



HAL
open science

Surface uplift and topographic rejuvenation of a tectonically inactive range: Insights from the Anti-Atlas and the Siroua Massif (Morocco)

R. Clementucci, P. Ballato, Lionel Siame, M. Fox, R. Lanari, A. Sembroni, C. Faccenna, A. Yaaqoub, A. Essaifi

► To cite this version:

R. Clementucci, P. Ballato, Lionel Siame, M. Fox, R. Lanari, et al.. Surface uplift and topographic rejuvenation of a tectonically inactive range: Insights from the Anti-Atlas and the Siroua Massif (Morocco). *Tectonics*, 2023, 42 (2), pp.e2022TC007383. 10.1029/2022TC007383 . hal-03966926

HAL Id: hal-03966926

<https://hal.science/hal-03966926>

Submitted on 1 Feb 2023

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 Surface uplift and topographic rejuvenation of a tectonically inactive range: Insights
2 from the Anti-Atlas and the Siroua Massif (Morocco)

3 Clementucci R.^{1,2*}, Ballato P.¹, Siame L.², Fox M.³, Lanari R.^{1,4}, Sembroni A.¹, Faccenna C.^{1,5},
4 Yaaqoub A.⁶, Essaifi A.⁶

5 ¹ *Dipartimento di Scienze, Università Roma Tre, Largo San Leonardo Murialdo 1, 00146 Rome, Italy.*

6 ² *Aix-Marseille Univ., CNRS, IRD, INRAE, Collège de France, CEREGE, Aix-en Provence, France.*

7 ³ *Department of Earth Sciences, University College London, Gower Street, London, WC1E 6BT, United
8 Kingdom.*

9 ⁴ *Dipartimento di scienze, Università di Firenze, Italia.*

10 ⁵ *GFZ-German Research Centre for Geosciences, Potsdam, Germany.*

11 ⁶ *Département de Géologie, FSSM, B.P. 2390, Université' Cadi Ayyad, Marrakech, Morocco.*

12 *Corresponding author: romano.clementucci@uniroma3.it*

13

14 **Key Points:**

- 15 - The transient state of the stream network and the high-standing erosional surface document a
16 recent phase of topographic rejuvenation
17 - Uplift started at ~14-10 Ma and increases from the western (0.5 km) to the central Anti-Atlas
18 (1.1 km) and the Siroua Massif (1.5 km)
19 - Regional surface uplift is most likely due to astenospheric upwelling, locally combined
20 (Siroua Massif) with magma injection and faulting

21

22 **Abstract**

23 The Atlas-Meseta intracontinental orographic system of Morocco experienced recent, large-scale
24 surface uplift as documented by elevated late Miocene, shallow-water marine deposits exposed in the
25 Middle Atlas Mountains. The Anti-Atlas Mountains do not present any stratigraphic records that
26 document regional vertical movements, however, the presence of a high-standing, erosional surface,
27 and the transient state of the river network, provides insights into the uplift history of the belt and the
28 mechanisms that drove it. Here, we combine geomorphic and stream profiles analyses, celerity of
29 knickpoints and linear inverse landscape modelling with available geological evidence, to decipher
30 the spatial and temporal variations of surface uplift in the Anti-Atlas and the Siroua Massif. Our
31 results highlight the presence of a transient landscape and document a long wave-length topographic
32 swell (~ 100 x 600 km) with a maximum surface uplift of ~1500 m in the Siroua Massif and ~1100

33 m in the central Anti-Atlas most likely starting from ~14-10 Ma. Surface uplift occurred in association
34 with the onset of late Miocene magmatism in the Siroua and Saghro Massif and contractional
35 deformation in the High Atlas. Regional surface uplift was most likely due to deep-seated mechanism,
36 such as asthenospheric upwelling. Additional processes such as magma injection and faulting
37 contributed to the surface uplift of the Siroua Massif. Overall, our approach allows to quantitatively
38 constrain the transient state of the landscape and the contribution of regional surface uplift on
39 mountain building processes.

40

41 **1. Introduction**

42 The interplay between exogenic and endogenic processes controls the growth and decay of
43 topography over diverse temporal and spatial scales. Topographic development is generated by
44 different mechanisms, such as crustal shortening and thickening, magmatic addition, flexural
45 rebound, and asthenospheric flow, where each process is associated with a typical wavelength (10^{-1}
46 to 10^3 km) and amplitude (10 to 10^3 m; *e.g.*, England and Molnar, 1990; De Celles et al., 2009;
47 Roberts and White, 2010; Faccenna and Becker, 2020). Topographic decay results from geomorphic
48 erosion, which is primarily controlled by climate, topographic gradients, and rock strength (*i.e.*,
49 Jansen et al., 2010; Scharf et al., 2013; Clementucci et al., 2022). When topographic decay prevails,
50 extensive low-gradient, low-topographic relief landscapes may form (Baldwin et al., 2003; Tucker
51 and van der Beek, 2013). These geomorphic low-relief features can be also found at the summit of
52 mountains areas (Miller et al., 2013; West et al., 2013; Calvet et al., 2015), where they have been
53 interpreted to have formed either in situ at high elevation (*e.g.*, Yang et al., 2015; Fox et al., 2020),
54 or to be part of a less elevated and extensive low-relief landscape that has undergone recent surface
55 uplift (*e.g.*, Calvet et al., 2015; Olivetti et al., 2012; 2016; Whipple et al., 2017). In the case of surface
56 uplift, such elevated paleo-topography, named relict landscape or paleo-surface, represents a transient
57 geomorphic marker that provides a reference frame for quantifying the spatial distribution of surface
58 uplift (Calvet et al., 2015; Fox, 2019). Thus, relict landscapes represent a unique opportunity to

59 decipher the wavelength and the amplitude of surface uplift, and consequently the mechanisms that
60 created such a topography, especially in areas devoid of stratigraphic constraints.

