

Review Article

The paradox of post-orogenic mountain longevity

O paradoxo da longevidade das montanhas pós-orogênicas

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Abstract: How landscapes evolve in the aftermath of mountain building has intrigued geoscientists for more than a century. Intraplate landscapes with pronounced relief, peaks exceeding 2000 m, and mean elevations above 1000 m persist even where tectonic crustal thickening ceased tens or hundreds of millions of years ago. The longevity of relief in these ancient orogens remains puzzling, and it is striking that, in some Brazilian orogenic belts, their Precambrian history is better understood than their post-orogenic evolution. This review synthesises current understanding of the paradox of post-orogenic mountain longevity. We examine: *i*) the global distribution of these mountains; *ii*) approaches used to estimate the timescale of topographic decay; *iii*) hypotheses proposed to explain post-orogenic landscape evolution; and *iv*) a simple numerical example illustrating the central role of the flexural-isostatic response in relief reduction. We argue that post-orogenic mountain evolution generally lies between two end-member scenarios, 'topographic rejuvenation' and 'long-term relief survival', with the relative importance of each varying among landscapes according to their geodynamic and climatic histories. We hope to stimulate further research on post-orogenic landscape evolution, using Brazilian settings as natural laboratories.

Keywords: geomorphology; post-orogenic landscapes; topography; topographic decay; landscape evolution.

Resumo: A evolução da topografia após o término dos processos orogênicos responsáveis pela formação de cadeias montanhosas tem intrigado as geociências há mais de um século. Paisagens intraplaca com relevos expressivos, picos acima de 2000 m e elevações médias superiores a 1000 m persistem mesmo onde o espessamento crustal tectônico cessou há dezenas ou centenas de milhões de anos. A longevidade do relevo nesses orógenos antigos permanece enigmática, e é notável que, em alguns cinturões orogênicos Brasileiros, sua história pré-cambriana seja mais bem compreendida do que sua evolução pós-orogênica. Esta revisão sintetiza o conhecimento atual sobre o paradoxo da longevidade das montanhas pós-orogênicas. Examinamos: *i*) sua distribuição global; *ii*) abordagens para estimar a escala temporal do decaimento topográfico; *iii*) hipóteses sobre a evolução das paisagens pós-orogênicas; e *iv*) um exemplo numérico simples que ilustra o papel central da resposta isostática-flexural na redução do relevo. Argumentamos que a evolução das montanhas pós-orogênicas se distribui ao longo de um contínuo entre dois cenários extremos, "rejuvenescimento topográfico" e "sobrevivência prolongada do relevo", cuja importância relativa varia entre paisagens de acordo com suas histórias geodinâmicas e climáticas. Esperamos estimular novas pesquisas sobre a evolução das paisagens pós-orogênicas, utilizando cenários brasileiros como laboratórios naturais.

Palavras-chave: geomorfologia; paisagens pós-orogênicas; topografia; decaimento topográfico; evolução de paisagens.

1. Introduction

Mountain building is ultimately driven by Earth's internal dynamics, which govern the motion of lithospheric plates and the tectonic processes associated with them (Dewey; Bird, 1970; England; Molnar, 1990; Summerfield, 1991; Burbank; Anderson, 2001). In continental settings, orogenesis is a primary mechanism for generating high relief because crustal shortening and magmatism commonly thicken the crust and, through isostatic adjustment, favour higher surface elevations (e.g., England; Molnar, 1990; Gilchrist; Summerfield, 1990, 1991; Burbank; Anderson, 2001; Watts, 2001; Clift; Vannucchi, 2004; Teixell et al., 2009). Yet in many mountain ranges observed today, the main phase of crustal thickening and associated rock uplift ended tens or hundreds of millions of years ago, while erosional processes remain active. The persistence of substantial relief in these 'post-orogenic mountains', landscapes that remain elevated after the main phase of orogenic construction has ceased, has challenged geologists and geomorphologists for more than a century. It was a central question in the early development of geomorphology and, despite decades of debate, still lacks a single widely accepted explanation (Bishop, 2007; Peifer, 2018), partly because different geodynamic settings can sustain elevated relief through different mechanisms.

Once tectonic rock uplift ceases, conceptual and numerical models predict long-term relief decay because geomorphic processes continue to weather rock and transport material downslope towards base level without being counterbalanced by tectonic uplift (e.g., Davis, 1899; Ahnert, 1970; Baldwin; Whipple; Tucker, 2003; Egholm; Knudsen; Sandiford, 2013). Assuming isostatic equilibrium between the crust and the denser underlying mantle, and adopting the mean densities used by Liu (2014), crustal density ($\rho_c = 2700 \text{ kg/m}^3$) and mantle density ($\rho_m = 3300 \text{ kg/m}^3$), denudational mass loss should drive compensatory vertical rock motion (Ahnert, 1970; Turcotte; Schubert, 1982). At the local Airy-isostatic limit, this rock uplift can compensate for approximately 82% of the thickness removed by denudation for the adopted densities (e.g., Ahnert, 1970; Turcotte; Schubert, 1982). Isostatic compensation therefore prolongs the lifetime of post-orogenic relief. Yet numerical modelling consistently predicts that post-orogenic topography should be lowered to sea level within a few tens of millions of years (e.g., Ahnert, 1970; Kooi; Beaumont, 1996; Baldwin; Whipple; Tucker, 2003; Egholm; Knudsen; Sandiford, 2013; Bishop, 2007).

Across a range of approaches, quantitative estimates of the time required to reduce post-orogenic relief to base level are generally one to two orders of magnitude shorter than the time elapsed since the last major phase of orogenic activity in many of these landscapes (e.g., Gilluly, 1955; Schumm, 1963; Judson; Ritter, 1964; Ahnert, 1970; Pinet; Souriau, 1988). This mismatch highlights the difficulty of existing quantitative approaches and topographic-decay models in reproducing both the inferred antiquity and the persistence of pronounced, still-elevated mountainous relief in ancient orogens. The core problem is that elevated and locally steep topography tends to promote higher denudation rates (e.g., Ahnert, 1970; Montgomery; Brandon, 2002; Portenga; Bierman, 2011; Harel; Mudd; Attal, 2016), whereas long-term relief survival requires very low denudation rates to persist over exceptionally long intervals (e.g., Baldwin; Whipple; Tucker, 2003; Bishop, 2007; Egholm; Knudsen; Sandiford, 2013; Peifer, 2018). Although isostatic compensation and other geomorphic mechanisms and feedbacks can substantially prolong relief persistence, published topographic-decay models ultimately converge on progressive lowering towards base level and the development of broadly low-relief, planar topography (Ahnert, 1970; Kooi; Beaumont, 1996; Baldwin; Whipple; Tucker, 2003; Egholm; Knudsen; Sandiford, 2013). The discrepancy therefore concerns not only the timescale of relief decay, but also the predicted end-state morphology: after comparable intervals, many ancient orogens remain elevated, locally steep and mountainous. This conflict between estimated decay timescales and observed post-orogenic topography underpins the 'paradox of persistent mountain belts' (Pelletier, 2008, p. 101) and has led to two end-member explanations: either (1) post-orogenic episodes of topographic rejuvenation are required to explain relief survival (e.g., Hack, 1982; Pazzaglia; Brandon, 1996; Quigley et al., 2007; Gallen; Wegmann; Bohnenstiehl, 2013; Liu, 2014; Gallen, 2018; Silva; Sacek, 2019; Calejari et al., 2021; Fonte-Boa et al., 2022), or (2) additional geomorphic processes that inhibit erosion or reduce erosional efficiency must be incorporated into decay estimates, thereby lengthening the timescale of topographic reduction (e.g., Baldwin; Whipple; Tucker, 2003; Egholm; Knudsen; Sandiford, 2013; Scharf et al., 2013; Vasconcelos et al., 2019; Peifer et al., 2021).

Post-orogenic and cratonic landscapes make up a large fraction of continental land area, yet their long-term dynamics and evolution remain comparatively poorly understood (Bishop, 2007). Here we provide a critical, integrated review of the persistence of mountainous topography in ancient orogens, organised around the paradox of persistent mountain belts and the principal observations, quantitative approaches, hypotheses and process

interactions proposed to explain their evolution. We emphasise that many elevated regions of Brazil are spatially associated with the Brasiliano orogenic systems of West Gondwana (e.g., Alkmim, 2015), but their present-day topography need not be a direct or exclusive expression of ancient orogenic inheritance. It may also record uplift and landscape reorganisation associated with South Atlantic rifting and subsequent post-rift evolution. By framing the problem in this way, we aim to encourage Brazilian geoscientists to investigate the geomorphic history of post-orogenic landscapes and contribute to resolving this enduring paradox. We also provide, particularly for postgraduate students, an entry point to key concepts such as flexural-isostatic compensation of denudation, topographic decay and the stream-power model of fluvial incision.

Specifically, we: *i*) review examples of post-orogenic landscapes worldwide; *ii*) explain the main quantitative approaches used to estimate post-orogenic relief reduction, with emphasis on the factors that control relief longevity; *iii*) examine the principal competing hypotheses for the geomorphic evolution of post-orogenic landscapes and their testable implications; *iv*) present an example from the Quadrilátero Ferrífero (QF) illustrating the fundamental role of the flexural-isostatic response in the topographic evolution of ancient orogens, using a numerical framework in which effective elastic thickness is an important parameter; and *v*) synthesise the geomorphic factors and process interactions likely to govern the long-term evolution of post-orogenic landscapes.

