence over rainfall rate (Reitano et al., 2022). An infinite CR means the system is completely dry, whereas a CR of zero indicates the system is not tectonically loaded. Note that here the CR number is only used to organize the models based on the relationship between rainfall and convergence and not for scaling to nature. We present these tests as representative "wet" ("low" and "high erosion") and "dry" CR scenarios. Throughout the dry experiment, we lightly misted the surface of the material with a spray bottle to keep the material hydrated and its infiltration capacity low. Misting must be done sparingly to maintain the rheology of the medium at its saturation threshold but avoid surface process initiation. Due to the highly differing boundary conditions between the dry experiment with limited misting and the wet experiments with constant misting, the wet experiments are more directly comparable.

We consider results robust if the initial fault geometries and the dimensions and general shape of the wedge are reproducible between models. Furthermore, we ensured that throughout the main stages, there was no connectivity between drainages propagating from the edge of the model and those within the wedge. The latter is necessary because the edge drainages provide a lower base level. Therefore, once connected with the wedge network, these outlets would localize and increase the erosion mass flux out of the wedge, dominating the topography and drainage network morphology (Leopold & Bull, 1979). Within the limitations of erosion-tectonic analog modeling (e.g., Paola et al., 2009), we show the reproducibility of erosion models by conducting additional end-member experiments under comparable rainfall and convergence rates (W_62321, W70221, Figure S1).

2.4 Scaling

To dynamically scale experiments to natural systems, we should follow the principles outlined by Hubbert (1937) and Ramberg (1981). However, given uncertainties about the physics of surface transport, it is not entirely clear how to upscale surface processes. We follow previous work by only considering geometric and kinematic similarity rather than full dynamic scaling. This approach might be appropriate if erosional processes are scale-invariant Paola et al., 2009).

In terms of geometric comparison with nature, we define a length scaling factor, $l^* = l_{model}/l_{nature}$. Given the approximate dimensions of transpressional mountain belts ($l = \sim10^5$ m, $w = \sim10^4$ m) and that of the wedges generated in the experiment ($l = 1$ m, $w = 10^{-1}$ m), $l^* = 10^{-5}$, meaning 1 cm in the model represents 1 km in nature. To derive the time scaling factor, t^*, we use the erosion number approach of Reitano et al. (2022), which assumes a steady state between the mass flux of accreted material and that of eroded material in both model and nature. This assumption allows one to solve for t^* as $t^* = 4l^*/v^*$, where v^* is the convergence rate scaling factor ($v^* = 10^4 \ldots 10^5$). From this equation, we estimate $t^* = 4 \times 10^{-10} \ldots 4 \times 10^{-9}$, suggesting that 1 hour of model time corresponds to 30…300 kyr, as in prior work (Graveleau et al., 2011; Mao et al., 2021; Reitano et al., 2022). In this approach, the t^* equation is the inverted form of that for erodibility, which has units of time^{-1}. Therefore, the difference in material erodibility, k, between the models and nature can similarly be evaluated as $k^* = v^*/4l^* = 10^9 \ldots 10^{10}$.

As discussed in previous works (Graveleau et al., 2011; Paola et al., 2009; Peakall et al., 1996), it is challenging to downscale geomorphic processes from nature to models. This difficulty arises because, at present, there is no way to rigorously reduce the dynamic interaction between fluid media and transported particles due to the range of scales in which surface processes act (Peakall et al., 1996). Therefore, rainfall rates corresponding to erosion in arid and wet natural settings cannot be downscaled directly to the experiments. Instead, we present "endmember" erosion models as those with CR numbers around the upper and lower bounds of what is possible, given the technical limitations of the material and misting system (Text S1). Considering the systematic geomorphic variability present in model results (Figure 3; Figure S1) and following the work of previous authors (Graveleau et al., 2011; Graveleau et al., 2012, 2015; Graveleau & Dominguez, 2008; Mao et al., 2021; Reitano et al., 2023; Reitano et al., 2022; Reitano et al., 2020; Strak et al., 2011) we thus consider it appropriate to qualitatively compare the effects of heightened erosion in the models with those possible in nature.

2.5 Analysis

We monitor the structural and surficial evolution by scanning the model incrementally with a laser scanner to create digital elevation models (DEMs) and conducting particle image velocimetry (PIV) from photos taken every minute. Vertical and horizontal resolutions for the laser are 0.07 mm and 0.05 mm, respectively. Scans are taken at 10 cm, 15 cm, 25 cm, and 35 cm of convergence. We chose the first increment based on our preliminary experiments, where we determined that 10 cm of convergence creates sufficient relief (~1 cm) for realistic drainages to develop. The subsequent increases provide 10 cm increments (+30% shortening) of experiment evolution up to a maximum of 35 cm (considered 100% shortening). These stages are appropriate given spatiotemporal constraints, including the influence of edge effects, which grow with rain time and total displacement. After 35 cm of convergence, especially for high rainfall experiments, the influence of drainages and faults propagating from the boundary cannot be neglected. We use the MATLAB software TopoToolbox (Schwanghart & Scherler, 2014) to analyze DEMs and the corresponding stream networks across the experimental stages. Structural interpretations are made of each stage by pairing DEMs with photographs, which more clearly display structures lacking sufficient vertical offset to be resolved by the laser scanner.

We derive the evolution of the velocity field using a 2-D cross-correlation technique, Particle Image Velocimetry (PIV, see Raffel et al., 2007), with the MATLAB toolbox PIVlab (Thielicke & Sonntag, 2021). However, some unavoidable limitations and high amounts of noise are associated with using this technique in the presence of a rain system, as surface transport is also partially tracked, and mist affects the quality of the images. Yet, with a 1-minute capture rate, image pre-processing, and velocity filtering, PIV can provide insight into the differences between end member erosional cases. We preprocess images using a contrast-limited, adaptive-histogram equalization filter and auto-contrast stretch. To generate velocity fields, we use a Fast Fourier Transform PIV algorithm across a region of interest of 90 × 30 cm with an initial interrogation area of 45 mm, three passes down to 12.5 mm, and a Gauss 2 × 3 point sub-pixel estimator (see Thielicke & Sonntag, 2021). Resultant velocity fields were calibrated using a photo reference and analyzed and plotted using Generic Mapping Tools (Wessel et al., 2019). We calculate the horizontal component of the velocity, u, the maximum horizontal shear strain rates, $\dot{\varepsilon}_s$, and the dilatational strain rate, $\dot{\varepsilon}_m$. The strain components $\dot{\varepsilon}_s$ and $\dot{\varepsilon}_m$ allow us to analyze the localization of strike-slip and compressional/extensional deformation, respectively. For ease of comparison between frames of each experiment, we normalize the values by the standard deviation in each frame and denote the normalized u, $\dot{\varepsilon}_s$, and $\dot{\varepsilon}_m$ as $\hat{u}$, $\hat{\dot{\varepsilon}}_s$, and $\hat{\dot{\varepsilon}}_m$, respectively.

3 Results

3.1 Structural evolution

Figure 3a shows the DEM results of the three presented experiments (dry, low erosion, high erosion), and Figure 3b shows the interpreted structural evolution of all models. For reference, we include contrast-enhanced images of the final model stages in Figure 3c-d. To describe our models, we use the wrench fault terminology of Naylor et al. (1986). Fault progression begins with the appearance of *en-échelon* synthetic shears (*R*, 15º – 30º) that initiate sub-parallel with the convergence direction and delineate rhomboidal packages within the wedge. Within the viewing frames, these features form at least three clear packages, which can be directly compared between models. *R*-shears are accompanied by antithetic shears (*R'*, 65º - 90º) and connecting splays (*S*, > 17º). The left-lateral displacement of material packages initiates the main inboard (on the fixed side of the model) thrust (bottommost thrust in Figure 3b), followed by the formation of an outboard (on the moving side of the model) back-thrust. Together these features form "*pop-up*" structures, which accommodate the uplift of the blocks. While somewhat obscure in erosional models due to erosion/sedimentation, a notable low-angle shear striking in the opposite direction

(P, 180º - 165º) forms in all models outlining the bottom-left portion of an elongate diamond-shaped or *"pug-nosed"* landform (Figure 3a).

The main inboard thrust feature moves only a few centimeters during the evolution of each model (2.3 cm, 3.4 cm, and 3.3 cm for dry, low erosion, and high erosion models, respectively). The outboard thrust belt propagates throughout the model at distances depending on the presence/amount of erosion. Thrust sheets nucleate at the tips of the R-shears that extend into the undeformed inboard and outboard sections of the model. With further convergence, lower angle R-shears (Y, 0º - 15º) form and coalesce with P-shears into an anastomosing velocity discontinuity-parallel master fault zone. In some cases, R' oriented fractures accommodate an extensional component. In either case, these fractures have an apparent clockwise rotation as they are offset by the left lateral R-shear system. R-shears may have an extensional component after they no longer accommodate strike-slip motion, especially when they optimally intersect with the evolving master fault and develop a releasing bend. This occurrence is prevalent in the later stages of the model. R-shears also tend to form arcs concaving into the velocity discontinuity and sometimes form sharp cusps at the velocity discontinuity-fault intersections (Figure 3b). Overall, the structural evolution of the experiments agrees with what is described by prior analog studies of wrench-dominated fault zones (e.g., Casas et al., 2001; Cloos, 1928; Leever et al., 2011a,b; Naylor et al., 1986; Pinet & Cobbold, 1992; Riedel, 1929; Schreurs & Colletta, 1998, 2002; Tchalenko, 1968; Wilcox et al., 1973). Though challenging to interpret, as the material is monochromatic, high-contrast photos of cross sections cut through the high erosion model (Figure 3f, g) show the complex internal deformation within the wedge, interpreted as a thrust-bounded, upward tulip-shaped structure (Figure 3g).