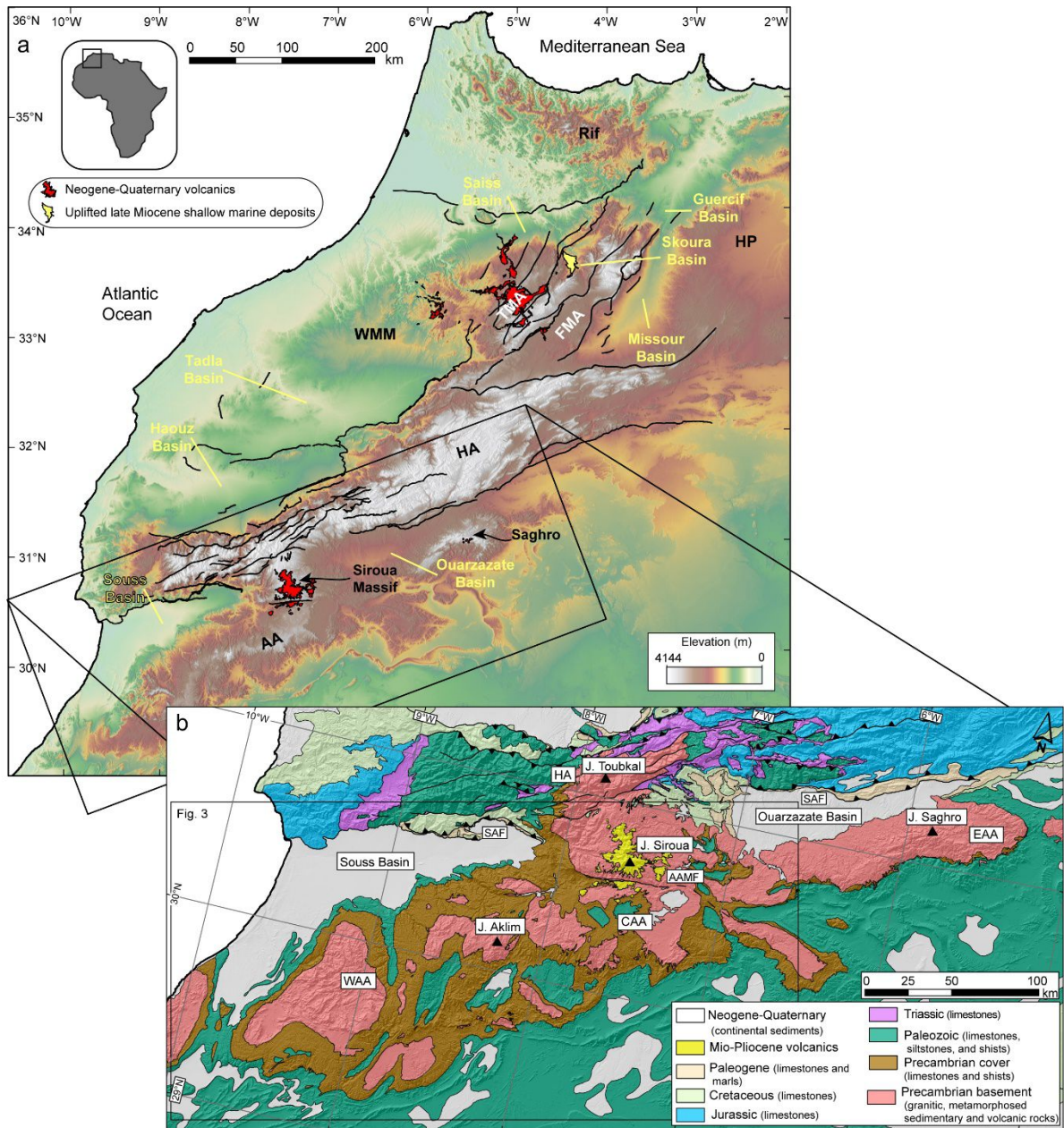
61 The Atlas-Meseta intracontinental orographic system of Morocco is composed of different morpho-
62 structural domains that from north to south include: the Moroccan Meseta (Western Meseta and
63 Eastern High Plateaus), the Middle Atlas, the High Atlas, the Siroua Massif, and the Anti-Atlas
64 (Figure 1a). The High and Middle Atlas are the main active orogen (*e.g.*, Sebrier et al., 2006), while
65 the other regions are characterized by a relatively high topography that formed without significant
66 crustal deformation (Teixell et al., 2003; Frizon de Lamotte et al., 2009; Pastor et al., 2015). This
67 configuration offers the opportunity to investigate rates, wavelength, magnitudes and causes of large-
68 scale topographic growth.

69 In this study, we focus on the Anti-Atlas Mountains, a Variscan orogen characterized by a complex
70 assemblage of intrusive, metamorphic, and sedimentary rocks. Similarly to the Appalachian, which
71 represent the American counterpart of the Variscan orogens in Africa and Europe, the topography of
72 the Anti-Atlas consists of a wide relict landscape composed of high-standing erosional surfaces
73 mostly sculptured on basement rocks (Choubert, 1952; Malusà et al., 2007; Guimera et al., 2011;
74 Clementucci et al., 2022). This basement has been slowly eroding at few tens of meters per million
75 years since the latest Cretaceous as documented by low-temperature thermochronology (Lanari et al.,
76 2020a; Clementucci et al., 2022). Overall, the occurrence of such a low-relief features at the
77 maximum topography together with the lack of Cenozoic deformation, supports the idea that the Anti-
78 Atlas experienced a relatively recent phase of surface uplift (Missenard et al., 2006). The causes of
79 this topographic rejuvenation, however, are still poorly understood. Some authors suggested that
80 uplift may be related to Cenozoic shortening (Giumera et al., 2011), whereas others proposed the
81 occurrence of deep dynamic processes (Missenard et al., 2006; Sebrier et al., 2006; Frizon de Lamotte
82 et al., 2009). In this context, the absence of tectonic activity, the steady protracted low denudation
83 rates, and the occurrence of an elevated relict landscape offer the possibility to quantify the regional

84 surface uplift in the Anti-Atlas and the Siroua Massif and to investigate possible causal uplift
85 mechanisms.

86 To achieve these goals, we combined quantitative analysis of regional and basin-scale topography
87 with stream profiles analysis (knickpoints characterization, longitudinal profiles and χ - z plot) and
88 recently published basin-wide denudation rates inferred from ^{10}Be concentrations (Clementucci et al.,
89 2022). Furthermore, we performed stream projections from the upper relict channel profiles and
90 reconstructed the paleo-landscape to constrain the amount of relative base level fall following the
91 increase in Cenozoic rock uplift rates. Finally, we constrained the timing of this geomorphic
92 rejuvenation to the middle-late Miocene by applying a knickpoints celerity model. This approach
93 provides a framework for discussing the contribution of large-scale, deep signals and shallower
94 processes in generating the regional topographic relief in areas where large wavelength uplift occurs.

95



96

97 **Figure 1.** (a) Topographic map of the Atlas-Meseta orographic system and his morpho-structural domains.
 98 FMA: Folded Middle Atlas. TMA: Tabular Middle Atlas. HP: High Plateaus. WMM: Western Moroccan
 99 Meseta. AA : Anti-Atlas. HA : High Atlas. (b) Simplified geological map of High and Anti-Atlas Mountains
 100 based on the 1:1.000.000 geological map of Morocco (modified from Maroc Service Géologique, 1985). EAA:
 101 Eastern Anti-Atlas, CAA: Central Anti-Atlas, WAA: Western Anti-Atlas, AAMF: Anti-Atlas Major Fault,
 102 SAF: South Atlas Fault. The main faults and mountain peaks are shown with black lines and triangles,
 103 respectively. The black lines with triangles indicate the thrust faults.

104

105

106

107

108 **2. Geological background**

109 2.1 **Geological setting**

110 The intracontinental Atlas-Meseta orographic system of Morocco is composed of two WSW-ENE
111 oriented mountain belts, the High Atlas (HA) and the Anti-Atlas (AA), and two elevated regions with
112 low-topographic relief, named Western and Eastern (or High Plateaus) Meseta, which are separated
113 by the SW-NE striking Middle Atlas belt (Figure 1a, TMA and FMA). This system of mountain belts
114 experienced a complex geologic history including multiple episodes of contractional deformation and
115 rifting (Froitzheim et al., 1988; Hafid, 2000; Frizon de Lamotte et al. 2000; Domènech et al., 2015).
116 Currently, active deformation is controlled by the plate convergence between Africa and Eurasia
117 (Froitzheim et al., 1988; Hafid, 2000; Frizon de Lamotte et al. 2000; Gomez et al. 2000; Teixell et
118 al., 2003; Lanari et al., 2020b).