2. Post-orogenic Mountains Worldwide

Post-orogenic mountains occur in many parts of the world, including the Cape Fold Belt (South Africa), the Ural Mountains (Russia), the Appalachian Mountains (USA), the Western Ghats (India), the Scandes (Scandinavia), sectors of the Great Escarpment of Australia, the Great Escarpment of Namibia, the Guiana Shield and the Great Escarpment of Sri Lanka (Figure 1A). These examples are central to the literature on post-orogenic landscape evolution because they combine still-elevated or mountainous relief with long intervals since the main phase of orogenic activity, thereby encapsulating the problem of relief longevity in ancient orogens (e.g., Twidale, 1976, 1999; Pazzaglia; Brandon, 1996; Baldwin; Whipple; Tucker, 2003; Bishop, 2007; Pelletier, 2008; Peifer, 2018). They differ, however, in several important respects: (1) the age of the last orogeny; (2) the geometry and density of structural discontinuities; (3) climatic conditions and their temporal variability; (4) post-orogenic tectonic history; and (5) effective elastic thickness (T_e), which reflects lithospheric flexural rigidity. Annual precipitation, for example, ranges from only a few hundred millimetres in Namibia and South Africa, to slightly more than 1000 mm in Brazilian post-orogenic landscapes and the Appalachians, and to 2000–10,000 mm in Sri Lanka and the Western Ghats (Figure 1B). The time elapsed since the last major tectonic crustal-thickening event, which we use to mark the onset of the post-orogenic period, also varies markedly among these ancient orogens. We use ‘ancient orogen’ as shorthand, recognising that the geotectonic units involved commonly record polyphase tectonic histories. Where later uplift or tectonic reactivation occurred, this temporal marker refers not to those younger events but to the end of the principal phase of tectonic crustal thickening that generated the inherited proto-topography. On this basis, relief may have persisted for approximately 280–300 Ma in the Ural Mountains, Russia (Berzin et al., 1996; Puchkov, 1997), 325–360 Ma in the Appalachian Mountains, USA (Hibbard, 2000; Wagner; Stewart; Metcalf, 2012), approximately 380 Ma in the Lachlan Fold Belt, Australia (Blackburn; Ferrier; Perron, 2018), approximately 400 Ma in the Scandes (Pedersen; Huisman; Moucha, 2016), and approximately 450–500 Ma in the Quadrilátero Ferrífero, Brazil (Chemale Jr.; Rosière; Endo, 1994; Alkmim; Marshak, 1998). Several of these examples lie within mature passive margins, however, where present-day topography may reflect not only inherited orogenic relief but also rift and post-rift processes, including flexure, isostatic compensation, tectonic reactivation and, in some cases, later magmatism. The landscapes considered here are therefore not geomorphologically equivalent; rather, they share the defining feature that elevated relief persists after the principal phase of tectonic crustal thickening has ended.

Despite their substantial geomorphic and geological differences, post-orogenic landscapes share several first-order characteristics that point to at least some common elements in their evolution after orogenesis. Most notably, parts of these landscapes retain steep channel and hillslope gradients comparable to those found in tectonically active settings (e.g., von Blanckenburg; Hewawasam; Kubik, 2004; Gallen; Wegmann; Bohnenstiehl, 2013; Scharf et al., 2013), although steep topography is far more widespread in young orogens. Ancient orogens also commonly expose a wide range of lithologies with contrasting properties, erodibilities and geological histories, including crystalline rocks, volcanic units, metasedimentary rocks and sedimentary covers of different ages and compositions (e.g., Calegari et al., 2021; Marques et al., 2021; Peifer et al., 2021; Fonte-Boa et al., 2022). This heterogeneity produces strong lateral and vertical contrasts in resistance to weathering and erosion, influencing

both the spatial pattern of denudation and the style and timescale of transient landscape response (e.g., Braun et al., 2014; Gallen, 2018; Peifer et al., 2021; Fadul; Oliveira; Val, 2022). The post-orogenic interval can also reasonably be viewed as one of relative tectonic quiescence, substantially more stable than the tectonic regime of active orogenesis (e.g., Twidale, 1976; Mandal et al., 2015). Finally, denudation rates in post-orogenic landscapes are orders of magnitude lower than in tectonically active settings (Bishop, 2007). Global compilations of ^{10}Be -derived denudation rates from detrital sediment, which integrate over approximately 10^3 – 10^5 years, together with apatite fission-track exhumation rates, which integrate over approximately 10^6 – 10^8 years, indicate typical rates of approximately 400–3000 m/Ma in young orogens but only approximately 1–100 m/Ma in ancient orogens (Portenga; Bierman, 2011; Harel; Mudd; Attal, 2016; Blackburn; Ferrier; Perron, 2018).

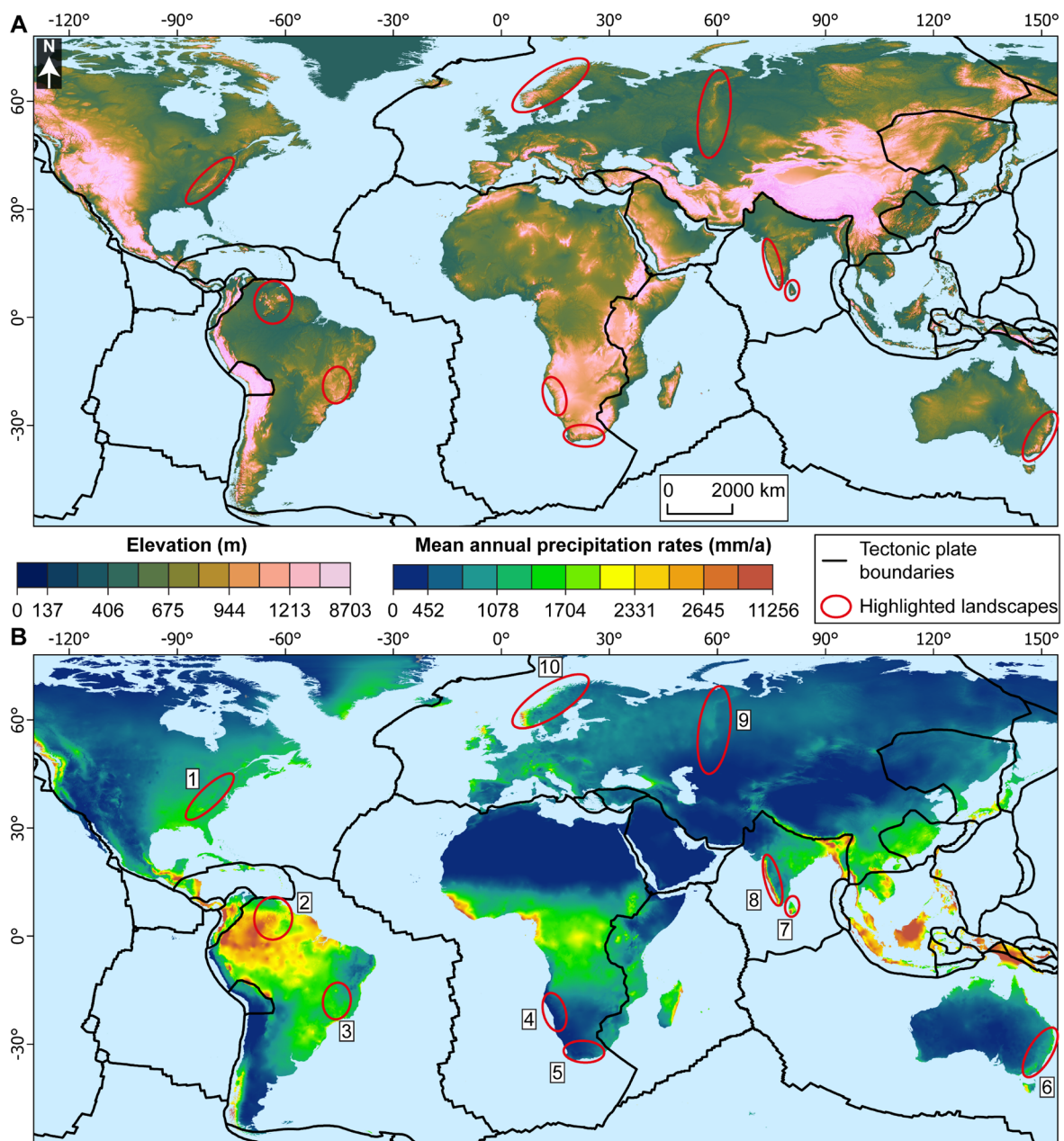


Figure 1. Elevation and mean annual precipitation across the principal examples of post-orogenic mountains worldwide. (A) Global elevation from GMTED2010 (Danielson; Gesch, 2011). (B) Mean annual precipitation from WorldClim v.2 (Fick; Hijmans, 2017). Black lines show tectonic plate boundaries from the PB2002 model of Bird (2003). Red ellipses highlight the post-orogenic landscapes discussed in the text: (1) Appalachian Mountains; (2) Guiana Shield; (3) southeastern Brazil; (4) Great Escarpment of Namibia; (5) Cape Fold Belt; (6) southeastern Australia; (7) Sri Lankan orogen; (8) Western Ghats; (9) Ural Mountains; and (10) Scandes.

Brazil contains several prominent post-orogenic landscapes, including the Espinhaço Range, Serra Geral do Tocantins Range, Chapada dos Veadeiros, Alto Paranaíba Arch, Quadrilátero Ferrífero, Mantiqueira Range, and Serra do Mar Range. Many are spatially associated with Brasiliano orogenic belts surrounding cratonic domains (e.g., Alkmim, 2015) and form regions of pronounced topography that share several geomorphic traits: relative Cenozoic tectonic quiescence, low seismicity, complex structural architecture, juxtaposition of lithologies with contrasting resistance, and strong spatial variability in topography, with subdued terrain adjacent to rugged, steep relief. At the same time, these landscapes differ markedly in their geotectonic and climatic histories, lithospheric effective elastic thickness, precipitation patterns, denudation rates and the resistance of rocks exposed at the surface. In some cases, particularly along the Atlantic margin of southeastern Brazil and in the Alto Paranaíba Arch, present-day topography also records uplift and landscape reorganisation associated with the separation of South America and Africa. Ancient orogenic inheritance should therefore not be treated as the sole control on modern relief. This combination of post-orogenic characteristics and pronounced geomorphic and geotectonic heterogeneity makes these Brazilian landscapes especially valuable natural laboratories for investigating the longevity of relief in ancient orogens.

