Supplementary Information for

Sensitivity of rift tectonics to global variability in the efficiency of river erosion

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Visco-elasto-plastic modeling of crust and mantle deformation

We model tectonic deformation in the vertical \((x, z)\) plane (Figure S1) with the 2-D thermo-mechanical code SiSiER (Simple Stokes Solver with Exotic Rheologies, 1). This code solves for conservation of mass, momentum and energy in an incompressible continuum without inertia:

\[
\frac{\partial \rho v_i}{\partial t} = 0 , \tag{1}
\]

\[
\frac{\partial s_{ij}}{\partial x_j} - \frac{\partial p}{\partial x_i} + \rho g_i = 0, \tag{2}
\]

and

\[
\frac{\partial T}{\partial t} + v_i \frac{\partial T}{\partial x_i} = \frac{\partial}{\partial x_i} \left( \lambda \frac{\partial T}{\partial x_i} \right) + \frac{H}{\rho c_p}, \tag{3}
\]

where \(v_i\) denotes components of the velocity field (along \(x_1 = x\) and \(x_2 = z\)), \(s_{ij}\) is the deviatoric stress tensor, \(p\) is pressure, \(g_i\) is the gravity field (9.8 m/s\(^2\) along the \(z\)-direction), \(\rho\) is density, \(T\) is temperature, \(c_P\) is heat capacity (1000 J/kg), \(\lambda\) is thermal conductivity (2 W/m/K), and \(H\) is the radiogenic heat production rate. Repeated indices imply summation. The above equations are discretized using finite differences on a staggered grid (2). Advection of various fields and material properties is handled through the marker-in-cell method (3).

The continuum is assigned a visco-elasto-plastic rheology in which the total strain rate \(\dot{\epsilon}_{ij} = \frac{1}{2} \left( \frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right)\), equal to the deviatoric strain rate is the sum of an elastic component (involving a shear modulus \(G = 30\) GPa), a viscous component representing high-temperature ductile creep, and a frictional-plastic component representing low-temperature brittle deformation at a finite yield stress:

\[
\dot{\epsilon}_{ij} = \frac{s_{ij}}{2G} + \frac{s_{ij}}{2\eta_{\text{creep}}} + \frac{s_{ij}}{2\eta_{\text{plas}}}. \tag{4}
\]

The elastic term in the above equation involves a time derivative of deviatoric stresses, which is approximated through backward-Euler finite difference. Elastic stresses are advected and rotated on markers at every time step, following the methodology described in ref. 2. Advection of various fields and material properties is handled through the marker-in-cell method (3).

In equation (4), \(\eta_{\text{creep}}\) denotes the effective viscosity enabled by dislocation creep processes. It is related to the second invariant of the deviatoric strain rate tensor \(\dot{\epsilon}_{il} = \sqrt{\frac{1}{2} \dot{\epsilon}_{ij} \dot{\epsilon}_{ij}}\) by the following equation:

\[
\eta_{\text{creep}} = A \frac{1}{\bar{n}} \dot{\epsilon}_{il}^{\bar{n}} n^\eta E_s, \tag{5}
\]

where \(A\), \(n\) and \(E_s\) are the experimentally-determined prefactor, exponent and activation energy of the flow law. \(R\) is the universal gas constant (8.314 J/mol/K).

Low-temperature brittle deformation is modeled by capping the deviatoric stress at a yield limit specified by the Mohr-Coulomb criterion:

\[
s_{il} = s_{\text{yield}} = C \cos \phi + p \sin \phi, \tag{6}
\]

where \(s_{il}\) is the second invariant of the deviatoric stress tensor \(s_{il} = \frac{1}{2} s_{ij} s_{ij}\), \(C\) is cohesion and \(\phi\) is the friction angle. In the assumed 2-D plane strain configuration (where the out-of-plane stress is equal to \(-p\)), equation (6) is equivalent to the Drucker-Prager criterion. When at yield, the material follows a non-associated plastic flow rule with no dilation. In the framework of visco-elasto-plasticity (equation 4) this can be achieved through an effective plastic viscosity of the form:

\[
\eta_{\text{plas}} = \frac{s_{\text{yield}}}{2kT}, \tag{7}
\]

where \(\dot{\epsilon}_{il}^{\text{plas}}\) is the second invariant of the plastic component of strain rate. Plastic strain accumulates at a rate determined by \(\dot{\epsilon}_{il}^{\text{plas}}\) in all areas that are at yield. In a mature fault, plastic strain roughly scales as total fault slip divided by fault width, which is typically the width of \(\sim 3\) grid elements. Strain softening of the brittle properties is implemented to facilitate strain localization (e.g., 4). Friction and cohesion, initially set to \(\mu = \tan(\phi) = 0.6\) and \(C = 30\) MPa decrease linearly as the locally accumulated plastic strain increases. When plastic strain reaches a critical value of 0.5, friction and cohesion remain at their weakened values of 0.3 and 0.01 MPa, respectively. This magnitude of...
friction and cohesion drop is consistent with values used in prior studies (4, 5). A critical plastic strain of 0.5 amounts to an offset of ~400 m on 3-element wide faults formed in the high-resolution region of our model domain (with 250×250 m wide cells, Figure S1). Patterns of strain localization were found by ref. 4 to weakly depend on critical offsets << 2 km. Our choice of softening strain thus seeks to minimize the dependence of our results to this poorly constrained parameter, while being of a similar order of magnitude to values commonly used in large-scale rift modeling studies (e.g., 5). Such strain weakening is typically thought to represent a whole range of weakening processes that take place as fault zones mature (e.g., precipitation of weaker mineral phases, evolving damage state...). Healing of plastic strain in regions that no longer yield (e.g., 6) is implemented as described by ref. 1, and takes place over a characteristic time scale of $5\times10^{14}$ s, comparable to the duration of our simulations.

At each time step, material properties carried by markers are linearly interpolated to the Eulerian grid, and used to construct fields of effective viscosity, density and elastically-accumulated deviatoric stresses. These fields are used to assemble and solve the discretized version of equations (1–3) with a direct solver. This operation is repeated until the strain rate that corresponds to the velocity solution becomes consistent (within a specified tolerance) with the non-linear effective viscosities used to solve for said velocity field.

Our model domain (Figure S1) spans 220 km laterally and 90 km vertically, and includes a 10-km thick "sticky air" layer of negligible density and constant Newtonian viscosity of $10^{18}$ Pa·s, which ensures that the air-rock interface is effectively a free surface (7). The underlying rock layer comprises an initially 30-km thick crust of density 2700 kg·m$^{-3}$, with a wet anorthite rheology characterized by $A = 1.4332\times10^{-14}$ Pa$^{-3}$·s$^{-1}$, $E_a = 356$ kJ/mol, and $n = 3$ (8). Sediments are assumed to follow the same flow law, but are assigned a lower density of 2200 kg·m$^{-3}$. The deeper mantle layer has a density of 3300 kg·m$^{-3}$ and a wet olivine rheology characterized by $A = 1.0378\times10^{-14}$ Pa$^{3.5}$·s$^{-1}$, $E_a = 480$ kJ/mol, and $n = 3.5$ (9). Rock densities decrease linearly with temperature according to their thermal expansion coefficient ($10^{-5}$ K$^{-1}$).