119 With a few mountain peaks above 4 km of elevation, the HA is the highest domain of the Atlas system
120 that strikes for approximately 600 km over a width of 50 to 100 km. The HA consists of a fold and
121 thrust belt composed of Precambrian, Paleozoic, and Mesozoic rocks (Figure 1). Mesozoic rocks
122 include mostly syn-rift (Jurassic and Triassic) and post-rift deposits (Cretaceous), related to the
123 opening of a rift basin developed during the breakup of Africa (Arboleya et al., 2004; Baudon et al.,
124 2009; Lanari et al., 2020b). Based on thermochronological and structural data the main exhumation
125 episode started in the middle/late Miocene and led to 4-6 km of exhumation (Balestrieri et al., 2009;
126 Barbero et al., 2007; Domènech, 2015; 2016; Ghorbal, 2009; Lanari et al., 2020a; 2020b; Leprêtre et
127 al., 2018; Missenard et al., 2008).

128 The HA and AA belts are separated by the Souss and the Ouarzazate basins, which have been
129 commonly interpreted as the southern Neogene foreland basins of the HA (Figure 1; Sebrier et al.,
130 2006; Arboleya et al., 2008). The Souss Basin has an eastward wedging geometry and is characterized
131 by a longitudinal, E-W oriented, drainage system flowing directly into the Atlantic Ocean (Figure
132 2a). The basin extends for a length of more than 150 km and has a rather flat, low-topographic relief
133 morphology that increases progressively toward the basin interior up to ~700 m of elevation. The

134 basin fill includes Pliocene to Quaternary fluvial, fluvio-lacustrine and aeolian sediments derived
135 from the adjacent uplifting HA and AA belts (Hssaine and Bridgland, 2009).

136 The Ouarzazate Basin lies between 1200 and 1800 m of elevation and extends over a length of more
137 than 150 km (Figure 1). The basin filling consists of a <1-km-thick-succession including Middle
138 Eocene marine deposits unconformably overlain by Mio-Pliocene terrestrial sediments marking the
139 development of endorheic conditions (Fraissinet et al. 1988; Gorler et al. 1988; El Harfi et al., 2001;
140 Teson and Teixell 2006; Teson et al., 2010). The basin is currently drained by the Draa River a
141 southward draining system that crosses the AA through a narrow bedrock gorge and then flows
142 parallel to the AA before entering in the Atlantic Ocean (Figure 2a). The transition from endorheic to
143 exoreic conditions most likely occurred in the Plio-Pleistocene, either through regressive erosion from
144 the southern AA flank (Stablein, 1988) or lake overspill (Arboleya et al., 2008).

145 The Siroua Massif represents a strato-volcano covering an area of approximately of 500 km² and
146 reaching maximum elevation of ~3300 m. Here, the Mio-Pliocene volcanic deposits directly overlie
147 the Precambrian basement, which is characterized by an elongated, dome-like morphology with a
148 low-topographic relief landscape (Missenard et al., 2008; Giumera et al. 2011).

149 The AA belt extends over approximately 600 km of length, has a width of 100 to 150 km, and rises
150 ~2 km above sea level. The orogen was built through a series of collisional events starting from the
151 Eburnean orogeny at around 2000 Ma (Ait Malek et al., 1998; Thomas et al., 2002; Walsh et al.,
152 2002). Afterward, it experienced two main orogenic events: the Neoproterozoic Pan-African orogeny
153 (Leblanc, 1975; Hefferan et al., 2000; Gasquet et al., 2008) with a subsequent extensional event
154 during the Late Neoproterozoic (Piqué et al., 1999; Soulimani et al., 2003) and the late Carboniferous
155 Variscan orogeny (Sebti et al., 2009; Soulimani et al., 2014; Sehart et al., 2018). Low-temperature
156 thermochronology and stratigraphic data indicate that the AA was not subjected to major Mesozoic
157 vertical movements, while denudation appears to have been rather uniform at rates of ~20 m/Myr
158 since at least the late Cretaceous (Lanari et al., 2020a; Charton et al., 2021; Clementucci et al., 2022).

159 The basement is composed of magmatic, metamorphic, and sedimentary rocks assembled during the

160 Eburnean and Pan-African orogenesis. These rocks are widely exposed along the axial sectors of the
161 range (Figures 1b and S1 in the supporting information) and are overlain by a late Precambrian and
162 Paleozoic sedimentary sequence with a thickness of almost 10 km in the Western AA decreasing to
163 less than 6 km in the eastern sectors of the belt (Piqué and Michard, 1989; Helg et al., 2004; Burkhard
164 et al., 2006). This sedimentary cover was predominantly deposited in a shallow-water marine
165 environment during the post Pan-African extensional event (Azizi Samir et al., 1990; Thomas et al.,
166 2002; Soulaimani et al., 2003), and includes Lower Cambrian carbonates, siltstones, and marls as
167 well as Middle Cambrian to Middle Devonian sandstones and shales. These strata were subsequently
168 deformed during the Variscan orogeny and unconformably covered by marine (mainly Cretaceous)
169 sediments (Figure 1a). The eastern Anti-Atlas includes another elevated (~2700 m) volcanic field, the
170 Saghro Massif. (Figure 1a). This is composed of multiple late Miocene tuff cones which, likewise the
171 Siroua Massif, overlie a wide erosional surface sculptured on the Precambrian basement.

172

173 **2.2 Geophysical and petrological data**

174 The peculiarity of the Atlas-Meseta orographic system is the lack of orogenic roots generated through
175 crustal shortening and thickening processes that would isostatically support the modern topography.
176 This applies to the HA (Beauchamp et al., 1999; Gomez et al., 2000; Teixell et al., 2003; Domènech
177 et al., 2016; Fekkak et al., 2018; Lanari et al., 2020b) and the MA (Gomez et al., 1998; Arboleya et
178 al., 2004; Pastor et al., 2015), where tectonic shortening has occurred during the Cenozoic, and to the
179 tectonically inactive AA, Western and Eastern Meseta (Babault et al., 2008; Frizon de Lamotte et al.,
180 2009; Pastor et al., 2015). Several geophysical and petrological studies indicate an anomalous
181 thinning of the lithosphere with a shallow, hot asthenosphere beneath the entire Atlas-Meseta
182 orographic system (El Azzouzi et al., 1999; Missenard et al., 2006; Duggen et al., 2009; Miller et al.,
183 2015). This includes geophysical modelling of gravity data, S-received function (Ayarza et al., 2005;
184 Missenard et al., 2006; Miller and Becker, 2014; Miller et al., 2015), heat flow measurements, deep
185 electrical resistivity (Rimi et al., 1999; Zeyen et al., 2005) and seismic tomography (Seber et al.,

186 1996; Palomeras et al., 2014; Bezada et al., 2014). The shallow asthenosphere is thought to support
187 the present-day topography of the Atlas-Meseta system, which should have been generated, at least
188 in part, by deep-seated, mantle driven processes. For example, a maximum of 2000 m of residual
189 topography has been estimated in the HA and MA considering a crustal thickness of ca. 35 km (Miller
190 and Becker, 2014), while more than 1000 m of elevation has been attributed to dynamic topography
191 (Frizon de Lamotte et al., 2009; Spieker et al., 2014).

192 Magmatic activity in the Siroua, Saghro and the MA Mountains presents an alkaline affinity
193 indicating a partial melting of sublithospheric mantle caused by heating generated by asthenospheric
194 flow (El Azzouzi et al., 1999; 2010; De Beer et al., 2000; Missenard et al., 2006). The geochemical
195 signature and trace element patterns of the MA lavas are very similar to those observed in the Canary
196 Islands, suggesting a possible link between the Canary mantle plume and continental intraplate
197 volcanism in the Atlas-Meseta system (Duggen et al., 2009). Interestingly, magmatism occurred in
198 two different pulses, in the Eocene and middle-late Miocene in association with renewed tectonic
199 activity. A more recent phase of magmatism, however, has occurred in the MA and Western Meseta
200 during the last 2 Ma, apparently without major acceleration in tectonic deformation (El Azzouzi et
201 al., 2010; Missenard et al., 2012).

