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1 Linking uplift, erosion, and sedimentation using landscape
2 evolution models: Madagascar since the Late Cretaceous

3 Ruohong Jiao ^{*1, 2}, Jean Braun¹, Antoine Delaunay^{3, 4}, Cécile Robin³, and François
4 Guillocheau³

5 ¹*Helmholtz Centre Potsdam, GFZ German Research Centre for Geosciences, Potsdam, Germany*

6 ²*School of Earth and Ocean Sciences, University of Victoria, Canada*

7 ³*Géosciences Rennes, Université de Rennes 1, France*

8 ⁴*Ali Al-Naimi Petroleum Engineering Research Center, King Abdullah University of Science and
9 Technology, Thuwal, Saudi Arabia*

*Correspondence: rjiao@uvic.ca

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Abstract

We present a study to estimate the large-scale landscape history of a continental margin, by establishing a source-to-sink volume balance between the eroding onshore areas and the offshore basins. Assuming erosion as the primary process for sediment production, we strive to constrain a numerical model of landscape evolution that balances the volumes of eroded materials from the continent and that deposited in the corresponding basins, with a ratio imposed for loss of erosion products. We use this approach to investigate the landscape history of Madagascar since the Late Cretaceous. The uplift history prescribed in the model is inferred from elevations of planation surfaces formed at various ages. By fitting the volumes of terrigenous sediments in the Morondava Basin along the west coast and the current elevation of the island, the landscape evolution model is optimized by constraining the erosion law parameters and ratios of sediment loss. The results include a best-fit landscape evolution model, which features two major periods of uplift and erosion during the Late Cretaceous and the middle to late Cenozoic. The model supports suggestions from previous studies that most of the high topography of the island was constructed since the middle to late Miocene, and on the central plateau the erosion has not reached an equilibrium with the high uplift rates in the late Cenozoic. Our models also indicate that over the geological time scale, a significant portion of materials eroded from Madagascar was not archived in the offshore basin, possibly consumed by chemical weathering, the intensity of which might have varied with climate.

Keywords

Landscape evolution model | Source to sink | Erosion | Sedimentary basin | Chemical weathering | Madagascar

1 Introduction

The topography of a continent can be built by dynamic evolution of the Earth through processes such as plate convergence and mantle convection. On the other hand, the topographic relief is reduced by relocating or consuming rocks through erosion and weathering by physical, chemical, and biological activities. As the topography controls the routing of freshwater and sediment, atmospheric circulation patterns, and distribution and evolution of the biota, estimating the landscape evolution of a region is valuable for understanding how these systems evolve and interact. Sediments accumulated in basins adjacent to uplifted areas can be used to provide information on the landscape history of the source regions. Paleoelevation and paleorelief can be estimated using isotope-based methods (e.g., Rowley and Garzione, 2007; Zhuang et al., 2019), but these empirical methods are limited by the availability of samples in the stratigraphy. Changes in sedimentation rate are also used for inferring variations in erosion rate in the provenance (e.g., Zhang et al.,

2001), to reflect modifications of the regional elevation or relief. However, sediment accumulation rates estimated from one-dimensional age-depth profiles are biased by the incomplete stratigraphy (Sadler, 1981), and therefore reliable estimates of the erosion rates of the source regions require stratigraphic records in two (across the basin margin) or three dimensions.

We demonstrate a case study in Madagascar, in which a numerical model of the landscape evolution is constrained using sediment volume data from basins on the continental margin. As a former part of Gondwana, Madagascar was separated from Africa at ~ 160 Ma, from Antarctica at ~ 120 Ma, and from the Seychelles and India at ~ 90 Ma (de Wit, 2003). Thereafter since the Late Cretaceous, Madagascar as an island in the Indian Ocean has its own geological history on a passive margin. Recovering the landscape history of the island will help understand the interaction between the Earth's tectonic, mantle, and surface processes (Emmel et al., 2012; Roberts et al., 2012; Stephenson et al., 2021). This long-term isolation of the island from other continental masses resulted in a relatively closed system for its crustal material, which was re-distributed by erosion, transport, and sedimentation processes following uplift of the island. If the input from magmatism (e.g., Storey et al., 1995) can be ignored and a closed system of sediment cycling can be assumed, the volume of siliciclastic sediments deposited in the basins should in general equal to the total onshore erosion after subtracting the component consumed by weathering. This simple balance allows us to link the two using a landscape evolution model (LEM) which could be constrained by empirical data. Here by using a formal inversion method, i.e., the Neighbourhood Algorithm (Sambridge, 1999), we attempt to optimize a LEM of Madagascar that reproduces volumetric sedimentary history in the basin. The results provide estimates of the erosion history of the island and portions of potential sediment loss, with implications about the factors that could affect erosion and weathering intensities in tectonically inactive regions.

2 Research background and data

2.1 Uplift history of Madagascar

The landscape of present-day Madagascar has been built by significant rock uplift accumulated over a long history. The Mesozoic uplift and exhumation histories of Madagascar were recorded by low-temperature thermochronology. The oldest apatite fission-track (AFT) cooling ages are in the range of ~ 440 – 200 Ma, and occur on the central plateau to the south of the Cenozoic volcanoes (Itasy and Ankaratra; Figure 1b) (Seward et al., 2004; Emmel et al., 2006, 2008; Jöns et al., 2009; Emmel et al., 2012; Stephenson et al., 2021). These ages indicate that since the Paleozoic, the total exhumation of this part of the plateau was less than the closure depth of the AFT system (~ 4 km). During the Late Cretaceous, the final break-up between Madagascar and the Seychelles-India block was possibly triggered by the passage of the Marion hot spot (Storey et al., 1995; Torsvik et al.,

1998). Mantle upwelling at the hot spot could have elevated the southeast of the island and tilted the island towards the northwest. Compared to the central plateau, AFT ages are younger (~119–68 Ma) along the east escarpment, but it is unclear whether these ages were thermally reset by the heat from the plume, or they reflect erosion of the uplifted rifting shoulder (Seward et al., 2004). Apatite (U-Th)/He cooling ages <60 Ma are also found on the coastal plain at the foot of the escarpment along the east coast of the island (Emmel et al., 2012), suggesting a possible retreat of the escarpment following the island’s rifting from India.

The uplift history of Madagascar during the Cenozoic was constrained by the current elevations of the carbonate platforms and geometries of river long profiles. Shallow marine Eocene sediments occur in some locations on the plateau surface at an elevation of 900 m (Analavelona Plateau near Toliara; Figure 1b) in the southwest and at 300–500 m at the northernmost tip of the island (Roig et al., 2012; Stephenson et al., 2019), suggesting that the island has been uplifted significantly during the Cenozoic. On the central plateau, dating of the overlying sediments suggests a period of uplift of the central island between the Late Eocene (prior to 34 Ma) and the Miocene-Pliocene boundary (5.5 Ma) (Bésairie, 1973). Based on the inversion of river profiles across the entire island, Roberts et al. (2012) inferred a 1–2 km-high dome-shaped tectonic uplift during the past ~15 Ma. The centers of the inferred doming spatially overlap the Late Cenozoic volcanic fields (Emerick and Duncan, 1982; Nougier et al., 1986; Cucciniello et al., 2011), suggesting that the high topography of central Madagascar may be dynamically supported by upwelling in the sub-lithospheric mantle (Roberts et al., 2012). Recent seismic tomographic studies (Pratt et al., 2017; Celli et al., 2020) have imaged a low-velocity zones at 50–150 km depths beneath the north, central, and southwestern regions of the island, supporting the hypothesis that asthenospheric upwelling is the source of the Late Cenozoic volcanism and uplift of Madagascar. Earthquake focal mechanisms and campaign and continuous GPS data from the island indicate an ongoing, east-west extension of the crust (Grimison and Chen, 1988; Bertil and Marc Regnault, 1998; Rindraharisaona et al., 2013; Saria et al., 2014; Stamps et al., 2018, 2021), suggesting the island being on a diffuse, extensional plate boundary as the southern extension of the East African Rift (Horner-Johnson et al., 2007; Kusky et al., 2010; Stamps et al., 2021).

2.2 Planation surfaces

We use planation surfaces as constraints on the history of uplift on the island. Planation surfaces are large areas of flat or gently tilted bedrock surfaces, which form as products of regional-scale erosional events (King, 1953; Orme, 2013; Guillocheau et al., 2018). In non-glacial settings, a planation surface may be created by the degradation of a drainage divide, the backwearing of an escarpment, or the wave-cutting of a marine platform. Although on the southeastern Tibetan Plateau, low-relief surfaces were also suggested to have formed *in situ* due to drainage area loss during dynamic

113 drainage network reorganization (Yang et al., 2015), such a mechanism does not apply to planation
114 surfaces widely preserved in continental interiors with no intense tectonic deformation, e.g., in
115 Australia, Central and Southern Africa, India, and both North and South America. In these
116 regions, high-elevation, low-relief plateaus formed due to relatively slow and long-wavelength uplift,
117 which may result from buckling of the lithosphere due to far-field forces (Cloetingh and Burov,
118 2011) or from upwelling currents in the sub-lithosphere mantle (Braun, 2010).