Extension is applied symmetrically on the freely-sliping sides of the domain at a uniform half-rate of 0.5 mm/yr. Inflow of air and rock through the free-slip top and bottom boundaries is prescribed to match the outflow through the sides. Plastic strain is initialized at zero everywhere except within a 1-km wide inclined slot cutting through the crust in the middle of the domain, where it is set to the critical softening strain of 0.5. This ensures full cohesion and friction weakening in a plastic "seed", which prompts the rapid localization of a unique master fault F1 with a slip rate close to 1 mm/yr at the beginning of our simulations. Temperature is maintained at 0°C along the top and within the entire air layer, and at 1250°C at the base of the domain. No conductive heat flow is allowed through the sides. Crustal rocks produce radiogenic heat at a rate $H$, which is varied between 0.35 and 1.4 μW/m$^3$ to model a range of thermal conditions. The temperature field is initialized as a 1-D steady-state geotherm (second-order polynomial) which verifies the above temperature boundary conditions, and accounts for a constant radiogenic heat source term $H$. We acknowledge that our choice of rheological and thermal parameters do not sample the full variability of crustal strength profiles. This would involve changing the initial crustal thickness, the creep laws, and the geotherm in a systematic manner, leading to a complex, multi-dimensional parameter space. We instead focused on varying the geotherm, in a proof-of-concept, broadly applicable approach to assessing the effect of strength profiles on the feedbacks between surface processes and brittle strain localization.

Overall, our approach is similar to other thermo-mechanical modelling studies of rift zones (e.g., refs. 5, 10, 11), but differs in its regional (fault-scale) focus. We assess the evolution of localized brittle strain in our simulations by tracking the position of areas where plastic strain exceeds the critical weakening strain, which we interpret as mature crustal faults (Figure S2). The location of new faults relative to F1 is used to define the regimes mapped in Figure 3. F1 is considered the master fault as it remains the only fault with a plastic strain in excess of the critical weakening strain at an arbitrary reference depth of 5 km below the surface. This definition allows us to measure the total extension accommodated by F1 as a master fault in a self-consistent manner across simulations. The throw accumulated on F1 is measured as the vertical offset of initially horizontal marker chains (e.g., gray "strata" shown in Figure 1B and Figure 2).
Landscape evolution model

One-dimensional surface topography \( h_{1D}(x) \) is tracked in the tectonic model as a chain of closely-spaced markers at the interface between the air and rock layers. A corresponding, two-dimensional representation of surface topography \( h_{2D}(x,y) \) is also tracked and subjected to surface processes (Figure S1), such that \( h_{1D}(x) \) corresponds to the average of \( h_{2D}(x,y) \) along the y-direction at all times (11). At every time step, the tectonic model produces a velocity field \((v_x, v_y)\), which we interpolate to the surface markers. These surface velocities are then uniformly extended over 73 km in the y-direction. The horizontal component is used to horizontally advect \( h_{2D}(x,y) \), and the vertical component is used as the uplift term \( U \) in a differential equation that describes the action of surface processes:

\[
\frac{\partial h_{2D}}{\partial t} = U(x,y) - K \sqrt{A_S} + DV^2 h_{2D}.
\]  

(8)

The second term on the right-hand side of equation (8) is a stream power law with drainage area \( A \) and slope \( S \) exponents of 0.5 and 1, respectively (12, 13). \( K \) denotes the erodibility coefficient. The third term describes hillslope diffusion with diffusivity \( D \). Equation (8) is solved over a time span that corresponds to the time step of the tectonic model. This is achieved through the Fastscape algorithm introduced by ref. 14. Side boundaries are closed \((y = 0 \text{ and } y = 73 \text{ km})\), and maintained at zero-altitude \((x = 0 \text{ and } x = 220 \text{ km})\). Once \( h_{2D}(x,y) \) has been updated, it is averaged along \( y \) to update the new air-rock interface \( h_{1D}(x) \) in the tectonic model. Rock markers are turned into air wherever erosion has taken place, and air markers are turned into sediments wherever the topography subsides, ensuring instantaneous infilling of basins with sediments (except in reference simulations without deposition). Here we focus on varying the \( K \) coefficient while keeping topographic diffusivity constant at \( D = 10^{-9} \text{ m}^2\text{s}^{-1} \) (3.2×10^{-2} \text{ m}^2\text{yr}^{-1}). With this parameterization of erosion, the impact of the \( K \) coefficient must be assessed relative to the characteristic uplift rate \( U_0 \). We do so through the dimensionless erosional efficiency number defined as \( EE = K \sqrt{A_0}/U_0 \), where \( A_0 \) is a reference area set to \( 10^6 \text{ m}^2 \).

Code availability

All the MATLAB scripts necessary to reproduce our simulations are available on the Zenodo repository at https://doi.org/10.5281/zenodo.5786473. The landscape evolution routines correspond to the files starting with "LEM-", and are based on a MATLAB implementation of the Fastscape algorithm (14) by B. Kaus and A. Popov. Each simulation corresponds to an input file in which the only parameters that are varied are the crustal radiogenic heat production \( H \), the erodibility coefficient \( K \), and whether sediment deposition is included or not. A summary of all simulation inputs is provided in Table S1. Simulations can be launched in a MATLAB prompt by running the command:

\[
\text{>> SiSTER\_RUN('SiSTER\_Input\_File\_name')}\]

where input file names match the simulation labels, e.g., SiSTER_Input_File_half_graben_H60.m corresponds to simulation H60. Further details on how to run the code and visualize the outputs are accessible at https://github.com/jaolive/SiSTER.

Determining erosional efficiency through river profile analysis

Theory. Following the approach laid out by ref. 15, let us consider the topographic evolution equation (8) at steady state, neglecting hillslope diffusion:

\[
U(x') - K \sqrt{A} = 0.
\]

(9)

Here we focus on the equilibrium between erosion and uplift along a river profile, where \( x' \) denotes upstream distance. We assume that erodibility is constant along this profile, and write the upstream drainage area at any point \( A(x') \) as the product of a reference area \( A_0 \) and a dimensionless shape function \( A'(x') \). We do the same for the uplift function: \( U(x') = U_0 \frac{U'(x')}{A'(x')} \). Since slope \( S \) is the derivative of topography \( h \) with respect to \( x' \), it is straightforward to integrate equation (9) upstream from a chosen base level \( (h = h_b \text{ at } x' = x_b') \). This yields:

\[
h(x') = h_b + \left( \frac{U_0}{K \sqrt{A_0}} \right) \int_{x_b'}^{x'} \frac{U'(x)}{\sqrt{A'(x)}} \, dx'.
\]

(10)
The integral term in equation (10) can be thought of as an upstream distance $\chi U$, which — unlike $x'$ — has been "corrected" for variations in drainage area and uplift along the river profile. River elevation above base level thus becomes a linear function of $\chi U$ with a proportionality coefficient that is exactly $1 / EE$. If the uplift function is spatially uniform ($U^*(x') = 1$), the corrected upstream distance is equivalent to the $\chi$-coordinate classically used in river profile analysis (15).