202

203 **3. Methods**

204 **3.1. River profiles and knickpoints discretization**

205 River networks represent a powerful reconnaissance tool to investigate the impact of rock uplift and
206 climate changes on landscape evolution (Hack, 1957; Kirby and Whipple, 2001; Wobus et al., 2006;
207 Kirby and Whipple, 2012; Whittaker, 2012). This is possible because channel steepness and erosion
208 rates along river profiles adjust to rock uplift, climatic conditions, and bedrock erodibility (Duvall et
209 al., 2004; DiBiase and Whipple, 2011). Channel slopes along the stream commonly exhibit an inverse
210 power-law scaling relationship with upstream contributing drainage area (Flint, 1974). Hence, the

211 stream power model describes the variation of channel elevation in time (dz/dt) (Howard and Kerby,
212 1993; Whipple and Tucker, 1999):

$$213 \quad \frac{dz}{dt} = U - KA^m S^n \quad (1)$$

214 where U is the rock uplift rate, K is the fluvial erodibility coefficient controlled by bedrock lithology,
215 climate and sediment load, A is the upstream drainage area, S is the local channel slope and m and n
216 are constants that depend on basin hydrology, channel geometry and erosional processes (Howard,
217 1994; Whipple and Tucker, 1999). Under steady-state conditions erosion and rock uplift rates (U) are
218 balanced (Willett and Brandon, 2002; Kirby and Whipple, 2012), and there is no change in elevation
219 of the channel bed over time ($dz/dt = 0$). Therefore, equation (1) can be rearranged as:

$$220 \quad S = \left(\frac{U}{K}\right)^{\frac{1}{n}} A^{-\left(\frac{m}{n}\right)} \quad (2)$$

221 where $(U/K)^{1/n}$ and m/n are the channel steepness index (k_s) is the concavity index (θ), respectively
222 (Flint, 1974). The relationship in equation (2) is only valid above a critical upstream drainage area of
223 0.1 to 5 km², where fluvial processes dominate over debris flow processes (Montgomery and
224 Fournelle, 1993; Stock and Dietrich, 2003; Wobus et al., 2006). k_s and θ can be extracted from DEMs
225 by a logarithmic regression of the local channel slope versus the contributing drainage area (Whipple,
226 2004; Wobus et al., 2006; Table S1 in the supporting information) or through an integral approach
227 (Perron and Royden, 2013; Mudd et al., 2014 see explanation below; Table S2 in the supplementary
228 information). k_s is sensitive to spatial variations in bedrock erodibility, climate and rock uplift rate
229 and hence represents a fundamental metric for geomorphic studies (Snyder et al., 2000; Wobus et al.,
230 2006). Because small variations in the concavity index produce large variations in k_s and considering
231 that DEM-noise can have a strong impact on θ and consequently on k_s , we used a typical reference
232 concavity index (θ_{ref}) of 0.45 (e.g., Wobus et al., 2006; Kirby and Whipple, 2012). The chosen θ_{ref}
233 is the most used value in the literature to date (e.g., Kirby and Whipple, 2012) and falls in the range
234 of θ estimates for every stream segment upstream and downstream of non-lithological knickpoints of
235 the Anti-Atlas (Snyder et al., 2000; Wobus et al., 2006; Tables S1 and S2 in the supporting

236 information). The use of a θ_{ref} allows calculating the normalized channel steepness index ($k_{sn} =$
 237 $(U/K)^{1/n}$) and hence to effectively compare longitudinal profiles with greatly varying drainage areas
 238 and infer relative variations in rock uplift rates. Specifically, in areas with similar climate and rock
 239 strengths, rivers with higher k_{sn} values will reflect higher uplift rates (e.g., Olivetti et al., 2012; Miller
 240 et al., 2013).

241 Furthermore, we performed a regional scale topographic and statistical analysis, using the integral
 242 approach calculating χ along the main river segments. The method is based on the transformation of
 243 the horizontal coordinates of a river profile to the χ reference frame assuming steady-state conditions
 244 and spatially invariant uplift rates and bedrock erodibility (Perron and Royden, 2013):

$$245 \quad z(x) = z(x_b) + \left(\frac{U}{KA_0^m}\right)^{\frac{1}{n}} \int_{x_b}^x \left(\frac{A_0}{A(x')}\right)^{\frac{m}{n}} dx = z(x_b) + k_s A_0^{-\left(\frac{m}{n}\right)} \chi \quad (3)$$

$$246 \quad \chi = \int_{x_b}^x \left(\frac{A_0}{A(x')}\right)^{\frac{m}{n}} dx \quad (4)$$

247 where χ is the independent variable of integral quantity in the upstream direction from a base level x_b
 248 to an observation point x , z is the elevation along the channel, k_s (or k_{sn} assuming a reference
 249 concavity) is the steepness index and A_0 is the reference drainage area, usually assumed to be 1 km²
 250 (Perron and Royden, 2013). To calculate χ , a value of θ must be specified. In results section we
 251 always refer to χ assuming $\theta_{ref} = 0.45$ as specified above. However, by collapsing the tributaries to a
 252 linear relationship in the χ - z plot, it is possible to estimate individual θ of the selected stream
 253 segments (Mudd et al., 2014). In this case, a linear regression of elevation against χ permits
 254 determination of k_{sn} for an equilibrated river for the upstream and downstream river segments of
 255 transient river profiles (Perron and Royden, 2013; Smith et al., 2022). Finally, the results of k_{sn} (using
 256 θ_{ref} of 0.45) and the optimized θ for each river segments of the Anti-Atlas obtained through the log
 257 slope -log area and the integral approach are reported in Tables S1 and S2 in the supporting
 258 information, respectively. Importantly, the integral approach allows visualizing k_{sn} variation along
 259 the stream segments and to discretize the knickpoints. Moreover, in the χ - z plot the slope is

260 proportional to the k_{sn} . In areas where erodibility and climate variability are marginal, the river
261 network analysis can be used to infer and visualize the pattern of relative rock uplift rate across the
262 region. Usually, a variation in rock uplift rate is identified by the presence of “transient knickpoints”
263 which separate portions of the landscape eroding at different rates (Kirby and Whipple, 2012; Royden
264 and Perron, 2013; Mudd et al., 2014). A detailed analysis of these features is fundamental to
265 understand the state of a landscape. Distinguishing the nature of the knickpoints requires a detailed
266 analysis of the geological and geomorphic characteristics (Kirby and Whipple, 2012). Transient and
267 lithological knickpoints were differentiated by looking at: (1) their position and distribution in the
268 χ - z plot (*i.e.*, rivers that experienced a similar rock uplift history should cluster in the χ - z plot; Gallen
269 and Wegmann, 2017; Figure S2 in the supporting information); (2) available geological maps (1:
270 200.000, 100.000 and 50.000, Service Géologique du Maroc); and (3) satellite imagery on Google
271 Earth.

272 A Shuttle Radar Topography Mission Digital Elevation Model (SRTM DEM, pixel size of 90 m) was
273 used to perform the topographic and the river network analysis described above. All steps were
274 conducted using ArcGIS tools, TopoToolbox (Schwanghart and Scherler, 2014), TAK (Forte and
275 Whipple, 2019) and a series of MATLAB functions (Goren et al., 2014; Gallen, 2017; Smith et al.,
276 2022).