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

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

3.2 Stream Evolution

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

Streams that initiate in the *R-shear* direction erode headward throughout the evolution of the model and follow the reorientation of *R-shear*s described above. With the left-lateral deflection of transverse drainages along *R-shear*s, these '*R*-streams' are captured, resulting in sharp cusps in the drainage topology. This pattern is apparent on both the main thrust and thrust belt sides of the wedge. However, on the main thrust portion, transverse streams dominate the evolving networks. In both models, major valleys capture the flow of several transverse streams following the *R*-shear structures in the low erosion model and the velocity discontinuity-parallel master fault in the high erosion model. The capture of this valley by transverse tributary drainages causes punctuated erosion events.

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

3.3 PIV analysis of velocities and strain-rates

From the normalized velocities, $\hat{u}$ (Figure 6a), and strain-rates, $\hat{\dot{\varepsilon}}_s$ (Figure 6b), and $\hat{\dot{\varepsilon}}_m$ (Figure 6c), derived through our PIV analysis, we recognize three main phases of strain-rate field evolution common for all models. In Figure 6d, we extract the structures corresponding to sharp gradients in $\hat{u}$ (Figure 6a) and related bands of high $\hat{\dot{\varepsilon}}_s$ (>1.5 $\hat{\dot{\varepsilon}}_s$, Figure 6a). Velocities are particularly useful in verifying shear zones and differentiating them from noise imposed by landsliding or mist interference (red blobs in Figure 6a).

Some early organization phases occur before the first panel in Figure 6, beginning with distributed deformation followed by shear strain localization along *R* faults at 6…9% shortening (not shown, stage 1 in Figure 6d). Shortly after, elevated $\hat{\dot{\varepsilon}}_s$ values are focused on the sides of the wedge (10 cm convergence or 30 % shortening), marking a phase of incomplete partitioning and oblique faulting on the major thrust structures (stage 2 in Figure 6d). The experiments then enter a transitional stage (stage 3 in Figure 6d). Synchronous with the structural evolution, strike-slip motion becomes increasingly localized on a narrow band of anastomosing strike-slip faults (Figure 6c). The high erosion experiment achieves near-complete strike-slip strain partitioning at 70% shortening (stage 4 in Figure 6d). We consider a system to be partitioned if there is a single continuous band of > 1.5 $\hat{\varepsilon}_s$. For the other experiments, a strain rate scenario resembling complete strike-slip partitioning is not reached until the final frame (35 cm or 100% shortening). This difference suggests that the strain partitioning evolution is accelerated in the high erosion model. The more prevalent noise from land sliding attests to more vigorous sediment routing out of the wedge. Balanced by compression at the boundaries, the interior of the wedge is under extension in each model and stage (Figure 6c). This band becomes localized to the master-fault zone as the models progress and compression becomes less organized. Additionally, more red anomalies (restraining bends?) within blue extensional bands occur along the master-fault zone in the dry model compared to wet models.

3.4 Topographic evolution

Initial topography forms along pop-up structures as rhomboidal slices (Figure 3). With the onset of thrust belt propagation, the topography develops transverse asymmetry, with one steep side corresponding to the main thrust and a broader side corresponding to the thrust belt. As can be seen from the differences between the dry and erosion models, hillslope diffusion and stream erosion drastically modify the topography by incising valleys and causing fault scarps to retreat inward toward the velocity discontinuity. Alluvial fans fill the recessed portions of ridges. As expected, there are broader and higher volume alluvial fans and more deeply incised channels in the high erosion model (Figure 3c-e). In the dry model, the final topography resembles an uplifted and broadly concave plateau. With increased erosion, the topography is more rugged and characterized by steeper peaks and more incised valleys. The topographic evolution of each model is captured in Figure 7a-c showing the changing maximum and mean width, elevation, and thrust belt gradient (elevation divided by distance from velocity discontinuity) across shortening stages. In all experiments, the maximum and mean values show similar trends. Thus, we will only discuss the mean, with the maximum serving as an upper bound. We calculated the mean width and distance from the velocity discontinuity by averaging the difference between 1,000 corresponding points on the inboard and outboard thrusts and velocity discontinuity, respectively.

Figure 7a shows the changing width of the wedges. The high erosion model shows an increase in width of ~4 cm after the first erosional stage, then a plateau with continued convergence. On the other hand, the low erosion model width increases by ~3 cm and continues to grow as the experiment continues. The dry model has a broader initial topography and shows slow and steady growth in the wedge width from 12 to 14 cm (Figure 7a). As the experiment evolves, it is marked by a higher curvature thrust belt and more salients and recesses. Furthermore, there is no channel incision, resulting in a broad wedge dissected only by strike-slip structures. The characteristic diamond or pug-nose shape of wrench-dominated systems is most evident in the dry model due to the intersection of *P* and *R-shear*s (Figure 3a). With increased erosion, structures that do not accommodate significant displacement become less apparent. For instance, in the erosion models, the scarp of the uplifted *P* and *R-shear*s that delineate the diamond structure is eroded in the outboard direction, nearly hiding the feature altogether. There are also differences in relief across strike-slip faults between models. For the high erosion model, *Y* structures have more relief. In the dry and low erosion model, relief is higher on *R* structures.

The surface uplift seems to progress similarly in all models, with only subtle differences in the rate and magnitude. From zero to 25 centimeters, each experiment shows an initial phase of more rapid uplift (3…3.5 cm at 25 cm of convergence) followed by a slow rise, perhaps approaching a limit of around 4 cm (Figure 7b). Figure 7c shows the elevation of the wedge divided by the distance of the thrust toe from the velocity discontinuity. The inverse tangent of the plotted values is effectively the slope of the thrust belt. Using a thrust belt dip of 25º…35º and the peak angle of internal friction, ϕ, of wet CM2 (ϕ = 25º…36º, from Reitano et al. 2020), the error window for the inverse tangent of the theoretical slope angle, α, from critical taper theory (Dahlen, 1990) is α = 15º…26º. The tangents of these alpha windows, α', are plotted in Figure 7c (α' = 0.27…0.49). Initially, the thrust belt slope in the dry and low erosion models increases into the α window. In contrast, the wedge slope in the high erosion model is stable at first, corresponding to the early propagation of a wide thrust sheet. All models reach a value of approximately 0.3 at 70% shortening. From 70% to 100% shortening, the wedge slope reaches a steady state in the wet models. In contrast, in the dry model, it continues to steepen.

A swath section of the evolution of the high erosion model is shown in Figure 7d, highlighting the occurrence of the velocity discontinuity-parallel valley and magnitude of stream incision. In this view, erosion by transverse streams is captured by the width of the colored area, while valleys along the section represent erosion by longitudinal streams. To provide a picture of the localization of exhumation across the profile, we isolated the swath profile at 100% shortening and found a line that captured local maxima and flat portions along the swath maximum (Figure 7e). This line may represent a rough estimate for rock uplift, assuming rock uplift is minimal. We then estimate the total exhumation driven by transverse and longitudinal stream incision as the difference between the rock uplift envelope and upper and lower swath bounds. There are four locations where a maximum exhumation value of 7 mm is reached. The minimum exhumation at some of these points ranges from 0…5 mm, depending on the presence of an incising longitudinal stream.

4 Discussion

4.1 Drainage evolution in response to transpressional tectonics

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

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

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

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

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

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

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

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

In general, the structural and strain partitioning results of our experiments agree with previous experiments using dry quartz sand (e.g., Leever et al., 2011a; Pinet & Cobbold, 1992; Schreurs & Colletta, 1998) and wet kaolin clay (Cooke et al., 2020). Variations in our results may be attributed to the tested convergence angle, the rheology of the material, erosion, and the approach to fault initiation. Using a similar velocity discontinuity approach, Leever et al. (2011a) built on the work of Pinet & Cobbold (1992), describing a 3-stage evolution of the strain field during transpression from distributed strain to full partitioning. Expanding the work of these authors, we describe the progression observed in our models (Figure 6) by combining wrench tectonics within the wedge (e.g., Naylor et al., 1986; Tchalenko, 1970; Wilcox et al., 1973) with the evolution described by Leever et al. (2011a).

Beginning with stage 1, following the period of distributed strain, strike-slip deformation is first accommodated along *R*-shear structures, as the principal infinitesimal strain axes are horizontal in wrench-dominated transpression (Tikoff & Teyssier, 1994). This order of fault formation (*en-échelon* *R*-shearing before thrusting) is also documented in previous experiments using differing rheologies under low-angle transpression (Cooke et al., 2020; Schreurs & Colletta, 1998). In stage 2, a slow-growing thrust forms on both sides of the velocity discontinuity, eventually resulting in a bivergent wedge (30% shortening in presented models). The complete formation of thrust structures bounding the material packages provides pervasive discontinuities in the system where oblique motion preferentially concentrates. The system then enters stage 3, a transitional stage (40% shortening in presented models), where low-angle structures (< 17º to the velocity discontinuity) and splay faults form, grow, and eventually link (see Naylor et al., 1986 for discussion). Stage 4 begins when a velocity discontinuity-parallel anastomosing "master fault" zone becomes apparent. Synchronously the zones of extension and principal shear narrow over the velocity discontinuity (70% and 100% shortening in high and low erosion models, respectively). Subsequent deformations are mostly independent, and bivergent thrusts now have a purely velocity discontinuity perpendicular dip-slip component, while the master fault system fully accommodates the strike-slip component of bulk strain.