**Assessing spatial variations of uplift rate in normal fault footwalls.** The theory outlined above suggests that erosional efficiency can be determined from river elevation vs. $\chi U$ plots, which can be thought of as standard $\chi$-plots corrected for spatial variations in uplift. The main difficulty in constructing such plots is to infer the spatial pattern of uplift in a given region. $U^*(x, y)$ is not well known in rift zones, beyond the fact that it should decrease away from the master fault into the footwall, in a manner that reflects local flexural parameters (16, 17). Following ref. 18, we assume that $U^*$ at a distance $x_F$ perpendicular to the strike of a master normal fault follows the functional shape predicted by a broken elastic thin plate model:

$$U^*(x_F) = U_{MIN} + (1 - U_{MIN}) e^{-x_F / \alpha} \cos \frac{x_F}{\alpha} \tag{11}$$

where $\alpha$ denotes a flexural length scale that should be longer in thicker / stronger brittle upper crust (1, 19). This allows us to construct a dimensionless uplift function which is maximized at the fault ($U = U_0$ at $x_F = 0$). The $U_{MIN}$ terms account for a uniform "background" uplift rate away from the master fault. In practice, we determine $\alpha$ by fitting the functional form of equation (11) to footwall topography averaged along strike, excluding the portion of the footwall where (steep) topographic slopes face the master fault. We thereby restrict our analysis to footwall rivers flowing away from the master fault, which have likely experienced an uplift function whose shape resembles the resulting topography. This is most likely to hold true if the erosion rates that sculpted this topography obey a near-linear scaling with slope, in which case the equilibrium relief should linearly scale with the decaying uplift function. Our numerical benchmarks (see next section) indeed show that the stream power incision law with a slope exponent of 1 does a good job at preserving the decay length scale of the uplift function and encoding it into the resulting footwall topography.

**Benchmarking on synthetic landscapes.** We benchmark the method outlined above on synthetic river networks that developed on the footwall of F1 in simulations M70–M50 (Table S1). To do so, we extract the topography $h_{2D}(x, y)$ shaped by slip on fault F1 (e.g., Figure 1B) as late as possible into the simulations before the localization of new faults alters the relief of the initial half-graben (See example from simulation M60 in Figure S3A). The uplift shape function $U^*$ is determined by fitting equation (11) to footwall relief averaged along the $y$-direction, excluding the steep fault-facing scarp, i.e., between distances of 5 and 100 km away from F1, into the footwall (Figure S3B). This is achieved through MATLAB’s non-linear least squares fitting toolbox. This uplift function does not vary along the strike of the fault (Figure S3A).

Major footwall rivers that drain away from the master fault are identified and analyzed through the TopoToolbox MATLAB package (20). This package also calculates upstream drainage area at any point in the landscape, which enables the computation of the integral in equation (10). An example from simulation M60 is shown in Figure S3C. It clearly shows that assuming $U^* = 1$ does not yield a linear trend in elevation vs. $\chi$ space. By contrast, using the uplift function described by equation (11) in the integral of equation (10) leads to a more linear and less scattered trend, suggesting that this $U^*(x_F)$ is a good approximation to the uplift pattern that self-consistently develops in the tectonic model. To estimate $EE$ from these plots, we group the elevation vs. $\chi U$ points into 100 bins of $\chi U$. In each bin, we compute the median elevation as well as its 10th and 90th percentiles, thereby bracketing the vast majority of the $\chi U$-plot. As these plots are never perfectly linear, we fit the trends outlined by each percentile with a second-order polynomial, and calculate the slope of this polynomial at base level. Our reasoning is that since equilibrium is progressively established as river knickpoints move upstream, portions of the landscape closest to base level should reach this equilibrium first. The slope of the lowermost portions of the $\chi U$-plots is therefore most likely to contain reliable information about erosional efficiency ($EE \approx 1 / \text{slope}$), as the whole method relies on the assumption of balanced uplift and erosion (equation 9).
For simulation M60, this method yields an estimated EE of 4.1 (median trend) with confidence bounds of 3.3–5.0 (10th and 90th percentiles) (Figure S3C). This simulation is characterized by an erodibility coefficient \( K = 10^4 \) yr\(^{-1}\). In all of our simulations, as long as F1 remains the dominant normal fault in the system (Figure 1B), its full vertical slip rate is \(~0.7\) mm/yr. This slip rate evolves through time as F1 flexurally rotates to shallower angles (21) and as new faults (F2 and possibly F3) begin to localize (Figure 2) around F1. In addition, vertical motion is not symmetrical between the blocks bounded by F1: hanging wall subsidence happens faster than footwall uplift (relative to the initial base level). Overall, we estimate that in all our simulations, peak footwall uplift due to slip on F1 (while it is still dominant) occurs at \(~0.25\) mm/yr (typically between 0.2 and 0.3 mm/yr). With this information, we can compute a "true EE", which would of course not be possible in real systems. For simulation M60, the true EE is 4 (3.3–5, if we account for temporal variability in footwall uplift rates). Our estimated EE from river profile analysis (4.1 (3.3–5.0)) is in excellent agreement with this true EE. In fact, repeating this exercise for simulations M70–M50 (Table S1) shows a good agreement between estimated and true EE for values greater than \(~2\) (Figure S4). Simulations M70 and M65 have lower erodibility coefficients, and therefore lower EEs. In these simulations, fluvial incision was not efficient enough (too slow) for a dynamic equilibrium between uplift and erosion to be attained while F1 was still the dominant master fault. Consequently, rivers had not yet steepened to their equilibrium slopes. Their shallow-dipping profiles thus yield an apparent EE that is greater than the actual EE, because erosional efficiency is related to the inverse of the \( \chi_u \)-plot slope.

In short, these synthetic tests show that the uplift function as parameterized by equation (11) is a good approximation to the more realistic uplift function that self-consistently develops through flexural-isostasy in visco-elasto-plastic simulations. This is in large part because stream power incision as parameterized in our simulations preserve the shape of the uplift function in the resulting equilibrium footwall relief (Figure S3B). The tests also suggest that river profile analysis can provide a reliable assessment of EE as long as (1) rivers have reached an equilibrium with the uplift field, (2) their evolution is controlled by a stream power-type law that resembles equation (8), and (3) that erodibility is constant over the half-graben footwall. Finally, we note that the value of EE alone is not sufficient to assess whether a system is likely to have reached equilibrium or not. For example, in the simple framework of equation (8), ignoring diffusion, the characteristic time to reach steady state will scale inversely with erodibility, but will not depend on the characteristic uplift rate. In other words, systems with low EE could well reach equilibrium rapidly if their erodibility coefficient is large. Their characteristic uplift rate must however be very large to ensure low EE.