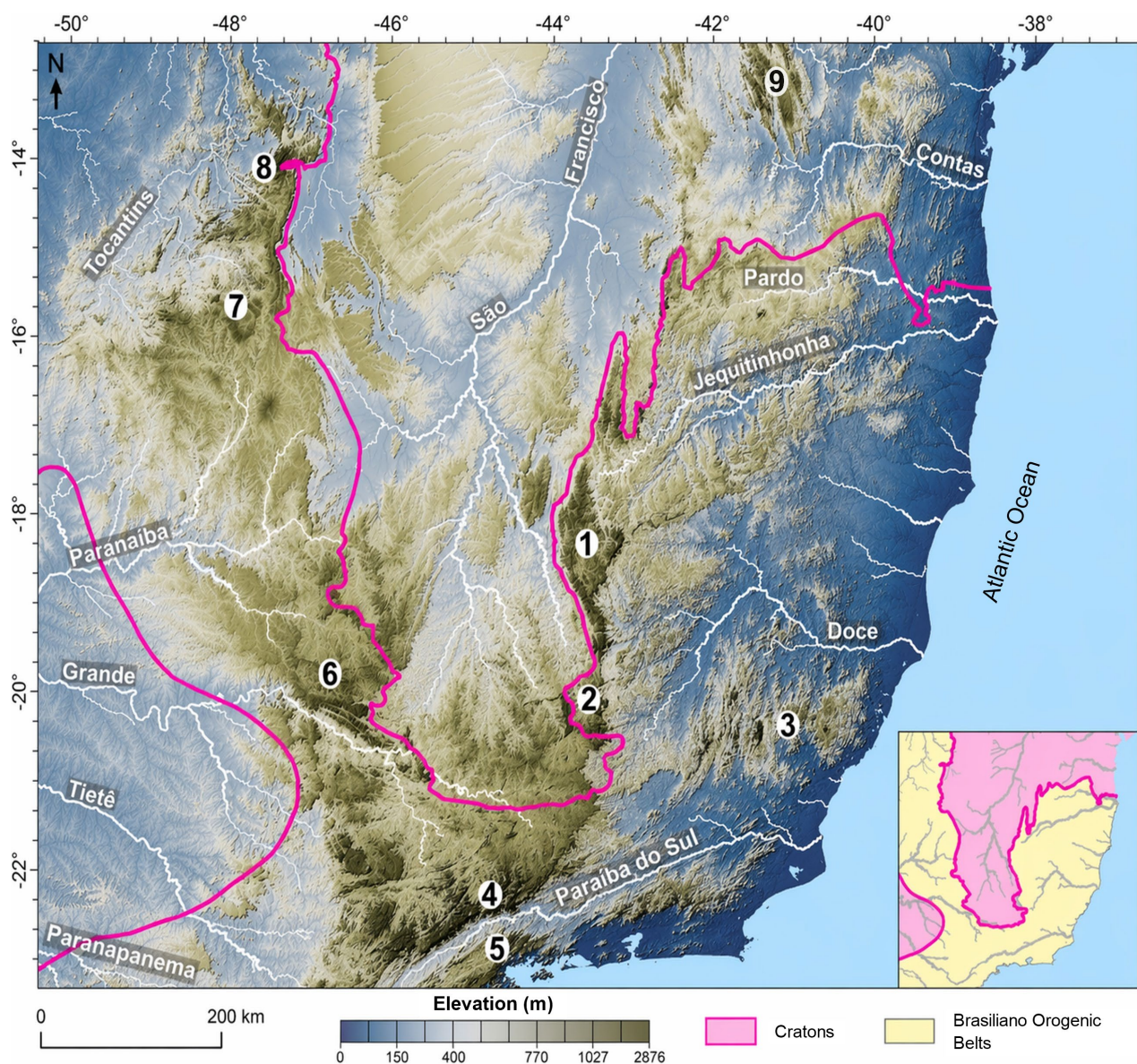


Figure 2. Elevation and tectonic setting of southeastern Brazil, with regional tectonic context shown in the inset. Pink lines mark craton boundaries. In general, areas associated with Brasiliano orogenic belts surrounding the cratons have pronounced topography, whereas cratonic interiors are topographically lower. Rivers with drainage areas greater than 100 km² are shown as white lines, with line width scaled to upstream drainage area. Numbers identify

post-orogenic landscapes: (1) Southern Espinhaço Range; (2) Quadrilátero Ferrífero; (3) Serra Capixaba Range; (4) Mantiqueira Range; (5) Serra do Mar Range; (6) Alto Paranaíba Arch; (7) Serra Geral do Tocantins Range; (8) Chapada dos Veadeiros; and (9) Chapada Diamantina. Elevation data: Copernicus DEM, 30 m resolution. Geological base: CPRM (2004). The craton boundaries follow this institutional cartographic database and should be regarded as one of several possible delineations in the literature, including more recent interpretations.

We also note a group of orogens in which the main phase of compressional deformation probably ended only approximately 25–50 Ma ago, including the southern Rocky Mountains (Colorado, USA), the Pyrenees (Spain and France), the Atlas Mountains (Morocco) and the Western Alps (e.g., Tucker; van der Beek, 2013). Unlike the classic ancient orogens discussed above, these systems entered the post-orogenic phase much more recently. Comparing their geomorphic characteristics with those of ancient orogens may therefore provide important insight into how mountain landscapes evolve after the main relief-generating phase has ended (Tucker; van der Beek, 2013).

3. Estimates of Post-orogenic Topographic Decay

Several analytical approaches have been used to estimate the time required for post-orogenic relief reduction. The simplest divides mean landscape elevation by the mean rate of denudational surface lowering. Early applications of this approach yielded decay times of 10–25 Ma for post-orogenic relief to be reduced to sea level (e.g., Gilluly, 1955; Schumm, 1963; Judson; Ritter, 1964), generally using sediment fluxes in modern rivers to estimate mean surface-lowering rates. These results were influential in establishing an almost consensual view that landscapes and landforms are geologically young, no older than the Neogene (e.g., Thornbury, 1969). Such estimates, however, have two major limitations: (1) they were derived using mean denudation rates faster than those representative of post-orogenic landscapes (Young, 1983; Bishop, 2007); and (2) they neglect both the role of isostasy in topographic decay and the decline in denudation rates as relief is reduced (e.g., Ahnert, 1970; Pinet; Souriau, 1988; Summerfield; Hulton, 1994; Bishop, 2007).

Alternative approaches were therefore developed. One of the most influential assumes that denudational efficiency depends on topography, so that denudation rates decline as topography is lowered through time (e.g., Ahnert, 1970; Pinet; Souriau, 1988). The topographic metrics used in these studies, however, are not identical. Ahnert (1970) related denudation rates to mean local relief, used as a proxy for mean slope, whereas Pinet and Souriau (1988) related mechanical denudation to mean basin elevation. Both formulations therefore express a dependence of denudation on topography, but the topographic variables are not strictly equivalent. In the equations below, H should be understood generically as the topographic metric used by each approach; when discussing the results of individual studies, we retain the authors' original definitions. If the rate of reduction of H is assumed to be proportional to H itself, topographic decay can be expressed by the following differential equation:

$$\frac{dH}{dt} = -\frac{1}{\lambda_d} \cdot H, \quad (1)$$

where H represents the topographic metric under consideration, t is time, and λ_d is the topographic decay constant, the time required for relief to decline to $1/e$ (~0.37) of its initial value or approximately 37% of H_0 , the initial value of the topographic metric H at the beginning of the decay interval. The solution to Eq. (1) is:

$$H = H_0 \cdot e^{-\frac{t}{\lambda_d}}. \quad (2)$$

If the mean denudation rate (D) varies approximately linearly with H , then:

$$D = a \cdot H + b, \quad (3)$$

where a is the slope of the linear regression and b is its intercept (Ahnert, 1970; Pinet; Souriau, 1988). In the simplified case where b is negligible relative to $a \cdot H$, the exponential form of Eq. (1) is recovered:

$$\lambda_d = \frac{1}{a}. \quad (4)$$

In this simplified form, the topographic decay time t_d , defined as the time required to reduce the topographic metric from an initial value H_0 to a final value H_f , is:

$$t_d = \lambda_d \ln \left(\frac{H_0}{H_f} \right), \quad (5)$$

such that, if H_f is equal to H_0/e , then $t_d = \lambda_d$.

In a related formulation, Ahnert (1970) showed that, without uplift, relief would decline to 10% of its initial value in approximately 11 Ma. Including isostatic compensation increased this time to at least 18.5 Ma. If the exponential form of these trajectories is extrapolated to 1% of the initial relief, the corresponding times are approximately 22 Ma without compensation and at least 37 Ma with isostatic compensation. These values are best viewed as order-of-magnitude estimates that depend on the assumed relationship between denudation and relief. To constrain that relationship empirically, Ahnert (1970) used mean denudation rates for mid-latitude drainage basins derived from fluvial sediment loads and related them to the mean local relief of each basin. Local relief was calculated as the difference between maximum and minimum elevation within 20×20 km windows, which were then averaged for each basin. Figure 3 schematically shows the decay trajectories proposed by Ahnert (1970) with and without isostatic compensation. Using a related approach, but relating mechanical denudation to mean basin elevation, Pinet and Souriau (1988) estimated a characteristic decay timescale of approximately 25 Ma for orogenic belts older than 250 Ma.

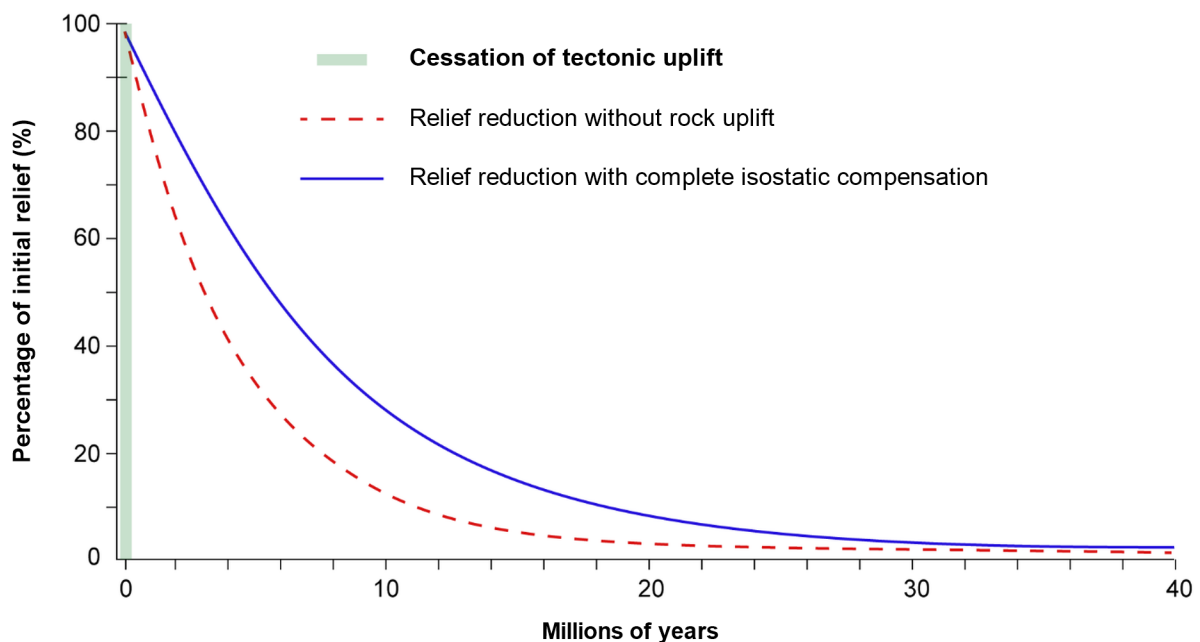


Figure 3. Schematic representation of post-orogenic relief decay through time following Ahnert (1970). The dashed red curve shows the decay trajectory without isostatic compensation, whereas the blue curve shows the trajectory when isostatic compensation of denudational unloading is included. According to Ahnert (1970), relief declines to 10% of its initial value in approximately 11 Ma in the first case and in at least 18.5 Ma in the second. Modified from Ahnert (1970).

