



Temporal and spatial evolution of dynamic support from river profiles: A framework for Madagascar

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[1] We present a strategy for calculating uplift rates as a function of space and time from large sets of longitudinal river profiles. This strategy assumes that the shape of a river profile is controlled by the history of uplift rate and moderated by the erosional process. We assume that upstream drainage area is invariant. The algorithm was tested on a set of ~ 100 river profiles which were extracted from a digital elevation model of Madagascar. This set of profiles was simultaneously inverted to obtain uplift rate as a smooth function of space and time. The fit between observed and calculated profiles is excellent and suggests that Madagascar was uplifted by 1–2 km at rates of 0.2–0.4 mm/yr during the last ~ 15 Myrs. The location of Madagascar suggests that its topographic elevation is maintained by convective circulation of the sub-lithospheric mantle. Residual depth anomalies of oceanic fragments encompassing the island show that the island straddles a dynamic topographic gradient which generates asymmetric Neogene uplift. Volcanism, warped peneplains and uplifted marine terraces corroborate the existence of youthful uplift. We suggest that sets of longitudinal river profiles contain useful information about the history of regional uplift which can be extracted by inverse modeling and calibrated by independent geologic observations.

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1. Introduction

[2] The physiography of Madagascar suggests that it is part of the basin and swell morphology of Africa, which is dynamically supported by sub-crustal processes (Figure 1) [Holmes, 1944; Burke, 1996; Burke and Gunnell, 2008]. It follows that the history of uplift of Madagascar may contain useful clues about spatial and temporal scales of sub-plate

convective circulation. Three sets of observations suggest that the topography of the island is supported by sub-crustal processes. First, dynamic support of the oceanic crust surrounding Madagascar can be estimated by comparing water-loaded bathymetry to a predicted ocean age-depth relationship [Crosby *et al.*, 2006; Winterbourne *et al.*, 2009]. The residual (i.e., dynamic) bathymetry of the oceanic crust surrounding Madagascar has a gradient of ~ 0.01 , from west to

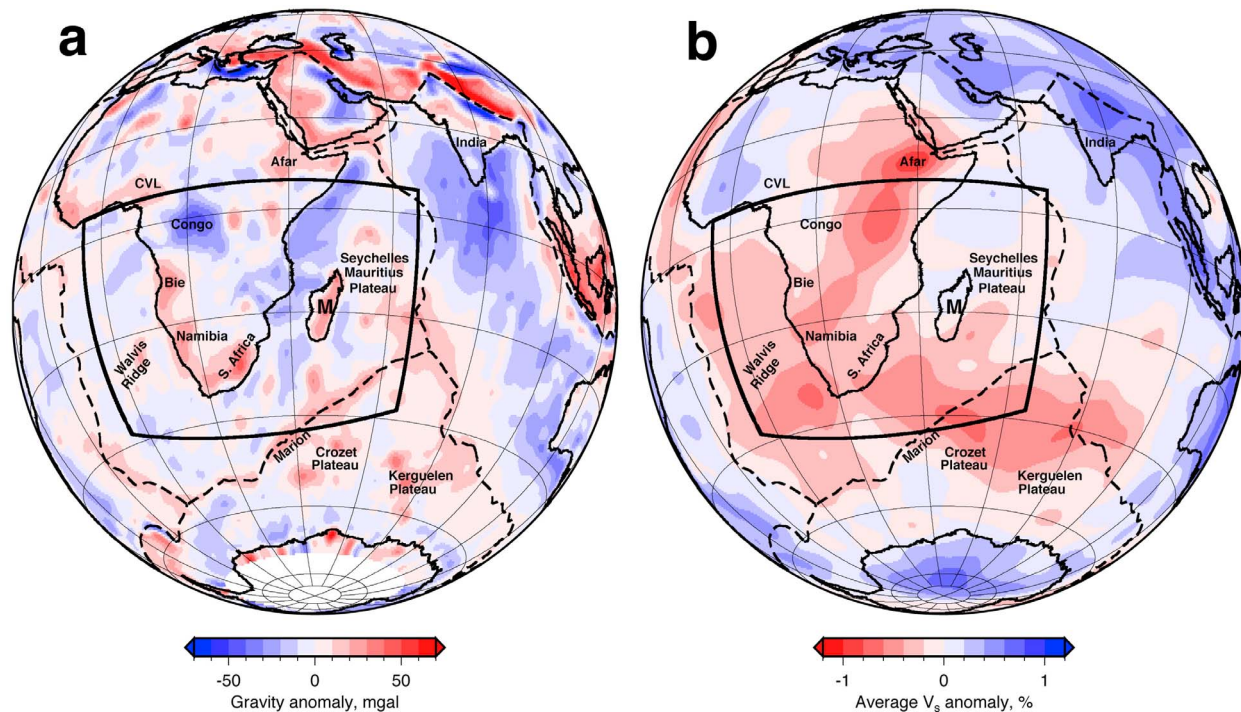


Figure 1. (a) Long wavelength (>800 km) GRACE free-air gravity anomaly map centered on Madagascar [Tapley *et al.*, 2005]. Red and blue = positive and negative anomalies respectively, contour interval = 10 mgal. A 30 mgal positive free-air gravity anomaly is centered on Madagascar (M), which suggests that the island is dynamically supported. Positive long wavelength free-air gravity anomalies are associated with the Cameroon Volcanic Line (CVL), Bié, Namibia, South Africa and Afar topographic swells, and with the Kerguelen, Crozet and Seychelles-Mauritius Plateaux. Dashed lines = plate boundaries. Box = location of Figure 2. (b) Mean velocity anomalies from S40RTS tomographic model projected to the Earth's surface (see equation (1)). Red and blue = slow and fast average velocity anomalies, respectively. Madagascar is on the northeastern edge of a large, seismically slow, region [Ritsema *et al.*, 2011]. Velocity anomalies beneath this region are probably caused by hotter than average mantle.

east (Figure 2). This gradient extends from the Mozambique Channel ~100 km west of Madagascar, which has up to 1 km of negative residual bathymetry, to the Mascarene Basin ~200 km east, which has 1–2 km of positive residual bathymetry. We interpret the asymmetry of Madagascar's topography as the onshore expression of this gradient (Figure 3a). Secondly, Madagascar has a long wavelength (~1000 km) free-air gravity anomaly with an amplitude of +30 mgal, which could be generated by density variations beneath the crust, and allows us to gauge the amplitude and wavelength of convectively supported topography (Figures 1a and 3b). This gravity anomaly is similar in amplitude and wavelength to anomalies associated with topographic domes in West Africa, which have 0.5–1 km of dynamic support [Al-Hajri *et al.*, 2009]. Thirdly, there is excellent evidence for a large, seismically slow region beneath southern Africa and its surroundings [e.g., Panning and Romanowicz, 2006; Simmons

et al., 2009]. The average shear-velocity variation of the mantle from Ritsema *et al.*'s [2011] S40RTS model, $\overline{V}_s$, is

$$\overline{V}_s = \frac{1}{S} \sum_{i=1}^S V_{s_i}, \quad (1)$$

where V_{s_i} is the average velocity perturbation from the Preliminary Reference Earth Model PREM [Dziewonski and Anderson, 1981] for each layer, i , of thickness 50 km. S is the number of layers. Madagascar is on the edge of a large, slow velocity anomaly (Figure 1b). It is generally agreed that slow seismic velocities in this region are likely to be generated by a thermochemical plume, which might extend from the core-mantle boundary to the base of the plate [Ni *et al.*, 2002]. The agreement between surface topography, long wavelength free-air gravity anomalies, seismic velocities and regional residual bathymetry suggests that Madagascar's

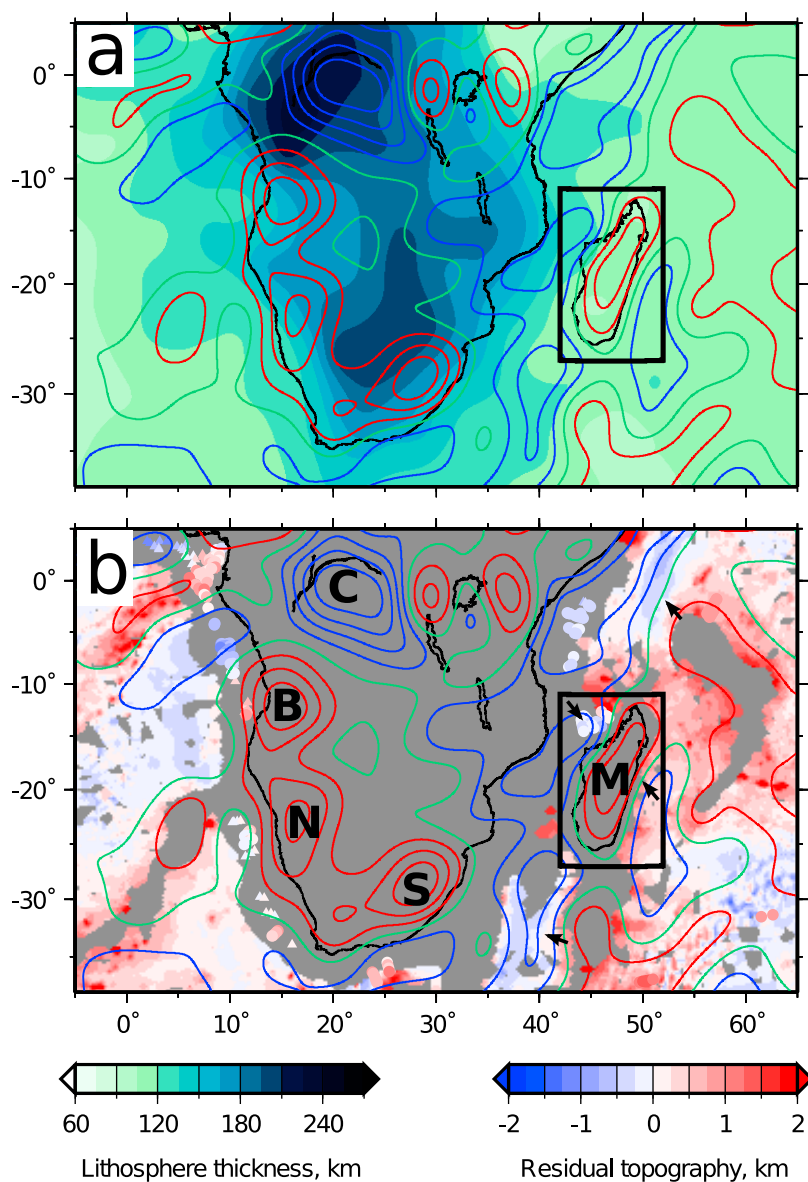


Figure 2. (a) Lithospheric thickness calculated from surface wave tomographic model [Al-Hajri *et al.*, 2009; Crosby *et al.*, 2010]. Seismic ray coverage across Madagascar is uniformly good (>100 paths per $3^\circ \times 3^\circ$ cell). Compared to southern Africa and the Congo basin, the lithosphere beneath Madagascar is not thick. Red, blue and green contours = positive, negative and zero long wavelength free-air gravity anomalies, respectively (Figure 1). Contour interval = 10 mgal. Box = location of Figure 3. (b) Residual topography in the oceans calculated relative to the age-depth curve of Crosby *et al.* [2006]. Filled circles = measurements where crustal thickness and sediment thickness are well constrained. Upward pointing and inverted triangles = minimum and maximum estimate of residual topography, respectively. Background grid = residual topography calculated using a global dataset of ship track bathymetry and the NDGC sediment thickness dataset [Divins, 2008]. Regions where sediment thickness is >1.5 km, and areas of known crustal thickening, have been excluded [see Winterbourne *et al.*, 2009]. Contours = gravity data; box = Figure 3. Note the ~ 0.01 west to east gradient in residual topography (arrows). M = Madagascar. C = Congo basin. B = Bié, N = Namibia, S = South Africa domes.

topography is convectively supported (Figures 1 and 2).