**Measuring erosional efficiency and structural characteristics in real half-graben systems**

**Methodology.** We analyze the elevation profiles of rivers draining away from the master fault into the footwalls of 9 half-grabens spanning 4 rift systems (Table S2). Our analysis is based on 30-m resolution SRTM topography acquired through the Open Topography online portal (opentopography.org). We use the TopoToolbox MATLAB package (20) to identify river networks that drain over large extents of the footwall, away from the tips of the master fault (22, 23), and down to a chosen reference altitude. We target ranges where the morphology is clearly dominated by fluvial as opposed to glacial valleys, and focus on portions of the range where relief averaged along strike smoothly decays away from the master fault (e.g., Figure S5B). We select two points on the master fault (UTM coordinates listed in Tables S2 and S3), which delineate the fault-normal topographic swath over which relief is averaged, non-dimensionalized and fitted with equation (11) to determine the shape of the uplift function, following the same methodology as in our synthetic tests. Maps of drainage area determined by the TopoToolbox scripts are then integrated upstream to produce standard \( \chi \)-plots and \( \chi_u \)-plots corrected for the dimensionless uplift function (e.g., Figure S5C), which only varies in the fault-normal direction (e.g., Figure S5A). Erosional efficiency and its confidence interval are determined from \( \chi_u \)-plots with the same approach as in our synthetic tests. While we have strived to select systems that are still actively growing, it is possible that the uplift rate on the master fault may have fluctuated through time, and over a range of time scales, adding some uncertainty to our analysis. In the case of long-term changes in fault activity, it is reasonable to expect that our method will sample an uplift pattern averaged over the characteristic time needed for the landscape to equilibrate.
**Paeroa Range, New Zealand.** The Paeroa fault, which bounds the Paeroa Range (Figure S5), is a NW-dipping fault located in the Taupo Rift zone of New Zealand. Its throw has been estimated between 560 and 900 m based on the vertical offset of ignimbrite layers (24). The fault is thought to offset a 6–8 km thick brittle upper crust based on seismicity surveys conducted over the Taupo volcanic zone (25).

**Sandia Mountains, New Mexico, USA.** The Sandia Mountains (Figure S6) are bounded by a complex W-dipping fault zone that marks the eastern edge of the Rio Grande Rift in its Albuquerque Basin segment (26). Ref. 27 estimated the vertical offset on the major basin bounding fault between 4600 and 6100 m in seismic reflection profiles. This estimate is consistent with gravity and flexural modeling of the basin geometry (28), and more recent minimum estimates of the vertical offset between the buried Santa Fe group and the top of the Sandias (29): ~4 km on average, 5.6 km at the most. In fact, ref. 29 proposed that uplift of the Sandia Mountains took place on two generations of faults that formed within ~5 km of each other, and together account for as much as ~7 km of vertical offset. Here we retain an estimate of throw between 5 and 7 km to account for the spread between various estimates. The thickness of the brittle upper crust offset by this fault system is not precisely known. Seismic imaging suggests that the basin bounding faults root onto a shallow-dipping structure at ~10 km depth (27), which could mark the brittle-ductile transition. Ref. 29 inferred that the brittle layer thickness varied between ~7 and ~10 km throughout the uplift of the Sandias, based on reconstructed fault geometries and the elevated local heat flow.

**Lemhi Range, Idaho, USA.** The Lemhi Range (Figure S7) is located in the northern Basin and Range province and is bounded by a major SW-dipping normal fault (30). Ref. 31 estimated a vertical offset of 5–6 km on this fault based on paleomagnetic measurements of tilted volcanic units (32) and previous flexural modeling (33). This estimate is consistent with the typical displacement-length scalings that characterize normal faults. The thickness of the brittle crust is not well known beneath the Lemhi Range. However, the seismicity that followed the 1983 Borah Peak earthquake on the nearby Lost River Fault (~25 km to the SW) was confined to depth of 12–16 km (34). Here we assume that this estimate applies to both the Lemhi and Beaverhead faults, which is consistent with the fact that their footwall blocks exhibit similar flexural strengths (33).

**Beaverhead Range, Idaho/Montana, USA.** The Beaverhead Range (Figure S8) is bounded by a major SW-dipping normal fault located ~25 km NE of the Lemhi fault (33). Paleomagnetic studies and flexural modeling cannot resolve a difference in throw between the Lemhi and Beaverhead fault (31). We also infer a similar brittle layer thickness across the two systems. Interestingly, our landscape analysis suggests a greater erosional efficiency over the Beaverhead Range (~2.7) than over the Lemhi Range (~1), even though they expose similar lithologies and are subjected to similar climatic conditions. This is however consistent with a slower slip rate on the Beaverhead fault — about half of the slip rate on the Lemhi fault (22)— which, with all other parameters kept constant would double the erosional efficiency.

**Wassuk Range, Nevada, USA.** The Wassuk Range (Figure S9) is bounded by an east-dipping normal fault located ~100 km east of the western edge of the Basin and Range province, within the Central Walker Lane Belt (35). Structural mapping and low-temperature thermochronology suggest that while the footwall block exposes units that were located at ~8.5 km depth prior to extension, exhumation occurred in two distinct stages (36). During the first phase (~15–12 Ma), the then range-bounding fault system accumulated ~6.5 km of offset. Slip on the fault ceased until extension associated with the development of the Walker Lane resumed at ~4 Ma. A second-generation range-bounding fault formed and to date has accumulated 2–3 km of throw. We retain this latter estimate as today’s rheological and surface processes conditions are likely more representative of the conditions that prevailed during the second phase of extension. Seismicity data suggests a present-day brittle-ductile transition depth between ~11 and 14 km below the Wassuk Range (37).

**Santa Rosa Range, Nevada, USA.** The Santa Rosa Range (Figure S10) is located in the northwestern Basin and Range province, and is bounded by a W-dipping normal fault. Ref. 38
inferred a total throw between 5 and 6 km on this fault, on the basis of structural mapping and tilt of volcanic units. The present depth to the brittle-ductile transition is not well known in this area. We estimate it between 10 and 15 km, a depth which encompasses 95% of earthquakes that occur over the state of Nevada (39).

**Warner Range, California, USA.** The Warner Range (Figure S11) is bounded by the east-dipping Surprise Valley Fault, which marks the northwestern margin of the Basin and Range Province (38). Ref. 40 used stratigraphic offsets to measure a vertical displacement of 4–5 km on the range bounding fault, which is possibly a low estimate. Ref. 41 interpreted the short-wavelength flexure of the up-warped footwall and the elevated heat flow to result from a somewhat shallow brittle-ductile transition which they estimated at ~9 km. Between 2014 and 2016, a seismic swarm occurred ~50 km east of Surprise Valley, with hundreds of small earthquakes within the uppermost ~13 km of the crust (42). Here we use 9 and 13 km as plausible brackets on the depth to the brittle-ductile transition in the area.

**Western shore of Lake Edward, Democratic Republic of the Congo.** The Edward lake basin (Figure S12) is one of the major half-graben systems of the East African Rift's Western branch. It is bounded by the E-dipping Lubero border fault which uplifted the Western Rift Mountains. The vertical offset between the base of the lacustrine sediment (inferred through seismic imaging and gravity/flexural modeling) and the top of the Lubero escarpment yields a minimum throw estimate of 7 km (43, 44). The thickness of the brittle crust in the area can be estimated between 25 and 30 km, from the depth distribution of small earthquakes recorded within the Rwenzori Mountains north of Lake Edward (45).

**Kipengere Range / north-eastern shores of Lake Malawi, Tanzania.** Lake Malawi marks the axis of the East African Rift in its southernmost portion. It comprises three basins shaped by three border faults of alternating polarity. We focus on the north basin, which is bounded by the SW-dipping Livingstone fault (Figure S13). It is flanked to the NE by the Kipengere Range, also referred to as the Livingstone Mountains, which formed through flexural-isostatic rebound to localized extension at the axis (46). Using high-resolution seismic imaging of the sedimentary pile deposited in the north basin, ref. 47 proposed a minimum estimate of throw on the Livingstone fault between 6.6 and 7.4 km. Recent seismological surveys indicate that the entire crust is seismogenic in the northern part of Lake Malawi (48). The depth distribution of earthquakes suggests that the base of the brittle faulted crust is located between depths of ~32 and ~37 km.