Baldwin, Whipple and Tucker (2003) later approached topographic decay in ancient orogens by assuming that the rate of bedrock river incision controls the timescale of relief reduction. Their formulation modifies Whipple's (2001) treatment of drainage-network transience following an abrupt cessation of tectonic uplift. The analytical solution of Baldwin, Whipple and Tucker (2003) is based on the stream-power model of fluvial incision, in which long-term incision rate (E) is a power-law function of upstream drainage area (A), local channel slope (S) and an erosional-efficiency coefficient (K) (Howard; Kerby, 1983):

$$E = K \cdot A^m \cdot S^n, \quad (6)$$

where m and n are positive exponents that reflect the dominant fluvial incision process, channel hydraulics and sediment flux (Whipple; Tucker, 1999). Peifer et al. (2022) provide a useful starting point for readers unfamiliar with the stream-power model of fluvial incision.

Following an abrupt cessation of tectonic uplift, the stream-power model predicts that the transient response of the drainage network begins at base level, where a reduction in local channel gradient generates a concave knickpoint (Whipple, 2001; Baldwin; Whipple; Tucker, 2003; Figure 4). Knickpoints are morphological discontinuities in river longitudinal profiles associated with localised changes in channel gradient. Although most commonly recognised as abrupt increases in gradient, concave knickpoints represent the opposite case, an abrupt decrease in local gradient (e.g., Whipple, 2001; Baldwin; Whipple; Tucker, 2003; Ruetenik et al., 2018; Gailleton et al., 2019). Their transient behaviour is analogous to that of convex knickpoints but opposite in direction. A concave knickpoint marks an abrupt reduction in local fluvial incision rate that propagates upstream through the drainage network as a wave of reduced erosion. Such a response can follow, for example, a decrease in rock-uplift rate or a rise in base level (e.g., Whipple, 2001; Baldwin; Whipple; Tucker, 2003; Chen; Shyu; Chang, 2015). After tectonic uplift ceases, the concave knickpoint separates a downstream reach that has already adjusted to the new uplift condition from an upstream reach that still retains the gradient inherited from the former steady-state profile (Figure 4). Importantly, the upstream reach is not fixed in elevation: while its gradient remains unchanged until the transient passes, its absolute elevation decreases through time because incision initially continues at the pre-perturbation rate while rock uplift has been reduced (Baldwin; Whipple; Tucker, 2003; Figure 4).

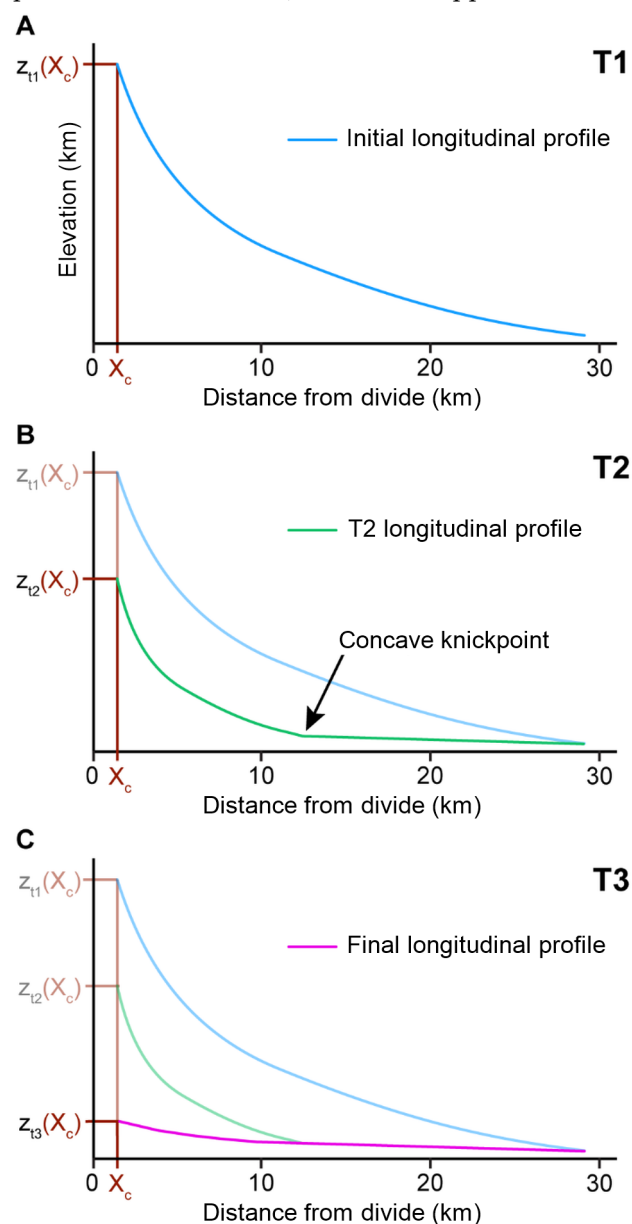


Figure 4. Simplified conceptual model of the transient response of a river profile to an abrupt decrease in tectonic rock uplift. At T1 (panel A), the channel displays its initial longitudinal profile, adjusted to the pre-perturbation

condition. At T2 (panel B), the transient response propagates upstream as a concave knickpoint that separates a downstream reach already adjusted to the new uplift condition from an upstream reach that still retains the gradient of the initial profile. Although the gradient of this upstream reach remains unchanged until the knickpoint passes through it, its absolute elevation decreases through time, such that $z_{T2}(x_c) < z_{T1}(x_c)$. At T3 (panel C), the final profile shown in the schematic represents a later stage of the transient response, after further upstream propagation, and channel-head elevation is reduced to $z_{T3}(x_c)$. In the example shown, the decrease in tectonic rock uplift is taken to the limiting case of $U = 0$. Adapted from Baldwin, Whipple and Tucker (2003).

If the landscape was at steady state before tectonic uplift ceased abruptly, the topographic-decay timescale can be estimated as the time required for the concave knickpoint to migrate from regional base level, where $x = 0$, to the channel head, where $x = x_c$. In this framework, the topographic-decay time is (e.g., Whipple; Tucker, 1999; Goren; Fox; Willett, 2014):

$$t_d = \int_0^{x_c} \frac{dx}{V_{kp}} = \int_0^{x_c} \frac{dx}{K \cdot A(x)^m \cdot S(x)^{n-1}}, \quad (7)$$

where V_{kp} is the celerity of upstream migration of the concave knickpoint, given by the nonlinear kinematic-wave equation $V_{kp} = K \cdot A(x)^m \cdot S(x)^{n-1}$ (Rosenbloom; Anderson, 1994; Whipple; Tucker, 1999). The formulation assumes that the tectonic rock-uplift rate immediately before tectonic activity ceased ($U_{initial}$) was balanced everywhere by long-term fluvial incision. $U_{initial}$, K , A , S , m and n therefore all influence the timescale of post-orogenic topographic reduction (Whipple; Tucker, 1999; Whipple, 2001; Baldwin; Whipple; Tucker, 2003). Two points are especially important: *i*) in the simplest case, $n = 1$, fluvial response time is independent of S and $U_{initial}$; and *ii*) low K values, commonly associated with rocks resistant to weathering and erosion (e.g., Peifer et al., 2021) or with low-precipitation settings (e.g., Ferrier; Huppert; Perron, 2013), slow knickpoint migration and substantially lengthen topographic-decay times. K is therefore a fundamental control on post-orogenic landscape evolution, governing both transient landscape response (Eq. 7) and fluvial erosion rates (Eq. 6) (Peifer, 2018; Peifer et al., 2021; Haag et al., 2025).

Using reasonable values of K , A , S , m , n and the initial uplift rate, Baldwin, Whipple and Tucker (2003) estimated that post-orogenic relief could be reduced to sea level within approximately 2–5 Ma, or approximately 10–30 Ma if the maximum flexural-isostatic response is included. Incorporating additional processes that inhibit fluvial incision, together with flexural-isostatic compensation, lengthens these timescales, although the simulated landscapes still evolve towards regionally extensive low-relief surfaces resembling classical planation surfaces (Baldwin; Whipple; Tucker, 2003). For example, relief-reduction times increase to approximately 35–90 Ma if the drainage network becomes transport-limited and persistent alluvial cover shields channel beds and margins from incision (Baldwin; Whipple; Tucker, 2003). Adding a critical incision threshold to this transport-limited landscape, so that incision occurs only during rainfall events that generate discharge above a critical value, extends estimated topographic-decay times beyond 100 Ma (Baldwin; Whipple; Tucker, 2003).

Sediment is itself an erosional agent, supplying tools that enhance bedrock incision through processes such as abrasion (Sklar; Dietrich, 2001). Post-orogenic transience has therefore also been modelled as a nonlinear function of sediment supply to river channels (e.g., Egholm; Knudsen; Sandiford, 2013). In this framework, the cessation of tectonic rock uplift substantially reduces landslide frequency, which in turn controls the volume and grain size of sediment delivered to channels (Larsen; Montgomery, 2012). Reduced sediment supply limits the availability of erosional tools, lowers fluvial incision rates and ultimately lengthens the timescale of post-orogenic relief reduction (Egholm; Knudsen; Sandiford, 2013). Under these conditions, simulated landscapes can retain peaks at elevations of approximately 2000 m for as long as 200 Ma after tectonic uplift ceases (Egholm; Knudsen; Sandiford, 2013). Mountainous relief can persist for up to approximately 300 Ma, after which the simulated landscape evolves towards a broadly planar configuration dominated by alluvial channels in which sediment supply exceeds transport capacity (Egholm; Knudsen; Sandiford, 2013).