We observe the above stages of strain evolution across all the presented models (Figure 6). However, the difference in shortening between when the high and low erosion models enter stage 4 suggests that strain partitioning is also dependent on the erosion/rainfall rate relative to the convergence rate. By 70% shortening in the high erosion model, a velocity discontinuity-parallel master fault system is evident (Figure 4b, c) with well-developed strike-slip partitioning (Figure 6). In the low erosion model, while there is an indication that these paired features are developing, the structure is geometrically and kinematically immature – in the context of a fully connected anastomosing master fault zone with localized strike-slip deformation. For the dry model, the velocity discontinuity-parallel strike-slip system is well-formed by 100% shortening with $\hat{\dot{\varepsilon}}_s > 1.5$, yet there is also $> 1.5\ \hat{\dot{\varepsilon}}_s$ on the outboard thrust.

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

We discuss two potential explanations for the earlier formation of a velocity discontinuity-parallel valley and earlier strike-slip partitioning: 1) focused mass removal by incision changing the stress balance in the material, thus exposing and localizing deformation earlier along actively developing *Y*- and *P*-shears, and 2) weakening of the fault by infiltration and water-induced friction reduction.

Considering the overburden removal mechanism, in a wedge loading scenario, there is a balance between tectonic forcing, fault friction, and the overburden due to topography. While local stress states may be complex, we can gain some insights by considering the surface transport-associated overburden modification, i.e., the unclamping due to the perturbance in pressure, $\Delta P = \rho g(h_0 - h)$, where ρ is the density of the material, g is the acceleration due to gravity, h_0 is the thickness of the wedge with no erosion, and h is the thickness of the wedge with erosion. Thus, given the density of the material is constant, the relative unclamping is simply, $h_0/(h_0 - h)$. To estimate this for the wet models, given variable across-trend incision patterns, we evaluate 5 cm wide swath profiles at 70% shortening, with a centerline across the midsection of the innermost rhomboidal package. We then consider the change in wedge thickness by erosion to be the difference in maximum and minimum elevations across the length of the swath profile. These values range from 0.5…8.5 mm and 0.25…6 mm for the high and low erosion cases, respectively. Thus, compared to the wedge thickness without erosion (maximum elevations across the swath on a 5 cm thick material package), there is an overall overburden reduction of up to 11%, with a maximum difference of ~3% between high and low erosion cases. While this change need not translate directly to fault stress mediation, Cooke et al. (2020) observe similar unclamping in transpression experiments using kaolinite clay, positing that increased slip on the wedge-bounding thrust results in extension in the hanging wall and unclamping of the strike-slip fault. Furthermore, since the initial slip style of the wedge-bounding thrusts is initially oblique, the reduction in vertical stress by erosion also assists in the rotation of the stress field so that the least principal stress is vertical, and the fault behaves as purely dip slip. The different stress states in the dry versus wet cases may also explain the observation that the models with erosion reach a steady state by ~70% shortening, yet the dry case does not until perhaps after ~100% shortening (Figure 7). While 11% is a relatively small fraction, the effect of unloading is sustained over the model run, and we consider a modified mechanical state due to unloading a plausible explanation of the observed strain localization.

The second mechanism, H_2O-related weakening, suggests that local differences in water content around faults in erosion models may reduce their strength. Faults in the model may provide pathways for fluids to enter, especially in the later stages, when extension along the master fault may provide additional space where water can more easily penetrate the material at depth. However, frictional weakening by fluid infiltration and pressurization only applies to a confined system (Terzaghi, 1943). Since faults in our model are surrounded by a permeable media, there should be no significant component of weakening by pore-fluid pressurization, as water can escape freely. A second H_2O-influenced consideration is that differences in internal friction by variable bulk water content may affect the angle at which faults initiate. According to the Coulomb-Mohr criterion, shears form at 45° - $\phi/2$ to the first principal stress, first with *R* shears, then later lower angle *Y* and *P* shears as the stress field rotates (Naylor et al., 1986). Therefore, if ϕ is lower, faults would initiate at a lower angle. Also, Burbidge and Braun (1998) suggest that ϕ affects the critical obliquity for partitioning, where a lower ϕ allows strain partitioning to develop at lower convergence obliquities. Thus, partitioning should develop more quickly when ϕ is low. However, given the low infiltration capacity of the material and that it is saturated at the outset of the experiments, there should be no substantial difference in water content for the bulk material package between erosion models. Thus, based on the fundamental material characteristics, we assume that the effect of the amount of water delivered by misters on differences in the evolution of strain partitioning in the models is minor.

4.3 Coupling between fault and stream networks to shape topography

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

The interplay between tectonic and erosional factors manifests in part as the trends we see in the lateral growth of the wedge between different models (Figures 3, 4, and 7; cf. Dahlen & Suppe, 1988; Steer et al., 2014). The high erosion model has fewer, yet wider, thrust sheets. The width of these sheets is a function of gradual thickening by syntectonic deposition of alluvium and more rapid and widespread erosion, leading to the preferential propagation of the basal thrust further away from the wedge and reduction of surface slope (Bonnet et al., 2007; Fillon et al., 2013; Malavieille, 2010; Reitano et al., 2022; Simpson, 2006; Stockmal et al., 2007). The formation of additional thrusts is also delayed due to the crustal thickening. Thus, further shortening is required to propagate deformation into the foreland. In contrast, the low erosion model continues to expand with the formation of thrust sheets and outpaces the growth of the high erosion model. The width of a wedge in an oblique system is further affected by lateral block motion along *R*- and *Y*-shears. The bookshelf-style faulting along these features generally reduces the width of the wedge, linking the width of the wedge to the degree of strain partitioning.

In Figure 7, for the models including erosion, the relationship between surface uplift and wedge propagation is well explained by the critical taper model (Dahlen, 1990). In these experiments, the wedge reaches a critical state by 70% shortening. At this point, material accretion is fully balanced by erosion out of the wedge and the wedge ceases to grow (Hilley & Strecker, 2004; Willett, 1999). The onset of a steady state condition coincides with the appearance of a fully strike-slip partitioned master fault with a component of dilation and the establishment of a subparallel main axial or *R*-shear drainage. Conversely, in the dry model, wedge widening stabilizes by 60% shortening (Figure 7a), yet a critical state is never fully attained (Figure 7c). This observation suggests that the stream-fault feedback is fundamental in achieving an erosional steady state condition.

The location and amplitude of salients and recesses in the thrust belt are also controlled by the along-trend distribution of erosion (Graveleau & Dominguez, 2008; Liu et al., 2020; Marshak, 2004). Erosion and sedimentation are localized where the tips of *R* shears intersect the outermost thrust. At these locations, the wedge is driven back to a supercritical state, potentially limiting thrust propagation. Such interactions may also help explain the relatively slow wedge growth in the high erosion model. Since structurally controlled drainages develop more rapidly, they concentrate sediment discharge and hamper wedge growth (Liu et al., 2020).

The swath profiles from the high erosion model shown in Figure 7d highlight how erosion, incision, and deformation modify the elevation and relief of the wedge at different shortening stages. The measurements given in Figure 7e capture the spatial exhumation patterns across the wedge by separating the contributions of transverse and longitudinal stream incision to exhume material. Maximum exhumation is reached in four locations across this profile, which, given an l^* of 1×10^{-5}, translates to 700 m of exhumation in nature and with a t^* of $4 \times 10^{-10} \ldots 4 \times 10^{-9}$, an exhumation rate of 0.5 … 5 km/Myr. This value is reasonable given the global range of long-term exhumation rates (e.g., Granger, 2007; Hecht & Oguchi, 2017, and references therein).

There are informative differences in the minimum values for the locations where exhumation reaches a maximum. A minimum of zero (e.g., position 4 in Figure 7d) suggests transverse stream incision is the dominant exhumation mechanism and is thus a local signal. Conversely, a value close to the maximum (e.g., position 2 in Figure 7d) means that exhumation is continuously high along strike. In other words, exhumation in a transpressional setting is greatest where the major trunk streams intersect the velocity discontinuity because both transverse streams and the longitudinal valley contribute to exhuming rock. Furthermore, the later capture of this main valley by transverse streams, as posited for transpressional systems by (Babault et al., 2012), would modify the sediment routing and transiently accelerate erosion (Bishop et al., 1995). As observed in Figure 5, the location of these transverse streams is likely controlled by the *R*-shear structure. Thus, in thermochronometric studies, we might expect the youngest dates at these intersections.

Our results indicate that the morphology of a transpressional wedge is linked to the systematics of the potential feedback between faulting and incision. Specifically, valley orientation and shape vary based on the amount of precipitation/erosion, the geometry of drainage networks, and the degree and duration of strike-slip partitioning. Figure 9 synthesizes the evolution of stream and fault networks, strain partitioning, and topography between high and low erosion systems. The synthesized stages described here follow an initial stage of distributed strain and correspond to the strain evolution stages in section 4.2.

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

Stage 2: With the onset of erosion, transverse streams form along uplifted rhomboidal packages. Streams are mainly offset along wedge-bounding thrusts since, at this stage, they accommodate both the strike-slip and velocity discontinuity-perpendicular components of deformation.