**Measuring erosional efficiency and structural characteristics in real horst systems**

We complement our analysis of half-graben systems with 3 horsts from the East African Rift and Basin and Range province (Table S3). For each of them, we identify the structurally and geomorphologically dominant border fault and apply the same methodology as outlined previously.

**Rwenzori Mountains, Democratic Republic of the Congo / Uganda.** The Rwenzori Mountains (Figure S14) stand out as a topographic and structural anomaly within the Albertine Rift, a portion of the Western branch of the East African Rift. They expose gneissic basement units which are otherwise buried under rift sediments (or volcanics) in the Lake Edward (to the south) and Lake Albert (to the north) half-grabens (49). The Rwenzoris are an asymmetric horst block bounded by antithetic N-S trending faults. The W-dipping Bwamba fault delineates the western front of the range and has been interpreted as the dominant border fault responsible for most of the uplift (50). Ref. 49 used stratigraphic offsets and scaling relations between gouge thickness and displacement to estimate throws on the major horst-bounding faults. This yielded vertical offsets in excess of 7 km on the Bwamba fault, significantly larger than typical offsets (~2–3 km) on the family of synthetically east-dipping faults that bound the range to the east (e.g., Bigo and Nyamwamba faults). The thickness of the brittle crust beneath and around the Rwenzori horst can be estimated between 25 and 30 km, from the depth distribution of seismicity recorded by ref. 45.

The extremely high relief of the Rwenzori Range (locally in excess of 4 km, ~2.7 km on average relative to the adjacent basins) makes it prone to significant glacial erosion at high altitude.
We therefore focused our analysis on lower altitude river networks draining away from the Bwamba fault, which likely play a strong role in shaping the morphology of the eastern part of the range.

**Ruby Mountains, Nevada, USA.** The Ruby Mountains (Figure S15) mark the southern extent of the Ruby Mountains-East Humboldt metamorphic core complex in the north-central Basin and Range province, where structurally deep units are now exposed at the surface (51). A large part of this exhumation occurred through ~7 km of vertical displacement on a major W-dipping fault termed the Ruby detachment, which initiated around 17 Ma and underwent significant eastward rollover (52) through ~12 Ma. Extension was later taken up by 2 antithetic fault systems formed within the last 10 Myrs, which now separate the Ruby Mountains from the Huntington Valley to the West and the Ruby Valley to the East. While both of these fault systems show evidence for Quaternary slip (e.g., faults 1714 and 1718 in the USGS Quaternary Faults Database: 53), the E-dipping Ruby Valley fault appears dominant in terms of both structural and topographic relief. Ref. 52 estimated its throw at ~2.4 km. Given the documented switches in polarity of the master normal faults and the coexistence of significant deformation on antithetic fault systems, we classify the Ruby Mountains as a horst structure, or in a broader sense, an area with a documented history of localized brittle strain migrating into the footwall of a master normal fault.

Lacking geophysical constraints on the thickness of the brittle crust in the area, we use the same estimate (10–15 km) as in the Santa Rosa range, which is based on seismogenic layer thickness measured over the entire state of Nevada (39).

**Toiyabe Range, Nevada, USA.** The N-S trending Toiyabe Range (Figure S16) is located in the central Basin and Range province. It is a horst block bound on its western and eastern sides by high-angle normal faults which separate it from the Reese River and Big Smokey Valleys (54, 55). Both fault systems show evidence of Quaternary activity (e.g., faults 1336c and 1337 in the USGS Quaternary Faults Database: 56, 57), but the E-dipping Toiyabe Range fault zone has been noted to have the dominant tectonic and geomorphic expression in the southern portion of the range (55), which we focus on in our river profile analysis. Structural mapping and thermochronology data are consistent with vertical offsets on the order of 4-5 km on the range-bounding faults. We infer a local depth to the brittle-ductile transition of 10–15 km following the reasoning applied to the Santa Rosa Range and Ruby Mountains.