119 Planation surfaces found in Madagascar are comparable to those identified in Africa, which
120 can be classified in two main types called pediplains (or pediments) and etchplains (Guillocheau
121 et al., 2018). Their formation mechanisms differ in whether chemical weathering played a significant
122 role or not. Pediplains (large in area) and pediments (smaller features) are mainly produced by
123 physical erosion, most likely by large-scale sheet floods or highly avulsionary rivers (Dohrenwend
124 and Parsons, 2009). Conversely, on top of an etchplain, strong chemical weathering associated with
125 hot and humid climate leads to the formation of a thick lateritic mantle, which can be partially
126 eroded later (Twidale, 2002). Overall, regardless of whether chemical weathering was involved, a
127 planation surface is generally considered to originate at base level and only preserved following a
128 period of uplift or base level drop. In Madagascar, considering the relative small area of the island
129 (compared to, e.g., the continent of Africa), it is reasonable to assume the sea level as the base level
130 of erosion.

131 In Madagascar, the planation surfaces appear as flat or slightly warped small patches or large
132 platforms (Figure 2), dissected by steeply incised river valleys. Assuming a relative stationary
133 base level over geological time scale, the current elevation of a surface records the total uplift
134 accumulated since its formation, i.e., the time when it was abandoned as the erosional base level.
135 Moreover, vertical offsets between two surfaces record the incremental uplift magnitude over a period
136 that is bounded by the ages of the two surfaces. Throughout the island, Delaunay (2018) mapped
137 planation surfaces formed at nine different ages, based on interpreting their spatial relationship
138 from field observations and a 90-m SRTM digital elevation model. In some places, these surfaces
139 are covered by sedimentary layers or volcanic deposits, the ages of which can be used as constraints
140 of the minimum ages of the underlying planation surfaces. Based on these age constraints and the
141 regional geological history, Delaunay (2018) proposed an age model for the planation surfaces in
142 Madagascar, as summarized below.

143 The oldest two surfaces (S_i and S_{ii}) are preserved only at very high altitudes in the center
144 of the island, with ages older than the separation of the island from India (Dixey, 1960; Bésairie,
145 1973); these two planation surfaces are not used in our model. After Madagascar became an
146 island, the oldest planation surface (S_{iia}) formed at ~ 90 Ma, and the next surface (S_{iib}) formed
147 after 84 Ma. Since the beginning of the Cenozoic, Madagascar has experienced a long history of
148 tectonic quiescence, with neither uplift nor planation surfaces formation until the late Eocene. In
149 the Cenozoic, the oldest planation surface (S_{iv}) formed between 38 and 31 Ma, concurrent with

150 the uplift of the carbonate platforms around the island (Roig et al., 2012; Stephenson et al., 2019).
151 Subsequently, two planation surfaces (S_{va} and S_{vb}) formed during the Miocene, 23–12 and 12–5
152 Ma, respectively. After Miocene, the planation surface S_{vi} formed at or after 3.6 Ma. The youngest
153 planation surface on the island formed since the Quaternary, associated with the most recent phase
154 of the uplift of the island.

155 **2.3 Regional climate history**

156 The relative intensity between chemical weathering and physical erosion is key to understanding
157 the long-term evolution of planation surfaces and hence the evolution of the Malagasy topography.
158 As shown by Gaillardet et al. (1995) for a modern system over a very short period, under humid
159 climate (i.e., Congo River over one year), close to 40% of the eroded volumes are due to chemical
160 weathering. According to the data compiled from 60 largest rivers in the world (Gaillardet et al.,
161 1999), the rock weathering rate is strongly dependent on the climate conditions (i.e., precipitation
162 and temperature). Other factors affecting the weathering rate include lithology and physical erosion
163 rate.

164 The climate history of Madagascar is still under debate and past studies have been based on
165 pollen analysis (e.g., Wells, 2003; Buerki et al., 2013) or numerical models (Ohba et al., 2016).
166 During the Late Cretaceous Madagascar was located between 30 and 40°S (van Hinsbergen et al.,
167 2015) in a zone of arid climate (Wells, 2003; Blakey, 2008). During the latest Cretaceous (85–65 Ma),
168 a transition to a very humid environment was recorded by changing vegetation types (Takhtajania
169 perrieri and Dilobeia; Marquinez et al., 2009; Buerki et al., 2013), and was recently confirmed in
170 wells offshore Mozambique (Ponte et al., 2019). Warm and humid conditions continued to prevail
171 through the Paleocene and Eocene, as the island moved northwards. According to the fossil records
172 from southern Africa, the climate changed to a more arid setting (Braun et al., 2014; Ponte et al.,
173 2019) at the beginning of the Oligocene, and back to very humid conditions at the mid-Oligocene
174 and humid conditions at the Oligo-Miocene boundary. Since the Miocene, the island has presented
175 a climate partitioning between the relatively arid side to the west and humid to the east. During
176 the middle Miocene (since ~16 Ma), the island overall was in a semi-arid condition, but since the
177 late Miocene (~11 Ma) alternation between dryer and wetter conditions started and continued to
178 the present.

179 **2.4 Sediment volume data**

180 We use the volume data of the siliciclastic sediment from the Morondava Basin (Delaunay, 2018) as
181 a main constraint for the landscape evolution models. The Morondava Basin is located southwest
182 of Madagascar (Figure 1a). Along its western margin, this deep marine basin is bounded by a
183 prominent bathymetric high, the Davie Ridge, which is commonly regarded as a transform boundary

184 formed during the break-up between Africa and Madagascar (e.g., Mahanjane, 2014); the latter was
185 still part of Antarctica at that time. As a consequence, the basin has stored most of the terrigenous
186 sediment that was shed off the southwest coast of Madagascar, first during the continental rifting
187 from Late Carboniferous to Early Jurassic (Wescott and Diggens, 1997, 1998; Piqué et al., 1999;
188 Geiger et al., 2004), and then in a passive margin setting since the Middle Jurassic (Geiger et al.,
189 2004; Geiger and Schweigert, 2006).

190 Delaunay (2018) estimated the incremental thickness and volume of sediment in the offshore
191 and onshore parts of the Morondava Basin, between the Archean and Proterozoic basement to the
192 present-day seafloor/surface. We use his estimates for the time period comprised between ~ 90 Ma
193 and the present as model constraints, as older sediment deposits were disrupted by the rifting pro-
194 cess and only partially preserved. According to Delaunay (2018), the total volume of terrigenous
195 siliciclastic sediments between ~ 90 and 0 Ma is estimated at $(131.7 \pm 29.4) \times 10^3$ km³ (Table 1).
196 Sediments accumulated predominantly during two periods, one in the Late Cretaceous (~ 90 –66
197 Ma) and the other in the Late Cenozoic (~ 34 –0 Ma). From ~ 90 to 66 Ma, the average volumetric
198 sedimentation rate was $(1.3 \pm 0.2) \times 10^3$ km³/Ma, and the sediments were mainly siliciclastic. During
199 the Paleocene and Eocene (~ 66 –34 Ma), clastic sedimentation was negligible. Late Cenozoic sed-
200 imentation started in the Oligocene, with a gentle increase in sedimentation rate during the early
201 Miocene. A marked acceleration of the sedimentation occurred during the mid-Miocene, when the
202 volumetric sedimentation rate jumped from $(1.8 \pm 0.4) \times 10^3$ to $(5.5 \pm 1.2) \times 10^3$ km³/Ma. Since the
203 Pliocene, the sedimentation rate has decreased slightly to $(4.8 \pm 1.0) \times 10^3$ km³/Ma.

204 3 Landscape evolution modeling

205 3.1 Forward model

206 To simulate the landscape evolution, we assume that the erosion of Madagascar was dominated
207 by fluvial erosion and hillslope processes. Instantaneous elevation change is computed by using an
208 adjusted version (Davy and Lague, 2009) of the stream-power law (e.g., Howard, 1994; Whipple
209 and Tucker, 1999) and a linear hillslope diffusion (Culling, 1960),

$$\frac{\partial h}{\partial t} = U - \max(0, K_f A^m (\nabla h)^n - \dot{e}_c) + K_d \nabla^2 h \quad (1)$$

210 where U is the uplift rate, h is the elevation, t is the time, A is the upstream drainage area, ∇h
211 is the local relief. K_f is the erosion coefficient whose value is dependent on factors such as rock
212 erodibility, and varies over a few orders of magnitude; we will optimize its value through the inverse
213 modeling process (Section 3.3). Exponents in the erosion law (m and n) dictate the degrees of
214 dependence of the predicted erosion rates on topographic parameters such as upstream drainage

215 area and local channel slope. The value of K_d is strongly dependent on the spatial resolution of the
216 model, and the ratio between K_f and K_d also determines the relative efficiency between long- and
217 short-distance processes as well as the resultant density of river networks. For the efficiency of our
218 inverse modeling, we fix the values of m , n , and K_d at 0.4, 1, and 0.1, respectively. These are within
219 the ranges of commonly estimated or assumed values (m/n , 0.35–0.6; n , 0–4; K_d , 0.01–10). A recent
220 study in southern Africa (Stanley et al., 2021) followed similar assumptions, and our adopted values
221 were among those used to simulate the landscape evolution of passive margins (Braun, 2018). In
222 addition, our initial experiments show that different values of m and n will affect the estimated
223 values of K_f , but have insignificant impacts on the sediment volumes predicted by the optimized
224 models. In addition to the classical stream-power law, we follow Davy and Lague (2009) and add a
225 threshold erosion rate, \dot{e}_c , such that fluvial erosion is only effective where the shear stress is above
226 a threshold. The value of \dot{e}_c is also constrained by inversion. We use the FastScape algorithm
227 (Braun and Willett, 2013) and the alternating direction implicit method (Peaceman and Rachford,
228 Jr., 1955) to solve the stream-power erosion and the hillslope diffusion equations, respectively. Both
229 methods are implicit and therefore stable for modeling landscape evolution using relatively large
230 time steps.