277

278 **3.2. River projections 1D and 2D**

279 The magnitude of fluvial incision, associated with changes in rock uplift, can be estimated from the
280 relict topography upstream of major non-lithological knickpoints (Berlin and Anderson, 2007;
281 Schildgen et al., 2012; Gallen et al., 2013; Miller et al., 2013). In particular, the reconstructed river
282 projection from the relict landscape permits determination of the paleo-base level and hence the
283 magnitude of minimum incision/surface uplift and paleo-relief before the development of the
284 knickpoints (Olivetti et al., 2016; Heidarzadeh et al., 2017; Fox et al., 2019). Particularly, the
285 estimates represent the incision into a specific surface which may be undergoing continued erosion

286 itself. Therefore, incision may not represent total erosion but the difference between the erosion
 287 across the low-relief landscape and that within the incising part of the landscape. We carry out two
 288 types of river projections: a) one dimensional projections where the elevation only varies as function
 289 of distance along a single channel and the model parameters are single values of surface uplift and
 290 paleo-channel steepness; and b) two dimensional projections where elevation varies as a function of
 291 space and the model parameters are maps of surface uplift and paleo-channel steepness. Both the
 292 methods require a fixed value of θ_{ref} , which has been assumed to be 0.45 as discussed in section 3.1.
 293 The one dimensional projections were carried out in χ space. To do this a linear model was regressed
 294 through the χ -elevation data from the relict part of the landscape. The intercept with y axis (i.e., where
 295 $\chi=0$) provides the elevation of the paleo-base level. This approach relies on several assumptions: that
 296 the paleo-river network has remained approximately the same through time; erosion rates across the
 297 relict landscape have remained constant; the channel steepness in the paleo-landscape was spatially
 298 uniform; and that the change in base level due to surface uplift is spatially uniform. The assumption
 299 that the erosion rates have remained overall steady across the Anti-Atlas relict landscape is justified
 300 by the close match between short and long-term erosion rates throughout the Cenozoic (i.e., erosional
 301 steady-state; e.g., Willett and Brandon, 2002; Clementucci et al., 2022). However, there is no
 302 guarantee that the projected rivers within the same catchment will predict the same amount of surface
 303 uplift. This has the benefit of providing a test of the underlying assumptions, however, the resulting
 304 landscape may be unrealistic because predicted elevations at confluences may be different.
 305 A two-dimensional approach, proposed by Fox (2019), allows reconstructing the paleo-topography
 306 ensuring tributaries and trunk streams share the same elevation at confluences. This approach was
 307 extended by Fox et al. (2020) and Smith et al. (2022) to account for spatial variability in surface uplift.
 308 The basis of this approach is that a discrete version of the stream power model can be written using
 309 a series of nodes along the main trunk river:

$$310 \quad z_i - B.L. = \sum_{j=1}^i (\Delta\chi_j) u_j^* + S.U._i \quad (5)$$

311 where the i th pixel is the upstream of the j th pixel and the lowest most pixel has the elevation of B.L.
312 (base level), $S.U.i$ is the surface uplift that the relict landscape has experienced following fluvial
313 dissection, u^* is the normalized rock uplift rate, which is proportional to the normalized channel
314 steepness that is preserved across the relict landscape. This expression provides a prediction of
315 elevation as a function of channel steepness and surface uplift. Provided both vary smoothly in space
316 over long length scales, the elevation of pixels within the DEM can be used to recover these values
317 using inverse methods. The results provide a reconstructed pre-incision landscape and allows
318 quantifying the amount of surface uplift for each node (Smith et al., 2022). Maps of u^* and $S.U.$ can
319 be calculated to minimize the misfit between predicted and observed topography of the relict
320 landscape and the roughness of the u^* and $S.U.$ maps. In this case, the χ values were calculated for
321 the entire drainage network of the Anti-Atlas but only the upstream knickpoints on the relict landscape
322 were used to estimate the u^* and $S.U.$. We used values of $m = 0.45$, $A_0 = 1 \text{ m}^2$. Because the topographic
323 reconstruction requires many more model parameters, smoothness constraints in the form of weighted
324 negative Laplacian operators are required. We used weighting parameters for the $S.U.$ parameters and
325 u^* (λ and α , respectively) equal to 10 and the grid size to 5 km, for the inversion method. The main
326 assumptions of the method are that the shape of the river network has remained constant in time, there
327 was no major drainage reorganization, and the erodibility coefficient (K) is considered constant.

328

329 **3.3. Knickpoint celerity model**

330 A celerity model was used to estimate the onset of knickpoints migration by using the stream model
331 shown in equation (1) and considering plucking as the primary erosion mechanism ($n = 1$; Whipple
332 et al., 2000). The horizontal migration of knickpoints along the river profiles in response to a relative
333 base level drop can be described as:

$$334 \quad \frac{dx}{dt} = KA^m \quad (6)$$

335 where, dx/dt is the knickpoint celerity, K is a dimensional coefficient of erosion (Whipple and Tucker,
336 1999; Whipple, 2004), A is upstream drainage area and m is a non-dimensional parameter that
337 depends on basin hydrology, channel geometry, and erosion process (Whipple and Tucker, 1999).
338 Particularly, we used a dynamic time-step approach as described in Crosby and Whipple (2006) and
339 Berlin and Anderson (2007). The misfit was the difference between the real knickpoint retreat
340 distance and the modeled one. K and m were found by a brute-force search: the best values of K and
341 m resulted in the lowest sum of squares of the misfits (R^2) over 100,000 iterations for each simulation
342 (Table S3; Stock and Montgomery, 1999; Crosby and Whipple, 2006; Berlin and Anderson, 2007;
343 Sembroni et al., 2016; Sembroni et al., 2020; Siravo et al., 2021). Specifically, we performed 100
344 simulations and calculated the mean initiation time and standard deviation of these best-fit values.
345 Then, to have a unique time value for each modeled knickpoint, we calculated the probability density
346 functions (PDFs) as described in Siravo et al. (2021).

347 In this specific case, m is assumed to be the concavity value, given that $m/n = \theta$ and considering that
348 $n = 1$ as demonstrated by Clementucci et al., (2022). Because the used approach searches the
349 parameters by minimizing the distance between the real and modelled position of knickpoints, we let
350 m vary linearly between 0 and 0.75. This range is motivated by the concavity values obtained from
351 the stream profiles analysis reported in Tables S1 and S2 in the supporting information. Similarly,
352 the erodibility parameter K was permitted to vary between 10^{-7} and 10^{-4} (Figure S4). This range of K
353 values results from a forced-origin linear regression in the plot k_{sn} versus denudation rates (E), using
354 the stream power incision model ($K = E/k_{sn}^n$). In particular, by using basin-averaged k_{sn} values
355 normalized for a θ_{ref} of 0.45 and assuming $n = 1$, Clementucci et al. (2022) documented that K values
356 in the Anti-Atlas are in the order of 10^{-6} - 10^{-7} . These values represent the minimum K values of our
357 search interval. For the maximum values we further calculate K by using the basin-averaged k_{sn}
358 normalized for a θ_{ref} of 0.17. We used this θ_{ref} because the axial zone of the Anti-Atlas is
359 characterized by a high-standing landscape with mean concavity value of ~ 0.17 , which is assumed to
360 be the most reliable value of the non-rejuvenated landscape (Clementucci et al., 2022; Tables S1 in

361 the supporting information). For the initiation point of the knickpoint propagation, we considered the
362 present-day river outlets.