*Figures in the main manuscript and Fig. S1 used colormaps from ref. 58.*
**Fig. S1.** Model setup and boundary conditions. The 10-km thick "sticky air" layer above the crust is not shown. The resolution of the tectonic model is 250×250 m in the active deformation area within +/- 50 km of the initially seeded fault, and coarsens to 1000×500 m (horizontal × vertical) on the outer regions of the model. The landscape is discretized on a regular grid with 80 and 200 elements along the y- and x-directions, respectively.
Fig. S2. Temporal evolution of strain localization in simulations shown in Figure 2: (A) F70, (B) F60, (C) H70, (D) H60. Black dots show the position (x-coordinate) of areas where plastic strain 5 km below the surface exceeds the critical softening strain, thereby indicating the position of major faults, relative to the initial position of fault $F_1$ initially at $x = 0$. New faults ($F_2, F'_2$) localize either in the footwall (FW, panels A & C) or hanging wall (HW, panel B) of $F_1$, or on both sides of $F_1$ (panel D).
Fig. S3. River profile analysis from synthetic landscapes. A. Model topography produced by prolonged slip on fault F1 in simulation M60 (with $K = 10^{-6}$ yr$^{-1}$). Major footwall rivers draining away from F1 are mapped and color-coded by elevation. Red and black lines are contours of the dimensionless uplift function $U^*$ used to compute the uplift-corrected $\chi$ plots (panel C). B. Dimensionless uplift function vs. distance to the fault (red) and footwall topography averaged along the strike of F1, normalized by its maximum value (blue). C. Elevation of simulated rivers vs. upstream distance corrected for changes in drainage area ($\chi$, blue dots), and vs. upstream distance corrected for changes in both drainage area and uplift ($\chi_U$, red dots). The 10th, 50th and 90th percentiles of the elevation vs. $\chi_U$ points (binned by $\chi_U$) are fitted with second-order polynomials shown as solid and dashed black curves. The thick gray line shows the expected $\chi_U$-plot if the maximum uplift rate at the fault was exactly 0.25 mm/yr, corresponding to an erosional efficiency $EE = 4$. Thinner gray lines show the same trends for maximum rates of 0.2 and 0.3 mm/yr ($EE = 5$ and 3.3).
Fig. S4. True erosional efficiency vs. EE assessed through river profile analysis of landscapes shaped by slip on fault F1 in simulations M50–M70. Error bars on true EEs account for variability in true uplift rate at the fault. Error bars on estimated EEs are based on fits to the 10th and 90th percentiles of the elevation vs. χu-plots (see Figure S3C). Dashed line is 1:1 line.
Fig. S5. River profile analysis of the Paeroa Range, New Zealand. (Top) SRTM topography and selected footwall rivers (draining to 400 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-bounding master fault, and along-fault width of averaging swath. Red dots are located at (x, y) = (438430, 5756700) and (431150, 5748700) meters in UTM zone 60H. Black bar = 10 km. (Middle) Dimensionless uplift function $U^*(x_F)$ (red) determined by fitting mean footwall relief (blue, normalized) between distances of 2 to 11.4 km to the master fault. (Bottom) Standard elevation above base level vs. $\chi$ plot (blue dots), and uplift-corrected $\chi_U$–plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
Fig. S6. River profile analysis of the Sandia Mountains, New Mexico, USA. (Top) SRTM topography and selected footwall rivers (draining to 2100 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-bounding master fault, and along-fault width of averaging swath. Red stars are located at (x, y) = (364230, 3897300) and (364520, 3886600) meters in UTM zone 13S. Black bar = 10 km. (Middle) Dimensionless uplift function $U^*(x_F)$ (red) determined by fitting mean footwall relief (blue, normalized) between distances of 6.6 to 17 km to the master fault. (Bottom) Standard elevation above base level vs. $\chi$ plot (blue dots), and uplift-corrected $\chi_U$-plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
Fig. S7. River profile analysis of the Lemhi Range, Idaho, USA. (Top) SRTM topography and selected footwall rivers (draining to 1805 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-boundary master fault, and along-fault width of averaging swath. Red dots are located at (x, y) = (255190, 4948600) and (287500, 4930500) meters in UTM zone 12T. Black bar = 10 km. (Middle) Dimensionless uplift function $U^*(x_F)$ (red) determined by fitting mean footwall relief (blue, normalized) between distances of 13 to 33 km to the master fault. (Bottom) Standard elevation above base level vs. $\chi$ plot (blue dots), and uplift-corrected $\chi_U$-plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
**Fig. S8.** River profile analysis of the Beaverhead Range, Idaho/Montana, USA. (Top) SRTM topography and selected footwall rivers (draining to 1900 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-bounding master fault, and along-fault width of averaging swath. Red dots are located at (x, y) = (295470, 4969300) and (310780, 4955400) meters in UTM zone 12T. Black bar = 10 km. (Middle) Dimensionless uplift function $U^*(x_F)$ (red) determined by fitting mean footwall relief (blue, normalized) between distances of 8 to 50 km to the master fault. (Bottom) Standard elevation above base level vs. $\chi$ plot (blue dots), and uplift-corrected $\chi_U$–plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
Fig. S9. River profile analysis of the Wassuk Range, Nevada, USA. (Top) SRTM topography and selected footwall rivers (draining to 1500 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-bounding master fault, and along-fault width of averaging swath. Red stars are located at (x, y) = (346430, 4296700) and (346890, 4276000) meters in UTM zone 11S. Black bar = 10 km. (Middle) Dimensionless uplift function $U^*(x)$ (red) determined by fitting mean footwall relief (blue, normalized) between distances of 6 to 21 km to the master fault. (Bottom) Standard elevation above base level vs. $\chi$ plot (blue dots), and uplift-corrected $\chi^U$-plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
Fig. S10. River profile analysis of the Santa Rosa Range, Nevada, USA. (Top) SRTM topography and selected footwall rivers (draining to 1450 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-bounding master fault, and along-fault width of averaging swath. Red dots are located at (x, y) = (436420, 4604600) and (434960, 4584800) meters in UTM zone 11T. Black bar = 10 km. (Middle) Dimensionless uplift function $U^*(x_F)$ (red) determined by fitting mean footwall relief (blue, normalized) between distances of 8 to 25 km to the master fault. (Bottom) Standard elevation above base level vs. $\chi$ plot (blue dots), and uplift-corrected $\chi_U$-plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
Fig. S11. River profile analysis of the Warner Mountains, California, USA. (Top) SRTM topography and selected footwall rivers (draining to 1430 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-bounding master fault, and along-fault width of averaging swath. Red dots are located at (x, y) = (732900, 4610700) and (741000, 4574400) meters in UTM zone 10T. Black bar = 10 km. (Middle) Dimensionless uplift function $U^*(x_F)$ (red) determined by fitting mean footwall relief (blue, normalized) between distances of 6 to 34 km to the master fault. (Bottom) Standard elevation above base level vs. $\chi$ plot (blue dots), and uplift-corrected $\chi_U$–plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
**Fig. S12.** River profile analysis of the Western shore of Lake Edward, Democratic Republic of the Congo. (Top) SRTM topography and selected footwall rivers (draining to 950 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-bounding master fault, and along-fault width of averaging swath. Red dots are located at (x, y) = (756540, 9935300) and (775610, 9975400) meters in UTM zone 35M. Black bar = 10 km. (Middle) Dimensionless uplift function $U^*(x_F)$ (red) determined by fitting mean footwall relief (blue, normalized) between distances of 15 to 100 km to the master fault. (Bottom) Standard elevation above base level vs. $\chi$ plot (blue dots), and uplift-corrected $\chi^U$-plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
**Fig. S13.** River profile analysis of the North-eastern shore of Lake Malawi / Kipengere Range, Tanzania. (Top) SRTM topography and selected footwall rivers (draining to 1200 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-bounding master fault, and along-fault width of averaging swath. Red dots are located at \((x, y) = (611280, 8951500)\) and \((668620, 8887100)\) meters in UTM zone 36L. Black bar = 10 km. (Middle) Dimensionless uplift function \(U^*(x_F)\) (red) determined by fitting mean footwall relief (blue, normalized) between distances of 40 to 200 km to the master fault. (Bottom) Standard elevation above base level vs. \(\chi\) plot (blue dots), and uplift-corrected \(\chi_U\)-plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
Fig. S14. River profile analysis of the Rwenzori Mountains, Democratic Republic of the Congo / Uganda. (Top) SRTM topography and selected footwall rivers (draining to 1100 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-bounding master fault, and along-fault width of averaging swath. Red dots are located at (x, y) = (800180, 24126) and (815030, 63931) meters in UTM zone 35N. Black bar = 10 km. (Middle) Dimensionless uplift function $U^*(x_F)$ (red) determined by fitting mean footwall relief (blue, normalized) between distances of 20 to 50 km to the master fault. (Bottom) Standard elevation above base level vs. $\chi$ plot (blue dots), and uplift-corrected $\chi^U$–plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
Fig. S15. River profile analysis of the Ruby Mountains, Nevada, USA. (Top) SRTM topography and selected footwall rivers (draining to 1850 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-bounding master fault, and along-fault width of averaging swath. Red dots are located at (x, y) = (632990, 4470400) and (644590, 4497600) meters in UTM zone 11T. Black bar = 10 km. (Middle) Dimensionless uplift function $U^*(x_F)$ (red) determined by fitting mean footwall relief (blue, normalized) between distances of 6 to 36 km to the master fault. (Bottom) Standard elevation above base level vs. $\chi$ plot (blue dots), and uplift-corrected $\chi_U$-plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
Fig. S16. River profile analysis of the Toiyabe Range, Nevada, USA. (Top) SRTM topography and selected footwall rivers (draining to 1950 m asl), color-coded by elevation, in m asl. Thin lines are contours of dimensionless uplift function (red: 1, black: decreasing increments of 0.2), inferred by averaging topography within a fault-normal swath. Thick red line shows approximate strike and location of range-bounding master fault, and along-fault width of averaging swath. Red dots are located at (x, y) = (478660, 4308400) and (490430, 4332400) meters in UTM zone 11S. Black bar = 10 km. (Middle) Dimensionless uplift function \( U^*(x_F) \) (red) determined by fitting mean footwall relief (blue, normalized) between distances of 7 to 22 km to the master fault. (Bottom) Standard elevation above base level vs. \( \chi \) plot (blue dots), and uplift-corrected \( \chi \)–plot (red dots), with parabolic fits to the median trend (black curve) as well as the 10th and 90th percentiles (dashed black curves).
Fig. S17. Characteristic decay length of uplift rate ($\alpha$) vs. brittle layer thickness. Data points from compilation of half-grabens (diamonds, Table S2) and horsts (triangles, Table S3). Correlation coefficient in log vs. log space: $R^2 = 0.85$. 
Fig. S18. Erosional efficiency compared to lithological, climatic and tectonic metrics. Erosional efficiency determined through river profile analysis (This study, Tables S2 and S3) plotted against relevant landscape and tectonic metrics compiled by ref. 22 at the Paeroa Range, Lemhi Range, Beaverhead Range, Wassuk Range, Santa Rosa Range (half-grabens) and at the Ruby Mountains and Toiyabe Range (horsts). Metrics include: (A) average precipitation rate (positive correlation, $R^2 = 0.41$); (B) fault vertical slip rate (positive correlation, $R^2 = 0.29$); (C) strength index of exposed lithologies (1: weak / highly erodible to 5: strong; negative correlation, $R^2 = 0.60$); (D) ratio of precipitation rate divided by slip rate multiplied by strength index (positive correlation, $R^2 = 0.49$).