[3] Madagascar's basement geology is dominated by a suite of Late Neoproterozoic and Cambrian terranes, whose structural fabric has undoubtedly

influenced younger fault orientations (Figure 3c) [Rechenmann, 1982; Bertil and Regnault, 1998; Schreurs *et al.*, 2010]. Two discrete phases of volcanism have occurred on Madagascar. Cretaceous volcanic rocks, dated by ^{40}Ar – ^{39}Ar step-heating, crop out across the island [Storey *et al.*,



1995; Torsvik *et al.*, 1998; Melluso *et al.*, 2005]. A second phase of volcanism occurred during Late Cenozoic times. The precise age of this phase of volcanism is uncertain since radiometric ages are scarce. Lavas from the Massif d'Ambre and Bobaomby Peninsula in the north of the island

have ^{40}Ar – ^{39}Ar ages of 11–12 Ma [Cucciniello *et al.*, 2011]. Outcrops of basalt on the Nosy Be archipelago, northern Madagascar, have K-Ar whole rock ages of 7–10 Ma [Emerick and Duncan, 1982; Nougier *et al.*, 1986]. Neogene-Quaternary alkaline basalts occur at several localities, notably

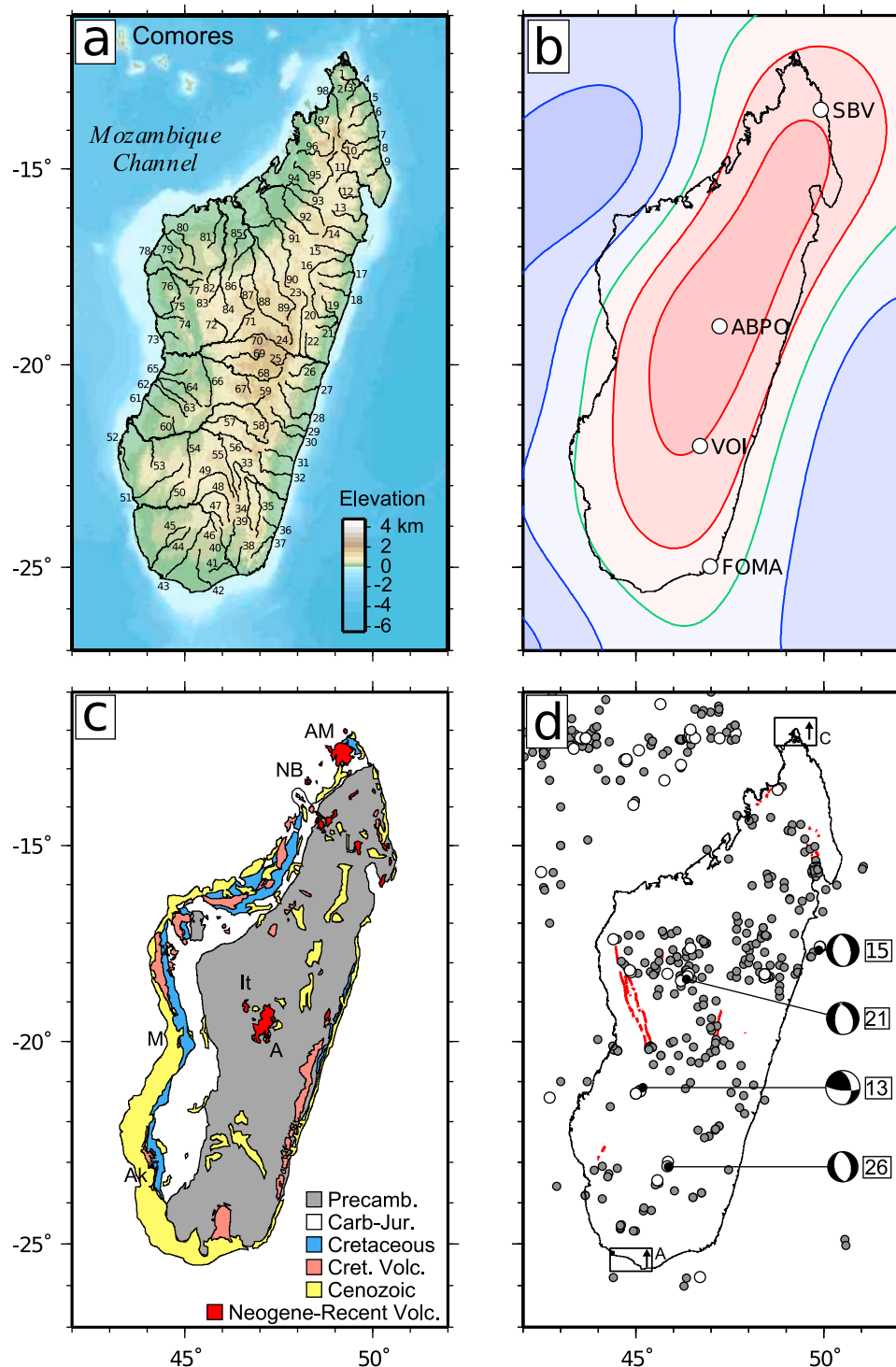


Figure 3

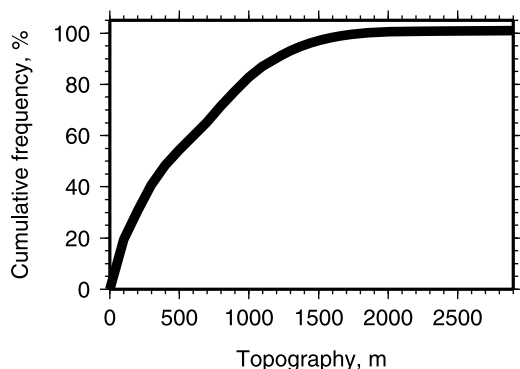


Figure 4. Cumulative frequency of Madagascar's topography. Topography estimated from SRTM dataset: $Q_1 = 150$ m, $Q_2 = 425$ m, $Q_3 = 850$ m.

in the Ankaratra highlands of central Madagascar [Bardintzeff et al., 2010].

[4] Madagascar is seismically active but most recorded events are shallower than 35 km and have magnitudes of less than 5 [Rambolamanana et al., 1997; Craig et al., 2011]. The focal mechanisms for these events indicate east-west extension. In Figure 3d, we show the location of faults mapped using the SRTM dataset and higher resolution Landsat imagery. Fault scarps on the eastern edge of the Morondava basin and in the Alatora-Ankay region (48°E, 18°S) have 10s of meters of relief, elsewhere scarp relief is lower [cf. Bertil and Regnault, 1998; Kusky et al., 2010].

[5] The maximum elevation of Madagascar is ~3 km and 75% of its topography is between 0 and 850 m (Figure 4). Four teleseismic receiver function analyses suggest that the crust beneath Madagascar is ~40 km thick [Chaheire et al., 2010] (Figure 3b). Although the vertical resolution of surface wave tomography at shallow mantle depths (50–125 km) in the continents is poor, velocity to plate thickness conversions show that the lithosphere beneath Madagascar is unlikely to be as

thick as that beneath southern Africa (Figure 2) [Priestley and McKenzie, 2006; Crosby et al., 2010]. Figure 5a shows elevation as a function of lithospheric thickness and mantle density. Elevations of up to ~2 km can be isostatically supported by a 40 km thick crust with a plate that has a thickness and mantle density of 90–160 km and 3.27–3.33 g cm⁻³ at standard temperatures and pressures, respectively. These different observations provide a useful framework for uplift models.

[6] Our principal aim is to address two questions. First, on what temporal and spatial scales did the topography of Madagascar develop? Secondly, what was the mechanism of uplift? Neogene-Quaternary volcanic activity, seismicity and deep gorges that cut into Cenozoic sedimentary rocks all indicate that the lithosphere beneath Madagascar has been modified since Paleogene times [Emerick and Duncan, 1982; Burke, 1996; de Wit, 2003; Emmel et al., 2004; Bardintzeff et al., 2010; Craig et al., 2011]. Uplifted coral reef terraces and beach deposits show that the north coast has been uplifted by several 10s of meters over the last ~1 Ma [Battistini, 1972; Pedoja et al., 2011]. The existence of deeply weathered and warped surfaces, which formed since Late Cretaceous times, implies that Cenozoic uplift is important [e.g., Storey et al., 1997; Cox et al., 2010]. Here, we are interested in exploring the constraints that drainage networks may provide. In principle, longitudinal profiles of rivers draining Madagascar contain clues about the history of uplift rate, which can be tested with coastal observations and with the history of magmatism.

2. Uplift History of Madagascar

2.1. Topographic Profiles

[7] Peneplains (i.e., erosional surfaces) are observed throughout Madagascar [Dixey, 1960; Bourgeat and

Figure 3. (a) Topographic map of Madagascar. Numbered lines = rivers extracted from SRTM dataset (see Appendix A). (b) Long wavelength free-air gravity map of Madagascar: red, blue and green = positive, negative and zero anomalies respectively; contour interval = 10 mgal. White circles labeled SBV, ABPO, VOI, FOMA = location of seismometers used to estimate crustal thickness from receiver functions [Chaheire et al., 2010]. (c) Geologic map of Madagascar [Choubert and Faure-Muret, 1990]. Key bottom right. Pink and red polygons = distribution of Cretaceous and Cenozoic magmatism, respectively: AM = Ambre Mountain, 2.0–1.8 Ma; NB = Nosy Be, 1.7–0.6 Ma; It = Itasy; A = Ankaratra, 28.6–2.8 Ma; M = Morondava 95.7–38.4 Ma; Ak = Ankililoaka, 9.7–8.7 Ma. K-Ar ages are from Bardintzeff et al. [2010]. (d) Seismicity: gray and white dots indicate earthquakes since 1973 with magnitudes $M_W < 5$ and $M_W \geq 5$, respectively (IRIS dataset from www.iris.edu). Mechanisms for earthquakes with magnitudes $M_W \geq 5$ indicate a dominance of normal faulting; boxes = centroid depths in kilometers [Bertil and Regnault, 1998; Craig et al., 2011]. Red lines = location of faults interpreted from SRTM dataset and Landsat imagery. Black boxes and arrows = Cape d'Ambre (C) coral reef terraces, which have been uplifted by ~30 m over the last 0.5–0.8 Ma; Androy-Mahafaly (A) where beach deposits are exposed at ≤ 7 m [Battistini, 1965, 1972; Pedoja et al., 2011].