231 3.2 Uplift model

232 We prescribe the uplift functions based on the vertical offsets between the planation surfaces mapped
233 by Delaunay (2018), through the following steps: 1) we divide the model space into a 16×32 grid,
234 and calculate the median elevation of the planation surfaces within each cell (of area 50×50 km²);
235 2) using the median elevations, we calculate the offset between each planation surface and the
236 subsequent surface (Figure 3), and use it as the proxy for the accumulated uplift during the period
237 between the times when the two surfaces formed; 3) we interpolate and extrapolate the estimated
238 uplift functions to the 400×800 model grid (of area 2×2 km²); 4) finally, we smooth the uplift
239 functions and filter out uplift signals with < 500 km wavelength (Figure 4). Based on ranges of
240 formation ages of the planation surfaces constrained in Delaunay (2018), we arbitrarily assign ages
241 of the seven planation surfaces, i.e., *iiia*, *iiib*, *iv*, *va*, *vb*, *vi*, and *vii*, at 90 Ma, 84 Ma, 34 Ma, 11 Ma,
242 5 Ma, 3 Ma, and 2.6 Ma, respectively; the Mesozoic uplift is set to stop at 66 Ma, corresponding
243 to the onset age of minimal sedimentation in the Morondava Basin (Piqué et al., 1999). We
244 acknowledge the existing uncertainties in the age constraints of the planation surfaces. However,
245 given the long time intervals (5 to > 30 Ma) over which the sediment volumes were estimated, we
246 argue that the inverse modeling directed by these estimates is not sensitive to the variations in ages
247 of the planation surfaces within the inferred constraints.

248 Our model of the landscape evolution of Madagascar starts in the late Cretaceous at 90 Ma,
249 when the planation surface *iiia* was abandoned as the erosion base level. This surface formed

250 after the break-up between Madagascar and India, and therefore recorded the oldest uplift history
251 of the island since its separation from other continents. At this time, the continent of western
252 Madagascar along the Majunga and Morondova Basins was submerged. The remainder of the island
253 areas presented as extensive, low-elevation plains directly connected to the ocean. An uncertainty
254 exists for the topography of the eastern margin of the island, where following the Madagascar-India
255 rifting, remnants of the continental shoulders could remain or have been degraded. Therefore, we
256 use two parameters to prescribe the initial topography of the model: H_0 is the elevation along
257 the western margin of the island and H_1 is that along the eastern margin, and a linear, eastward
258 increase is imposed for elevations between the two margins. Both H_0 and H_1 will be optimized in
259 the inversion (Section 3.3). The initial topography is imposed on a 400×800 grid and covers the
260 area of the continental lithosphere of Madagascar.

261 Each step of the landscape evolution model is 100 ka-long and correspondingly a total of 900
262 steps is simulated. We divide the model time into seven episodes of uplift (Figure 4), and each
263 episode is bounded by ages of two subsequent planation surfaces; the only exception is the second
264 episode, which starts at 84 Ma (age of the surface *iiib*) but terminates at the end of Cretaceous
265 (66 Ma). During each episode, the uplift rate of a given location is set to remain constant, i.e., the
266 vertical offset between planation surfaces divided by their age difference.

267 3.3 Inverse modeling

268 We use the Neighbourhood Algorithm (Sambridge, 1999) to optimize the parameters in the erosion
269 laws (K_f and \dot{e}_c in Equation 1), the elevations on the western and eastern margins of the initial
270 topography, and the ratios of sediment loss (see below in this section). The inversion method is
271 proven to be an efficient scheme for constraining the values of unknown parameters in a multi-
272 dimensional space (Sambridge, 1999). During each iteration, the parameter space is uniformly
273 divided into Voronoi cells defined by combinations of parameter values, and models in centers of
274 the cells are assessed using defined misfit functions. The subsequent iteration repeats the process
275 but within a reduced parameter space that is defined by models with lowest misfits from previous
276 iteration. Optimization of the forward model is achieved by performing a large number of iterations.
277 The sampling ranges and results for parameters in our inversion are listed in Table 2.

278 The inversion is constrained by the present-day topography and the sediment volume data
279 from the Morondava Basin. As the evolution of a natural landscape is a stochastic process, it is
280 unrealistic to reproduce the exact natural topography with the LEM. We choose to minimize the
281 difference between the cumulative density functions (CDFs) of the elevations between the present-
282 day Madagascar and the final topography predicted by the LEM.

283 As a portion of the eroded materials may be consumed by chemical weathering, to fit the
284 sediment volume data we impose a sediment loss ratio to represent the portion of the erosion

285 products that were not preserved in the offshore basin. Assuming a strong dependence of chemical
286 weathering intensity on climate (Gaillardet et al., 1999), we assign an independent sediment loss
287 ratio for each of the seven climate period (Section 2.3), i.e., c_1 to c_7 for 90–85 Ma, 85–34 Ma,
288 34–29 Ma, 29–23 Ma, 23–16 Ma, 16–11 Ma, and 11–0 Ma, respectively. In the inverse modeling, all
289 ratios are searched in the range of 0–0.5. Our approach ignores recycling of sediment, and therefore
290 sometimes could potentially underestimate eroded volume. However, the recycling of sediments
291 should be restricted to the relatively small areas in the lowlands along the west coast, and we
292 therefore consider sediment recycling as a secondary process.

293 During the inversion, we compute the misfits in elevation and sediment volume data separately,
294 and then combine them using the equation

$$\phi = aD + b \sum_{i=1}^N \log\left(\frac{v_{oi} - v_{pi}}{\delta_i}\right)^2 \quad (2)$$

295 where D is the Kolmogorov-Smirnov (K-S) statistics between the CDFs of natural and predicted
296 elevations, v_o and v_p and observed and predicted sediment volumes, respectively, and N is the
297 number of sediment volume data points (or time intervals). Two scaling coefficients, a and b , are
298 used to calculate a combined misfit. We tested the values of a and b by trial and error, attempting
299 to keep the two components of misfit at the same order of magnitude; both parameters were finally
300 prescribed at 1. The inversion presented here comprises a sum of 200 iterations. The first iteration
301 includes 1,000 forward runs, and each of the following iterations 250. We use a resampling ratio
302 of 0.8, i.e., a random walk in 200 out of the 250 cells from the previous run, to reach a balance
303 between convergence rate and a reasonable exploration of the parameter space.

304 4 Results

305 The inversion processes showed good convergence for all sampled parameters except H_0 . The
306 performance of the modeling is presented here as frequencies of the sampled sediment loss ratios
307 (c_1 – c_7) in the seven climate periods (Figure 5) and projections of a thinned (20%) assemblage of
308 sampled erosion law parameters onto 2D planes (Figure 6). We also show the estimated density
309 functions for the exhumation rates at three locations in the north, center, and south of the island,
310 in comparison with their uplift functions prescribed at the respective sites (Figure 7).

311 The modeling results suggest that throughout the island, the predicted erosion rates are mostly
312 in consistent with the rock uplift rates, until at least the late Miocene (<10 Ma) (Figure 7). During
313 the Late Cretaceous, the island experienced slow erosion with rates most likely <0.07 mm/yr; large
314 uncertainties for the initial period (90–85 Ma) of the model exist, likely due to the uncertainty
315 in the initial topography of the model (Figure 7). The Late Cretaceous uplift and erosion ceased

316 prior to ~ 50 Ma. The Cenozoic erosion started in response to the rejuvenated uplift of the island
317 near the Eocene–Oligocene boundary, but the rates had remained low (Figure 7). For all parts
318 of the island, uplift rates increased in the late Miocene (~ 12 Ma). However, associated with this
319 event the models predict no significant increase in the erosion rates, except in the north where the
320 models suggest a delayed feedback of erosion rate increase. A more marked acceleration in uplift
321 rates occurred at 5 Ma, when the rates at the selected sites increased from <0.05 to >0.2 km/Ma
322 (Figure 7). In response to this event, substantial increases in erosion rates are predicted throughout
323 the island. By the end of the model time (with the exception of the north of the island) erosion
324 rates have not adapted to the same magnitudes as the most rapid uplift rates prescribed for the
325 Pliocene period (Figure 7).

326 Figure 8 shows the topographic history predicted by a best-fit model. The initial topography
327 of the model is made of a low-relief surface gently tilted to the west. During the Late Cretaceous,
328 the model predicts an increase in onshore area associated with the uplift of the island. Since the
329 Cenozoic, the onshore area has had a similar shape as the present-day island, but the elevation
330 remained low (<500 m) prior to 10 Ma. During the last ~ 10 Ma, the modeled elevation of the
331 island increased significantly. However, the best-fit model cannot reproduce some details from
332 the landscape of the modern Madagascar, e.g., distribution of the low-relief central plateau and
333 the asymmetry in the drainage areas between basins draining to the west and those to the east
334 (Figure 1a), which indicates the potential limitation of the current model (Section 5.1).