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

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

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

4.4 Comparison with natural systems

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

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

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

Curiously, the early-stage macroscopic topographic features we observe only exist in some of these orogens. Notably, orogen-scale repeated *R*-shear structures occur in few locations of wrench-dominated deformation (e.g., Tchalenko & Ambraseys, 1970). Based on experiments and natural observations, Keller et al. (1997) posit that *P-shears* dominate over *R-shear*s in zones of oblique convergent deformation. The rarity of *R*-shear structures is likely linked to the varying kinematic modes of bulk strain accommodation in zones of transpressional deformation due to lithological complexity, pre-existing structural anisotropy, convergence angle, poor relief on faults or nearfield sedimentation, and climatic control.

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

For the evolution and later topographic expression of transpression in nature to strongly resemble our experiments, the region must also have a nearly single-phase tectonic history with a limited amount of inherited structural anisotropy. The convergence angle also plays a significant role. At low obliquities, there is little to no strike-slip partitioning, while at high obliquities, such as along the San Andreas fault system (Figure 10c), there are high degrees of strike-slip partitioning (Teyssier et al. 1995). The first-order differences in the structural style, degree of partitioning, and morphology between each of these examples are likely related to this point, as the estimated convergence angles are different for each of the examples presented in Figure 10 at 60º for the central and western Colombian Andes, 25º for the Merida Andes (Mora-Páez et al., 2019) and 30º for the central Transverse Ranges (McCaffrey, 2005). Though these estimates are from kinematic block models using current GPS velocities and do not account for complexities related to comparing long-term and current convergence vectors, such as block rotations. Lastly, salt tectonics can modify the partitioning state and lead to highly complex structures and landforms (e.g., Archer et al., 2012; Lohr et al., 2007).

The two natural prototypes that most resemble the models presented here are the Merida Andes of Venezuela and the central Transverse Ranges of the San Andreas fault system (Figure 10b, c). Though these systems are presently exposed to different climatic regimes (San Andreas – semi-Arid; Merida Andes – tropical), we focus mainly on the general results of our experiments because of other confounding variables present in natural systems (lithological heterogeneity, climatic gradients, preexisting structures). These transpressional ranges predominantly exhibit the bivergent wedge structure bounding an uplifted zone of internally deformed topography that we observe in our models.

The Merida Andes (Figure 10b) is a roughly 350 km long × 100 km wide dextral transpressional mountain range that is thought to have begun significant deformation in the Late Miocene (Audemard, 1992; Colletta et al., 1997; Stephan, 1982). Strike-slip deformation is highly partitioned to the Bocono fault, a 500 km dextral strike-slip system, since 15 ± 2 Ma with slip rates of 7.3 – 10.7 mm/a (Audemard, 2003). River systems in the Merida Andes exhibit similar patterns as those described in the laboratory models, including irregular or rectangular drainages, prevalent wind gaps, beheaded or diverted rivers, and densely dissected fault scarps (Audemard, 1999). River channels exhibit deep incision with valley walls as high as 200 – 300 m (Audemard & Audemard, 2002). The high incision and active drainage reorganization result from the highly erosive setting with around 200 cm of yearly rainfall (Martin et al., 2020). The main valley cutting through the Merida Andes follows the trace of the Bocono fault, as we observe in the high erosion model. Considering the exhumation estimates derived from our models, the trace of the Bocono fault, especially in the center of the orogen, should also correspond to the highest amounts of exhumation. Furthermore, the jog in the Bocono fault is appropriately aligned with the ideal *R*-shear orientation and is likely influenced by pre-existing *R*-shear structures. At the tips of the Bocono fault, large alluvial fans reflect its role in sediment routing from the internal portion of the orogen. The ~160 km long by ~40 km wide triangular feature visible in the northwestern part of the Merida Andes is likely formed by the intersection of the master fault with *R* and *P* shear structures. These similarities with model results suggest that the Merida Andes is in stage 4 (Figure 9) of the development of the transpressional system. At this point, deformation and exhumation may be localized to the Bocono fault system by the stream fault feedback. Additionally, pull-apart basins along this structure (Audemard and Audemard, 2002) express the concentrated extension we note in erosion model $\hat{\varepsilon}_m$ maps. Projecting into the future, we expect the rhomboidal landforms cut by the Bocono fault to offset left-laterally, as in Figure 9e.

The central Transverse Ranges of the San Andreas fault system (Figure 10c) are composed of two distinct lenticular mountain ranges, the San Gabriel and San Bernadino mountains, separated by the main strand of the San Andreas fault. Both mountain ranges are roughly 35 km wide and 100 km long. Beginning with the activation of the San Gabriel fault around 12 Ma, which presently bounds the San Gabriel mountains to the south, the transverse ranges were uplifted to elevations > 2000 m and vastly reconfigured. The San Bernadino block was translated as much as 200 km to the southwest by motion along the main San Andreas fault strand starting as early as 5 Ma (Blythe et al., 2002; Matti & Morton, 1993). The evolution of the Transverse ranges strongly resembles the presented model for the topographic and deformational development of transpressional wedges with the present configuration beyond stage 4 (Figure 9e). However, slight differences include more radial drainage patterns around the uplifted San Gabriel and San Bernadino blocks, which are instead bounded by the master fault (San Andreas) and a single wedge-bounding thrust (Figure 10c). These variations may relate to the structural and drainage evolution characteristics specific to local transpression along a restraining bend rather than a more continuous continental transform.

Considering model observations, valleys corresponding to the master fault system should be the locus of exhumation in transpressional ranges (Figure 8). Where erosion/denudation estimates are available, the presented natural prototypes support this claim. In the central Transverse ranges (Figure 10c), low-temperature thermochronometric ages (Buscher & Spotila, 2007) and denudation rates from radionuclide dating (Binnie et al., 2008) support the hypothesized trends with erosion/denudation rates increasing toward the main San Andreas fault strand. The same trends are apparent further to the northwest in the San Emigdio and Mount Pinos regions, where the western Transverse ranges accommodate most transpressional deformation. There, a low-temperature north-south thermochronometric transect shows a substantial decrease in thermochronometric dates from 19.4 ± 2.4 Ma to 4.4 ± 0.7 Ma across ~10 km. The youngest dates are at higher elevations than the older dates within a hanging stream valley, similar to that observed in our models (Niemi et al., 2013). Though, as discussed by Niemi et al., these differences may be partially related to rheological heterogeneity across the San Andreas Fault, which may also apply to the central Transverse Ranges. In the Merida Andes (Figure 10a), though only higher temperature $^{40}Ar/^{39}Ar$ thermochronometric data are available, the youngest Muscovite dates (approximately 135 … 200 Ma) lie along the Bocono fault on the edges of the Chama valley near the city of Mérida. Muscovite $^{40}Ar/^{39}Ar$ dates are older outside this valley at ~ 250 … 425 Ma (van der Lelij et al., 2016). Cosmogenic radionuclide dating (Ott et al., 2023) and low-temperature thermochronology (Pérez-Consuegra et al., 2022) from the central and western Colombian Andes (Figure 10c) show the highest erosion/denudation rates in the lofted Cauca valley (1 km elevation) along the Romeral-Cauca fault systems. Though perhaps coincidentally, we note that the trend of the main valley in each natural prototype supports the observed differences between the final configurations of the presented high and low erosion models. In the more arid central Transverse Ranges, the wedge and main valley trends are dissimilar. Meanwhile, in the wetter central-western

Colombian Andes and Merida Andes, the main valley–master fault feature is subparallel to the wedge trend.

5 Conclusions

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

The results of our experiments assist in understanding patterns of uplift and exhumation in natural transpressional systems. The proposed feedbacks between incision and strike-slip strain localization suggest that, in nature, deeply incised valleys should form along the master fault. The location of this valley is influenced by the concentration of erosion energy due to crustal weakening along fault strike. Maximum exhumation occurs along the wedge axis, which roughly aligns with the velocity discontinuity. Therefore, in natural systems, neglecting confounding variables, including lithology, and pre-existing structure, the intersection of the velocity discontinuity with the throughgoing master fault–main valley feature should yield the highest exhumation rates. The Merida Andes, Transverse Ranges, and central-western Colombian Andes each show patterns demonstrating this trend.

We infer that fault and drainage network development are linked to deformation and exhumation patterns in a transpressional system. However, work is needed to fully understand the complexities of the stream-fault feedback. Numerical models that pair the thermomechanical evolution of the wedge with surface processes would be beneficial to clarify the physics of the system and more deeply explore some of the interactions. Additional tectonic-geomorphic field studies focused on continental transpressional systems could provide the data necessary to interpret model results more rigorously in the context of natural systems. Despite the work still to be done, the results of this study provide insight into the range of possible questions, complexities, and future research directions related to the dynamic interactions between processes at the surface and at depth.