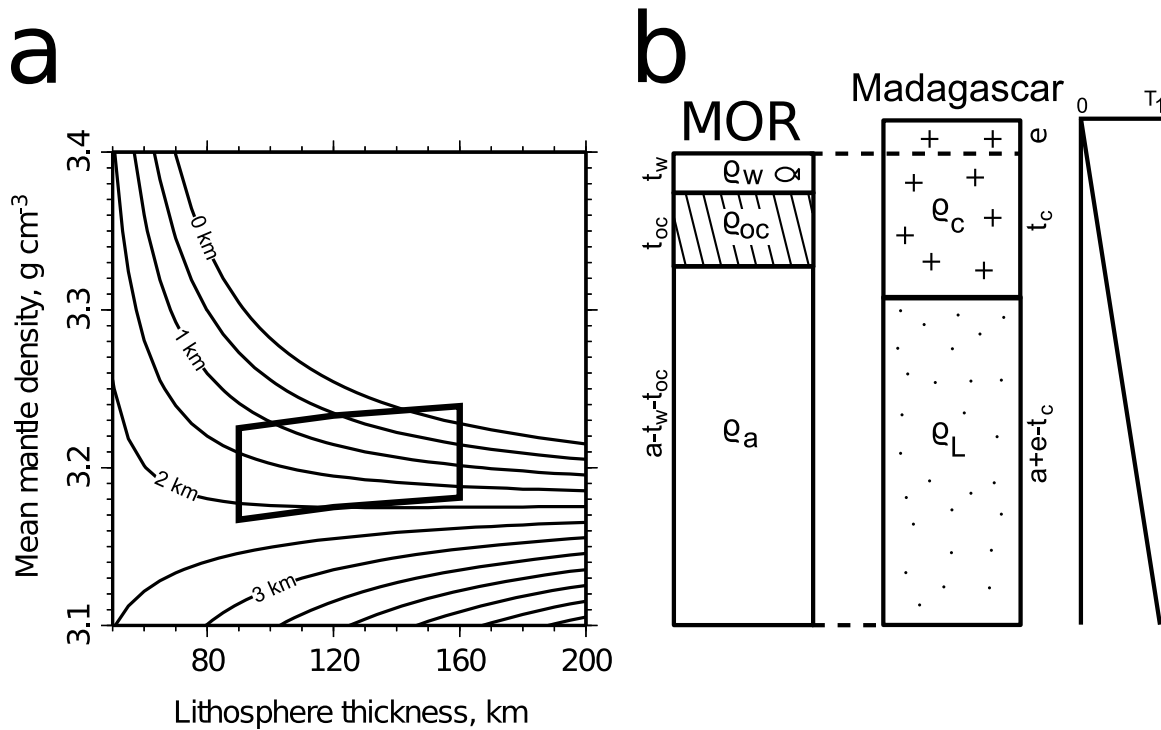


Figure 5. Isostatic balance of continental lithosphere against mid-ocean ridge to calculate elevation (e) as a function of lithospheric thickness ($a + e$) and density. We assume a linear geotherm: 0°C and 1330°C at the top and base of plate, respectively. Density of sea-water, $\rho_w = 1.03$ g cm⁻³; oceanic crust, $\rho_{oc} = 2.80$ g cm⁻³; asthenosphere, $\rho_a = 3.18$ g cm⁻³; continental crust, $\rho_c = \rho_{co}(1 - \alpha\bar{T}) \approx 2.8$ g cm⁻³; lithospheric mantle, $\rho_L = \rho_{Lo}(1 - \alpha\bar{T})$. $\alpha = 3.28 \times 10^{-5}$ °C⁻¹; $\rho_{co} = 2.85$ g cm⁻³; $3.20 \leq \rho_{Lo} \leq 3.60$ g cm⁻³. Crust is 40 km thick [Chaheire et al., 2010]. (a) Contours = $e(a + e, \rho_L)$. For example, if $a + e = 120$ km and $\rho_L = 3.21$ g cm⁻³ then $e = 1$ km. Black box shows range of elevations when $3.27 \leq \rho_{Lo} \leq 3.33$ g cm⁻³ and $90 \leq (a + e) \leq 160$ km. (b) Cartoon showing model set-up. Left panel = column at mid-oceanic ridge (MOR); central panel = continental column; right panel = continental geotherm.

Petit, 1969; Burke and Gunnell, 2008]. The most extensive peneplain can be mapped on a digital elevation model (DEM) of Madagascar, which was generated from the Shuttle Radar Topographic Mission (SRTM) dataset [see Farr et al., 2007]. Three topographic profiles reveal the main topographic features of the island (Figure 6). The main peneplain dips gently towards the coast about a central axis, which is broadly parallel to the strike of the island. This warped surface has a maximum elevation of ~ 1 km, cuts across Paleogene strata, and is overlain by Miocene marine sedimentary rocks [Dixey, 1960]. Two higher peneplains also occur. The highest, and therefore the oldest one, crosses the central Hauts Plateaux ~ 2 km above sea level. A second peneplain cuts across Late Cretaceous sedimentary rocks and lava flows at an elevation of ~ 1.5 km [Bourgeat and Petit, 1969]. Peneplains have also been recognized at elevations of ~ 0.1 – 0.5 km. These lower surfaces post-date development of the most extensive surface and a simplified topographic analysis implies that they

were produced by dome-shaped uplift events (Figure 6).

2.2. Drainage Networks

2.2.1. Data

[8] Longitudinal river profiles, where elevation is plotted as a function of distance along a river, record temporal and spatial variations in surface and rock uplift [e.g., Whipple and Tucker, 1999; Schoenbohm et al., 2004; Roberts and White, 2010]. We exploit longitudinal profiles of 98 Malagasy rivers of Strahler order ≥ 4 [Strahler, 1952]. These profiles were extracted from the SRTM dataset. First, we removed anomalous spikes and sinks from the digital topography. Secondly, we determined a drainage network by calculating flow directions across Madagascar (Figure 3a). Finally, we extracted longitudinal river profiles from this drainage network (Appendix A). Away from narrow channels, the vertical and horizontal resolution

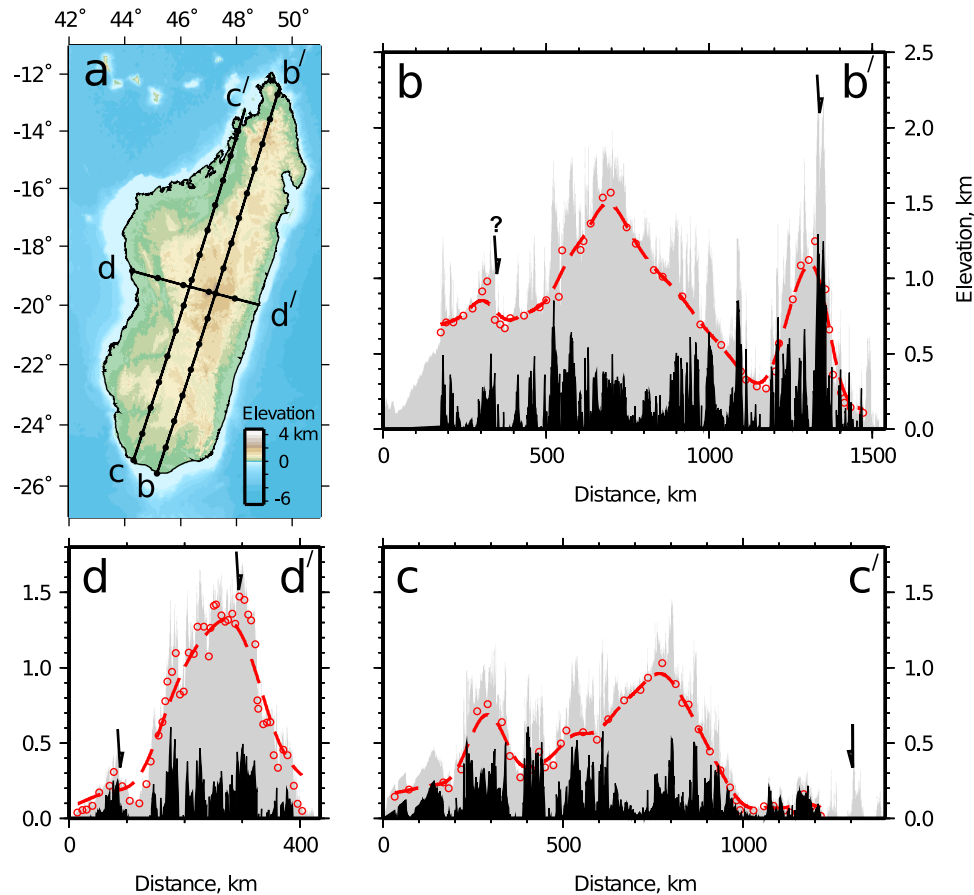


Figure 6. Erosion surfaces: estimating temporal and spatial variation in uplift rates. (a) Black lines = topographic profiles extracted from SRTM data. Distance between black dots = 100 km. (b–b') Strike-parallel topographic profile. Gray = topography. Red circles = *Dixey's* [1960] mid-Cenozoic erosional surface picked from DEM. Red dashed line = 150 km wide cosine-filter of picked erosion surface. Morphology of this most extensive erosional surface reflects the wavelength and amplitude of recent uplift. Black = residual topography calculated by flattening this erosion surface. Recent phase of uplift has a broadly domal shape. Arrows = location of faults interpreted from Landsat data. (c–c') Strike-parallel topographic profile. (d–d') Strike-perpendicular topographic profile.

is ± 90 m [Hancock *et al.*, 2006] (Figure 7). Almost all of the rivers that drain Madagascar have irregular, convex-upward, profiles. Some rivers have a broad (~ 200 km wide) knickzone, which separates zones of low relief (e.g., Figure 8). Numerous sharp (< 10 km) knickpoints also occur.

[9] The erodibility of a river channel can be inferred from the compressive strength of the substrate, which is estimated using a Schmidt hammer [e.g., Whittaker *et al.*, 2008]. Laboratory experiments have shown that tensile strength, σ_t , is a proxy for rock erodibility [Sklar and Dietrich, 2001]. Structural anisotropy may also exert an important influence on drainage patterns [e.g., Schreurs *et al.*, 2010]. However, the time and length scales on which lithology exerts a control on river shape are poorly understood. Figure 3c shows

that rivers draining Madagascar cross different lithologies. We have estimated the change in tensile strength, $\Delta\sigma_t$, along each river using Sklar and Dietrich's [2001] approach and in Figure 8 we plot $\Delta\sigma_t$ as a function of both slope and curvature. At wavelengths greater than ~ 2 km, the largest slopes and changes of slope exhibit little correlation with $\Delta\sigma_t$. In many instances, slope and curvature appear to correlate negatively with $\Delta\sigma_t$ (i.e., stronger rocks are downstream of weaker rocks). These observations indicate that shapes of river profiles at wavelengths of greater than several km are not primarily controlled by lithology or by a change in lithology. Instead, the size and location of knickzones may be more strongly determined by the timing and amplitude of regional uplift [e.g., Rosenbloom and Anderson, 1994]. Regional uplift can produce gradient changes along rivers which

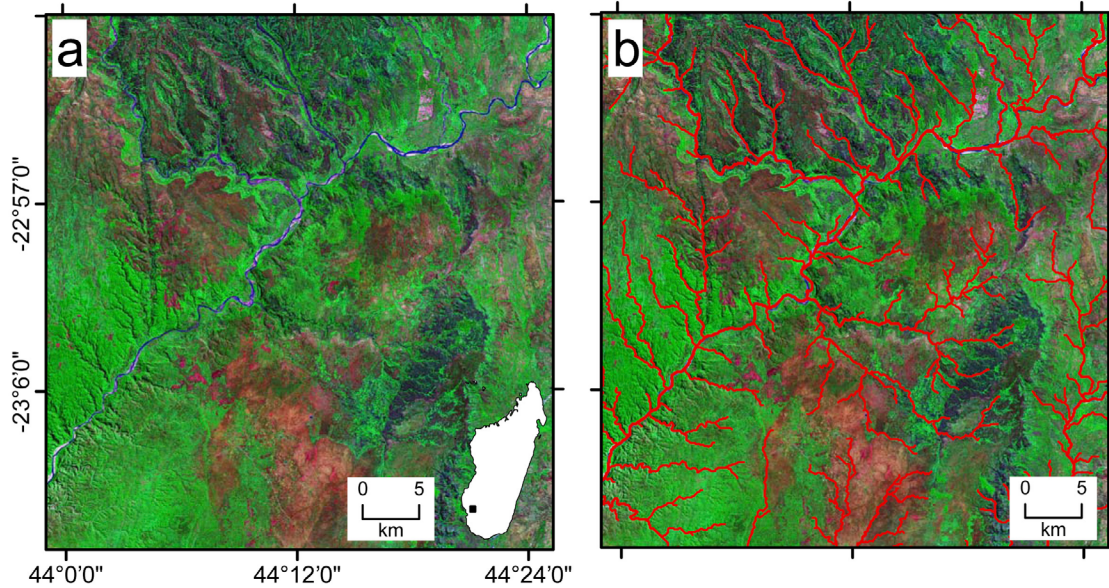


Figure 7. SRTM-derived drainage compared to Landsat data. (a) Landsat image (bands 2, 4, 7). Blue lines = tributaries of Onilahy River. Black box on inset map of Madagascar = location of Landsat data. (b) Red lines = SRTM derived drainage. Maximum discrepancy of the planform network in this region is <300 m. Horizontal resolution decreases by several hundred meters in very low relief regions.

then propagate upstream at rates determined by advective velocities which are related to integrated discharge [e.g., *Whipple and Tucker, 1999*].