335 5 Discussion

336 5.1 Limitations of the modeling

337 It is possible that during the early Cenozoic, Madagascar experienced a period of subsidence, as
338 evidenced by the carbonate platform preserved along the west coast. This event could decrease the
339 land area and hence reduce the sediment flux into the Morondava Basin. As the planation surfaces
340 could not record the magnitude of the surface downwarping, we did not implement any subsidence
341 (i.e., negative uplift) in the modeling. Instead, we prescribed a period of zero uplift rate, which
342 effectively resulted in a time span with minimum erosion and sediment flux (Figures 9a and 11a).
343 As the net uplift for a time period is decided by the offset between planation surfaces, ignoring the
344 subsidence could lead to potential underestimates of the uplift magnitude and rate over the period
345 immediately following the subsidence, i.e., 34–12 Ma. However, with the current setup our model
346 agrees well with the uplift history of the island inferred from river profile inversion (Roberts et al.,
347 2012; Stephenson et al., 2021), in which the average uplift rates were low (~ 0.01 km/Ma) before
348 20 Ma and rapid uplift did not start until mid- to late Miocene. Therefore, we suggest that ignoring
349 the early Cenozoic subsidence is a reasonable simplification of the uplift model.

350 The modeling results do not fully replicate the modern topography of Madagascar (Figure 9b).
351 Particularly, the numerical model cannot reproduce the bimodal distribution of the elevation data,
352 which represents an extensive, elevated low-relief plateau in the center of the island and extensive
353 plains connected to the coast (Figure 1a). We suspect that this discrepancy is at least in part due
354 to the limitation of the forward erosion model: a combination of the fluvial incision and hillslope
355 diffusion cannot fully reproduce the high-elevation, low-relief surface. Adding a threshold on the
356 stream power could help preserve some of the relict surfaces (e.g., Stanley et al., 2021). However, as
357 the physical process of planation (e.g., sheet floods) is not simulated by the LEM, the models cannot
358 generate an expansive pediplain or etchplain at the base level, and thus the low-relief surfaces also
359 do not occur extensively on the high plateau on the final landscape (Figure 8). This imperfection
360 may be improved by incorporating into the forward model additional components that simulates
361 more surface processes, such as chemical weathering (Braun et al., 2016) or sediment deposition
362 (Yuan et al., 2019). However, in order to focus on the relationship between uplift and sedimentation
363 and reduce the dimension of the inverse problem, we opted for a simple LEM rather than a complete
364 recovery of the natural landscape.

365 Our misfit function based on the K-S statistics of the DEM (Equation 2) is constructed with an
366 aim to reconstruct the general range of the elevations of the island, and thus is insensitive to the
367 pixel-wise shape of the topography. Therefore, specific features of the landscape, such as drainage
368 basin geometry and location of the escarpment on the east coast, are not monitored during the
369 inversion. As our uplift function is prescribed and initial topography is very simple, such detailed
370 features of the topography are almost impossible to reproduce without imposing spatial variability
371 in the rock erodibility or the precipitation intensity.

372 5.2 Uplift and erosion model

373 The fluvial incision coefficient was sampled in the logarithmic space, and $\log(K_f)$ converged to the
374 range of -5.5--5 (Figure 6a), which are within the range of values commonly adopted by landscape
375 evolution studies (e.g., Whipple and Tucker, 1999). The logarithm of the erosion threshold, $\log(\dot{\epsilon}_c)$,
376 is constrained to the range of -5--4.5 (Figure 6a). The optimized values of the two parameters
377 show an apparent positive correlation. This can be expected, as we impose the sediment yield and
378 therefore the mean erosion rate of the landscape; thus, a higher threshold for the minimal erosion
379 rate in the headwaters or low-relief regions requires to be balanced by higher efficiency in the fluvial
380 incision rate in lower reaches or steep sections of rivers. The same correlation between the fluvial
381 incision efficiency and erosion threshold was also observed in the inverse modeling presented in
382 Stanley et al. (2021), which were used to inspect the uplift and erosion history of southern Africa.
383 Moreover, the optimized value of K_f is most consistent with the results ($\log(K_f) \approx -5.5$) of the
384 “hybrid early scenario” model of Stanley et al. (2021), which in addition to the Cretaceous uplift

385 involves a phase of early Cenozoic uplift due to the mantle process; the similar, two phases of uplift
386 history are also used in our models. Therefore, if the same erodibility can be assumed for the
387 bedrocks from southern Africa and Madagascar (both predominantly Precambrian basement), our
388 results support a consistency in the uplift and erosion scenarios between models of southern Africa
389 and Madagascar. This general synchronization implies that the geodynamic processes responsible
390 for the Cenozoic uplift of the two regions were connected, perhaps both subject to the development
391 of the East African Rift system (Ebinger and Sleep, 1998; Cucciniello et al., 2018).

392 By inverting the geometry of long river profiles, Roberts et al. (2012) and Stephenson et al. (2021)
393 estimated the fluvial erosion coefficient K at $2 \times 10^{-4} \text{m}^{0.6} \text{yr}^{-1}$ and $4 \times 10^{-6} \text{m}^{0.3} \text{yr}^{-1}$, respectively.
394 Based on their modeling results, both Roberts et al. (2012) and Stephenson et al. (2021) suggest
395 a period of Cenozoic uplift to construct the high plateau of Madagascar. Based on basin averaged
396 erosion rates estimated from cosmogenic ^{10}Be concentration, Wang et al. (2021) calibrated the
397 erosion coefficient K at $2 \times 10^{-6} \text{m}^{0.3} \text{yr}^{-1}$ ($\text{m}/\text{n}=0.35$) or $4 \times 10^{-7} \text{m}^{0.1} \text{yr}^{-1}$ ($\text{m}/\text{n}=0.45$). The
398 low erosion rate inferred from ^{10}Be is consistent with a model of continuous retreat of the great
399 escarpment on the east coast since the Cretaceous rifting of the India-Madagascar margin, without
400 necessity for a renewed phase of uplift during the Cenozoic. The estimated K value of our best-fit
401 model is about $6 \times 10^{-6} \text{m}^{0.2} \text{yr}^{-1}$, lower than the calibration of Roberts et al. (2012) but higher
402 than Stephenson et al. (2021) and Wang et al. (2021). Considering that our erosion model includes
403 a threshold for minimum erosion, it is reasonable for our estimates to be slightly higher; we also
404 demonstrate that the estimate of K value tends to reduce with a decreasing erosion threshold $\dot{\epsilon}_c$.
405 The uplift function in our model is extracted from the distribution of planation surfaces of Delaunay
406 (2018), which also invoke a Cenozoic period of uplift, consistent with inferences of Roberts et al.
407 (2012) and Stephenson et al. (2021). However, an observation emerges from our 2D planview model:
408 if spatial variability in the uplift rates existed, dynamic reorganization of the river networks would
409 have been active near the main drainage divide, including captures of stream channels between the
410 west- and east-draining rivers. Therefore, as pointed out by Wang et al. (2021), assumptions of
411 stationary upstream drainage areas, which the inversion of river long profiles are based on, could
412 be problematic. On the other hand, as the misfit function in our inverse modeling is insensitive to
413 the geometry of the drainage basin, our results are not suitable for negotiating between the models
414 of Cenozoic uplift and low erosion rates on the east coast.

415 Nevertheless, our models suggest that along the west coast and on the central plateau of Mada-
416 gascar, the total magnitudes of exhumation since the India-Madagascar rifting have been low, and
417 mostly occurred after the Miocene. By assuming a near-surface geothermal gradient of $\sim 25^\circ\text{C}/\text{km}$,
418 we used a 1D heat diffusion model to calculate the cooling paths of rocks in the north and center
419 of the island. The results show that these rocks were at temperatures $< 50^\circ\text{C}$ since the Late Creta-
420 ceous (Figure 10). These predictions are consistent with the thermal history models derived from
421 apatite fission-track and (U-Th-Sm)/He data (Stephenson et al., 2021), confirming that the post-

422 rifting magnitudes of exhumation on the central plateau of Madagascar could not “reset” the apatite
423 low-temperature thermochronology systems (Seward et al., 2004; Emmel et al., 2012; Stephenson
424 et al., 2021).

425 **5.3 Volume balance between erosion and sedimentation**

426 The results of our inverse modeling of the landscape evolution allow a semi-quantitative assessment
427 of the sediment balance between the source and sink. In southern Africa, Rouby et al. (2009)
428 documented a good volume balance between the terrigenous sediments in the basins along the
429 west coast and the onshore erosion estimated from low-temperature thermochronology for the past
430 150 Ma, suggesting that over geological time scales the volume data can effectively reflect the evo-
431 lution of the sediment provenance. Using the same proxies along the south coast of South Africa ,
432 Tinker et al. (2008) found much lower volumes of offshore sedimentation than onshore denudation,
433 suggesting a substantial portion of the erosion products being removed from the system. Stanley
434 et al. (2021) constrained landscape evolution models of the southern African plateau using topo-
435 graphic, sediment flux, and low-temperature thermochronological data, and found that the erosion
436 volumes predicted by the best-fit models are significantly lower than the observation. After careful
437 consideration of the possible factors, they attributed most of the sediment loss to chemical weath-
438 ering though some materials might also be transported out of the system by tectonic (transform
439 faulting and rifting), oceanic (deep and surface currents), or eolian processes.