363 Finally, we set the onset of knickpoints migration between 3.8 and 18.6 Ma. This temporal range was
364 estimated by using the maximum excavation time required to erode the missing rock volume from
365 the river catchments (Norton et al. 2008; Gallen et al., 2013; Siame et al., 2015). The excavation time
366 is expressed as:

$$367 \quad T = \left(\frac{V}{A}\right) E^{-1} \quad (7)$$

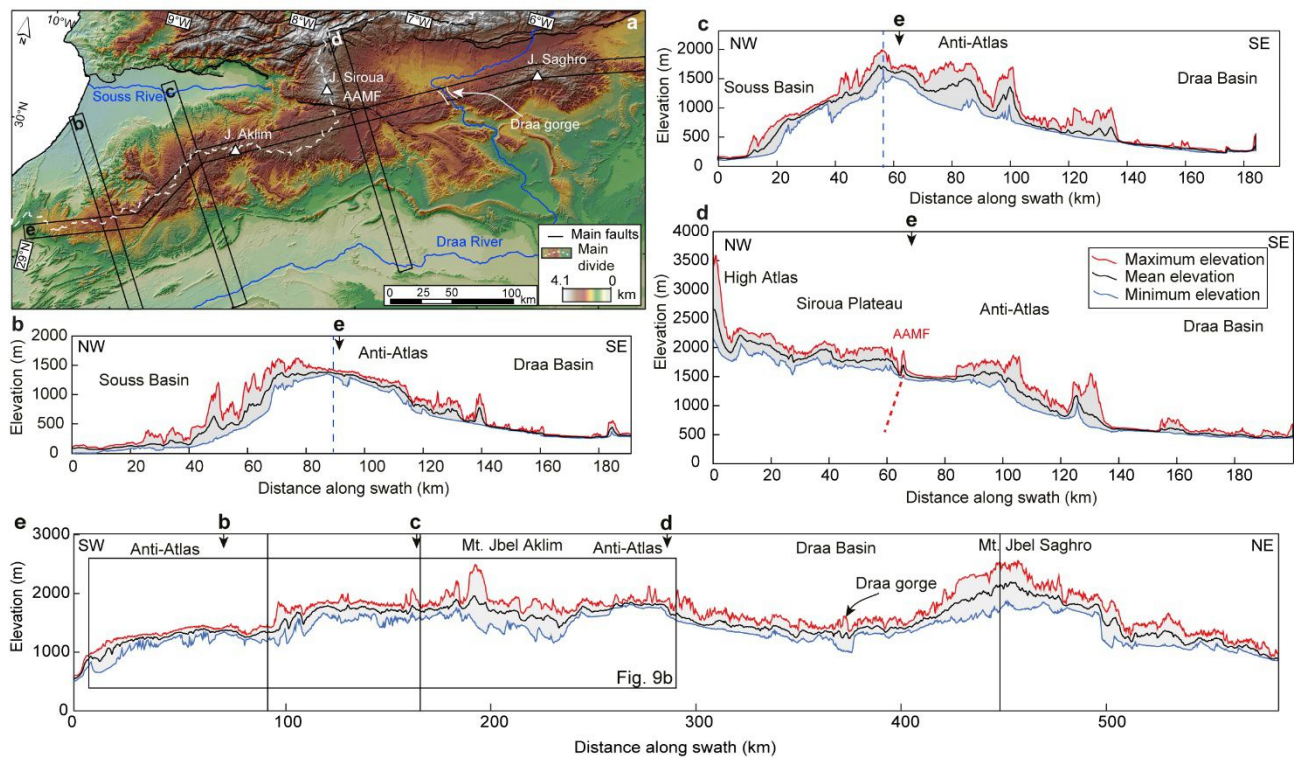
368 where T is the time required to erode the missing rock volume in the catchments, V is the rock volume,
369 A is the drainage area of the catchments and E is the basin-wide denudation rate. V can be estimated
370 by the difference between a surface connecting the highest points in the current landscape and the
371 current topography itself.

372

373 **4 Results**

374

375 **4.1 Topographic analysis**



376

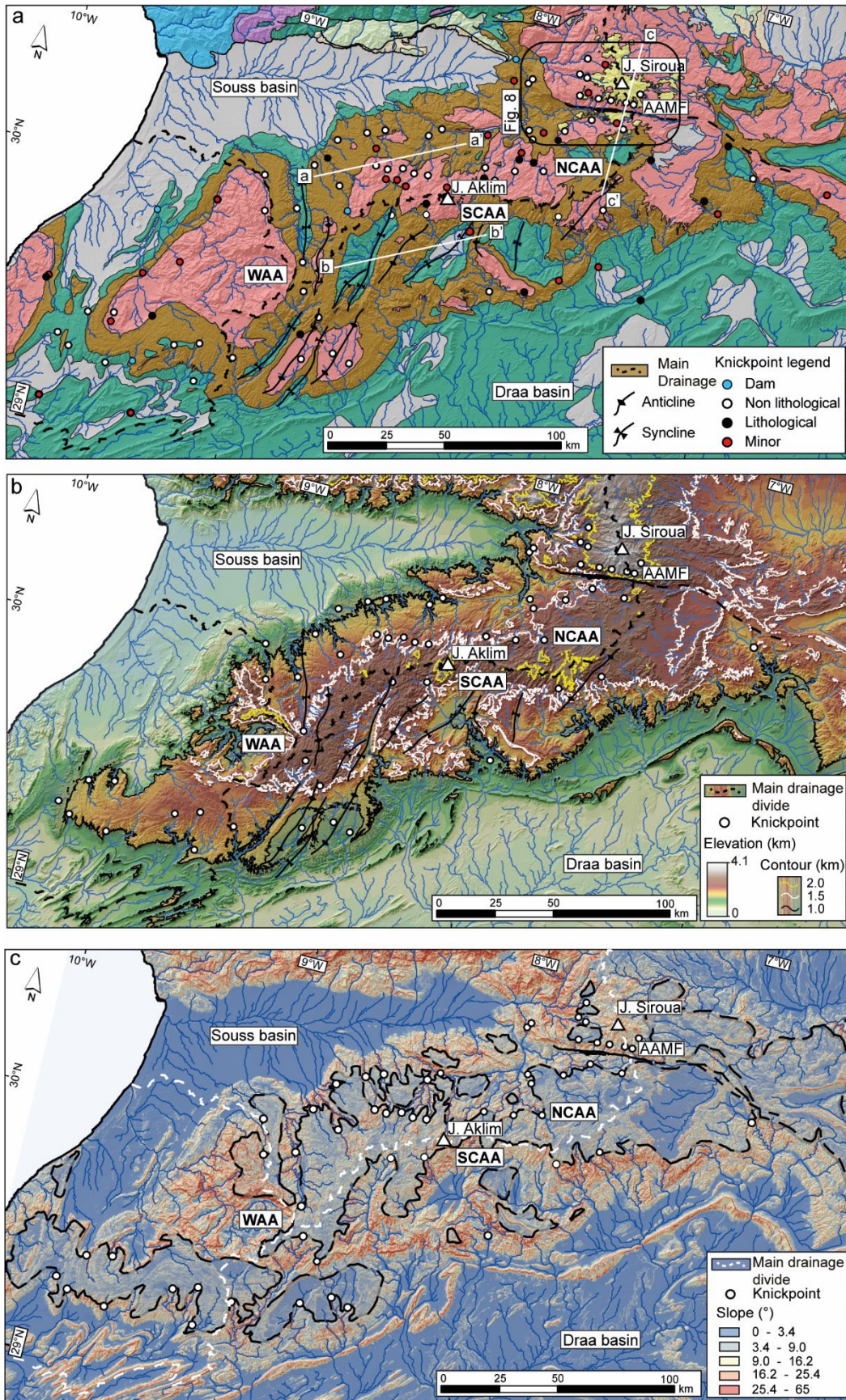
377

378 **Figure 2.** (a) Topographic map of the Anti-Atlas and the Siroua Massif (SRTM DEM database) with location
 379 of the swath profiles. The dashed white line indicates the position of the main drainage divide. (b, c, d) are
 380 swath profiles orthogonal to the main drainage divide of the Anti-Atlas, while (e) is parallel to the main
 381 drainage divide. The dashed blue line indicates the position of main drainage divide on swath profiles b, c.