2.2.2. Uplift Histories From Drainage Patterns

[10] River profiles can be used to extract uplift rate histories [*Pritchard et al., 2009; Roberts and White, 2010; Hartley et al., 2011*]. The rate of change of elevation (i.e., surface uplift), $\partial z/\partial t$, along a river profile is given by

$$\frac{\partial z}{\partial t} = U(x, t) + E(x, t), \quad (2)$$

where x is distance along the profile. U and E are the rates of rock uplift and erosion, respectively. To extract information about uplift rate from observed river profiles, the rate of erosion is required. Following *Whipple and Tucker [1999]* and *Rosenbloom and Anderson [1994]*, we assume that the rate of erosion varies as a function of upstream drainage area, A , and time, t , so that

$$E(x, t) = -vA^m \left(\frac{\partial z}{\partial x} \right)^n + \kappa \frac{\partial^2 z}{\partial x^2}, \quad (3)$$

where n and m are constants that affect the concavity of a river profile. A^m is related to the average discharge along a river [*Hack, 1957*]. n is probably ~ 1 since there is no evidence for shockwave behavior when knickpoints retreat [*Pritchard et al.,*

2009]. If $m = 0$, then v is the advective (i.e., knickpoint retreat) velocity. If $m > 0$, the advective velocity is a non-linear function of upstream drainage area. m typically varies between ~ 0.3 – 1.2 [e.g., *Schoenbohm et al., 2004*]. Thus v and m control the value of the advective term, which governs the transient form of a river profile and the knickpoint retreat velocities. *Crosby and Whipple [2006]* show that knickpoint retreat velocities may vary by at least five orders of magnitude (i.e., 1 – 10^4 cm a $^{-1}$). This term represents the detachment-limited component of erosion [e.g., *Whipple and Tucker, 1999*]. *Rosenbloom and Anderson [1994]* observed that channel convexities relax with distance upstream and time, and that rivers may erode synchronously along their length. This observation prompts the inclusion of a diffusive term, which is a simplified version of the transport-limited component of erosion [e.g., *Whipple and Tucker, 1999*]. Here κ is erosional ‘diffusivity’, which may vary by several orders of magnitude (10 – 10^7 m 2 Ma $^{-1}$), without significantly affecting our results. v and m are the most important constants since their values directly control the velocity of knickpoint retreat. Note that these two values trade-off negatively against each other and different combinations of v and m yield similar distributions of knickpoint retreat. v and m were calibrated in a study of the domal swells of west and southern Africa, which are at a similar latitude

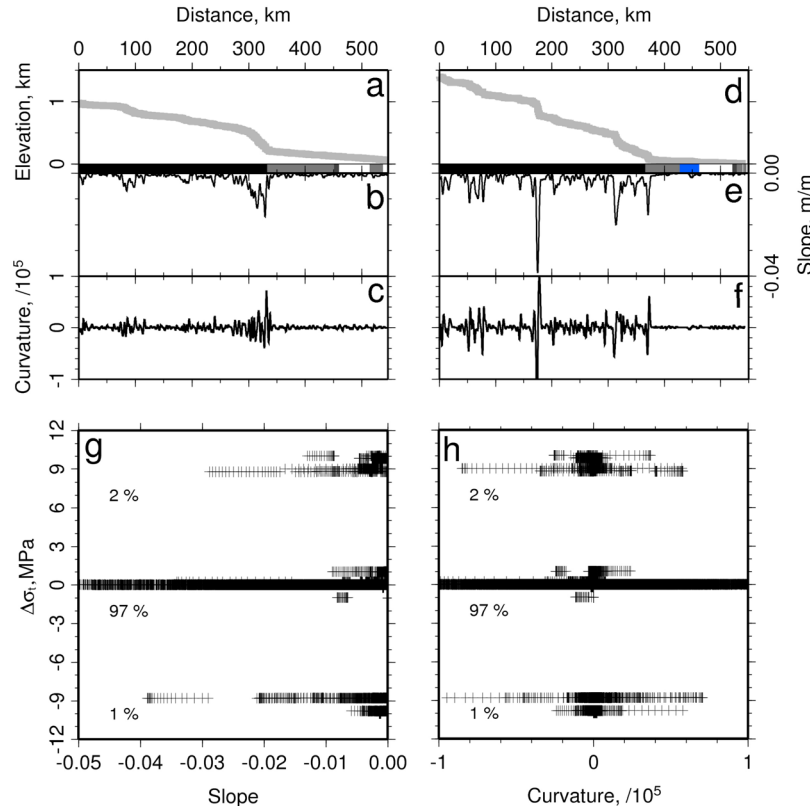


Figure 8. Difference in tensile strength of rocks along rivers in this study, $\Delta\sigma_t$, plotted as function of slope and curvature along river profiles. σ_t is a measure of rock erodibility [Sklar and Dietrich, 2001]. $\Delta\sigma_t$ estimated within a 2 km moving window and its maximum value recorded. (a–c) Example of lithology, slope and curvature measurements along Mangoky river (Figure 3a: river 56). Gray line = river profile, $z(x)$. Black, gray, blue, white and red bars = Precambrian basement, Paleozoic/Mesozoic sedimentary rocks, Cretaceous limestone, Cenozoic sedimentary rocks and magmatic rocks, respectively [Choubert and Faure-Muret, 1990]. Bedrock was divided into five categories based on likely erosional resistance. At 340 km, note change in lithology, high slopes and curvature. On Madagascar, this observation is generally atypical. (d–f) Tsiribihina river (Figure 3a: river 68). At 60, 170 and 320 km, note high slope but no lithological change (i.e., $\Delta\sigma_t = 0$). (g) Crosses = measurements of tensile strength difference for a given slope. Highest slopes do not correlate with large values of $\Delta\sigma_t$. There are numerous locations where slope positively correlates with $\Delta\sigma_t$ (i.e., weaker rock is upstream of stronger rock). (h) Crosses = measurements of tensile strength for a given curvature.

to Madagascar and have independently constrained uplift histories [Roberts and White, 2010]. These swells have drainage basins that are similar in area and relief to those on Madagascar. The misfit between observed and calculated river profiles is smallest when $v \sim 200 \text{ m}^{1-2m} \text{ Ma}^{-1}$ and $m \sim 0.2$.

2.2.3. Landscape Response Time

[11] If $\kappa = 0$, the time taken for an uplift event to propagate along Malagasy rivers, τ_G , is approximated by

$$\tau_G = \int_0^x \frac{dx}{vA^m} \approx \sum_{i=1}^x \frac{\Delta x_i}{vA_i^m}. \quad (4)$$

Solving equation (4) shows that Madagascar's rivers are unlikely to record uplift older than $\sim 30 \text{ Ma}$. In Figure 9a we show the time taken for knick-points to propagate upstream. In this way, commonalities in slope and response time can be mapped (e.g., Figure 9b).

2.2.4. Uplift as a Function of Time

[12] Equations (2) and (3) imply that if we know the shape of a river profile and if we can parameterize the rate of erosion, uplift rate histories can be calculated as a function of space and time. Equation (3) is not easily solved by analytical means. However, Pritchard et al. [2009] showed that if $\kappa = 0$ and

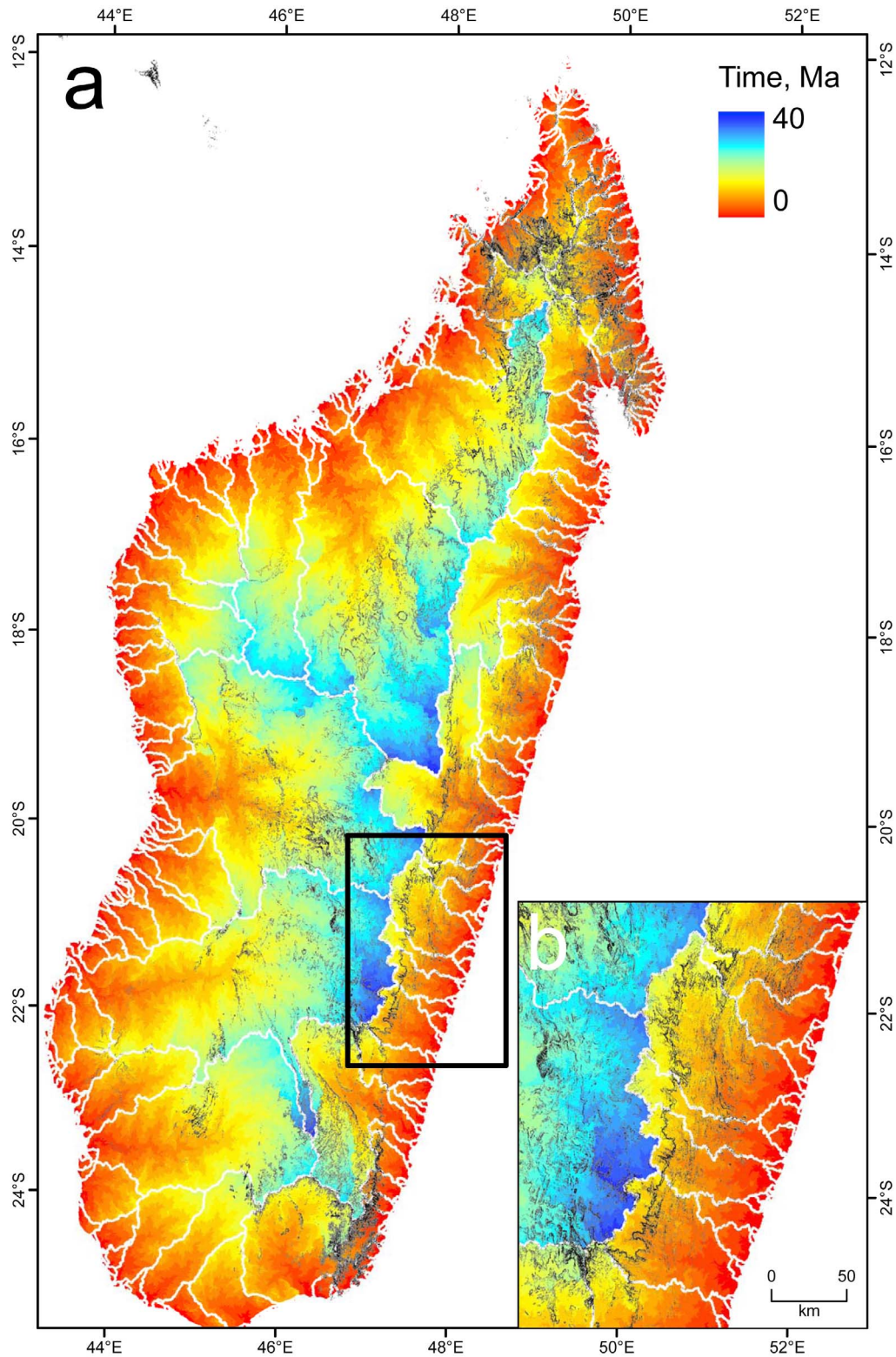


Figure 9. (a) Landscape response time of Madagascar for knickzone created at coastline. Contours every 2 Ma. Gray shading = slope between 10–25°; black shading = slope > 25°. Highest slopes occur within the central and northern highlands. (b) Enlarged section of east coast. High slopes between isochrons of 15–10 Ma (orange to yellow contours) record effects of headward migration of base-level change.