These models also generally do not account explicitly for the fact that, in many ancient orogens, rocks exposed at the surface are not limited to units originally deformed during orogenesis; they also include lithologies exhumed from deeper crustal levels. This is particularly relevant in parts of southeastern Brazil, where some present-day relief is underlain by gneissic domes and other metamorphic and igneous units exposed after erosion removed higher structural levels of the orogen. Differential erosion therefore acts on a lithological architecture inherited

from the internal structure of the orogen, and progressive exhumation of harder, more resistant rocks may reduce erosional efficiency through time and further prolong relief persistence (Summerfield; Hulton, 1994; Bishop, 2007; Braun et al., 2014; Peifer et al., 2021).

All of these estimates, albeit to different degrees, underpredict the longevity of post-orogenic relief in landscapes such as the Scandes, where tectonic uplift probably ceased approximately 400 Ma ago (e.g., Pedersen; Huismans; Moucha, 2016), and the Quadrilátero Ferrífero, where the last major tectonic event probably occurred approximately 450–500 Ma ago (e.g., Chemale Jr.; Rosière; Endo, 1994; Alkmim; Marshak, 1998). Within these models, the extremely low erosion rates required for long-term relief survival can be sustained only if the residual landscape becomes essentially flat, resembling the final simulated topography produced by the different decay estimates (e.g., Baldwin; Whipple; Tucker, 2003; Egholm; Knudsen; Sandiford, 2013). Yet this conceptual end-member, broadly analogous to a planation surface (e.g., Davis, 1899), is a poor description of the erosional, still kilometre-scale relief that characterises many present-day post-orogenic mountains (Bishop, 2007).

4. Explaining Post-orogenic Topography

Despite a long history of debate, the dynamics and evolution of post-orogenic landscapes remain incompletely understood (Bishop, 2007; Pelletier, 2008; Egholm; Knudsen; Sandiford, 2013; Blackburn; Ferrier; Perron, 2018; Gallen, 2018; Vasconcelos et al., 2019; Peifer et al., 2021; Fadul; Oliveira; Val, 2022). Explanations for the persistence of mountainous relief in intraplate settings generally fall into two broad groups: (1) long-term survival of landforms with an ancient origin, as erosional remnants of ‘fossil’ landscapes (e.g., Twidale, 1976; Baldwin; Whipple; Tucker, 2003; Bishop, 2007; Egholm; Knudsen; Sandiford, 2013; Scharf et al., 2013; Pedersen; Huismans; Moucha, 2016; Vasconcelos et al., 2019; Peifer et al., 2021); or (2) post-orogenic topographic rejuvenation (e.g., Hack, 1982; Pazzaglia; Brandon, 1996; Quigley et al., 2007; Gallen; Wegmann; Bohnenstiehl, 2013; Liu, 2014; Calegari et al., 2021; Marques et al., 2021; Fadul; Oliveira; Val, 2022; Fonte-Boa et al., 2022; Fonseca et al., 2023; Crelier et al., 2025; Fonseca; De Grave, 2025). These competing hypotheses make different predictions about how lithospheric dynamics, lithological variability, climate and denudation interact to shape topography over million-year timescales.

Long-term subaerial survival of relief generated by an ancient orogeny, meaning continuous exposure without burial, requires a combination of factors that suppress denudation over tens to hundreds of millions of years under conditions of relative tectonic stability. Proposed mechanisms include the exposure of highly resistant lithologies that are effectively ‘inert’ to physical and chemical breakdown (e.g., Twidale, 1976; Scharf et al., 2013; Peifer et al., 2021); limited delivery of coarse detrital sediment to channels, reducing the supply of abrasive tools and thereby slowing fluvial erosion (e.g., Egholm; Knudsen; Sandiford, 2013); transport-limited conditions in which sediment supply greatly exceeds transport capacity, maintaining alluvial cover over bedrock channels and inhibiting incision (e.g., Baldwin; Whipple; Tucker, 2003); and Crickmay’s (1976) concept of ‘unequal activity’, in which differential weathering between elevated, well-drained surfaces and adjacent wetter valleys causes relief to increase rather than decline through time (e.g., Twidale, 1976). Isostatic compensation of denudation also necessarily promotes relief persistence (Gilchrist; Summerfield, 1990; Bishop; Brown, 1992; Bishop, 2007), although it can itself drive topographic rejuvenation (e.g., Bishop; Goldrick, 2010; Silva; Sacek, 2019). Section 5 illustrates how this isostatic response can prolong post-orogenic relief.

The long-term relief survival hypothesis predicts a slow, progressive decline in both topography and denudation rates after orogenesis ends (Baldwin; Whipple; Tucker, 2003). Under this view, the morphology of the erosional remnant reflects a long history of passive geological controls, expressed as a close adjustment between inherited features, including exposed lithologies and geological structures, and topographic form, rather than simply the processes operating today (Hack, 1975; Bishop, 2007). Denudation rates should therefore vary spatially, remaining low where erosion is inhibited (Figure 5), so that high, steep terrain underlain by resistant, relatively unfaulted or unfractured rocks can remain effectively ‘out of reach’ of erosional processes (e.g., Crickmay, 1976; Twidale, 1976). This prediction is empirically testable. Significant contrasts in ^{10}Be -derived denudation rates between uplands and valleys have, for example, been documented in several settings (e.g., Portenga; Bierman, 2011; Bierman et al., 2014). Geochronological evidence from supergene materials and preserved surfaces likewise shows that parts of post-orogenic landscapes in India, South Africa, Australia and Brazil were exposed and undergoing chemical weathering tens of millions of years ago, supporting their geomorphic antiquity (e.g., Beauvais et al., 2008; Monteiro et al., 2014; Bonnet et al., 2016; Beauvais et al., 2016; Vasconcelos et al., 2019; Jean et

al., 2020). These constraints come mainly from (U-Th)/He and $^{40}\text{Ar}/^{39}\text{Ar}$ dating of iron oxides and hydroxides such as haematite and goethite, together with K-Mn oxides in weathering profiles, and generally record phases of intense chemical weathering that began tens of millions of years ago (e.g., Beauvais et al., 2008; Monteiro et al., 2014; Bonnet et al., 2016; Beauvais et al., 2016; Vasconcelos et al., 2019; Jean et al., 2020). The long-term relief survival hypothesis further predicts smooth, concave-up river longitudinal profiles in which elevation decreases progressively downstream and knickpoints are absent (e.g., Hack, 1975; Young, 1989; Baldwin; Whipple; Tucker, 2003). These are also characteristic attributes of steady-state rivers in tectonically active settings.

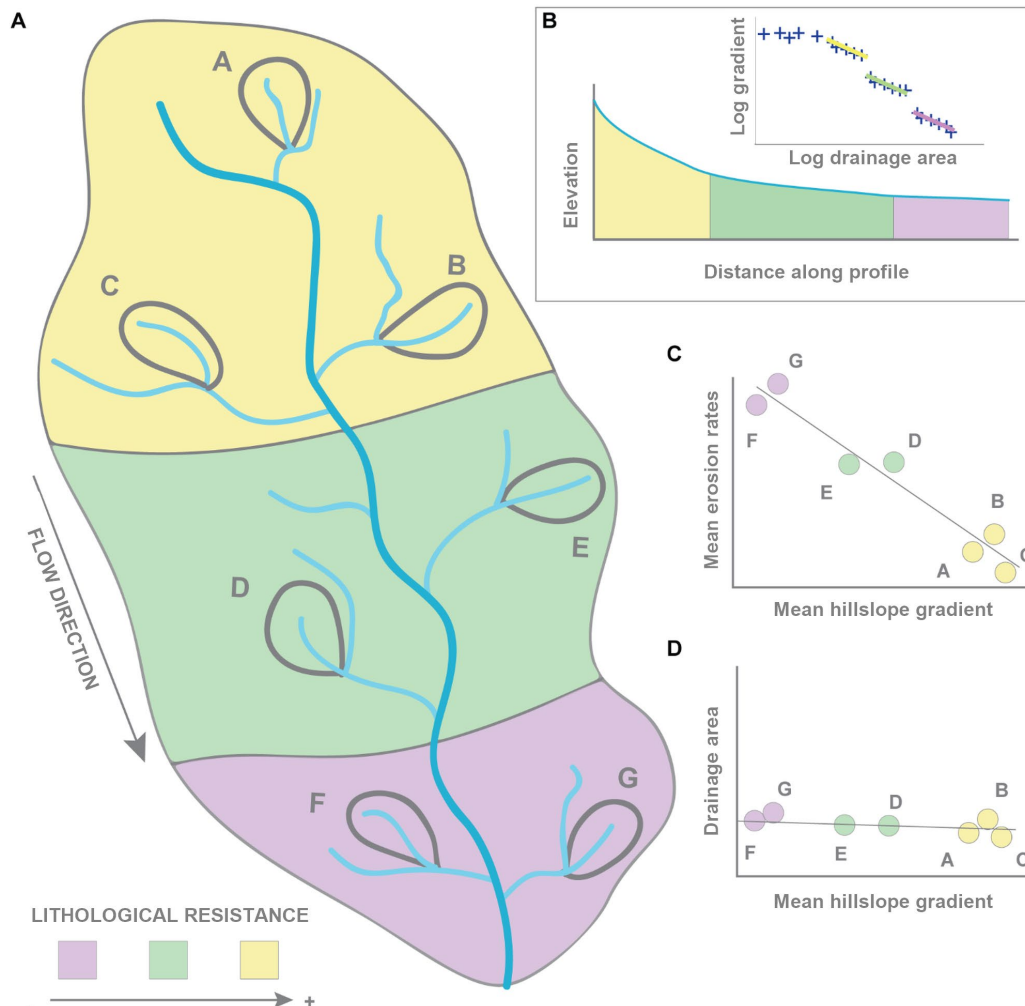


Figure 5. Schematic landscape consistent with the long-term relief survival hypothesis. (A) Plan view of a hypothetical drainage basin underlain by lithologies of contrasting resistance. (B) The trunk river has a concave-up longitudinal profile in which local channel slope increases with bedrock resistance, so that steeper reaches occur over more resistant rocks. The inset slope-area plot in panel B shows the same relationship between channel morphology and rock type. (C) Catchment-averaged erosion rates are inversely related to lithological resistance, while catchments underlain by more resistant rocks have steeper mean hillslopes than those underlain by more erodible rocks. (D) Sub-catchments A–G have similar contributing areas. Uplands underlain by resistant rocks are therefore higher because they have undergone less denudation, and at lower rates, than the surrounding terrain underlain by less-resistant rocks.