382

383 The AA is characterized by an extensive, high-standing (~2000 m), axial zone with low-topographic
 384 relief delimited by steep and dissected flanks (Figure 2). The westernmost sectors of the belt have a
 385 dome-shape geometry with a rather symmetric topography across the drainage divide (Figure 2b).
 386 The central sectors of the western AA have an asymmetric topography with a gentle southern and a
 387 steep northern flank grading toward the Draa and the Souss basins, respectively (Figure 2c). The
 388 central AA has also a flat top but is bounded to the north by the Siroua plateau through a topographic
 389 step in proximity of the Anti-Atlas Major Fault (AAMF; Figure 2d). The along-strike swath profile
 390 highlights the geometry and the extent of the elevated axial surface, with a few mountain peaks (*e.g.*,
 391 Mt. Jebel Aklim, 2531 m and Mt. Jebel Saghro, 2592 m) and deeply incised valley (the Draa gorge;
 392 Figure 2a and 2e). The high-standing landscape is mostly composed of Paleozoic limestones and

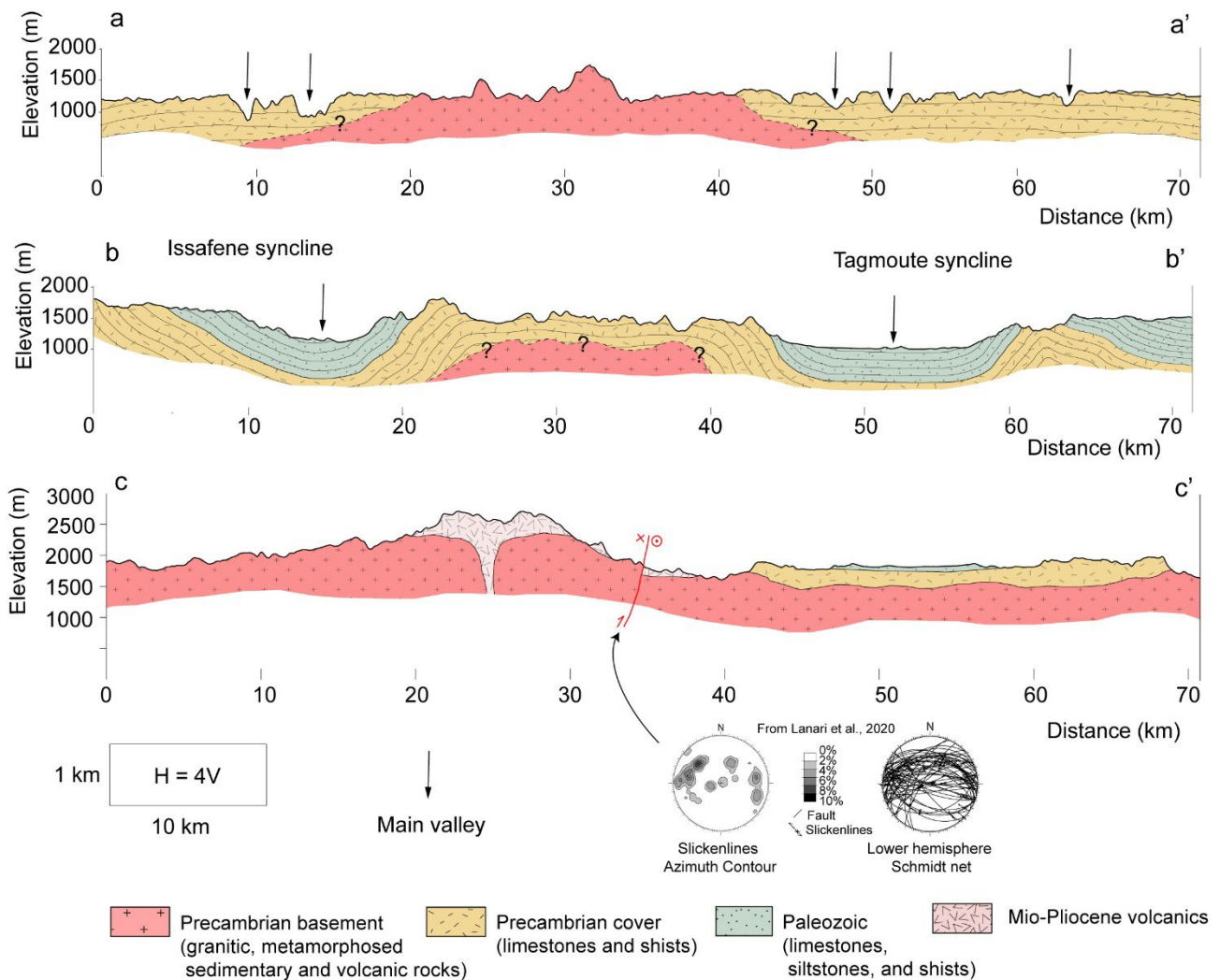
393 Precambrian basement rocks (Figure 3a) and is located upstream of non-lithological knickpoints.
394 Downstream of these knickpoints, the topography is steeper and includes late Precambrian and
395 Paleozoic sedimentary rocks (Figure 3). The northern flank of the belt has a slope that is sub-parallel
396 to the bedding of late Precambrian carbonates and consists of narrow and deep valleys perpendicular
397 to the bedding (Figure 4, profile a-a'). Conversely, the southern flank includes wide valleys mostly
398 located along the core of the synclines with Paleozoic sediments (Soulaimani and Burkhard, 2008;
399 Figure 4, profile b-b'). Finally, the swath profiles indicate that the high-standing, elevated surface
400 plunges along the strike of the mountain belt (Figure 2e), and perpendicularly to it (Figure 2b, 2c).



401

402 **Figure 3.** (a) Geological map of the study area with a detailed analysis of the knickpoints for the northern flank
 403 of the central Anti-Atlas (NCAA), western (WAA), and southern flank of central Anti-Atlas (SCAA) and the
 404 Siroua Massif. The main fold axes are associated with the Variscan orogeny and are shown with black lines
 405 (from Soullaimani and Burkhard, 2008). (b) Topographic map (SRTM DEM database) of the study area with

406 the non-lithological knickpoints and the fold axes. The black, white, and yellow dashed lines demark the 1000,
 407 1500 and 2000 m contours, respectively. (c) Slope map of the study area with non-lithological knickpoints.
 408 The dashed line indicates the low-slope area of the axial zone of the Anti-Atlas. Location of the maps is shown
 409 in figure 1b.