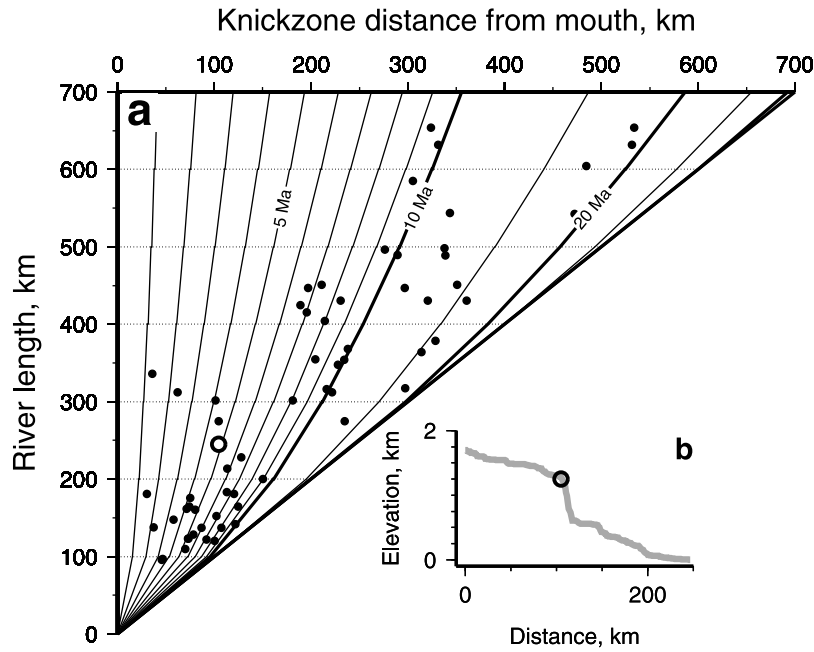


Figure 10. Look-up chart for estimating timing of uplift events from knickzone distances along river profile. Amplitude of event can be estimated from knickzone height. Knickzone retreat rate = $50x^{0.5}$ m Ma⁻¹ [Roberts and White, 2010]. (a) River length plotted as a function of knickzone distance from mouth. Numbered lines are isochrons in Ma which yield timing of uplift event for a given river length-knickzone distance pair. Open circle represents knickzone estimate for the Mangoro River shown in panel (b). Black dots = knickzone estimates of uplift timing from other Malagasy rivers. 80% of analysed rivers have knickzone ages between 0–15 Ma.

if A^m is approximated by $x^{0.5}$ and v is scaled appropriately [e.g., Hack, 1957; Rosenbloom and Anderson, 1994], $U(t)$ can be recast as

$$U(\tau) = 50x^{0.5} \left(-\frac{\partial z}{\partial x} \right), \quad (5)$$

where

$$\tau = -\frac{\sqrt{L} - \sqrt{x}}{25}. \quad (6)$$

τ is the time taken for a knickpoint to retreat to a given position along a river after an uplift event, L is the total length of the river in meters. x is the distance, in meters, from the source of the river to the knickpoint. For example, the Simianona River is 180 km long and there are two prominent knickzones, at ~ 50 and ~ 130 km downstream, which yields $\tau = 8$ and 2.5 Ma, respectively (Figure 12: River 14). These values of τ are merely approximations but they do imply that uplift has occurred during the last 10–15 Ma [cf. de Wit, 2003]. Similar results are obtained by plotting knickzone distance as a function of river length on a template which was created by solving a simplified version of equation (3) for a set of idealized longitudinal river profiles.

This scheme assumes that knickpoint retreat velocity is governed by the relationship $50x^{0.5}$ and that knickpoints are formed at the mouth of a river (Figure 10). Equations (2) and (3) were solved using a combination of finite-difference techniques [see Roberts and White, 2010].

[13] Approximations are a helpful guide but it is also useful to solve the generalized inverse problem. What is the smoothest uplift rate history which yields the best fit between calculated and observed river profiles? In the first instance, we will assume that uplift rate only varies as a function of time. To invert for U , we devised a trial function with suitable penalties which measures the difference between an observed river profile and a river profile calculated for any given U . This trial function is minimized by systematically varying U . Our goal is to find the smoothest distribution of U which produces the best-fitting synthetic river profile (Appendix B).

[14] Figure 11 shows the results of an inverse model in which uplift rate is allowed to vary only as a function of time along five major rivers. Inverse modeling of the Mangoro River, which enters the Indian Ocean, suggests that one major phase of uplift, since ~ 15 Ma, has produced the large-scale

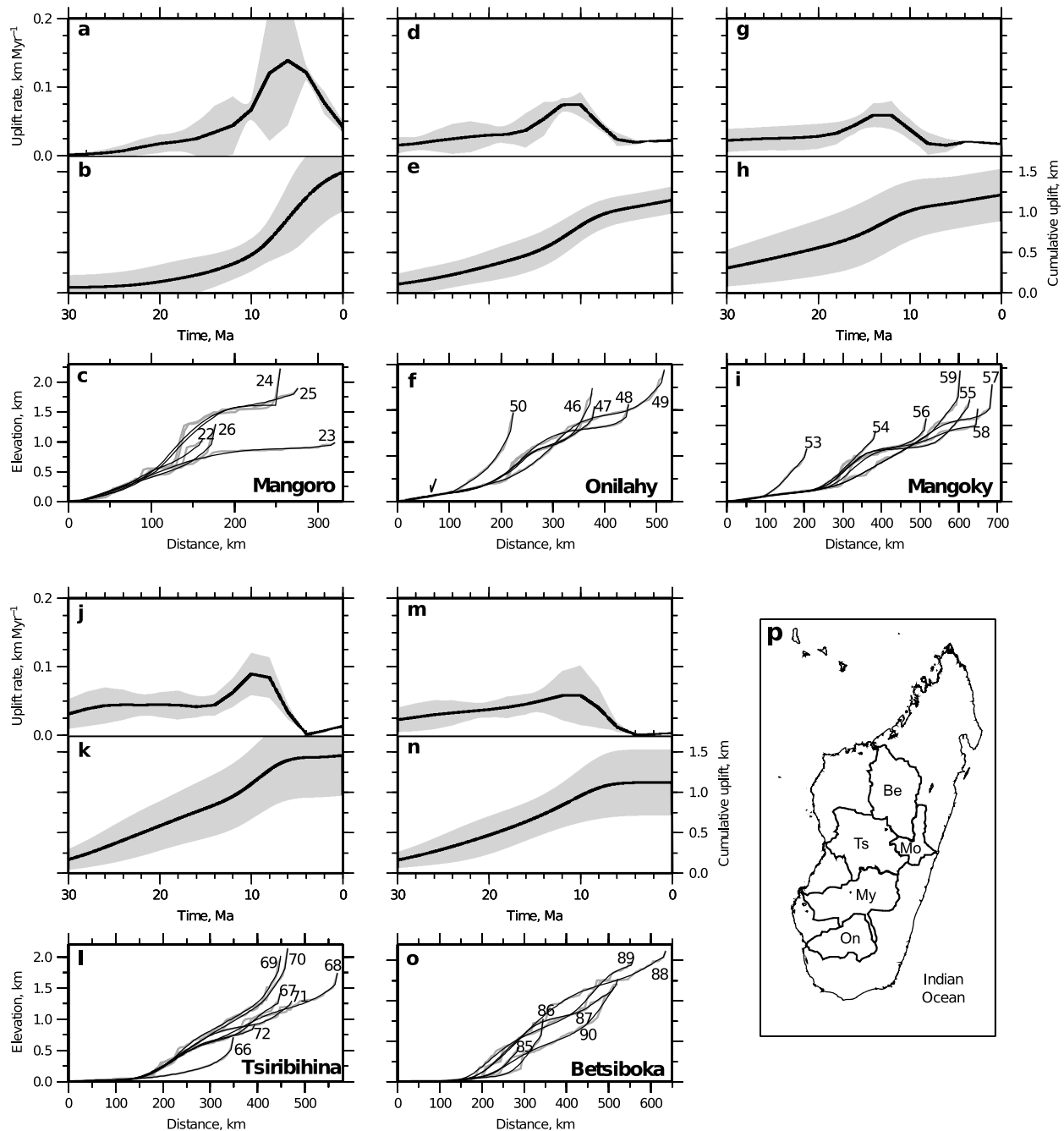


Figure 11. River profile modeling. (a) Uplift rate history of the Mangoro River. Black line = uplift rate history calculated from river profiles; gray band = 1σ uncertainty reflecting variation in erosion parameters: $200 \leq v \leq 210 \text{ m}^{1-2m} \text{ Ma}^{-1}$, $0.19 \leq m \leq 0.21$, $1.0 \leq n \leq 1.05$, $5 \times 10^2 \leq \kappa \leq 7 \times 10^2 \text{ m}^2 \text{ Ma}^{-1}$. (b) Cumulative uplift history, $\int_0^t U dt$, of the Mangoro river. Black line = cumulative uplift history; gray band = 1σ uncertainty. (c) River profile modeling. Gray numbered lines = profiles of the Mangoro River and its tributaries [see Figure 3 for location]; black lines = calculated river profiles obtained by varying uplift rate history. (d–f) Onilahy River profile analysis. Black arrows = location of faults interpreted from Landsat data [see Figure 3d]. (g–h) Mangoky River profile analysis. (j–l) Tsiribihina River profile analysis. (m–o) Betsiboka River profile analysis. (p) Map of Madagascar showing analyzed drainage basins: Be = Betsiboka, Ts = Tsiribihina, My = Mangoky, On = Onilahy, Mo = Mangoro.

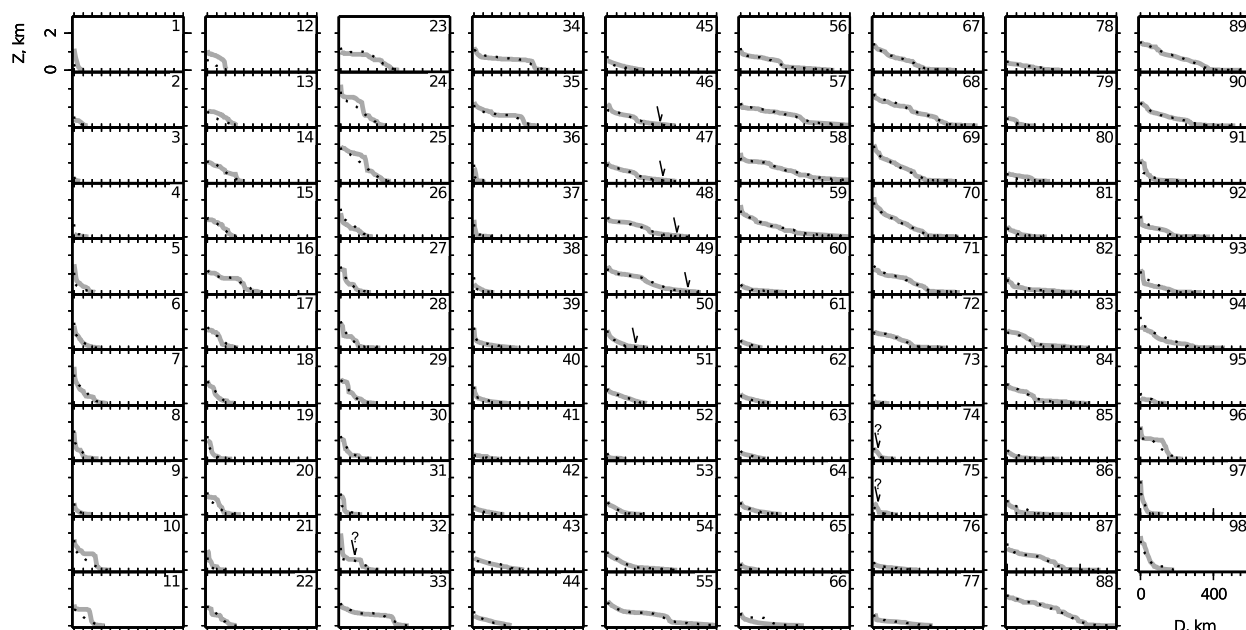


Figure 12. Inverting for uplift as a function of space and time (see Appendix B for method). 98 rivers were jointly inverted for a smooth, temporal and spatial history of uplift (see Figure 3a for location). Gray lines = observed rivers; dotted lines = calculated river profiles obtained by varying uplift rate history (Figure 14). Fits between observed and calculated rivers are excellent: $\chi^2 = 4.6$. See Appendix A for river statistics. Arrows = location of faults interpreted from Landsat imagery (see Figure 3d). Note that faults generally do not coincide with knickzones.

knickzone observed. The maximum calculated rate of uplift is $\sim 0.15 \text{ mm a}^{-1}$, which yields a cumulative uplift of $\sim 1.5 \text{ km}$. Inverse modeling of the Onilahy, Mangoky, Tsiribihina and Betsiboka rivers, which terminate on the west coast, yields uplift rates of $0.06\text{--}0.08 \text{ mm a}^{-1}$. Peak uplift rates of the west draining rivers occur between 15–10 Ma. The disparity between uplift rates and ages, calculated from east and west draining rivers, arises because uplift rate was only allowed to vary as a function of time. In reality, uplift rate history must vary as a function of both space and time, $U(x, y, t)$.