440 Our best-fit models suggest that most of the sediment loss occurred from the Late Cretaceous to
441 Eocene and since the Pliocene; during the two periods, inversions of the sediment loss ratios both
442 converge to the upper boundaries of the sampling spaces (Figure 5). The high sediment loss ratio
443 during late Cretaceous is consistent with the best-fit models of Stanley et al. (2021), which predict
444 that only one third of the erosion products accumulated in the basin. The high weathering intensity
445 is supported by the very humid climate conditions from Late Cretaceous to Eocene recorded both
446 globally (e.g., Pearson et al., 2001; Jenkyns et al., 2004; Cramwinckel et al., 2018) and in southern
447 Africa (Braun et al., 2014; Ponte et al., 2019), as the weathering rate is strongly dependent on the
448 precipitation rate and temperature (e.g., White and Blum, 1995; Murphy et al., 2016). In central
449 and southern Africa, extensive weathering surfaces seem to form during the Cretaceous and Eocene
450 (Guillocheau et al., 2018; De Putter and Ruffet, 2020). In Madagascar, similar planation surfaces
451 with weathering profiles also formed during the early Cenozoic until Eocene (Delaunay, 2018). For
452 the period since the Pliocene, the modeled high sediment loss ratio could also be a consequence of
453 intense weathering. In the north of Madagascar, a 12-meter thick weathering profile is observed
454 at the surface of a late Cenozoic igneous complex (Estrade et al., 2019). In central Africa, the
455 youngest peaks in the age distribution of oxidized manganese ores occur from the late Miocene to
456 the Quaternary. In the modern Congo River basin, the chemical weathering is estimated at 40% of

457 the total budget of erosion (Gaillardet et al., 1995).

458 Despite the oscillating humidity of the climate model between Oligocene and mid-Miocene (34–
459 11 Ma), our models predict generally low sediment loss ratios for this period (Figure 11c). The
460 lack of variability in the modeled ratios is primarily due to the coarse temporal resolutions in both
461 sediment volume data and imposed uplift function, and thus cannot be interpreted in further details
462 regarding the geological or climate histories.

463 **6 Conclusions**

464 We presented a case study that establishes a source-to-sink balance in the sediment volumes between
465 the onshore erosion and offshore sedimentation, based on constraining a landscape evolution model
466 using empirical data. With a focus on the landscape of Madagascar since its separation from
467 India, we modeled the uplift and erosion histories of the island for the past 90 Ma. The uplift
468 function imposed in our models is based on the distribution and heights of the planation surfaces
469 mapped from the island, and the estimated ages of these surfaces. Parameters in the erosion
470 models are either prescribed to commonly used values or constrained by inverse modeling. Data
471 used for constraining the inverse modeling include the cumulative density function of the elevation
472 of the current topography and the incremental volume estimates of sediments from the Morondava
473 Basin off the west coast since the Late Cretaceous. During the inversion, we also examined the
474 potential temporal variation in balance between erosion and sedimentation, by optimizing the ratios
475 of sediment loss for periods of different climate. The results show that the reconciliation between the
476 uplift history and sediment volume data requires a significant fraction of the erosion products being
477 moved out of the source-to-sink system. The models suggest relatively high sediment loss ratios
478 during the warm and humid periods from Late Cretaceous to Eocene and from the late Miocene
479 to the present-day, implying chemical weathering as an important mechanism for consuming the
480 sediments in regions with no strong tectonic deformation.

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482 Data used in this paper are present in Table 1 and available from Delaunay (2018) [[https://](https://tel.archives-ouvertes.fr/tel-01865476)
483 tel.archives-ouvertes.fr/tel-01865476]. The inverse modeling was performed on the Linux
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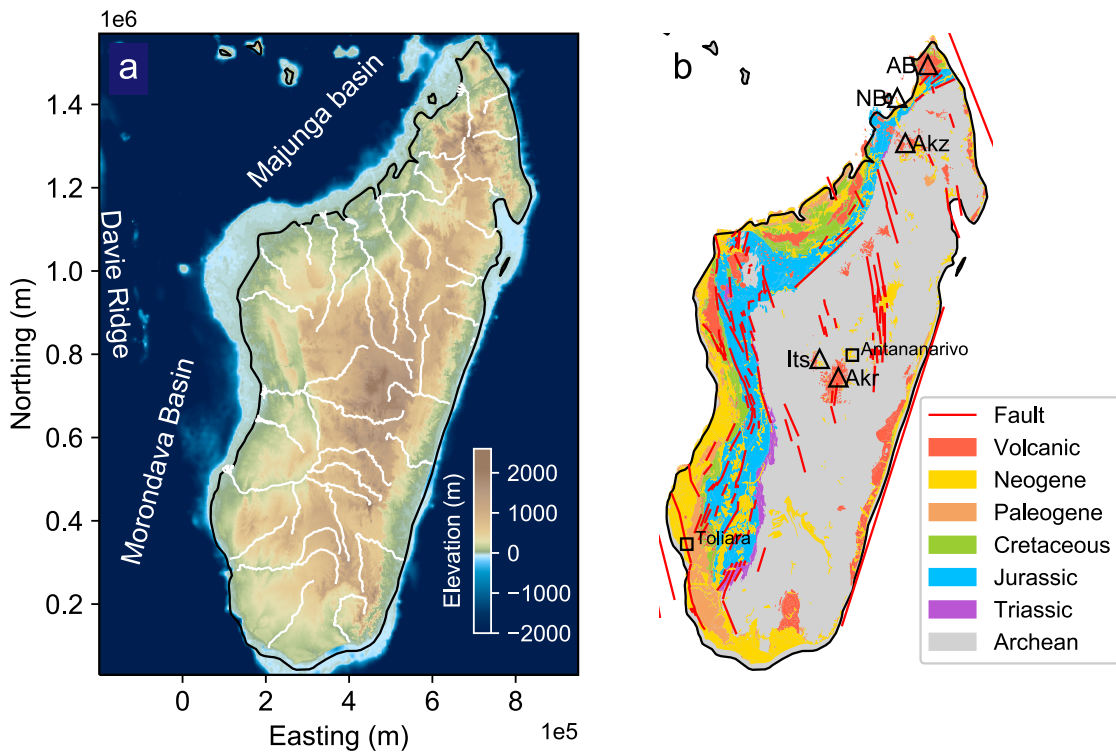


Figure 1: (a) Topography and bathymetry of Madagascar. White lines indicate the courses of major rivers. (b) Simplified geology of Madagascar (after Roig et al., 2012). Triangles indicate Quaternary volcanoes: AB, Ambre-Bobaomby; NB, Nosy-Be; Akz, Ankaizina; Its, Itasy; Akr, Ankaratra.

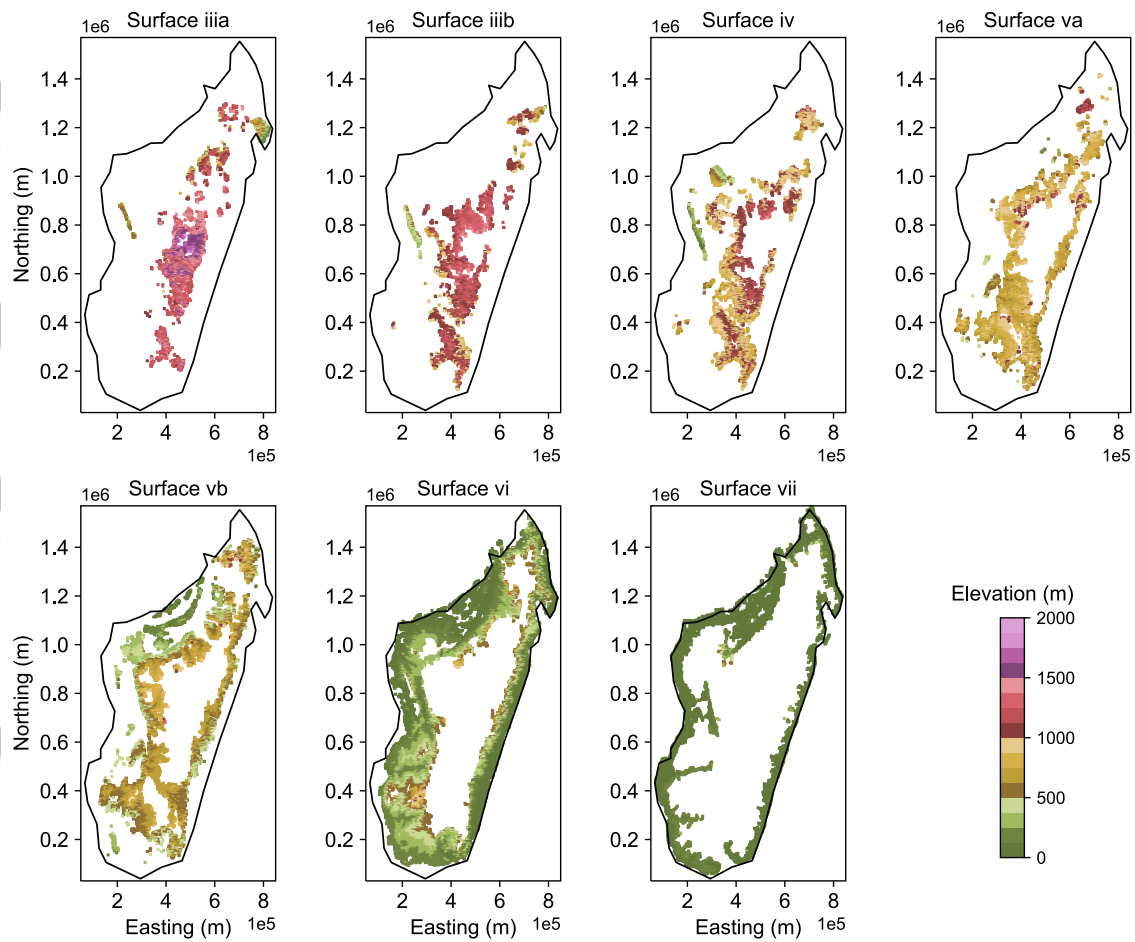


Figure 2: Remnants of pediment surfaces in Madagascar mapped by Delaunay (2018). Only surfaces formed after 90 Ma are shown. Formation ages of surfaces were estimated at (iiiia) 90–84 Ma, (iiib) 84–66 Ma, (iv) 41–29 Ma, (va) ~11 Ma, (vb) ~5 Ma, (vi) <3.6 Ma, and (vii) the Quaternary, and prescribed in the our model at 90 Ma, 84 Ma, 34 Ma, 11 Ma, 5 Ma, 3 Ma, and 2.6 Ma, respectively.