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Open Research

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

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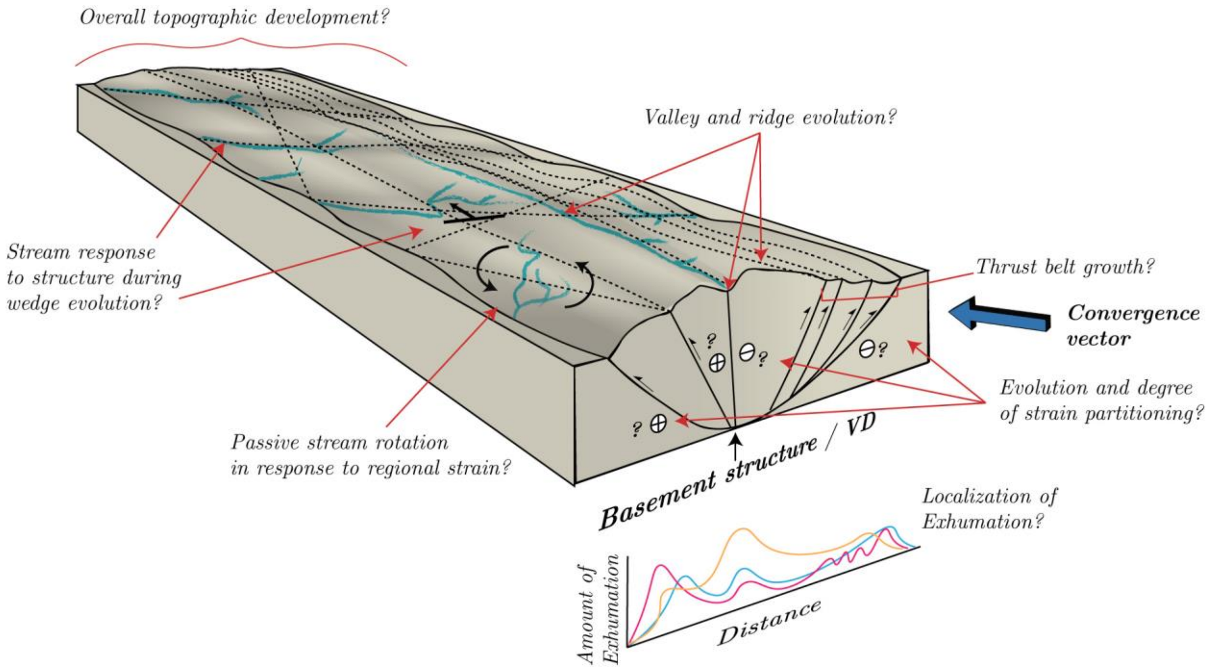

Figure 1. Illustration highlighting the unresolved components of transpressional (left-lateral) wedge growth addressed in this paper. The velocity discontinuity (VD) and convergence vector are in bold font. Dashed lines on topography show fault traces. The x-y plots show hypothetical exhumation patterns (colored 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.

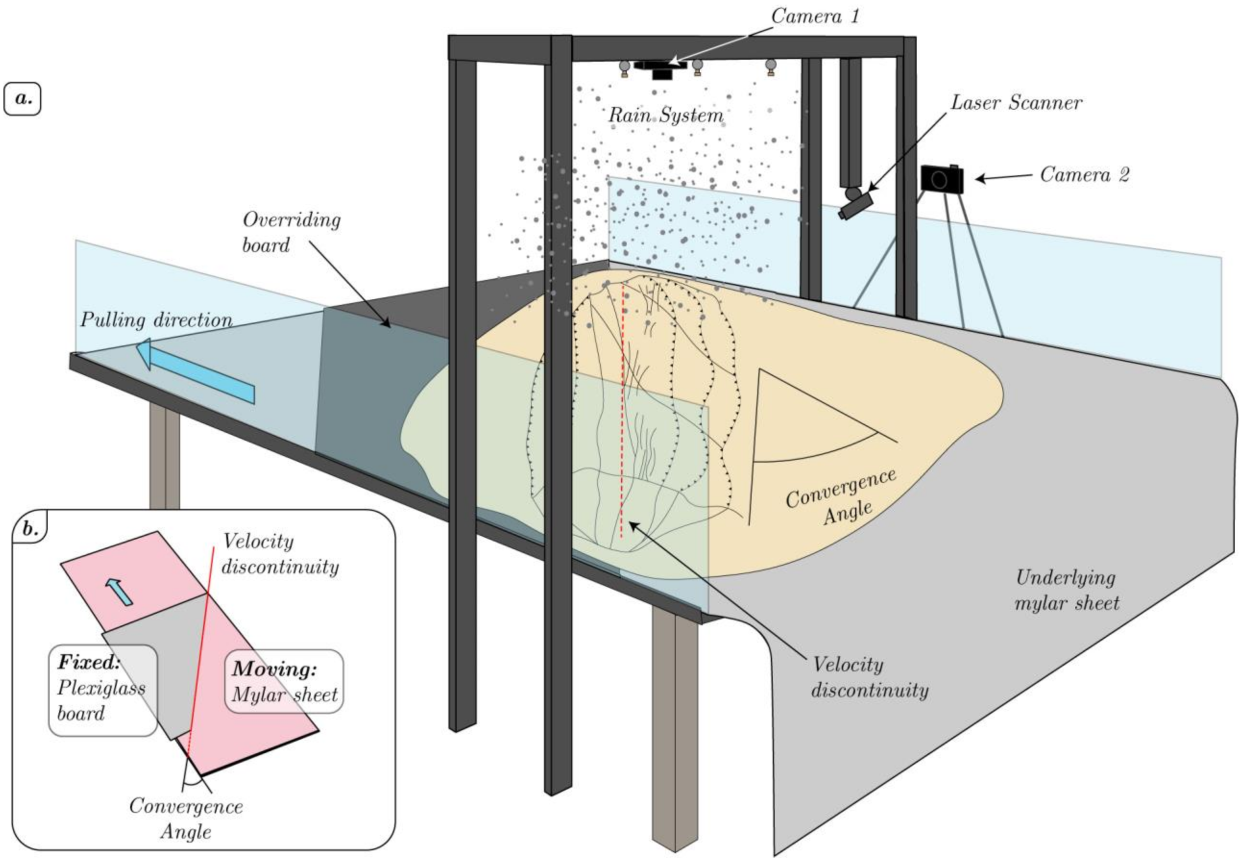

Figure 2. Experimental setup (a) Cartoon showing the full sandbox model set-up. The experiment comprises a 2 × 1 × 1 m plexiglass box set on a tabletop. The beige feature represents the material loaded on top. The red line shows the velocity discontinuity, and the blue arrow on the left is the pulling 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 (25 × 100 × 60 × 96 × 0.2 cm). 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.

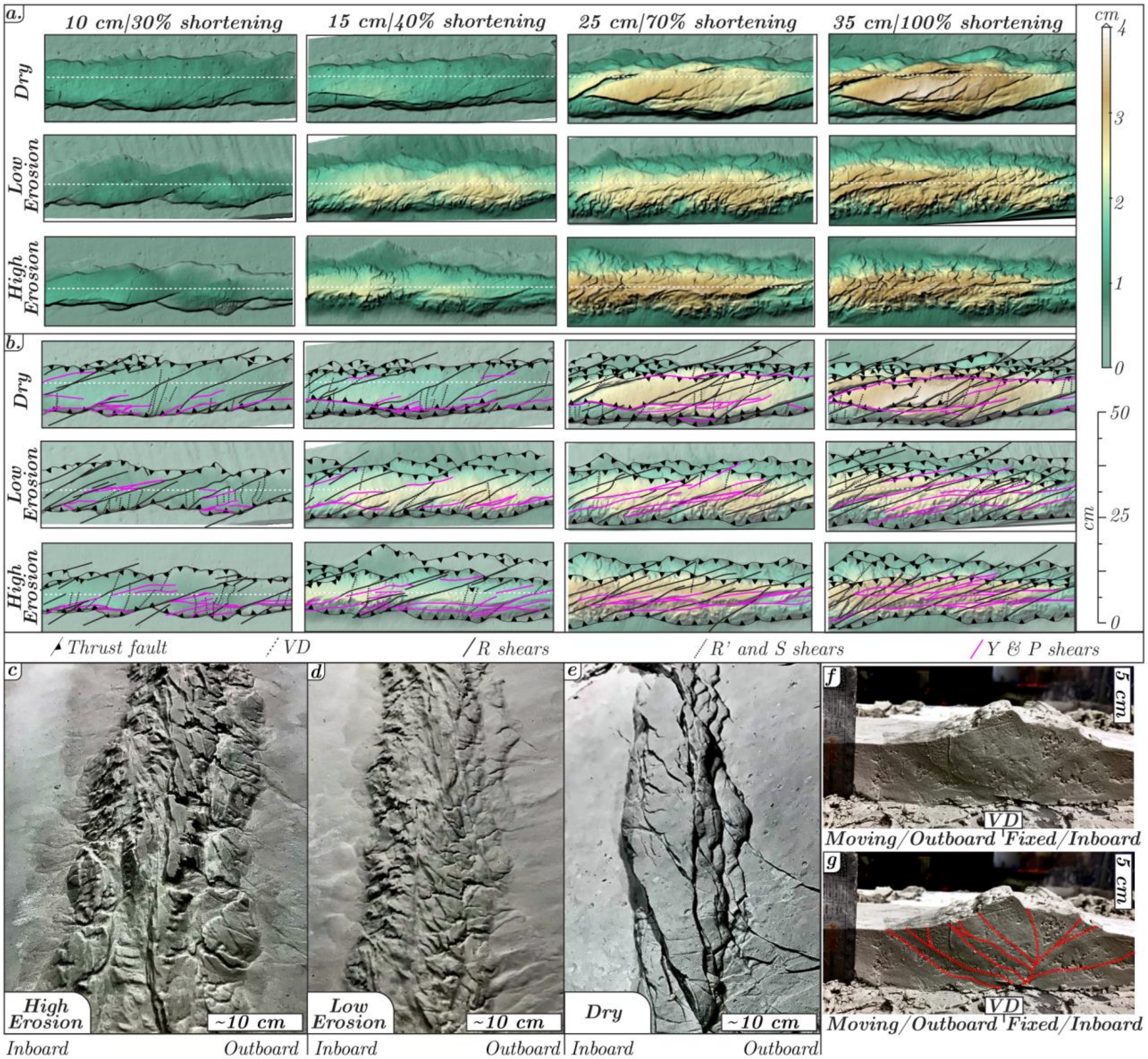

Figure 3. Overview of the results of selected experiments. (a) Digital elevation models (DEMs) across shortening stages. Rows show increasing erosion and columns show increasing convergence/shortening. White lines represent the location of the velocity discontinuity (VD). (b) Interpreted structures showing thrusts (decorated lines), the VD (white dashed line), riedel (*R*) shears (black lines), anti-riedel (*R'*) shears and connecting splay (*S*) faults (dotted black lines), and low-angle faults (*Y* and *P* shears, see text for description, magenta 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) interpreted cross sections of the high erosion model.