2.2.5. Uplift as a Function of Time and Space

[15] Our strategy for determining the pattern of uplift rate through time and space has three parts. First, we define the position of vertices in an $X \times Y \times T$ grid, where $X \times Y$ and T are the number of spatial and temporal vertices, respectively (Figure B1 in Appendix B). Secondly, an initial uplift rate history is input into this model; the uplift history along each river is then interpolated from the grid. Finally, the smoothest uplift rate history that yields the smallest misfit between calculated and observed rivers is sought (Appendix B). We have simultaneously

inverted 98 river profiles (Figure 12). The residual misfit, χ^2 , between observed and calculated river profiles decreases from 22 to 4.6 (Figure 13). Uplift rates of $\sim 0.2\text{--}0.4 \text{ mm a}^{-1}$ occurred between 15 and 0 Ma. A calculated history of cumulative uplift, $\int_0^t U(x, y, t)dt$, is shown in Figure 14. It is important to emphasize that no *a priori* assumptions about the

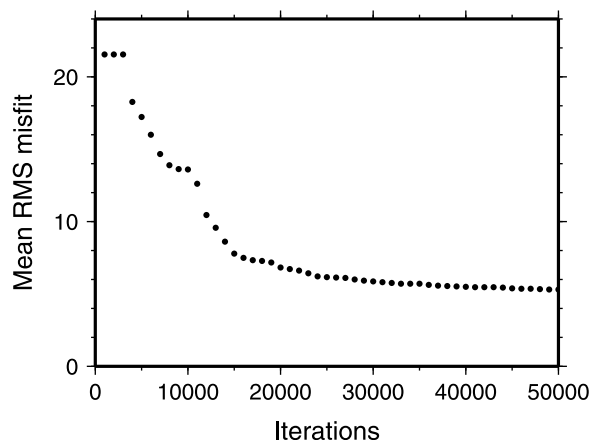


Figure 13. RMS misfit plotted as a function of number of iterations of inverse model. Optimal solution is found by minimizing misfit (and penalty functions) by varying the history of uplift rate. Black dots = misfit every 1000 iterations. Residual misfit in this study ~ 4.6 (Figure 14).

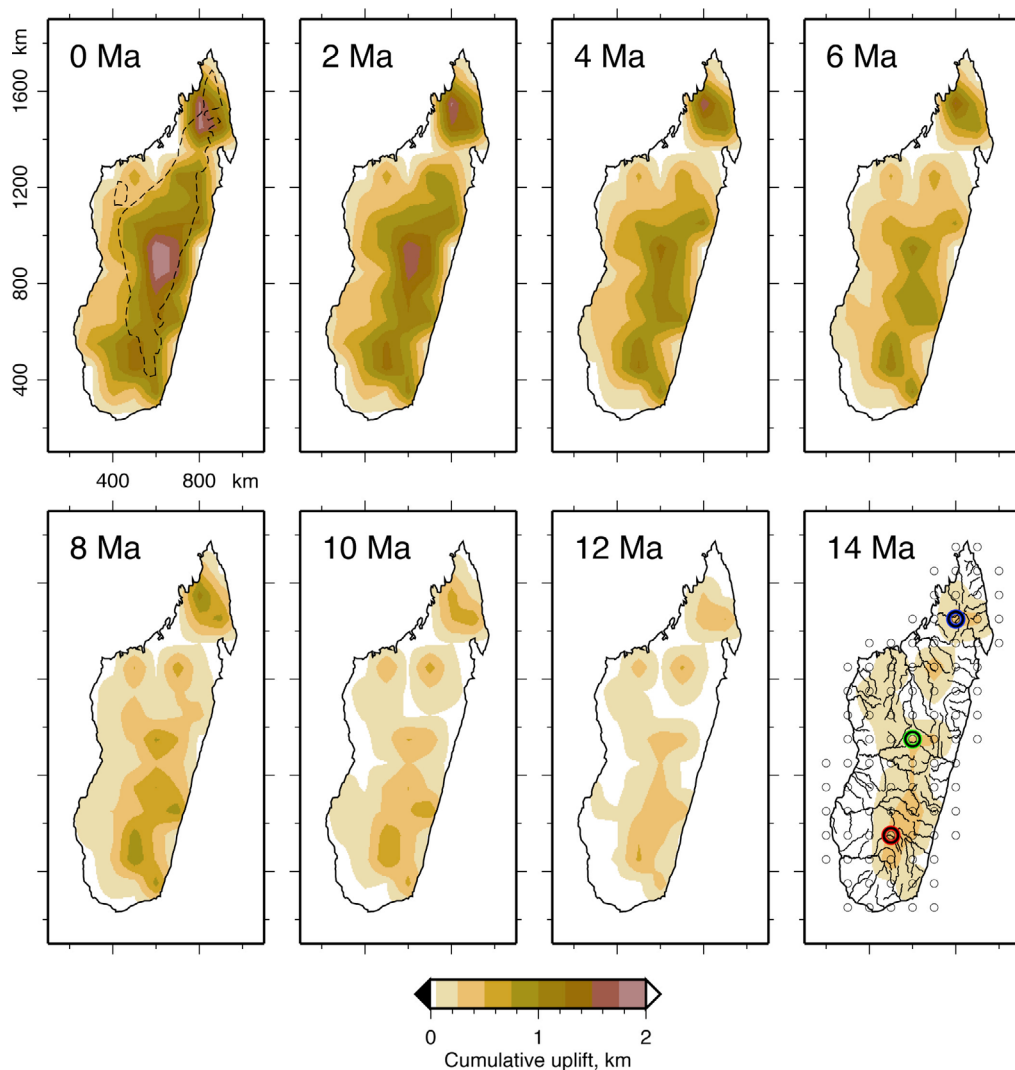


Figure 14. Inverting for uplift as a function of space and time. Calculated cumulative uplift history, $\int_0^t U(x, y, t) dt$. Black lines = rivers; black circles = vertices where uplift rate is input; colored circles = location of vertices shown in Figures 15 and 16; dashed line = lavaka-prone area [e.g., Cox *et al.*, 2010]. Madagascar has been uplifted by 1–2 km since 15 Ma.

spatial and temporal pattern of uplift rate were included. Instead, the inversion algorithm has the flexibility to choose a smooth uplift rate history which is compatible with all river profiles. Cross sections of calculated cumulative uplift suggest that two thirds (~ 1.5 km) of Madagascar's topography was generated in the last ~ 10 Ma. Prior to ~ 15 Ma Madagascar was probably relatively low-lying (Figure 15).

3. Independent Calibration

[16] Four independent sets of observations corroborate the results of our inverse modeling. First,

Quaternary coral reef terraces have been observed at elevations of ~ 4 m and ~ 30 m above sea level on the north coast of Madagascar, which have ^{14}C ages of ~ 0.1 Ma and 0.5 – 0.8 Ma, respectively (Figure 3d) [Battistini, 1972]. On the south coast, ancient beach deposits occur at ~ 0.5 – 7 m and have been assigned an age of ~ 0.1 Ma based on a correlation with Marine Isotopic Stage 5e [Battistini, 1965; Pedoja *et al.*, 2011]. These observations suggest that the northern tip of Madagascar has been uplifted at a rate of ~ 0.02 – 0.06 mm yr^{-1} over the last ~ 0.1 – 0.8 Ma and that uplift rates along the south coast are lower. These rates are consistent with our calculated uplift rate history (Figure 14).

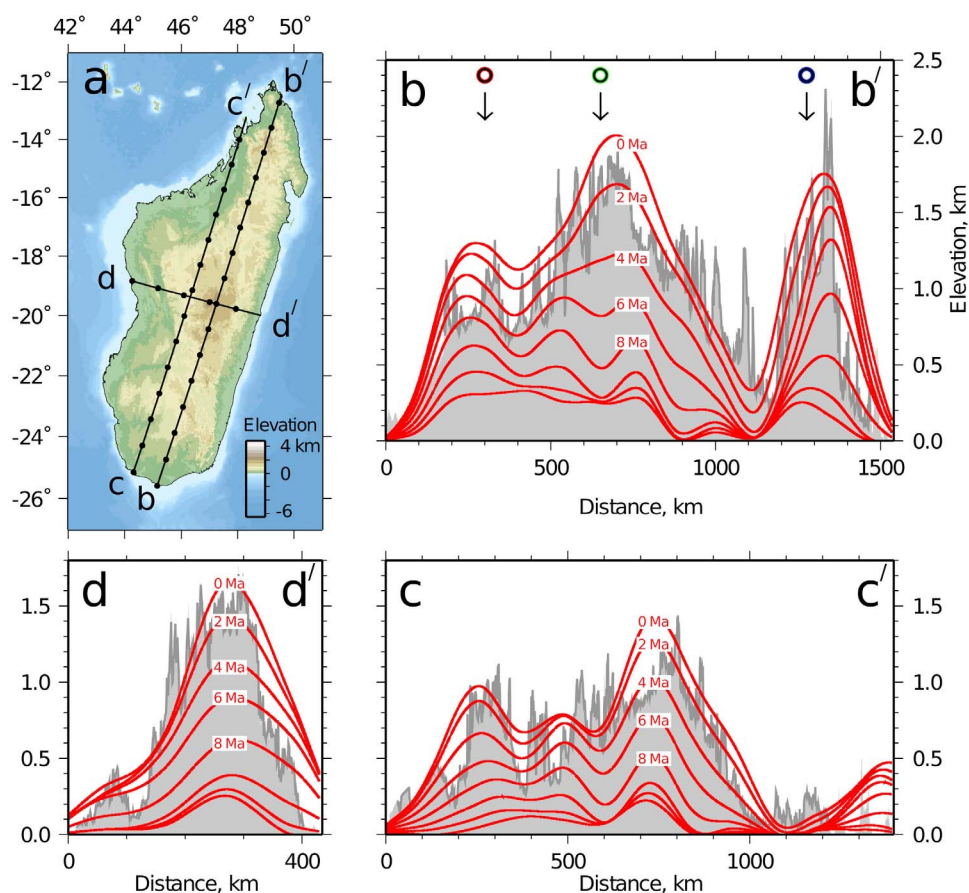


Figure 15. Cross-sections through calculated cumulative uplift history of Madagascar. Gray = modern topography. Red lines = profiles extracted from cumulative uplift history at ages shown (Figure 14). $\sim 2/3$ of topography was generated from ~ 15 – 0 Ma at a rate of ~ 0.2 – 0.4 mm a^{-1} . Discrepancies between uplift history estimated from inverse modelling and from erosional surfaces (e.g., at 1300 km on profile c–c') are partly due to spatial resolution of uplift grid used in inverse model. In Figure 15b, circles indicate location of vertices shown in Figures 14 and 16.

[17] Secondly, the distribution and morphology of peneplains helps to constrain the history of uplift. Globally, rates of laterite/saprolite formation vary widely from 1 – 50 m Ma^{-1} [Wright, 1994; Mathieu et al., 1995; Hill et al., 2000; Braun et al., 2009]. In Madagascar, Cretaceous basalt has been reduced to saprolite and laterite, which is up to 100 m thick [Storey et al., 1997]. These production rates suggest that Madagascar was probably low-lying during Paleogene times. If so, any topography generated during separation from India at ~ 85 Ma could have been eroded away by Late Cretaceous times. Deeply weathered surfaces are now cut by steep cliffs, by erosional gullies (lavakas), and by river channels [Wells and Andriamihaja, 1993; Cox et al., 2010]. Canyons cut into Cenozoic sedimentary rocks and Miocene marine rocks have been recorded at elevations >1.1 km [Dixey, 1960; de Wit, 2003]. These observations suggest that

Madagascar has been uplifted by at least 1 km since Early Miocene times.