| Simulation name | H (μW/m²) | sedimentation | K (yr⁻¹) | Initial BDT depth (km) | Time elapsed with F1 as master fault (low estimate, Myr) | Time elapsed with F1 as master fault (high estimate, Myr) | Location of new master fault(s) relative to F1: footwall (FW), hanging wall (HW), or both. | Throw on F1 as master fault (low estimate, km) | Throw on F1 as master fault (high estimate, km) | EE |
|----------------|-----------|----------------|----------|------------------------|--------------------------------------------------|--------------------------------------------------|------------------------------------------------------------------------------------------|-----------------------------------------------|-----------------------------------------------|----|
| C70NS          | 0.35      | no             | 1.00E-07 | 22                     | 8.62                                             | 10.5                                             | HW                                                                                       | 4.9                                           | 5.3                                           | 0.4 |
| C70            | 0.35      | yes            | 1.00E-07 | 22                     | 10.4                                             | 13.6                                             | HW                                                                                       | 6.6                                           | 7.7                                           | 0.4 |
| C65            | 0.35      | yes            | 3.16E-07 | 22                     | 9.65                                             | 13                                               | HW                                                                                       | 6.4                                           | 7.8                                           | 1.26|
| C60            | 0.35      | yes            | 1.00E-06 | 22                     | 10.6                                             | 13.6                                             | HW                                                                                       | 7.3                                           | 8.6                                           | 4   |
| C55            | 0.35      | yes            | 3.16E-06 | 22                     | 11.5                                             | 14.4                                             | HW                                                                                       | 8.1                                           | 9.4                                           | 12.6|
| C50            | 0.35      | yes            | 1.00E-05 | 22                     | 13.6                                             | 16.5                                             | HW                                                                                       | 9.2                                           | 10.4                                          | 40  |
| F70NS          | 0.55      | no             | 1.00E-07 | 19                     | 7.09                                             | 8.53                                             | both                                                                                     | 4.7                                           | 5.1                                           | 0.4 |
| F70            | 0.55      | yes            | 1.00E-07 | 19                     | 11.8                                             | 15                                               | FW                                                                                       | 7.1                                           | 7.8                                           | 0.4 |
| F65            | 0.55      | yes            | 3.16E-07 | 19                     | 11.2                                             | 14.6                                             | FW                                                                                       | 7.1                                           | 7.6                                           | 1.26|
| F60            | 0.55      | yes            | 1.00E-06 | 19                     | 16.2                                             | 19.7                                             | HW                                                                                       | 9.5                                           | 10.7                                          | 4   |
| F55            | 0.55      | yes            | 3.16E-06 | 19                     | 16.5                                             | 19.8                                             | HW                                                                                       | 10.2                                          | 11.1                                          | 12.6|
| F50            | 0.55      | yes            | 1.00E-05 | 19                     | 15.8                                             | 19.1                                             | HW                                                                                       | 10.2                                          | 11.1                                          | 40  |
| H70NS          | 0.75      | no             | 1.00E-07 | 16                     | 6.33                                             | 8.82                                             | both                                                                                     | 4.1                                           | 4.7                                           | 0.4 |
| H70            | 0.75      | yes            | 1.00E-07 | 16                     | 8.96                                             | 11.9                                             | FW                                                                                       | 5.9                                           | 6.7                                           | 0.4 |
| H65            | 0.75      | yes            | 3.16E-07 | 16                     | 8.91                                             | 12.5                                             | FW                                                                                       | 6.1                                           | 7.1                                           | 1.26|
| H60            | 0.75      | yes            | 1.00E-06 | 16                     | 15.3                                             | 19.1                                             | both                                                                                     | 8.8                                           | 9.5                                           | 4   |
| H55            | 0.75      | yes            | 3.16E-06 | 16                     | 19.3                                             | 21.5                                             | HW                                                                                       | 10.5                                          | 10.6                                          | 12.6|
| H50            | 0.75      | yes            | 1.00E-05 | 16                     | 17.8                                             | 20.2                                             | HW                                                                                       | 10.5                                          | 10.7                                          | 40  |
| M70NS          | 1         | no             | 1.00E-07 | 14                     | 6.55                                             | 9.14                                             | both                                                                                     | 4.1                                           | 4.9                                           | 0.4 |
| M70            | 1         | yes            | 1.00E-07 | 14                     | 6.46                                             | 9.44                                             | FW                                                                                       | 4.7                                           | 5.5                                           | 0.4 |
| M65            | 1         | yes            | 3.16E-07 | 14                     | 9.17                                             | 12.6                                             | FW                                                                                       | 5.7                                           | 6.5                                           | 1.26|
| M60            | 1         | yes            | 1.00E-06 | 14                     | 13                                               | 17                                               | FW                                                                                       | 7.6                                           | 8.6                                           | 4   |
| M55            | 1         | yes            | 3.16E-06 | 14                     | 16.9                                             | 19.2                                             | FW                                                                                       | 9.1                                           | 9.2                                           | 12.6|
| M50            | 1         | yes            | 1.00E-05 | 14                     | 18.5                                             | 21.3                                             | HW                                                                                       | 9.9                                           | 10.3                                          | 40  |
| V70NS          | 1.4       | no             | 1.00E-07 | 11                     | 5.59                                             | 7.57                                             | FW                                                                                       | 3.3                                           | 4.7                                           | 0.4 |
| V70            | 1.4       | yes            | 1.00E-07 | 11                     | 6.44                                             | 8.94                                             | FW                                                                                       | 4.2                                           | 4.8                                           | 0.4 |
| V65            | 1.4       | yes            | 3.16E-07 | 11                     | 8.78                                             | 11.3                                             | FW                                                                                       | 5.5                                           | 5.5                                           | 1.26|
| V60            | 1.4       | yes            | 1.00E-06 | 11                     | 10.4                                             | 12.9                                             | FW                                                                                       | 5.9                                           | 6.5                                           | 4   |
| V55            | 1.4       | yes            | 3.16E-06 | 11                     | 16.1                                             | 19.5                                             | both                                                                                     | 9.1                                           | 9.9                                           | 12.6|
| V50            | 1.4       | yes            | 1.00E-05 | 11                     | 17.4                                             | 21                                               | both                                                                                     | 9.7                                           | 10.6                                          | 40  |