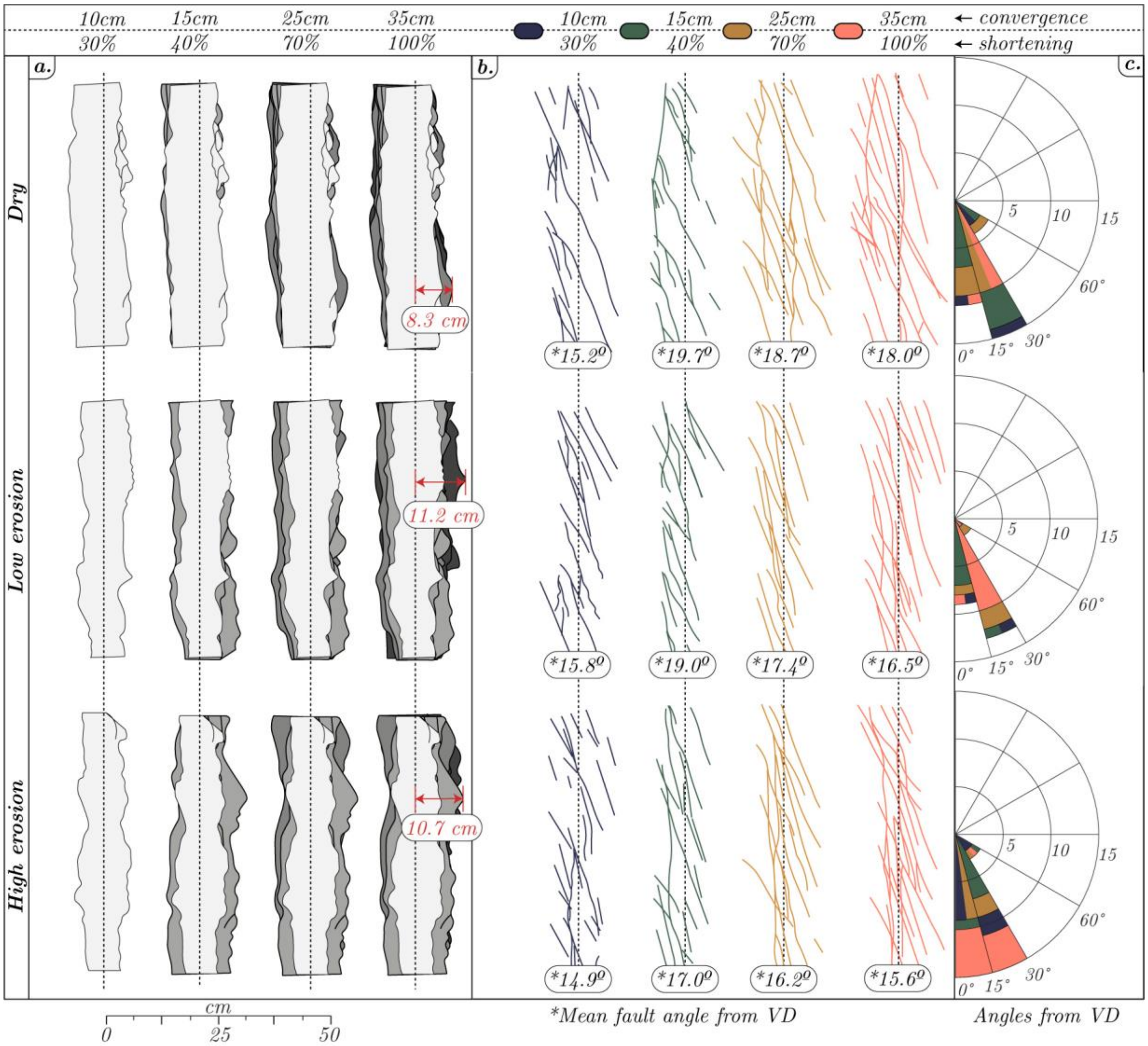

Figure 4. The evolution of thrust sheets and internal strike-slip faults with increasing shortening. The upper, middle, and lower rows show the dry, low erosion, and high erosion models, respectively. (a) Change in thrust belts with increasing convergence. Each grey shade represents a horizontal slice 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 15º bins from 0 to 60º. 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 > 60º and < 0º were included in the nearest bins because they are uncommon and deviate only small amounts from those angles.

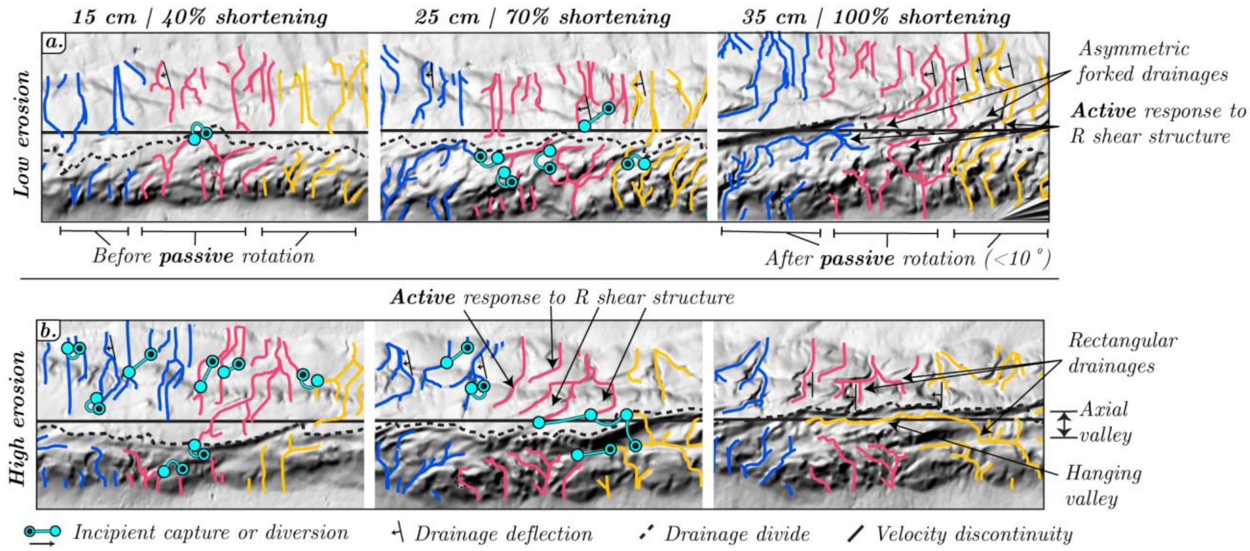

Figure 5. Snapshots of stream evolution between erosion models. Streams are cut to the lowest thrust sheet and colored as a visual aid to tracking reorganization mechanisms between frames. (a) Low erosion model. The active drainage response to structure is more delayed, while the passive rotation is better expressed (preserved). Drainage networks exhibit asymmetric forked patterns in the final stage. (b) High erosion 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.