The opposing view begins from the marked discrepancy between predicted and observed topographic-decay timescales in post-orogenic mountains (Bishop, 2007; Peifer, 2018). This mismatch has often been interpreted as evidence that post-orogenic topography cannot be as old as previously proposed (cf. Bishop, 2007), and the hundreds-of-millions-of-years antiquity inferred for some intraplate landscapes remains only partly constrained by geochronology (Bishop, 2007). Moreover, denudation rate measurements show that elevated areas continue to erode, albeit slowly, rather than remaining literally ‘out of reach’ of erosion (e.g., Belton et al., 2004). A denudation

rate of 10 m/Ma, entirely plausible for an ancient orogen, would remove 5 km of rock over 500 million years (Bishop, 2007). These observations have led many authors to argue that post-orogenic relief is comparatively young and was generated by rock uplift long after the ancient orogeny itself had ended (e.g., Hack, 1982; Pazzaglia; Brandon, 1996; Quigley et al., 2007; Gallen; Wegmann; Bohnenstiehl, 2013; Liu, 2014; Calegari et al., 2021).

Proposed mechanisms for post-orogenic topographic rejuvenation include far-field stress, whereby tectonic stresses generated at distant plate boundaries are transmitted into plate interiors and drive rock uplift (e.g., Hack, 1982; Quigley et al., 2007; Cogné; Gallagher; Cobbold, 2011; Calegari et al., 2021; Campos et al., 2023); rock uplift driven by mantle dynamics (e.g., Gallen; Wegmann; Bohnenstiehl, 2013; Liu, 2014; Sacek, 2017; Silva; Sacek, 2019); flexural-isostatic rock uplift (e.g., Gilchrist; Summerfield, 1990; Pazzaglia; Gardner, 1994; Bishop; Goldrick, 2010; Ruetenik; Jansen; Sandiford, 2024); permanent addition of material to the base of the lithosphere through underplating (e.g., Young, 1989; Matmon; Bierman; Enzel, 2002); and exposure of rocks with contrasting erodibility (e.g., Gallen, 2018; Fadul; Oliveira; Val, 2022). Because erosion is strongly coupled to topography (Ahnert, 1970; Montgomery; Brandon, 2002; Portenga; Bierman, 2011; Harel; Mudd; Attal, 2016), the rejuvenation hypothesis predicts a geomorphic differentiation in which rejuvenated areas are steeper and erode faster than areas that were not rejuvenated. A substantial thermochronological, sedimentological and geomorphic record indicates that the post-orogenic histories of at least some ancient orogens were not uniformly slow through time and space, but included episodes of rapid upward advection of rock through the shallow crust towards the surface (e.g., Gallagher; Hawkesworth; Mantovani, 1994; Pazzaglia; Brandon, 1996; Gunnell; Fleitout, 1998; Quigley et al., 2007; Japsen et al., 2012; Tucker; van der Beek, 2013; Liu, 2014; Gallen, 2018; Calegari et al., 2021; Marques et al., 2021; Fadul; Oliveira; Val, 2022; Fonte-Boa et al., 2022). Yet the geometry, magnitude and duration of the processes responsible for rejuvenation, and their interactions with factors such as lithology and effective elastic thickness, remain poorly constrained. Independent evidence that a particular mechanism actually operated is also often lacking (Gallen, 2018). Importantly, rejuvenation alone does not explain observations central to the relief survival hypothesis, such as the close adjustment between topographic form and the nature of the exposed bedrock (Hack, 1960).

A third hypothesis places post-orogenic mountains in a state of 'topographic equilibrium', in which surface morphology remains approximately invariant through time because denudation rates are spatially uniform despite variations in lithology or topographic form (Hack, 1960; Howard, 1965). This requires rock uplift to balance erosion everywhere in the landscape, with steeper topography where more resistant rocks are exposed (Hack, 1975). The concept is problematic in post-orogenic settings because continued rock uplift is required to maintain such equilibrium (Hack, 1960; Howard, 1965; Bishop, 2007). Nevertheless, several ancient orogens, including the Appalachian Mountains (e.g., Matmon et al., 2003), the Cape Fold Belt (e.g., Scharf et al., 2013) and the Western Ghats (e.g., Mandal et al., 2015), have been interpreted as being in topographic equilibrium, although in these cases rock uplift may arise largely from the flexural-isostatic response to denudation. These uncertainties underscore how much of the post-orogenic history of ancient landscapes remains unresolved (Bishop, 2007; Peifer et al., 2021).

5. An Example of the Role of Isostasy in Topographic Decay

Because crust and mantle have different densities, changes in crustal loading drive isostatic adjustment through vertical rock motions (Watts, 2001). Added loads, such as sediment deposition or ice loading, cause subsidence, whereas unloading through glacier melting or denudation drives isostatic rock uplift (Gilchrist; Summerfield, 1990; Bishop; Brown, 1992; Bishop, 2007). The response to loading and unloading depends on lithospheric flexural rigidity and may be distributed over an area far larger than the footprint of the load itself (Watts, 2001). Because flexural rigidity is not an especially intuitive quantity, it is commonly expressed in terms of effective elastic thickness (T_e), the thickness that a purely elastic plate would require to reproduce the flexural behaviour of the real lithosphere (Watts, 2001; Turcotte; Schubert, 1982). For a load of a given spatial scale, lower T_e implies weaker flexural support and a greater degree of isostatic compensation, whereas higher T_e implies stronger flexural support and a response distributed over greater distances (Watts, 2001). In the idealised post-orogenic setting considered here, we assume that long-term tectonic uplift is negligible relative to denudation and its isostatic response. Continued denudational mass loss therefore imposes a negative load that is partly compensated by the flexural-isostatic response of the lithosphere. Isostasy is consequently an unavoidable component of the geomorphic evolution of ancient orogens and acts to prolong post-orogenic relief persistence (Gilchrist; Summerfield, 1990; Bishop; Brown, 1992; Bishop, 2007).

To illustrate how isostasy influences topographic decay, we use a simplified empirical example from the Quadrilátero Ferrífero (QF), Minas Gerais. The analysis draws on catchment-averaged denudation rates derived from in situ ^{10}Be concentrations in fluvial sediment reported by Peifer et al. (2021). These concentrations integrate catchment denudation over the nuclide integration timescale; they are not direct measurements of surface lowering over million-year intervals. Denudation in the QF is generally slow, ranging from 0.6 ± 0.1 to 22.2 ± 1.9 m/Ma, but is strongly spatially variable. Extremely low rates (<3.5 m/Ma) occur in steep quartzite terrain, whereas rates reach 22.2 ± 1.9 m/Ma over more friable rocks in areas of subdued relief (Peifer et al., 2021). We use the topographic metric that best explains the estimated denudation rates, local relief ($R^2 = 0.73$; $p < 0.01$; Peifer et al., 2021), defined here as the elevation range within a circular window 2 km in diameter, to construct a spatially continuous modelled surface of denudation rates. The spatial mean of this surface is 10.49 m/Ma. This value is therefore the spatial average of the modelled denudation-rate surface for the QF, not the arithmetic mean of the sampled catchment rates reported by Peifer et al. (2021).

We first estimate the local-isostatic end-member associated with denudational mass loss using an Airy-isostasy model, in which topographic variations are balanced by changes in crustal thickness according to (Liu, 2014):

$$h_{\text{Airy}} = \frac{\rho_c}{\rho_m} \cdot H_{\text{den.ep}}, \quad (8)$$

where h is isostatic uplift, expressed as vertical rock advection; ρ_c is crustal density (2700 kg/m^3); ρ_m is asthenospheric density (3300 kg/m^3); and $H_{\text{den.ep}}$ is the spatially weighted mean denudation rate of the QF (10.49 m/Ma) multiplied by the time interval used in the calculation. For these densities, the local Airy end-member compensates approximately 82% of the thickness removed by denudation.

Lithospheric rigidity, however, means that the response to unloading need not reach the local Airy-compensation limit. In this illustrative exercise, we represent this first-order effect using the degree of compensation C , following the periodic-load solution of Turcotte and Schubert (1982). In that formulation, C is the ratio between the amplitude of actual lithospheric deflection and the maximum hydrostatic deflection expected for the same load. C therefore ranges from 0 to 1: $C = 0$ is the limit in which the load is supported entirely by lithospheric elastic rigidity, with negligible deflection, whereas $C = 1$ represents complete hydrostatic compensation. We apply this relationship, as a first-order approximation, to the negative load produced by denudational mass removal in the QF.

Approximating the lithosphere as a thin elastic plate of constant thickness overlying a fluid substrate (e.g., Turcotte; Schubert, 1982; Gilchrist; Summerfield, 1990, 1991; Bishop; Brown, 1992; Baldwin; Whipple; Tucker, 2003), we calculate C for a periodic load using (Turcotte; Schubert, 1982):

$$C = \frac{\rho_m - \rho_c}{\rho_m - \rho_c + \frac{D}{g} \cdot \left(\frac{2\pi}{\lambda}\right)^4}, \quad (9)$$

where D is the flexural rigidity of the elastic plate, g is gravitational acceleration, taken here as 9.82 m/s^2 , and λ is the wavelength of the applied negative load, representing the characteristic spatial scale of mass removal in the model. Flexural rigidity D is given by (Turcotte; Schubert, 1982):

$$D = \frac{Y \cdot T_e^3}{12 \cdot (1 - \nu^2)}, \quad (10)$$

where Y is Young's modulus, a measure of material stiffness, taken here as 70 GPa (e.g., Turcotte; Schubert, 1982; Pelletier, 2004); ν is Poisson's ratio, a dimensionless measure of deformation perpendicular to the applied load, taken here as 0.25 (e.g., Turcotte; Schubert, 1982; Bishop; Brown, 1992; Pelletier, 2004; Watts, 2007; Wickert, 2016); and T_e is effective elastic thickness, estimated at approximately $30\text{--}40 \text{ km}$ for the QF (Bizzi et al., 2003; Tassara et al., 2007).

In Eq. (9), λ is the wavelength of the applied periodic load, whereas T_e influences flexural rigidity D through Eq. (10). Following Turcotte and Schubert (1982), λ and T_e are therefore treated as independent parameters. For the illustrative reference scenario, we adopt $\lambda = 475 \text{ km}$ and $C = 0.38$. With $Y = 70 \text{ GPa}$, $\nu = 0.25$, $\rho_c = 2700 \text{ kg/m}^3$, $\rho_m = 3300 \text{ kg/m}^3$ and $g = 9.82 \text{ m/s}^2$, Eqs. (9) and (10) together correspond to $T_e \approx 37 \text{ km}$, within the approximately $30\text{--}40 \text{ km}$ range estimated for the QF. The values of λ , C and T_e are used only to define an illustrative reference scenario, not as unique or tightly constrained estimates of QF flexural behaviour. Figure 7A shows the sensitivity

of C to independent variations in λ and T_e . In the exercise, the effective vertical response to denudation is approximated by applying this degree of compensation to the local Airy response in Eq. (8).