**Table S1.** Summary of all simulation inputs and key results.
### Table S2. Compilation of structural and morphometric characteristics of fluvially-eroded half-grabens. *a* indicates relief measurements from ref. 22. See Supplementary Text for references related to fault throw and brittle layer thickness. B&R: Basin & Range. EAR: East African Rift.

| Half-graben                        | Rift          | Relie (m) | Dip direction of master fault | Throw on master fault (km) | Thickness of brittle upper crust (km) | Base altitude for χ plots (m) | Master fault UTM coordinates (x1, y1) and (x2, y2) (m) + UTM zone | Limits of topographic fit (distance to fault, km) | Far-field uplift rate U* (dimensionless) | Uplift decay length (km) | Erosional efficiency: median estimate and range |
|-----------------------------------|---------------|-----------|-------------------------------|----------------------------|--------------------------------------|-----------------------------|-----------------------------------------------------------------|-------------------------------------------|-----------------------------------------------|-------------------------------|------------------------------------------------|
| Taupo Rift, New Zealand          | Paeroa Range  | 413*      | NW                           | 0.56–0.9                   | 6–8                                  | 400                         | (4.3843x10^4, 5.7567x10^4) (4.3115x10^4, 5.7487x10^4) UTM 60H | 2–11.4                     | 0.0733                                      | 4.0                           | 2.81 (1.99–3.02)                                            |
| Rio Grande, New Mexico, USA      | Sandia        | 1178      | W                            | 5–7                        | 7–10                                 | 2100                        | (3.6423x10^4, 3.8973x10^4) (3.6452x10^4, 3.8866x10^4) UTM 13S | 6.6–17                     | 0.2013                                      | 6.6                           | 1.28 (1.06–2.32)                                            |
| Nevada, USA                      | B&R           | 1273*     | SW                           | 5–6                        | 12–16                                | 1805                        | (2.6519x10^4, 4.9486x10^4) (2.875x10^4, 4.9305x10^4) UTM 12T | 13–33                     | 3.736x10^4                                  | 23.4                          | 0.98 (0.63–2.08)                                            |
| E                            | Beaverhead Range, Idaho / Montana, USA | 916*     | SW                           | 5–6                        | 12–16                                | 1900                        | (2.9547x10^4, 4.9693x10^4) (3.1078x10^4, 4.9554x10^4) UTM 12T | 8–50                       | 0.3039                                      | 15.7                          | 2.66 (2.63–5.87)                                            |
| E                            | Wassuk Range, Nevada, USA | 1166*     | E                            | 2.5–8.5                     | 11–14                                | 1500                        | (4.3642x10^4, 4.6046x10^4) (4.3496x10^4, 4.5848x10^4) UTM 11S | 6–21                       | 0.1765                                      | 8.9                           | 1.15 (0.87–2.57)                                            |
| E                            | Santa Rosa Range, Nevada, USA | 1071*     | W                            | 5–6                        | 10–15                                | 1450                        | (4.3496x10^4, 4.5848x10^4) UTM 11T | 8–25                       | 0.0954                                      | 9.1                           | 0.62 (0.51–1.11)                                            |
| E                            | Warner Mountains, California, USA | 946     | E                            | 4–5                        | 9–13                                 | 1430                        | (7.329x10^4, 4.6107x10^4) (7.410x10^4, 4.5744x10^4) UTM 11T | 6–34                       | 0.0724                                      | 15.1                          | 1.34 (0.94–2.19)                                            |
| E                            | W. shores of Lake Edward, Democratic Republic of the Congo | 1535 | E | ≥7 | 25–30 | 950 | (7.6478x10^4, 9.9477x10^4) (7.7653x10^4, 9.9768x10^4) UTM 35M | 15–100 | 0.0794 | 48.5 | 1.21 (0.84–2.37) |
| E                            | Kipengere Range / N.E. shores of Lake Malawi, Tanzania | 1695 | W | ≥6.6–7.4 | 32–37 | 1200 | (6.1128x10^4, 8.9515x10^4) (6.6862x10^4, 8.8871x10^4) UTM 36L | 40–200 | 0.3093 | 124.9 | 3.03 (1.90–4.52) |
| Horst                          | Rift  | Relief (m) | Name and dip direction of dominant fault | Throw on dominant fault (km) | Thickness of brittle upper crust (km) | Base altitude for plots (m) | Dominant fault UTM coordinates (x1, y1) and (x2, y2) (m + UTM zone) | Limits of topographic fit (distance to fault, km) | Far-field uplift rate U*min (dimensionless) | Uplift decay length (km) | Erosional efficiency: median estimate and range |
|-------------------------------|-------|------------|------------------------------------------|-----------------------------|--------------------------------------|----------------------------|---------------------------------------------------------------------|-----------------------------------------------|---------------------------------------------|------------------------------------------|-------------------------------------------------|
| Rwenzori Mountains, Democratic Republic of the Congo / Uganda | EAR   | 2769       | Bwamba fault, W-NW                        | 6.3–9.1                     | 25–30                                | 1100                       | (8.0018×10⁵, 2.4126×10⁵) (8.1503×10⁶, 6.3931×10⁶) UTM 35N | 20–50                                         | 0.0551                                       | 26.0                                     | 0.28 (0.19–0.46)                                |
| Ruby Mountains, Nevada, USA   | B&R   | 1068       | Ruby Valley fault zone, E-SE              | 2.4                         | 10–15                                | 1850                       | (6.3299×10⁵, 4.4704×10⁵) (6.4459×10⁶, 4.4976×10⁶) UTM 11T | 6–36                                          | 0.0563                                       | 17.1                                     | 1.08 (0.72–1.35)                                |
| Toiyabe Range, Nevada, USA    | B&R   | 1051       | Toiyabe Range fault zone, E              | 4–5                         | 10–15                                | 1950                       | (4.7866×10⁵, 4.3084×10⁵) (4.9043×10⁶, 4.3324×10⁶) UTM 11S | 7–22                                          | 0.1603                                       | 10.4                                     | 0.92 (0.90–2.60)                                |

**Table S3.** Compilation of structural and morphometric characteristics of fluvially-eroded horsts. a indicates relief measurements from ref. 22. See Supplementary Text for references related to fault throw and brittle layer thickness. B&R: Basin & Range. EAR: East African Rift.
Site | Throw rate (mm/yr) | K (10^4 yr⁻³), median estimate and range
--- | --- | ---
Paeroa Range, New Zealand | 1.5 | 1.5 (1.1–1.6)
Lemhi Range, Idaho, USA | 0.5 | 0.18 (0.11–0.37)
Beaverhead Range, Idaho / Montana, USA | 0.25 | 0.24 (0.24–0.53)
Wassuk Range, Nevada, USA | 0.6 | 0.25 (0.19–0.56)
Santa Rosa Range, Nevada, USA | 0.45 | 0.10 (0.08–0.18)
Ruby Mountains, Nevada, USA | 0.24 | 0.09 (0.06–0.12)
Toiyabe Range, Nevada, USA | 0.3 | 0.10 (0.10–0.28)

**Table S4.** Erosion coefficient $K$ determined at study sites where fault throw rate has been estimated ($K = EE \frac{U_0}{\sqrt{A_0}}$, with $A_0 = 10^6$ m²). Throw rates are from ref. 22. $U_0$ is assumed to be 36% of the fault throw rate based on our numerical simulations. Range of $K$ values entirely reflects uncertainty on $EE$. 


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