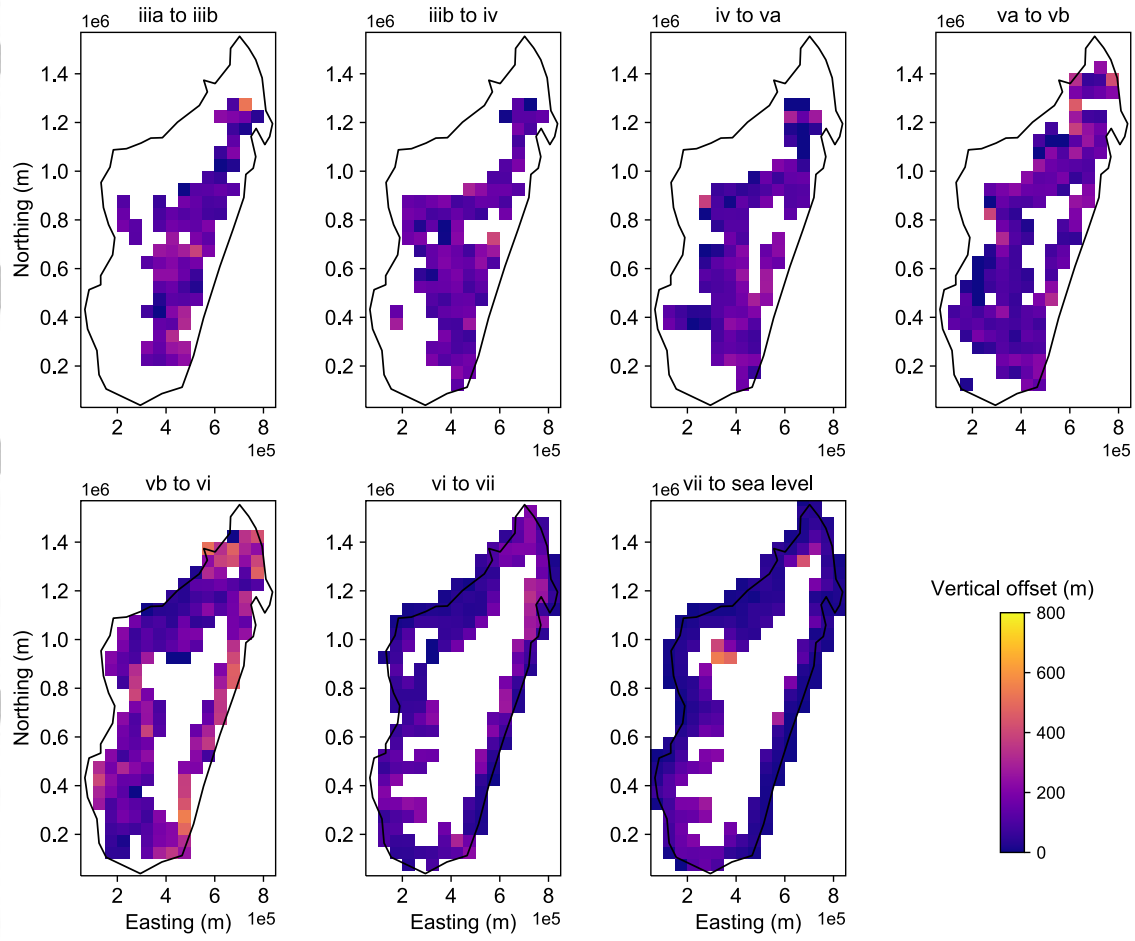


Figure 3: Vertical offsets between two planation surfaces. The offsets are calculated on a 16×32 grid of $50 \times 50 \text{ km}^2$ cells using the median elevation of the surfaces within each cells.

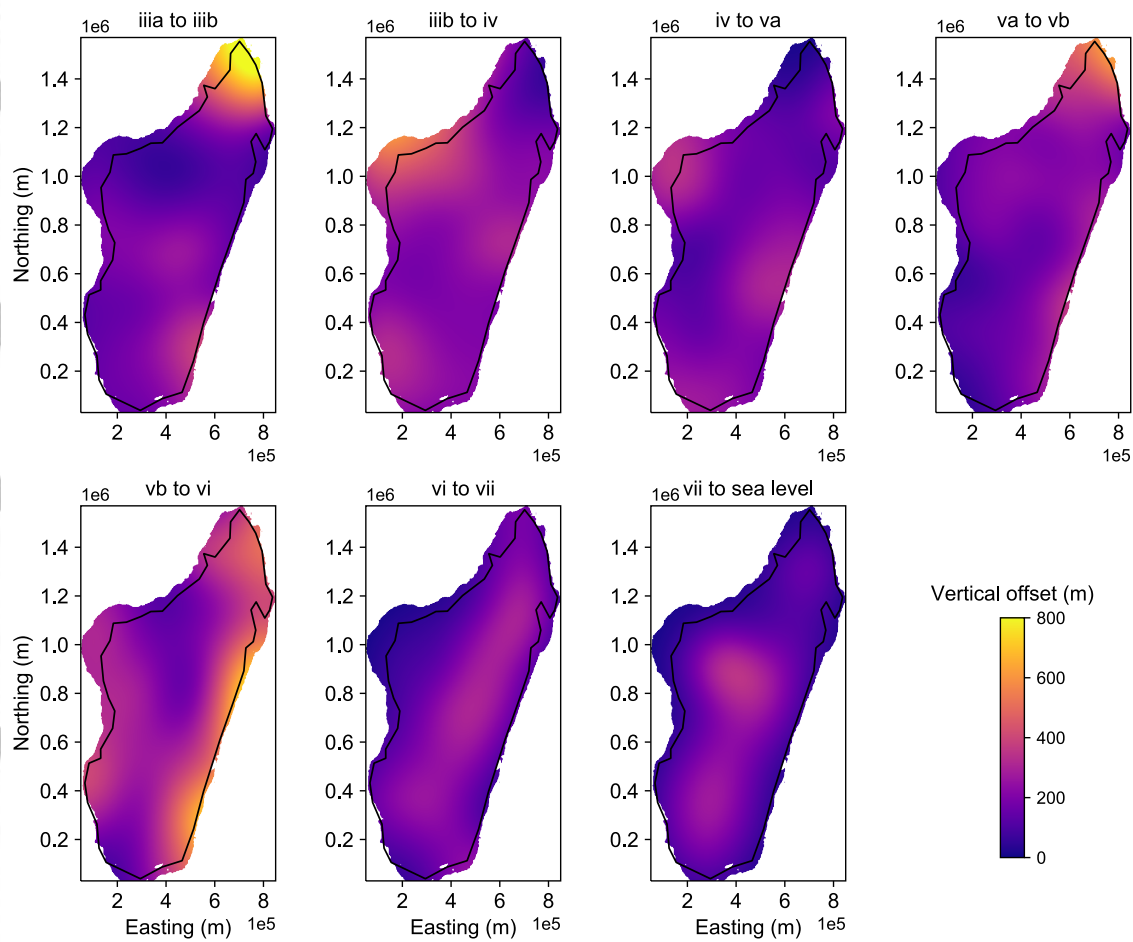


Figure 4: Interpolated and extrapolated uplift magnitudes between two planation surfaces. The uplifts are calculated on a 400×800 grid of 2×2 km² cells and used in the landscape evolution model as uplift magnitudes accumulated for the episodes of 90–84 Ma, 84–66 Ma, 34–11 Ma, 11–5 Ma, 5–3 Ma, 3–2.6 Ma, and 2.6–0 Ma, respectively.