Because lateral lithological variations control the spatial pattern and rates of denudation in the QF (Peifer et al., 2021), we simplify the exposed geology into three lithological-erodibility groups. Quartzites and itabirites form the 'resistant-rock' group; schists, phyllites, dolomitic units and predominantly mafic metavolcanic rocks form the 'intermediate-resistance' group; and crystalline rocks, chiefly gneisses and granitoids including those exposed in gneiss domes, form the 'low-resistance' group. Using the data of Peifer et al. (2021), we then calculate a mean denudation rate for each lithological-erodibility group (Figure 6).

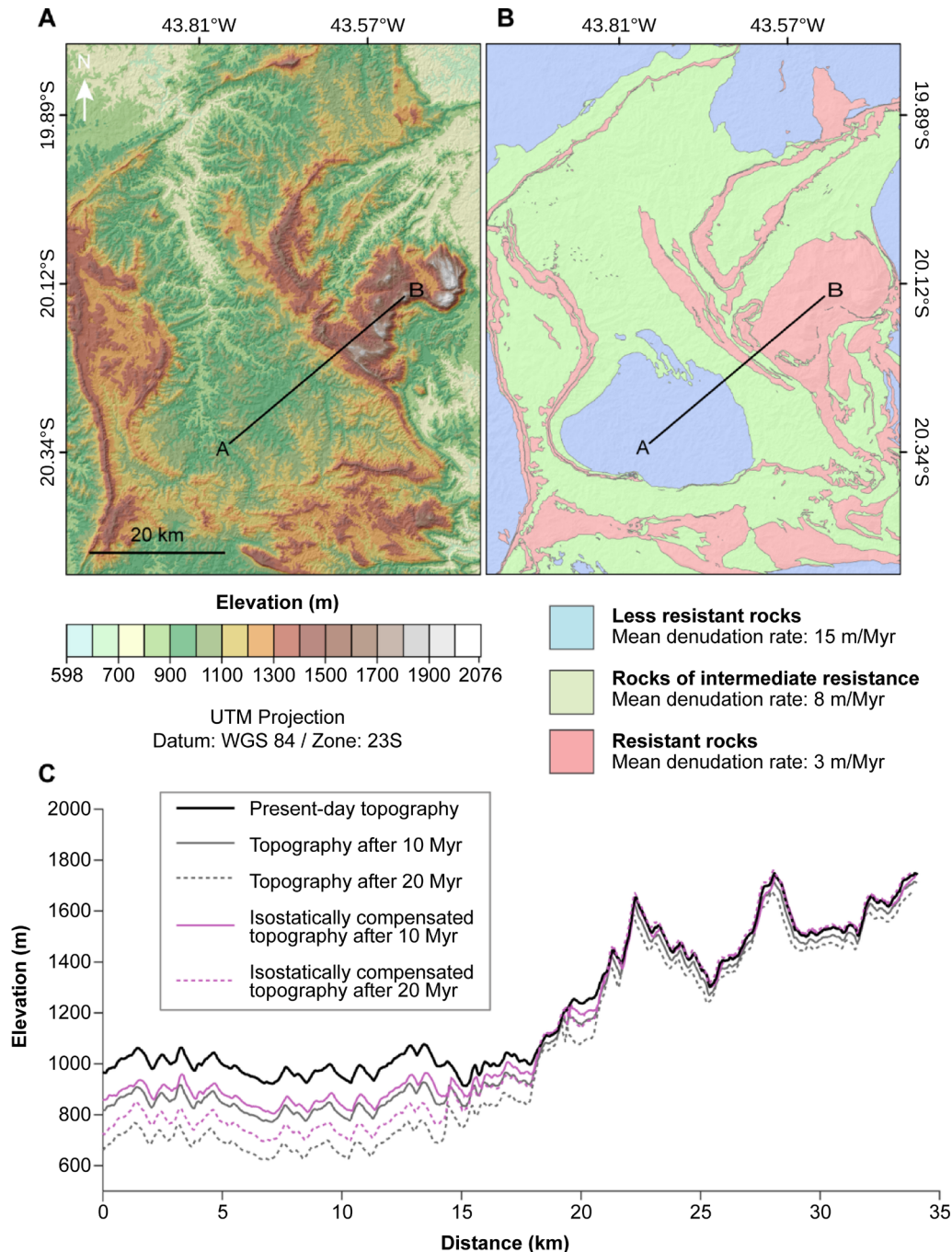


Figure 6. Extrapolated effect of estimated denudation rates on the topographic relief of the Quadrilátero Ferrífero. (A) Topographic map coloured by elevation above sea level. (B) Simplified map of lithological resistance, grouped into resistant, intermediate-resistance and low-resistance rocks, with the mean denudation rate calculated for each group. (C) Swath profiles showing mean elevation along line AB and adjacent terrain (swath width = 1 km): (i) present-day topography (thick black line); (ii) topography after applying the mean denudation rates for 10 and 20

Myr; and (iii) topographic profiles after flexural-isostatic compensation of denudational unloading over 10 and 20 Myr. For the illustrative scenario in (C), we adopt $\lambda = 475$ km and $C = 0.38$, corresponding to $T_e \approx 37$ km for the physical parameters described in the text.

To explore the potential implications of this spatial denudation pattern for relief evolution, we extrapolate the mean rates assigned to the three lithological-erodibility groups over 10 and 20 Ma (Figure 6). This does not imply that ^{10}Be -derived rates directly measure surface lowering over such timescales. Rather, the exercise deliberately assumes, solely for illustration, that the present-day spatial pattern of denudation persists for millions of years. We calculate the flexural-isostatic response to spatially weighted denudation across the QF using $C = 0.38$ for the reference scenario. The aim is to illustrate the potential influence of isostasy on post-orogenic topographic decay, not to provide a quantitative reconstruction or forecast of the future evolution of the QF. Cosmogenic-nuclide-derived rates integrate much shorter intervals, typically of the order of 10^3 – 10^5 years, and cannot be directly extrapolated over tens to hundreds of millions of years. We also neglect changes in geology with depth, which are likely to become important over the timescales considered. Finally, our simplified representation of surface lowering does not explicitly resolve the dynamics of fluvial, hillslope and weathering processes through which erosional landscapes lose mass over time.

As shown by Peifer et al. (2021), rocks with contrasting erodibility exposed across the QF produce strong spatial variations in denudation rates. Extrapolating this spatially unequal pattern through time causes relief to grow rather than decay because of the spatial distribution of lithological erodibility and the resulting denudation-rate gradients (Figures 6 and 7B). The flexural-isostatic response to denudation, however, operates over a much broader spatial scale than local lithological variations (Gilchrist; Summerfield, 1990; Watts; McKerrow; Fielding, 2000; Watts, 2001). Elevated areas underlain by resistant rocks and eroding slowly therefore experience broadly similar isostatic uplift to surrounding areas underlain by more erodible rocks and denuding more rapidly. The net result is a slow decline in regional mean elevation through time, even as the mean elevation of resistant-rock areas increases slightly (Figure 7B). This behaviour is consistent with earlier arguments that flexural-isostatic compensation of spatially unequal denudation can increase relief in post-orogenic landscapes (e.g., Stephenson, 1984; Lambeck; Stephenson, 1986; Bishop; Brown, 1992; Bishop; Goldrick, 2010), a mechanism more widely associated with the formulation of Molnar and England (1990).

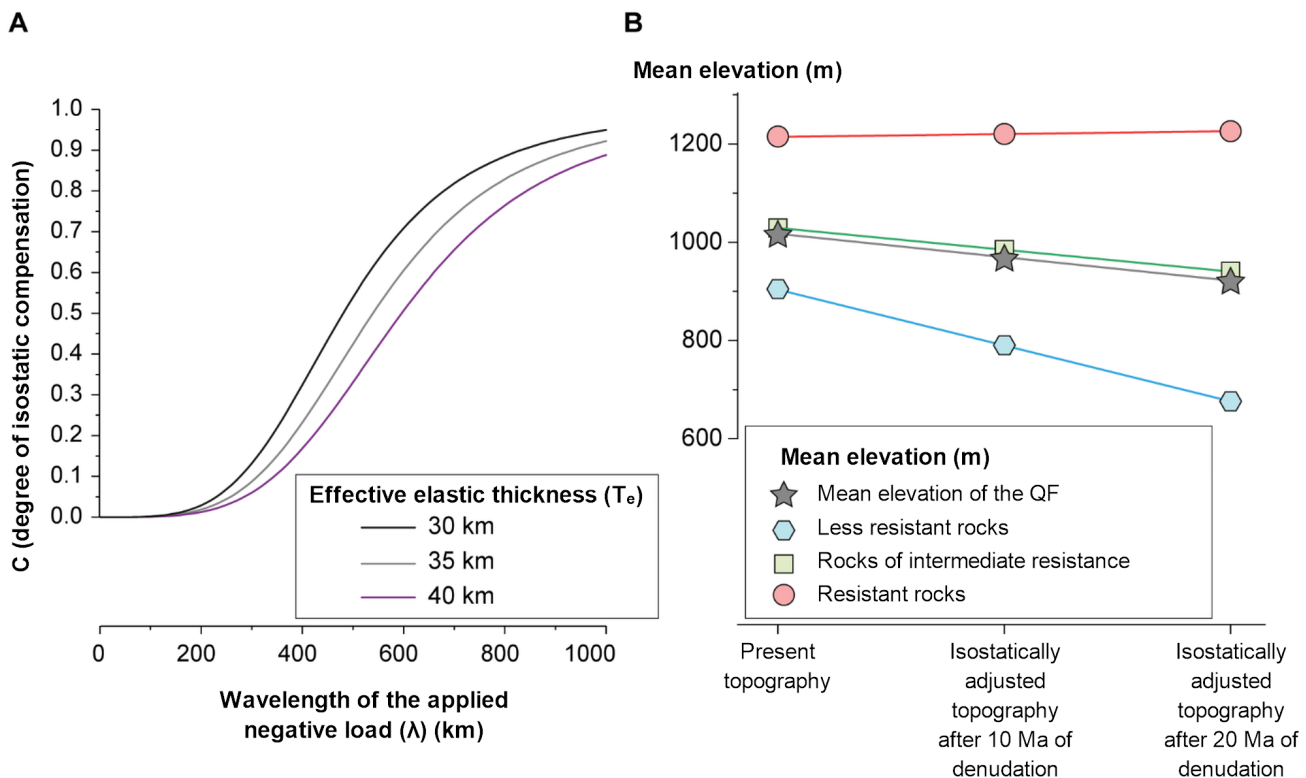


Figure 7. (A) Degree of isostatic compensation (C) for different values of effective elastic thickness (T_e) and wavelength of the applied negative load (λ); and (B) change in mean elevation through time after accounting for the flexural-isostatic response to denudational unloading in the QF. For the illustrative scenario in (B), we adopt $\lambda = 475$ km and $C = 0.38$, corresponding to $T_e \approx 37$ km for the physical parameters described in the text. Panel (A) shows the sensitivity of C to independent variations in λ and T_e . Colours in panel (B) denote areas underlain by rocks of contrasting resistance, mapped in Figure 6B; grey stars show the mean elevation of the QF.