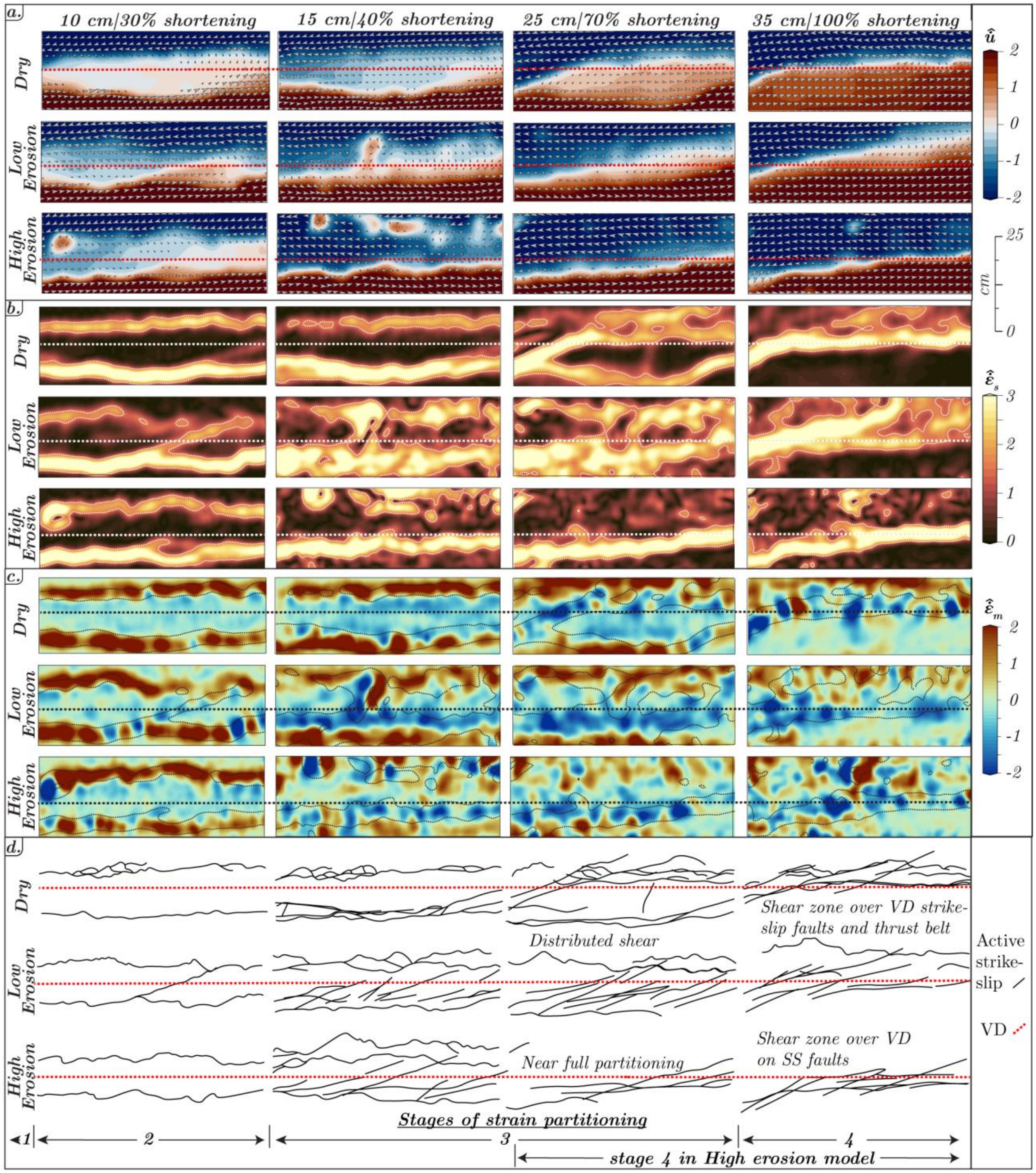

Figure 6. Evolution of strain partitioning from PIV at the same stages shown in Figures 2, 3, and 4. (a) The normalized horizontal velocity components ($\hat{u}$). Sharp color contrast indicates locations of strike-slip deformation. Irregular red blobs show locations of land sliding or

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 ($\hat{\dot{\varepsilon}}_s$). White dotted lines are superimposed $\hat{\dot{\varepsilon}}_s = 1.5$ contours. The dashed white line indicates the VD. (c) The normalized dilatational strain-rate, ($\hat{\dot{\varepsilon}}_m$). Positive values (red) indicate compression and negative values (blue) indicate extension. Black dotted lines are superimposed $\hat{\dot{\varepsilon}}_s = 1.5$ contours. The dashed black line indicates the VD. (d) Active faults are determined by interpreting fault locations with respect to $\hat{u}$ and $\hat{\dot{\varepsilon}}_s$. The dashed red line indicates the VD. The evolution is divided into four stages of strain partitioning: 1) distributed strain and *en échelon R*-shear formation, 2) oblique slip on bivergent thrusts, 3) transitional strain partitioning, 4) full partitioning to throughgoing structure (s).

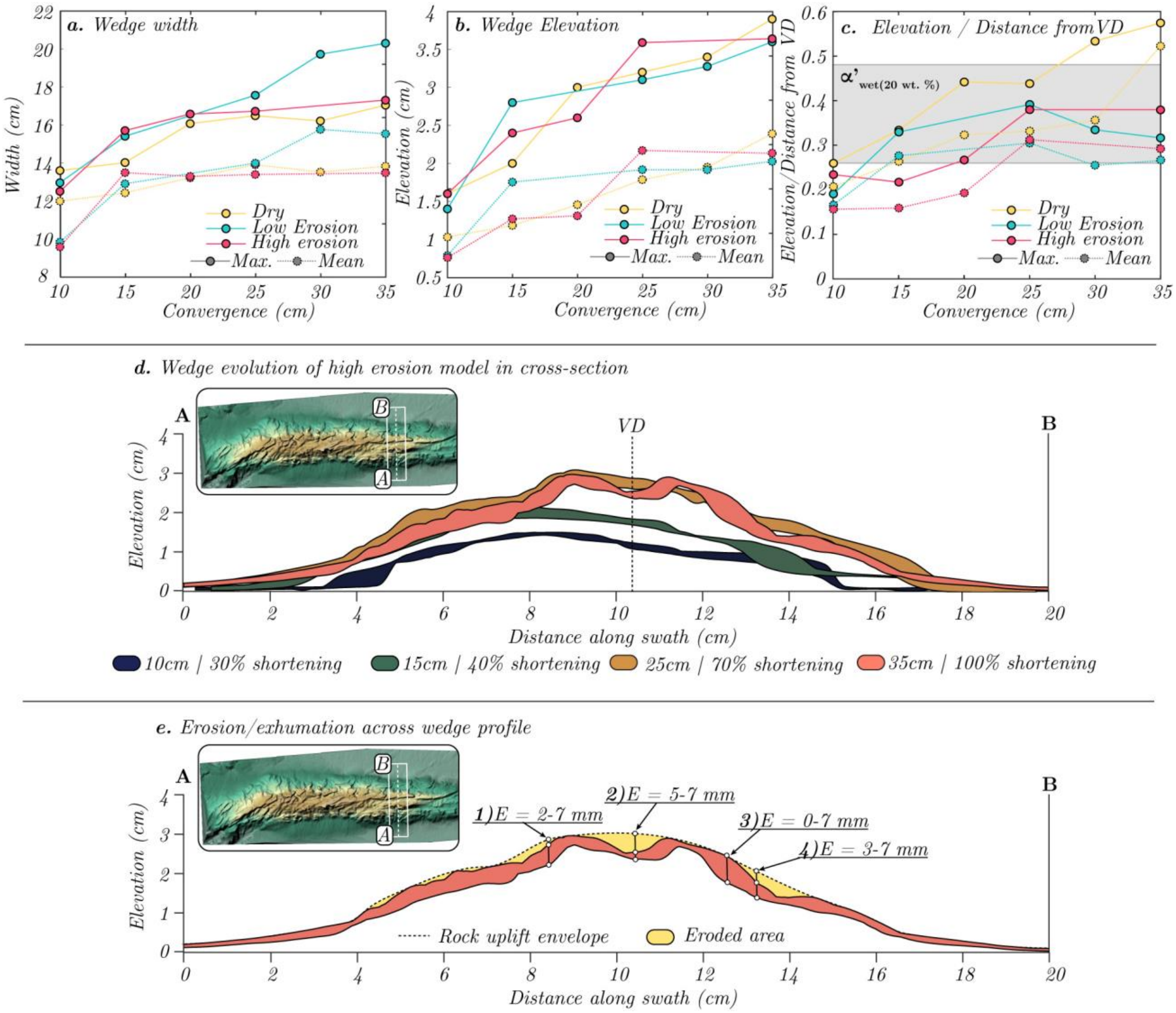

Figure 7. The topographic evolution of the presented models. (a-c) maximum and mean (a) wedge width, (b) elevation, and (c) elevation divided by the distance from the velocity discontinuity (VD). The gray bar is the error window for the theoretical wedge slope calculated from critical taper theory (see Dahlen, 1990) using the material parameters for wet (20 wt. % H_2O) CM2 from Reitano et al. (2020), α'wet. (d) Swath section showing the wedge evolution of the high erosion model. The inset shows the location of the swath at the final stage (convergence = 35 cm/shortening = 100%). (e) Estimates of exhumation across the swath taken at the final shortening stage (100%). The eroded area (yellow) is the difference

between the rock uplift envelope (dashed line) and the surface uplift (upper and lower swath bounds). Erosion/exhumation estimates, E, are the difference between rock and surface uplift (Molnar and England, 1990).

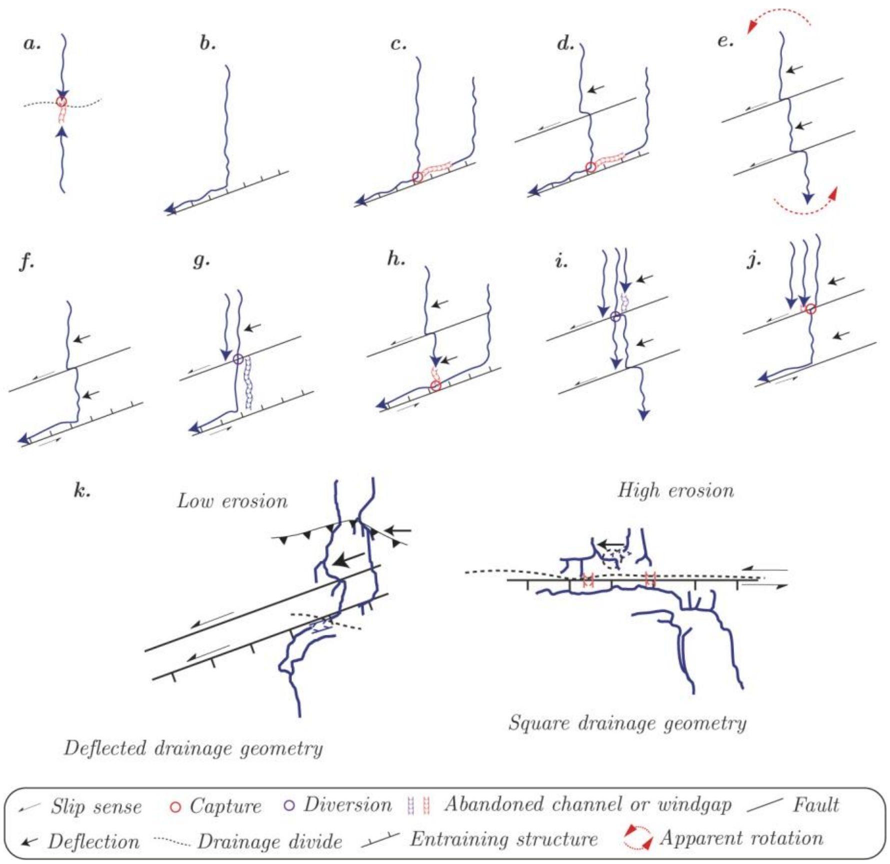

Figure 8. Examples of drainage reorganization mechanisms leading to the stream network geometries in the high and low erosion models. (a) across divide capture, (b) structural entrainment, (c) structural entrainment and capture upstream along *R*-shear, (d) deflection, entrainment, and capture upstream along *R*-shear, (e) deflection, and rotation, (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. It is evident that the structural evolution impacts the drainage geometry forming more rectangular drainages in the high erosion model.