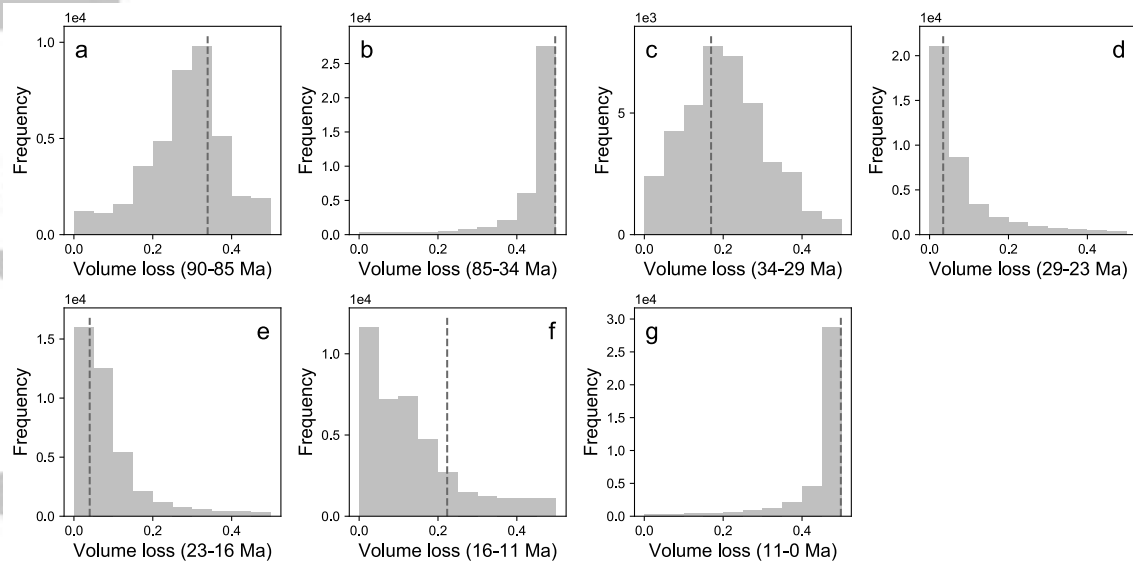


Figure 5: Inverse modeling results of the sediment volume loss. Distributions of sampled values are shown for prescribed climate periods of (a) 90–85 Ma, (b) 85–34 Ma, (c) 34–29 Ma, (d) 29–23 Ma, (e) 23–16 Ma, (f) 16–11 Ma, and (g) 11–0 Ma, respectively. Dashed vertical lines indicate the values of the best-fit model.

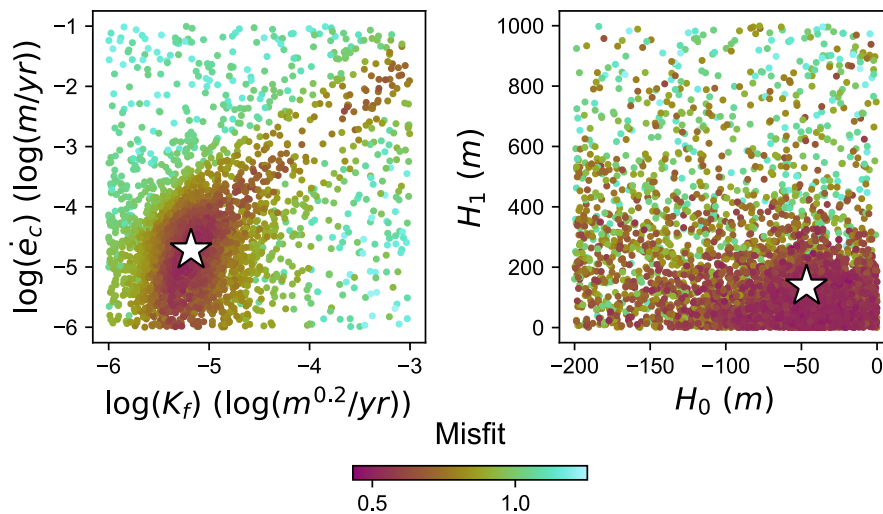


Figure 6: Inverse modeling results of K_f , \dot{e}_c , H_0 , and H_1 . Results are shown as a thinned (20%) ensemble of sampled forward models projected onto planes defined by pairs of parameters. Color indicates the misfit of a forward model, and star depicts the “best-fit” model.

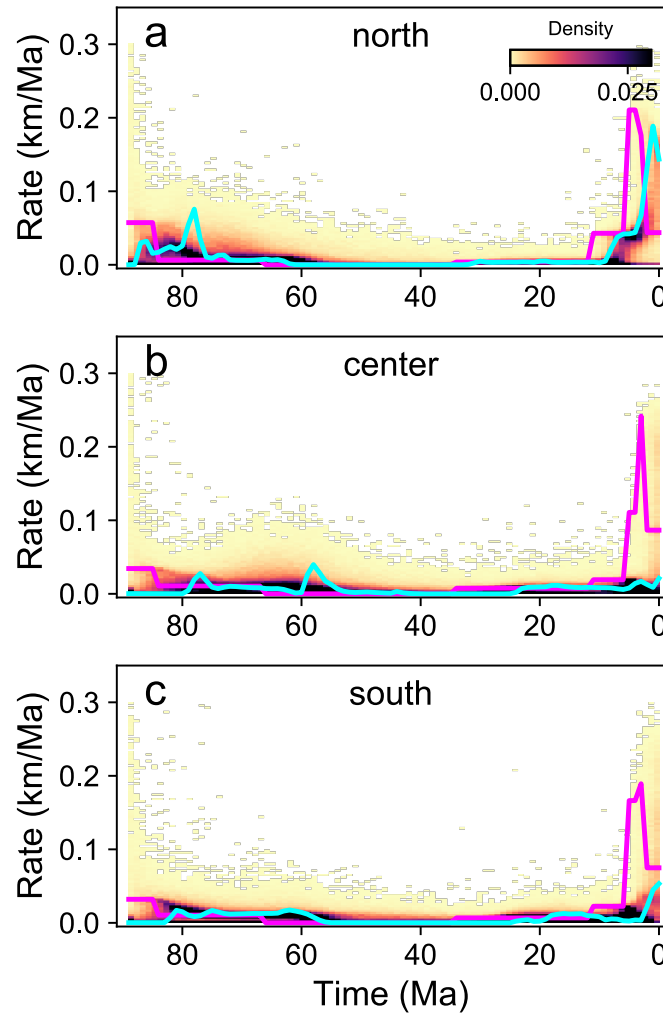


Figure 7: Rock uplift and predicted erosion rates at three representative sites in the (a) north, (b) center, and (c) south of Madagascar. Site locations are shown in Figure 8. Magenta lines depict the uplift model. Modeled erosion rates are shown as probability density functions calculated from a thinned assemblage (20%) of the sampled forward models; cyan lines depict erosion rates of the best-fit model.

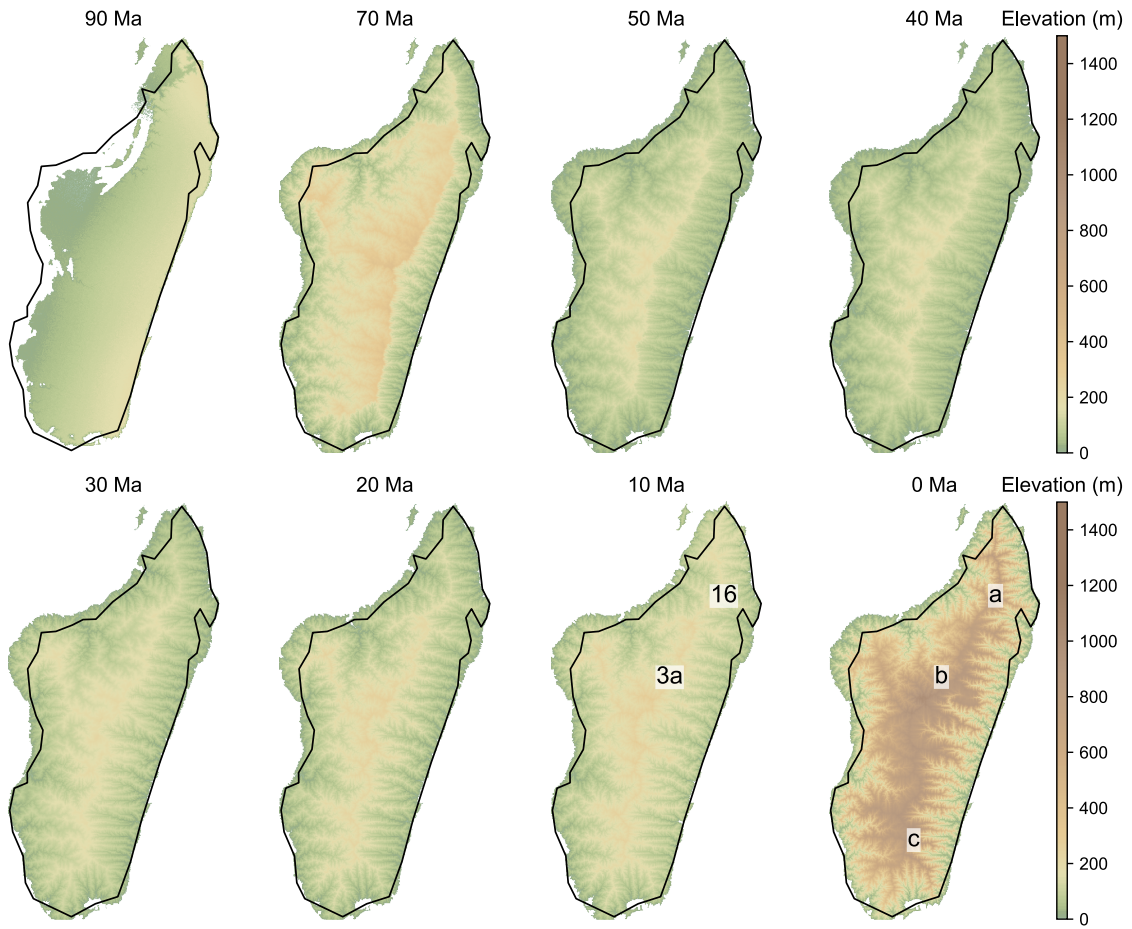


Figure 8: Evolution of the topography predicted by the best-fit model. Sites a, b, and c indicate locations where uplift and erosion histories are calculated (Figure 7). Sites 3a and 16 are locations where rock cooling models are calculated (Figure 10).