Topography cannot, however, increase indefinitely because regional mean elevation continues to decline (Figure 7B). Retaining all simplifying assumptions of the exercise, extrapolating the denudation pattern and its flexural-isostatic compensation yields a timescale of the order of 200 Ma for regional mean elevation to reach sea level. In areas underlain by resistant rocks, mean elevation increases slightly over the interval shown in Figure 7B, so the exercise does not yield a meaningful decay timescale for that domain. The approximately 200 Ma estimate for regional mean elevation is therefore not a chronological prediction for the QF, but an order-of-magnitude illustration of how differential denudation coupled with flexural-isostatic response can greatly prolong relief persistence. Even in the absence of active tectonic uplift, a landscape with strong lithological contrasts may therefore retain pronounced relief over very long intervals. This does not imply that topographic rejuvenation has not occurred or cannot occur; rather, it shows that processes intrinsic to an erosional post-orogenic landscape can themselves favour long-term relief survival.

The exercise is not intended to resolve all of the complexities of isostasy and lithospheric flexure in post-orogenic mountains. In the case considered here, the scale of the topography is far smaller than the wavelength of the applied negative load ($\lambda = 475$ km), so peripheral lithospheric flexure is neglected. Such flexural deflection is particularly important in other settings, notably elevated passive margins, where denudational unloading of the uplands is coupled with depositional loading in adjacent sedimentary basins (e.g., Weissel; Karner, 1989; Gilchrist; Summerfield, 1990, 1991; Tucker; Slingerland, 1994; Sacek; Braun; van der Beek, 2012; Silva; Sacek, 2019). For more complete treatments of isostasy and lithospheric flexure, see Turcotte and Schubert (1982) and Watts (2001, 2007).

6. Synthesis and Conclusions

The term ‘post-orogenic landscapes’ describes landscapes that have remained comparatively tectonically stable for tens to hundreds of millions of years yet still preserve mountainous relief despite prolonged denudation (Bishop, 2007). As this review has shown, a wide range of methods and datasets has been used to reconstruct their post-orogenic histories, but no consensus exists on how such landscapes evolve through time. The Appalachian Mountains, arguably the world’s most intensively studied post-orogenic setting, illustrate the problem particularly well. Hack (1960) and Matmon et al. (2003) support interpretations consistent with long-term relief survival, whereas Hack (1982), Pazzaglia and Gardner (1994), Gallen, Wegmann and Bohnenstiehl (2013), Gallen (2018) and Stokes et al. (2023) emphasise later topographic rejuvenation. This disagreement persists despite extensive application of thermochronology (e.g., McKeon et al., 2014), cosmogenic-nuclide analysis (e.g., Matmon et al., 2003; Portenga et al., 2013; Stokes et al., 2023), quantitative analysis of river longitudinal profiles and many other modern approaches (e.g., Liu, 2014; Gallen, 2018; Stokes et al., 2023). Broadly, post-orogenic landscape evolution has therefore been framed through two perspectives often treated as mutually exclusive: either relief was generated by ancient orogenesis and survived because erosion was strongly inhibited, or present-day relief is comparatively young and reflects geodynamic or climatic processes that rejuvenated the topography.

This is a false dichotomy. Relief inherited from an ancient orogen can subsequently be rejuvenated by rock uplift generated by different mechanisms acting alone or in combination. Mountain-building processes operating hundreds of millions of years ago, commonly within polyphase tectonic histories, simultaneously created mountainous topography and a complex structural and lithological framework in which rocks of contrasting composition, age and resistance are juxtaposed. The resulting ‘proto-topography’ stood above surrounding lowlands and imposed a fundamentally erosional evolutionary regime characterised by a net-loss mass balance. Denudation has governed the long-term evolution of these landscapes ever since, and progressive denudational mass loss necessarily drives a flexural-isostatic response. Isostasy is therefore a fundamental component of their geomorphic history and acts to prolong relief persistence (Gilchrist; Summerfield, 1990; Bishop; Brown, 1992; Bishop, 2007; Ruetenik; Jansen; Sandiford, 2024). The spatial distribution of effective elastic thickness is consequently a key control on this flexural-isostatic evolution. Lithological heterogeneity is equally important:

extremely resistant rocks such as quartzites commonly occur alongside low-resistance rocks such as granites in warm, humid climates, creating strong erodibility gradients that exert a first-order influence on landscape-scale denudation patterns (e.g., Peifer et al., 2021).

Together with other characteristics of intraplate settings, including low denudation rates, low seismicity, declining landslide frequency after orogenesis and limited delivery of coarse sediment to channels, these factors favour the long-term persistence of inherited proto-topography. This 'persistence tendency' does not require topography to remain static. Rather, geomorphic change may proceed extremely slowly, particularly where resistant rocks are exposed, so that complete lowering of relief to sea level requires hundreds of millions of years of continued denudation. As these landscapes are progressively denuded, deeper metamorphic levels may also be exhumed, locally exposing rocks that are denser, harder and more resistant (Braun et al., 2014), further slowing surface lowering (Bishop, 2007). Post-orogenic evolution can therefore include relief growth, rather than simple monotonic decay towards widespread planation, as a natural consequence of strong spatial contrasts in lithological erodibility. Importantly, published topographic-decay estimates generally do not incorporate this spatial variability in erodibility explicitly, even though it may exert a first-order control on the distribution of denudation rates (e.g., Peifer et al., 2021).

A possible expression of this 'persistence tendency' is the spatial coincidence between many Brazilian post-orogenic landscapes and the Brasiliano orogenic belts surrounding cratonic domains (e.g., Alkmim, 2015). These belts probably formed an elevated proto-topography at the end of Brasiliano compression, whereas the neighbouring cratons formed comparatively lower terrain. Present-day Brazilian topography preserves, at first order, the same broad pattern of high and low regions (Figure 2) (e.g., Alkmim, 2015). We therefore consider it more plausible that this large-scale morphological continuity reflects persistence of inherited proto-topography than a complex sequence of geographical cycles interrupted by epeirogenic uplift and represented by planation surfaces at discrete elevations (e.g., King, 1956). It is difficult to reconcile the modern pattern with repeated complete destruction of relief during the post-orogenic period followed by recent reconstruction of a large-scale topographic configuration closely resembling that expected at the end of Brasiliano compression. Such repeated destruction and reconstruction lacks clear physical or empirical support.

At the same time, a substantial body of empirical evidence shows that tectonic, climatic, mantle and lithological processes have produced topographic rejuvenation and spatial and temporal variability in erosion rates within several ancient orogens during their post-orogenic evolution (e.g., Pazzaglia; Brandon, 1996; Quigley et al., 2007; Gallen; Wegmann; Bohnenstiehl, 2013; Tucker; van der Beek, 2013; Liu, 2014; Gallen, 2018; Calegari et al., 2021; Marques et al., 2021; Fadul; Oliveira; Val, 2022; Fonte-Boa et al., 2022; Fonseca et al., 2023; Campos et al., 2023; Crelier et al., 2025; Fonseca; De Grave, 2025). Although the evidence for individual mechanisms is case-specific, such sources of transient forcing should be expected to affect post-orogenic mountains more generally, with different amplitudes and over different spatial and temporal scales. Far-field tectonic stresses, for example, can be transmitted into plate interiors and concentrated along pre-existing zones of crustal weakness such as shear zones and major lineaments. Depending on their orientation and geometry relative to the regional stress field, progressive stress accumulation along these discontinuities can generate seismicity and rock uplift (e.g., Zoback, 1992; Talwani, 2017). Cenozoic climate has likewise oscillated repeatedly between wetter and drier states over a range of timescales (e.g., Zachos et al., 2001). Over the geological timescales relevant to post-orogenic evolution, lithology also changes with depth as exhumation proceeds, so that rocks with contrasting erodibility can generate complex transient responses in drainage networks and adjacent hillslopes (e.g., Forte; Yanites; Whipple, 2016; Perne et al., 2017; Gallen, 2018; Fadul; Oliveira; Val, 2022). Other relevant mechanisms include: *i*) flexural-isostatic compensation of denudation, which can trigger regional transient adjustment (e.g., Bishop; Goldrick, 2010) and reactivate pre-existing structures (e.g., Silva; Sacek, 2019); *ii*) dynamic topography, which generally modifies landscape elevation at long wavelengths and modest amplitudes, although changes of the order of 1 km can occur in some settings (Braun, 2010); and *iii*) sea-level change, which can drive regional landscape transience (e.g., Bishop et al., 2005).

All of these processes act on an inherited 'proto-topography' characterised by strong lateral and vertical geological contrasts, which influence transient landscape-response times, and by a particular configuration of effective elastic thickness. The same combination of mechanisms should not be expected to operate everywhere, or with the same intensity. In Brazil, preservation of elevated relief through ancient orogenic inheritance appears more plausible in regions less affected by Cenozoic post-rift deformation, such as the Espinhaço Range, than along

parts of the Atlantic margin, where present-day topography more directly reflects the superposition of inherited orogenic relief and later tectonic rejuvenation. We therefore argue that post-orogenic mountain evolution is best viewed as a continuum between the two end-member hypotheses, with the balance varying among landscapes according to their geodynamic and climatic histories. Even so, the post-orogenic evolution of ancient orogens remains incompletely understood, and further work is needed to resolve the histories of many intraplate landscapes. Brazil offers particularly promising natural laboratories for this effort. We hope that this review stimulates further investigation of post-orogenic relief evolution, especially in Brazilian settings.

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