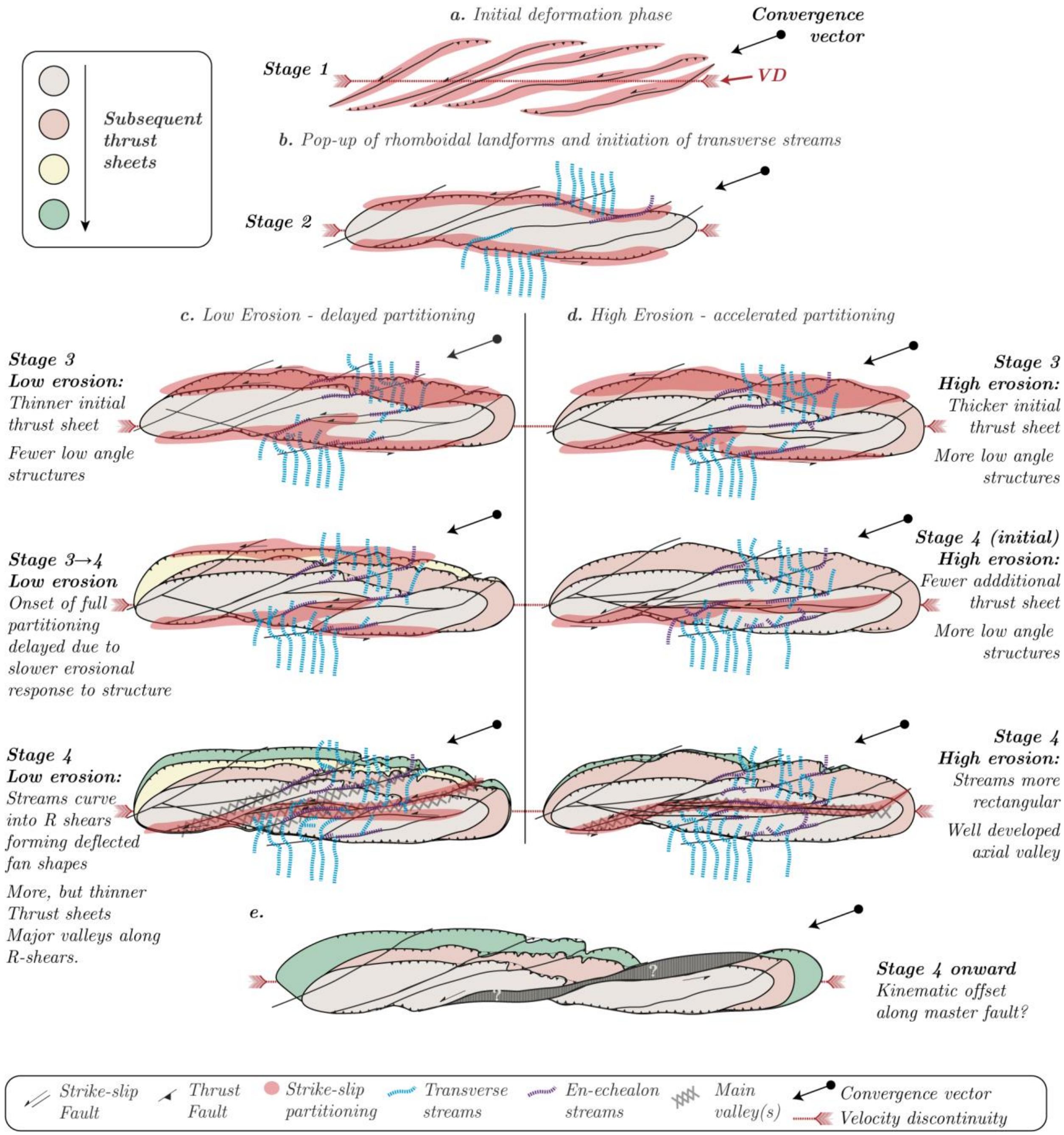

Figure 9. Illustration highlighting the differences between the evolution of fault and stream networks, strain partitioning, and topography. (a) The initial deformation phase. (b) Beginning stages of wedge development. (c) Stages 3 (transitional) – 4 (full partitioning) for low erosion systems. (d) Stages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

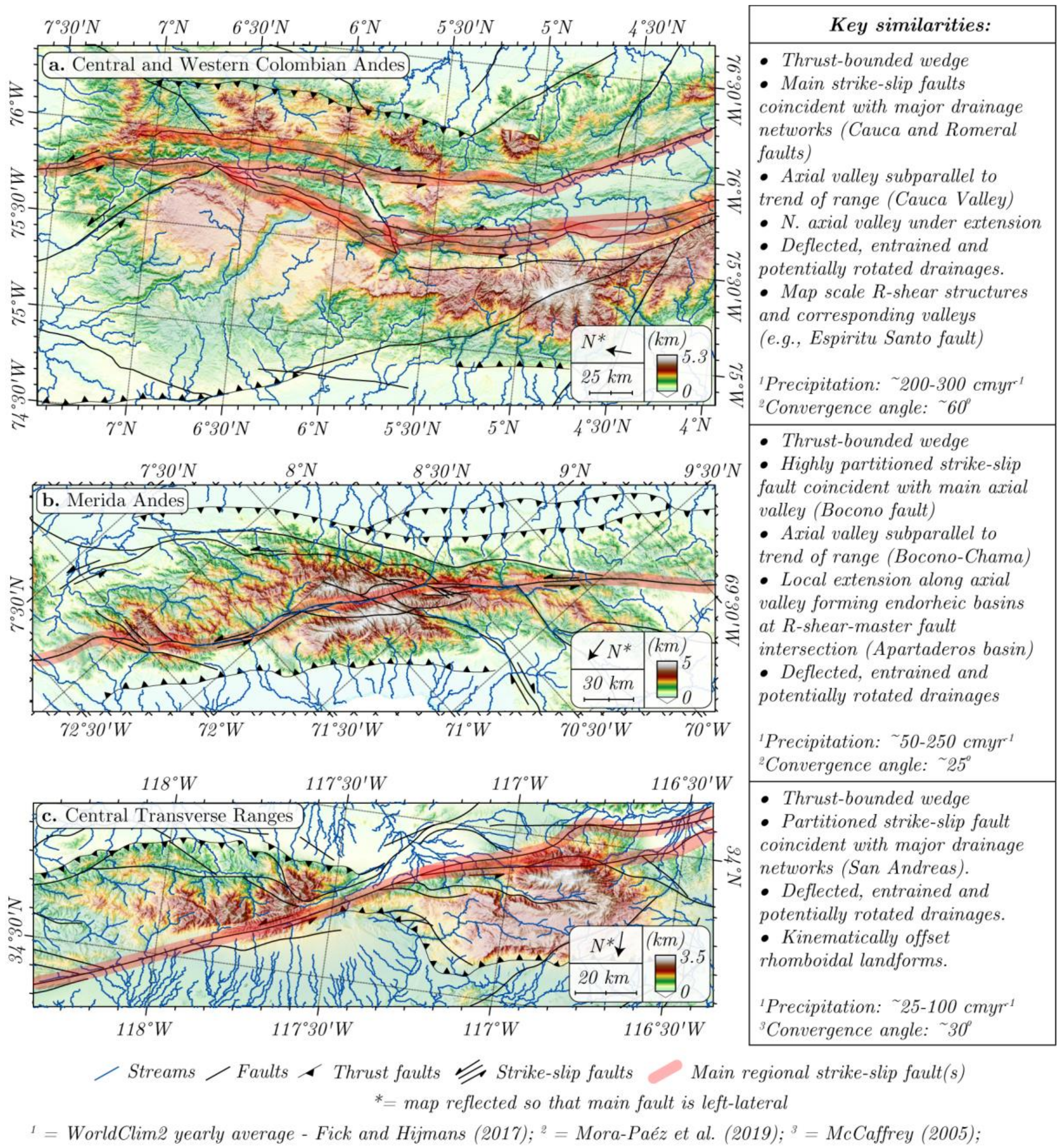

Figure 10. Natural prototypes with similar morphostructural characteristics as our models. (a) The Merida Andes in Venezuela. (b) The Central Transverse Ranges along the San Andreas Fault north of Los Angeles, California. (c) The Central and Western Colombian Andes. The similarities between each setting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

2017), and convergence angles are estimated from GPS velocity-derived kinematic block models by Mora-Paéz et al. (2019) for the northern Andes and from McCaffrey (2005) for the central Transverse Ranges.

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

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.