[18] Thirdly, the magmatic history of Madagascar helps to constrain the timescale of dynamic support. Cretaceous basalts of the Volcan de l'Androy ($\sim 46^\circ E$, $24^\circ S$) and the southern Morondava Basin ($\sim 44^\circ E$, $22^\circ S$) have similar trace element data [Storey et al., 1997]. This observation prompted de Wit [2003] to conclude that these lavas flowed across large areas of the island and that it was probably low-lying during Cretaceous times. The composition of Neogene–Recent igneous rocks at Nosy Be is similar to those observed on the Comore islands, a chain of oceanic islands ~ 350 km northeast of Madagascar (Figure 3a). Here, Neogene–Recent volcanic rocks have a spectrum of compositions which include those derived from deep plume material, as well as those formed by variable degrees

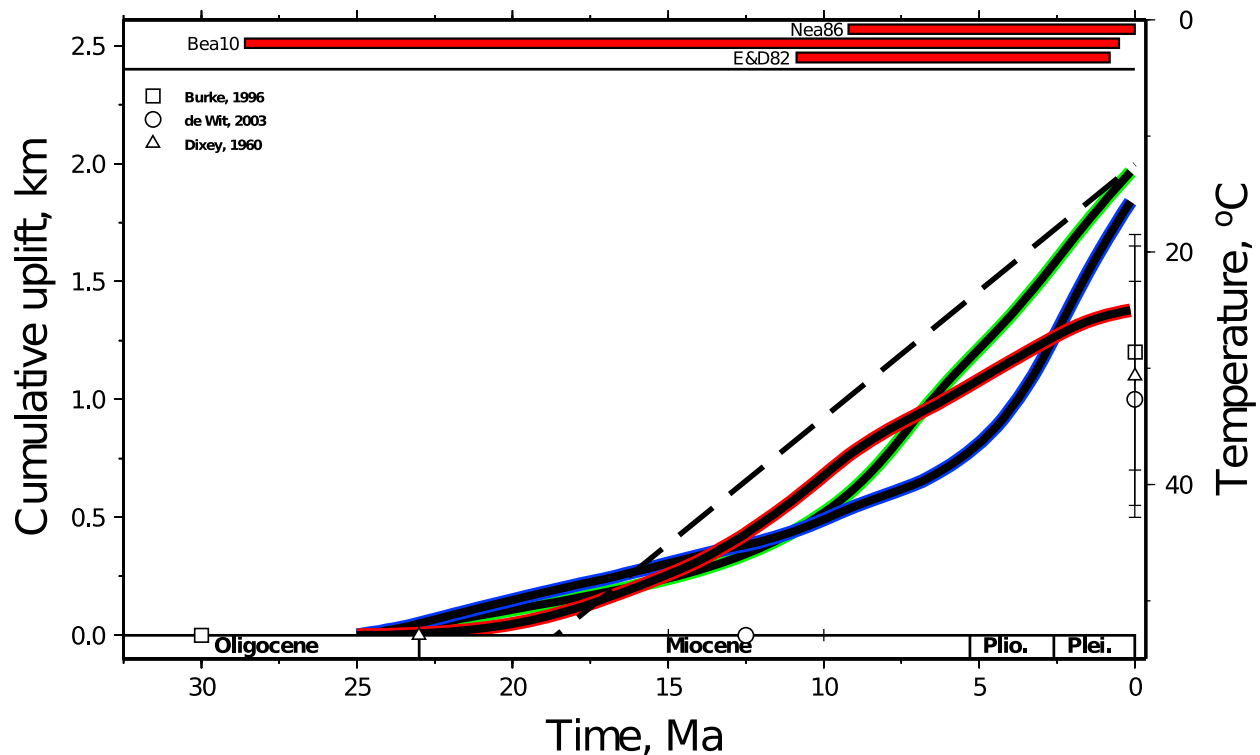


Figure 16. Comparison between predicted and published cumulative uplift histories of Madagascar. Black lines with red, blue and green outline = cumulative uplift history of vertices in southern, northern and central regions, respectively (see Figures 14 and 15 for location). Red polygons indicate periods of magmatic activity at Nosy Be, Nosy Mitsou, Nosy Komba, Diego Suarez and the Massif d’Ambre in north of the island (E&D82 [Emerick and Duncan, 1982]; Nea86 [Nougier et al., 1986]) and at the Ankaratra and Itasy volcanic fields in the center of the island (Bea10 [Bardintzeff et al., 2010]). Dashed line = cooling history from titanite and apatite fission track analysis of basement rocks [Emmel et al., 2004].

of shallow-level melting during interaction with plume-derived melts [Deniel, 1998; Melluso and Morra, 2000]. Hence, the regional picture is consistent with an emergent mantle plume, which pooled beneath Madagascar during the last ~15 Ma [de Wit, 2003]. Finally, analysis of titanite and apatite fission tracks indicate that basement rocks have cooled by at least 60°C during the last ~30 Ma (Figure 16) [Emmel et al., 2004; Seward et al., 2004]. These thermochronometric studies constrain denudation rather than uplift although both are linked. Our estimated history of uplift is consistent with cooling over the last ~30 Ma.

4. Discussion

[19] If Madagascar was low-lying at ~15 Ma, it has since been uplifted by up to ~1.5 km at a maximum rate of ~0.2–0.4 mm a⁻¹. These rates of uplift are

similar to those calculated in other regions of predicted dynamic support, for example the Bié dome (~0.4 mm a⁻¹ [Al-Hajri et al., 2009]) and the North Atlantic Ocean (~0.6 mm a⁻¹ [Hartley et al., 2011]). Explanations of Madagascar’s uplift history must account for a range of disparate observations: long wavelength free-air gravity anomalies of +30 mgal; the location, age, and composition of magmatism across the island; the apparent youthfulness of Madagascar’s landscape; widespread, but localized, seismicity; and elevated, warped peneplains.

[20] A positive free-air gravity anomaly centered on Madagascar suggests that there is some component of sub-plate support. Thin lithosphere may lie beneath the Quaternary volcanic fields of Ankaratra and Itasy in the island’s center, where seismicity is more intense (Figures 3c and 3d [Rakotondraompiana et al., 1999]). These observations are consistent with the notion that the asthenosphere beneath central

Madagascar is upwelling [*de Wit*, 2003]. Tomographic images indicate that this upwelling is related to a much larger convective feature [e.g., *Ritsema et al.*, 2011] (Figure 1b). Our calculated uplift rate history suggests that dynamic support of Madagascar grew over the last ~15 Ma.

[21] How does the erosional process affect our results? How do changes in discharge and lithology affect our results? Our starting point is that fluvial erosion moderates the effects of regional uplift. In Figure 8, we show that change in lithology does not correlate with either slope or curvature. On sufficiently long time and length scales (i.e., 1–10s Ma, 10–100s km), changes in river substrate may therefore play a less significant role in determining the shape of a river profile (Figure 8). Instead, the history of uplift is the predominant control on the transient form of a river profile (Figure 14). It is also possible that the erosional process is controlled by integrated, rather than instantaneous, discharge. The time period over which discharge should be integrated is not known but it may be of the order of 1 Myrs. Thick, deeply weathered peneplains, which probably formed during Paleogene times, are consistent with negligible pre-existing topography. Movement on normal faults has had no obvious effect on the shapes of the rivers that we have inverted (Figures 11–14). The reference level can also change since sea level fluctuates, although this effect is usually small when compared with cumulative uplift [cf. *Miller et al.*, 2005].

5. Conclusions

[22] We have estimated the uplift rate history of Madagascar over the last 15 Ma by simultaneously inverting nearly 100 longitudinal river profiles. Our inversion algorithm is predicated upon an empirical model which exploits an advective-diffusive formulation to match river profile shape. The fit between observed and calculated profiles is surprisingly good. Calculated uplift rate histories imply that Madagascar was uplifted during the last 15 Ma, at a maximum rate of ~0.2–0.4 mm a⁻¹. Despite uncertainties in the values of erosional constants, our history of uplift is also consistent with increased volcanic activity during the last 15 Ma and with the erosion of Cenozoic sedimentary rocks. We conclude that Madagascar's eroding landscape contains important clues about the temporal and spatial evolution of regional uplift. Neogene convective upwelling appears to have played

an important role in creating and maintaining the present-day physiography of Madagascar.

Appendix A: Drainage Inventory

[23] Longitudinal profiles of 98 Malagasy rivers have been extracted from the SRTM dataset (Figure 3a). Their attributes are shown in Table A1.

Appendix B: Numerical Methods

[24] Equations (2) and (3) were solved by finite-difference techniques. Distance from the head of a river, x , and time, t , are discretized so that

$$\begin{aligned} x_j &= x_0 + j\Delta x, & j &= 0, 1, \dots, J \\ t_i &= t_0 + i\Delta t, & i &= 0, 1, \dots, I. \end{aligned} \quad (\text{B1})$$

We used a Crank-Nicolson scheme, which is second-order accurate and stable for any Δt , to solve the diffusive term

$$\frac{z_j^{i+1} - z_j^i}{\Delta t} = \frac{\kappa}{2} \left(\frac{(z_{j+1}^{i+1} - 2z_j^{i+1} + z_{j-1}^{i+1}) + (z_{j+1}^i - 2z_j^i + z_{j-1}^i)}{(\Delta x)^2} \right). \quad (\text{B2})$$

An upwind differencing scheme is used to solve the advective term

$$\frac{z_j^{i+1} - z_j^i}{\Delta t} = -vA^m \left(-\frac{z_{j+1}^i - z_j^i}{\Delta x} \right)^n, \quad (\text{B3})$$

where the stability criterion is

$$\frac{\left| vA^m \left(-\frac{z_{j+1}^i - z_j^i}{\Delta x} \right)^{n-1} \right| \Delta t}{\Delta x} \leq 1. \quad (\text{B4})$$

The steepest slope, $(z_{j+1}^i - z_j^i)/\Delta x$, and maximum drainage area are used to set the timestep required for stability. Both numerical schemes are combined using operator-splitting techniques [*Press et al.*, 1992; *Roberts and White*, 2010].

[25] During inversion, different distributions of $U(x, y, t)$ are used to calculate M river profiles which are then compared to M observed rivers. In our starting model $U(x, y, t) = 0$ and $z(x) = 0$. We set Dirichlet boundary conditions, $U(x, y, t) = 0$, at the coast. $U(x, y, t)$ is systematically varied until calculated and observed river profiles agree within error. To implement this scheme, $U(x, y, t)$ is parameterized by selecting $X \times Y \times T$ discrete