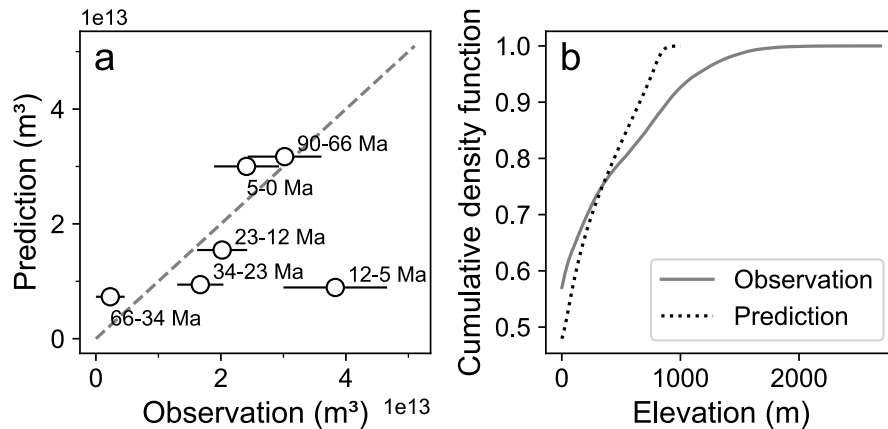


Figure 9: Predictions of the best-fit model compared to observations. (a) Observed versus predicted sediment volumes in the Morondava Basin. (b) Cumulative density functions (CDFs) of the present-day topography (observation) and the DEM at the final stage of the landscape evolution model (prediction).

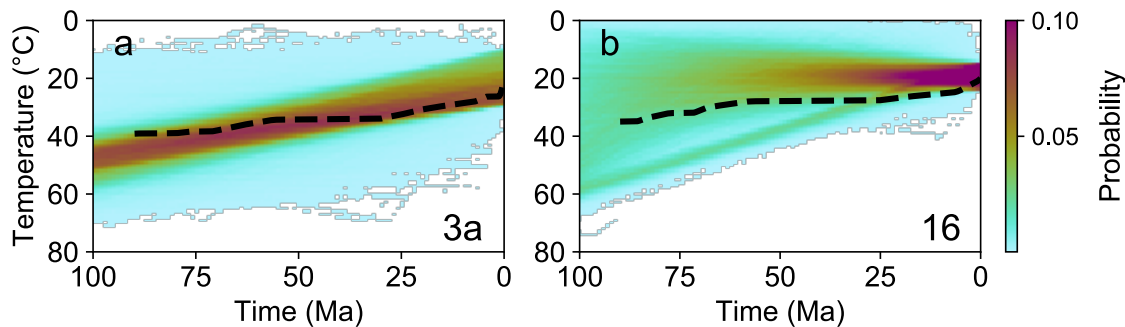


Figure 10: Rock cooling histories predicted by the best-fit landscape evolution model (dashed line) compared to results from inversion of thermochronological data (heat map). Dashed lines depict the prediction of the landscape evolution model, assuming a surface temperature at 20°C and a near-surface geothermal gradient at 25°C/km. Thermal history models reported by Stephenson et al. (2021) are reproduced, using the same data, approach, and setup. The modeling method is Bayesian transdimensional Markov Chain Monte Carlo (Gallagher, 2012) and results are shown as the probability density function of the 400,000 post-burn-in models following the initial 100,000 burn-in. (a) and (b) present results using apatite fission-track and (U-Th-Sm)/He data of two samples 3a (a) and 16 (b) in Stephenson et al. (2021). Locations of samples are shown in Figure 8.

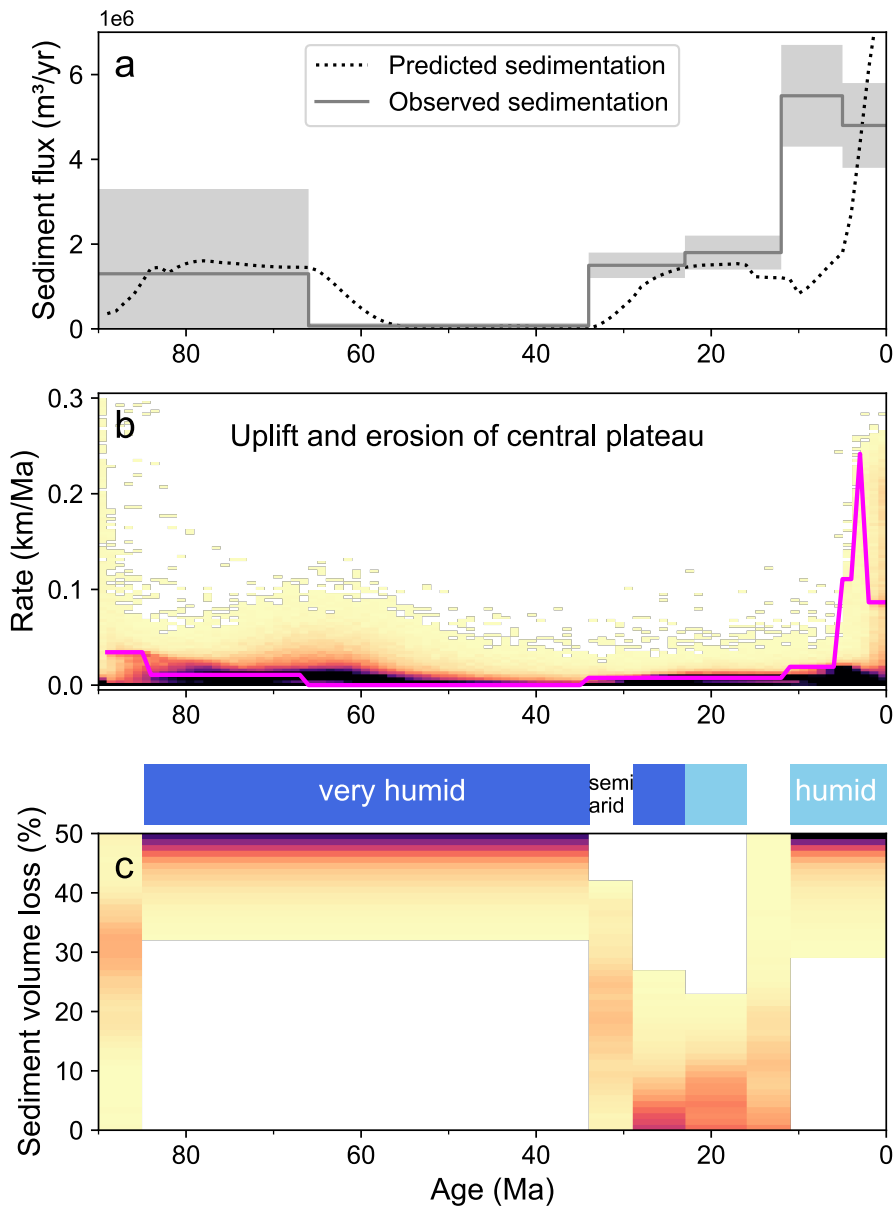


Figure 11: Summary of the uplift, predicted erosion and sedimentation histories in Madagascar. (a) Observed versus predicted sediment fluxes into the Morondava Basin. Grey shading indicates the uncertainty (σ) of the estimates. (b) Calculated uplift and modeled erosion rates on the central plateau of the island, same as Figure 7b. (c) Predicted volume fraction of sediment loss from the total erosion. Color scale indicates the probability density function calculated from a thinned ensemble (20%) of the sampled models. Dark blue, light blue, and white bars on top of the figure indicate the very humid, humid, and semiarid climate settings during geological times. Sediment loss ratio is imposed to be invariant within each climate period.

Table 1: Sediment volume in the Morondava Basin since ~ 90 Ma Delaunay (2018)

Time interval (Ma)	Volume ($\times 10^{12} m^3$)	σ ($\times 10^{12} m^3$)
90–66	30.2	5.9
66–34	2.3	2.3
34–23	16.7	3.7
23–12	20.2	4.0
12–5	38.3	8.3
5–0	24.1	5.2

Table 2: Model parameters sampled in inversion

Parameter	Code	Range	Unit	Best-fit
Logarithm of erosion coefficient	$\log(K_f)$	-6–-3	$\log(m^{0.2}yr^{-1})$	-5.195
Logarithm of erosion rate threshold	$\log(\dot{e}_c)$	-6–-1	$\log(m/yr)$	-4.819
Initial elevation in the west	H_0	-200–0	m	-32
Initial elevation in the east	H_1	0–1,000	m	28
Sediment loss ratio (90–85 Ma)	c_1	0–0.5		0.34
Sediment loss ratio (85–34 Ma)	c_2	0–0.5		0.49
Sediment loss ratio (34–29 Ma)	c_3	0–0.5		0.19
Sediment loss ratio (29–23 Ma)	c_4	0–0.5		0.02
Sediment loss ratio (23–16 Ma)	c_5	0–0.5		0.04
Sediment loss ratio (16–11 Ma)	c_6	0–0.5		0.15
Sediment loss ratio (11–0 Ma)	c_7	0–0.5		0.50