Table A1. Attributes of Madagascar's Rivers Numbered Clockwise From North^a

Number	Name	Longitude ^b	Latitude ^c	Elevation ^d (m)	Length (km)	Basin ^e (km ²)
1–3	Irodo	49.549	–12.643	1160	77	1211
4	Lokia	49.658	–12.801	133	80	1365
5	Manambato	49.924	–13.198	1516	115	1228
6	Fanambana	50.026	–13.493	1190	144	1820
7	Bemarivo	50.161	–14.149	1986	186	4507
8	Lokoho	50.178	–14.426	1433	134	2084
9	Ankavia	50.267	–14.866	623	93	1118
10–11	Antainambalana	49.763	–15.457	1662	198	5379
12	Rantabe	49.626	–15.688	1085	108	1449
13	Manaro	49.755	–16.157	828	166	2380
14	Simianona	49.697	–16.871	1076	198	3735
15–16	Maningory	49.445	–17.219	1173	298	12531
17	Onibe	49.495	–17.644	978	166	2459
18	Mahatsara	49.366	–18.264	1240	158	2886
19–21	Rianila	49.105	–18.975	1086	185	7247
22–26	Mangoro	48.79	–19.997	2224	321	16717
27	Fanantara	48.537	–20.583	1270	168	1946
28	Mananjary	48.35	–21.253	1362	217	6486
29	Namorona	48.219	–21.655	1344	196	2101
30	Faraony	48.162	–21.804	1078	168	2608
31	Sandrananta	47.921	–22.421	1052	119	4127
32	Manampatrana	47.845	–22.791	1977	212	3805
33–35	Mananara	47.697	–23.349	1275	414	15927
36	Mamanbondro	47.441	–24.049	709	61	1115
37	Manampanihy	47.342	–24.285	915	106	2042
38–41	Mandrare	46.433	–25.18	1085	240	13685
42	Manambovo	45.717	–25.463	375	166	5432
43	Menarandra	44.518	–25.295	662	272	8198
44–45	Linta	44.06	–25.02	531	205	5832
46–50	Onilahy	43.76	–23.563	1420	515	30365
51	Fiherenana	43.635	–23.336	797	217	7255
52	Kililaly	43.279	–21.96	126	94	1852
53–60	Mangoky	43.458	–21.445	1717	686	51460
61	Ankelifolo	43.919	–20.84	389	117	1816
62	Maharivo	44.135	–20.481	530	164	3353
63–64	Morondava	44.271	–20.312	706	225	5795
65	Andranomena	44.378	–20.156	408	100	965
66–72	Tsiribihina	44.42	–19.653	2133	394	55773
73	Manambolo	44.370	–19.317	111	1419	1419
74–75	Ankiliroa	44.211	–18.693	639	131	4054
76–77	Manambaho	43.983	–17.723	570	324	7329
78	Ranobe	44.159	–17.167	405	297	4101
79–80	Sambao	44.434	–16.606	432	236	8144
81–84	Mahavavy	45.903	–15.773	1086	457	17959
85–90	Betsiboka	46.302	–15.826	1626	635	53698
91–94	Sofia	47.04	–15.506	1246	463	31708
95	Ankofia	47.978	–14.790	301	144	3607
96	Maevarano	47.75	–14.616	1767	234	5030
97	Sambirano	48.498	–13.545	1858	123	2695
98	Mananjeba	48.876	–13.019	1838	186	3069

^aSee Figure 3 for location.

^bLongitude of river mouth in decimal degrees.

^cLatitude of river mouth in decimal degrees.

^dMaximum elevation.

^eArea of drainage basin.

values, U_k . $X \times Y$ vertices are arranged in a regular grid, where $\Delta x = \Delta y$, typically 50–200 km (Figure B1). There are T parameters at each vertex, with a time interval Δt , where Δt is typically 1–

5 Ma. Optimal values of U_k are interpolated at each vertex to determine $U(t)$. $U(x, t)$ for each river is bilinearly interpolated from the grid of

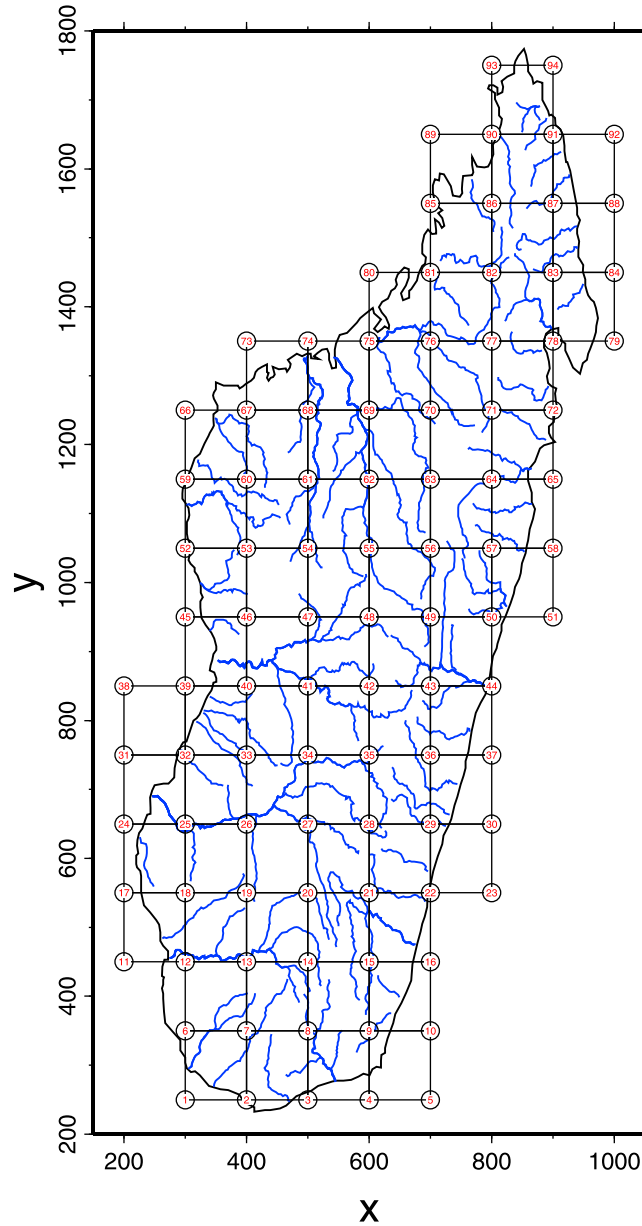


Figure B1. Distribution of vertices in inverse model. Blue lines = modeled rivers; black numbered circles = position of vertices where uplift is input into the model. x and y are plotted in kilometers. Uplift rate at each point along a river is bilinearly interpolated from four surrounding vertices.

$X \times Y$ vertices (e.g. Figure B2). We chose to minimize a trial function, H , by varying U_k , where

$$H = \left[\frac{1}{NM} \sum_{n,m=1}^{N,M} \left(\frac{z_{nm}^o - z_{nm}^c}{\sigma_n} \right)^2 \right]^{1/2} + P, \quad (\text{B5})$$

where z_{nm}^o and z_{nm}^c are the observed and predicted rivers, respectively. N is the number of data points along each river. σ_n is the uncertainty in elevation. To stabilize the inversion algorithm, a

set of penalty functions, P , is included. P ensures that the first and second derivatives of U are smooth and that U is positive. Our spatial and temporal smoothing penalties were given weighting coefficients of 0.00001 and 50, respectively. Our positivity penalty function was given a weighting coefficient of 5. Our results are not significantly affected by changing these weightings by several orders of magnitude. A conjugate gradient method is used to minimize H [Press *et al.*, 1992]. Figure 13

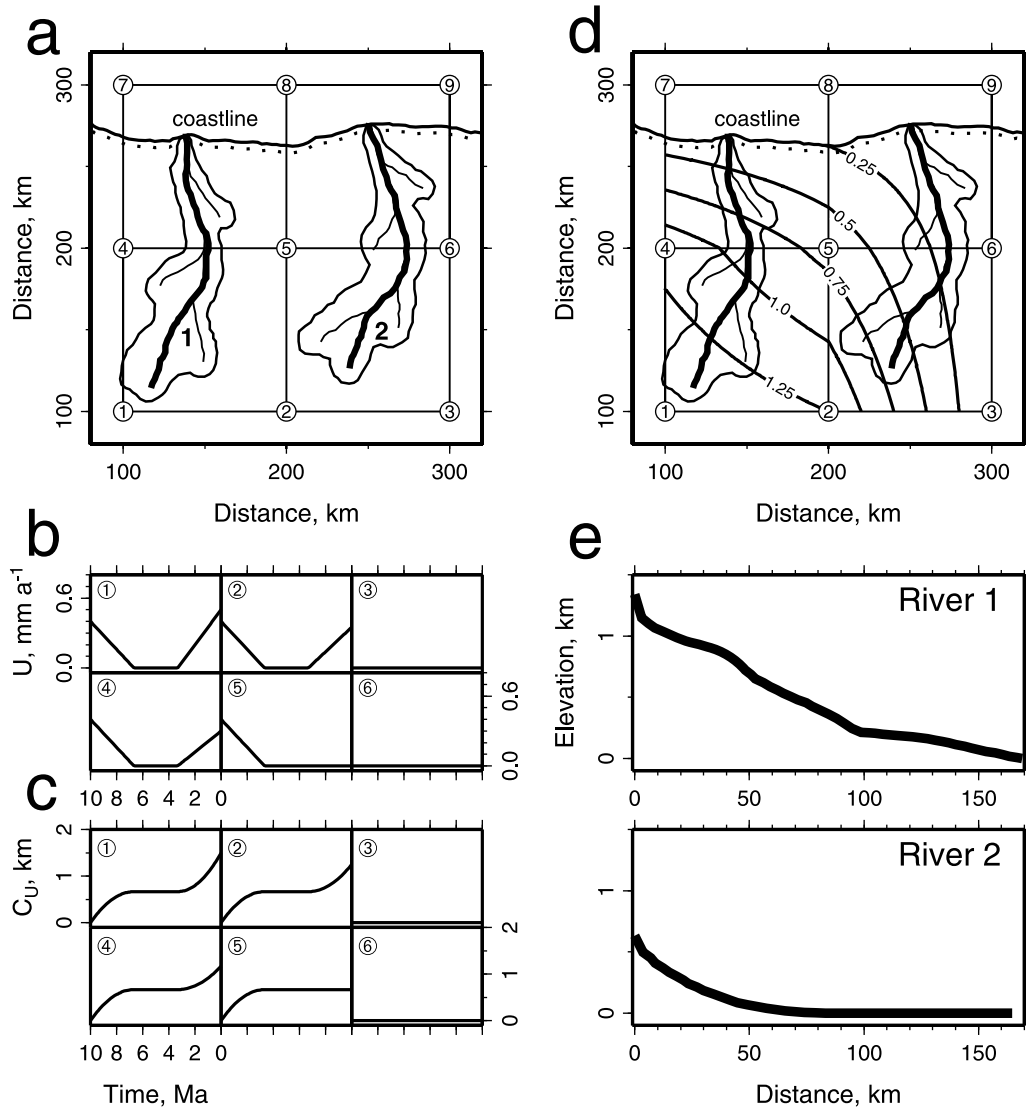


Figure B2. Forward model calculated using synthetic data and an idealized history of uplift. (a) Map showing idealized drainage network. Black numbered lines = rivers (draining to the north); thin black lines = drainage divides. White numbered circles = vertices where uplift is input into the model. Dirichlet boundary conditions are set at vertices 7–9: $U(t) = 0$. Straight black lines = edges. Dotted line = coast. Uplift as a function of distance along a river, $U(x, t)$, is bilinearly interpolated from uplift history of surrounding vertices, $U(x, y, t)$. (b) Idealized uplift rate history of each vertex, $U(t)$. Location of each vertex is shown in panel a. In this example, there were two phases of rapid uplift. (c) The history of cumulative uplift (C_U) of each vertex is $\int_0^t U dt$. (d) Map showing cumulative uplift at final timestep (i.e., 0 Ma). Contoured surface = cumulative uplift; contour interval = 250 m. (e) Calculated longitudinal river profiles, $z(x)$, at final timestep. Location of rivers is shown in panels a and d. Both rivers have been uplifted and eroded. In this example: $v = 200 \text{ m}^{1-2m} \text{ Ma}^{-1}$, $m = 0.2$, $n = 1$, $\kappa = 10^2 \text{ m}^2 \text{ Ma}^{-1}$.

shows misfit as a function of number of iterations for the x, y, t solution given in this study, which is shown in Figure 14.

Notation

z elevation of river [L].
 L river length [L].

x distance [L].
 y distance [L].
 A upstream drainage area [L^2].
 t time [T].
 τ knickpoint retreat time [T].
 τ_G knickpoint retreat time [T].
 U uplift rate [$L \text{ T}^{-1}$].
 E erosion rate [$L \text{ T}^{-1}$].

v	advective coefficient of erosion [$L^{1-2m} T^{-1}$].
m	area exponent, dimensionless.
n	slope exponent, dimensionless.
κ	diffusive coefficient of erosion [$L^2 T^{-1}$].
σ	variance of data [L].
$X \times Y$	number of spatial dimensions, dimensionless.
T	number of temporal dimensions, dimensionless.
M	number of rivers, dimensionless.
N	number of data points along each river, dimensionless.
P	penalty functions, dimensionless.
$\overline{V}_s$	mean shear velocity [$L T^{-1}$].
V_{s_i}	average layer velocity [$L T^{-1}$].
S	number of layers in velocity calculation, dimensionless.

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