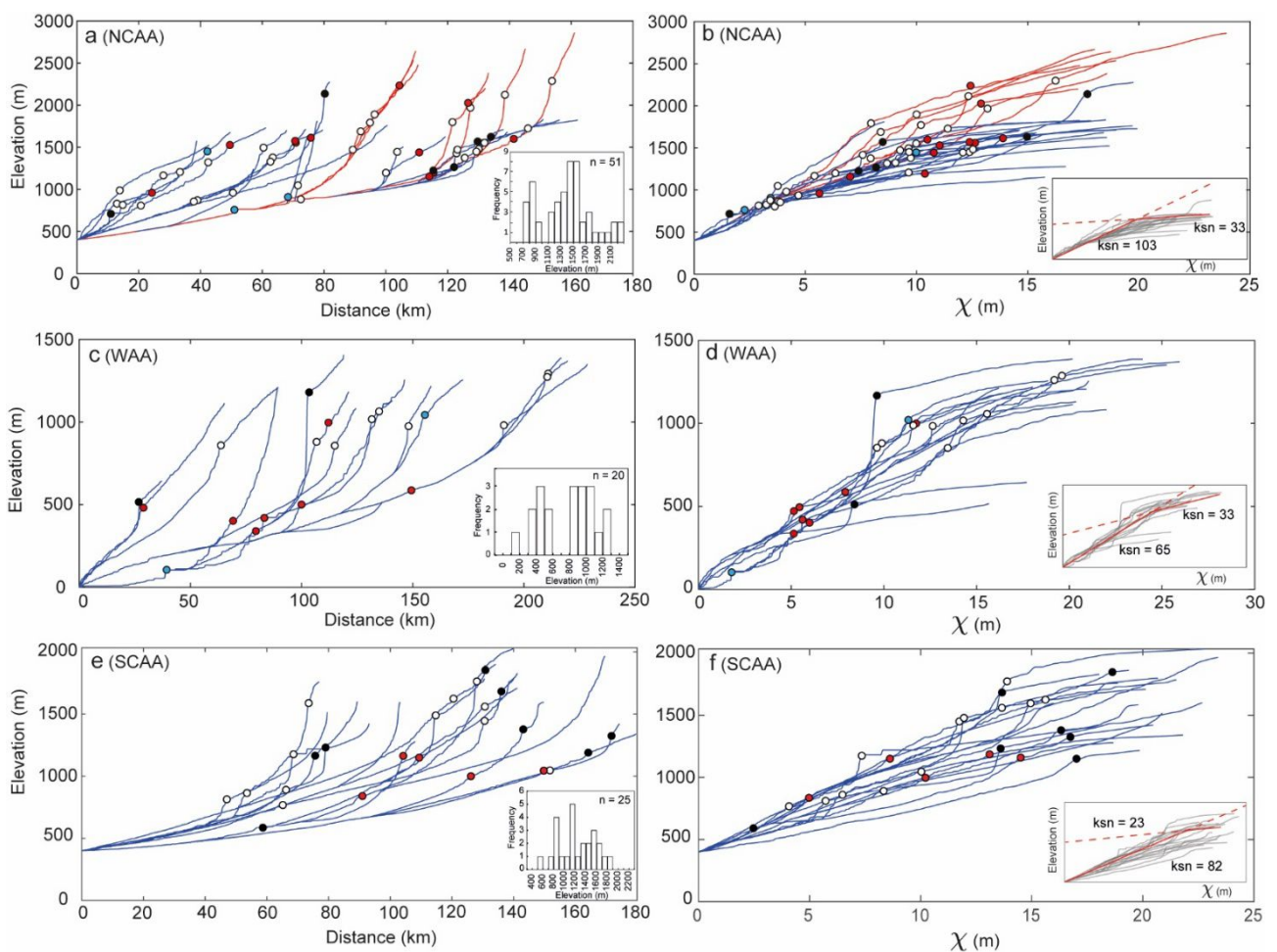
410
 411 **Figure 4.** a-a' geological cross section of the northern flank of the Anti-Atlas. b-b' geological cross section of
 412 the southern flank of Anti-Atlas. c-c' geological cross section in the Siroua Massif. Location of the cross-
 413 sections is indicated in figure 3a.

414

415 4.2 River morphology

416 Seventeen stream profiles from the northern flank of the central Anti-Atlas (NCAA) together with
 417 seventeen profiles from its southern flank (SCAA) and fourteen from the western side of the orogen
 418 (WAA) were analysed individually. The NCAA is mostly characterized by transient longitudinal river
 419 profiles. In particular, we recognized 51 knickpoints with two main patterns of non-lithological
 420 knickpoints standing between 700 and 1000 m and 1300 and 1500 m of elevation and several minor

421 lithological knickpoints (Figures 3a, 3b and 5a). The two patterns of non-lithological knickpoints
 422 cluster approximately at values of χ of 3-5 and 9-11, respectively (Figure 5b). The steepest and
 423 concave up segments of the NCAA rivers are downstream of the highest knickpoints (Figure 5a, 5b),
 424 whereas, the upstream segments are mostly rectilinear, have a shallow gradient and contain minor
 425 knickpoints (Figures 3a and 5a, 5b). Another minor group of knickpoints separating gentle upstream
 426 river segments from steeper downstream portions is in the Siroua Massif, over 2000 m of elevation
 427 (see red profiles in figure 5a and 5b), upstream of the Anti-Atlas Major Fault (AAMF, Figure 6).



428
 429 **Figure 5.** Longitudinal river profiles and χ plots with major knickpoints for (a-b) NCAA, (c-d) WAA, and (e-
 430 f) SCAA. The rivers of NCAA and SCAA sectors are extracted from the bedrock-alluvial transition, which in
 431 the study area is around 400 m. The knickpoints legend is in figure 3a. The red profiles indicate the rivers
 432 flowing from the Siroua Massif. The inset plots in the χ space show the mean regional k_{sn} values upstream and
 433 downstream of the highest knickpoints. The inset histograms in the longitudinal profiles show the frequency
 434 distribution of the elevation of the knickpoints.

435
 436 The WAA sector presents also transient river profiles with approximately 20 knickpoints. Most of
 437 the non-lithological knickpoints are located within a relatively narrow elevation band, of about 800

438 to 1100 m while the minor knickpoints are at about 400 to 500 m (Figure 5c, 5d). The steeper
439 downstream portion of the major non-lithological knickpoints exhibit a concave-up geometry with a
440 few minor knickpoints (Figure 5c, 5d). The upstream segments are mostly rectilinear with a typical
441 equilibrium profile (straight segment in χ space; Figure 5d). Here, the two patterns of knickpoints at
442 ~ 500 m and ~ 1000 m, occur approximately at χ values of 5-6 and 10–15, respectively (Figure 5d).
443 The SCAA sector is characterized by both, equilibrated and transient longitudinal river profiles and
444 contain at least 25 knickpoints distributed across a wide range of elevation (Figure 5e). Most of the
445 non-lithological knickpoints stand at about 800 to 1000 m of elevation and over 1500 m of elevation,
446 while the lithological knickpoints are at an elevation range of 1000 to 1500 m (Figure 5e, 5f).
447 Overall, most of the rivers are characterized by transient conditions with mean k_{sn} values of 84 ± 3.8
448 ($\sigma = 19.7$) $m^{0.9}$ and 30 ± 2.9 ($\sigma = 14.8$) $m^{0.9}$, and θ values of 0.58 ± 0.04 ($\sigma = 0.22$) and 0.17 ± 0.03 (σ
449 $= 0.18$) in the downstream and the upstream segments, respectively (Table 1 and Table S1 in the
450 supporting information).

451 **Table 1**

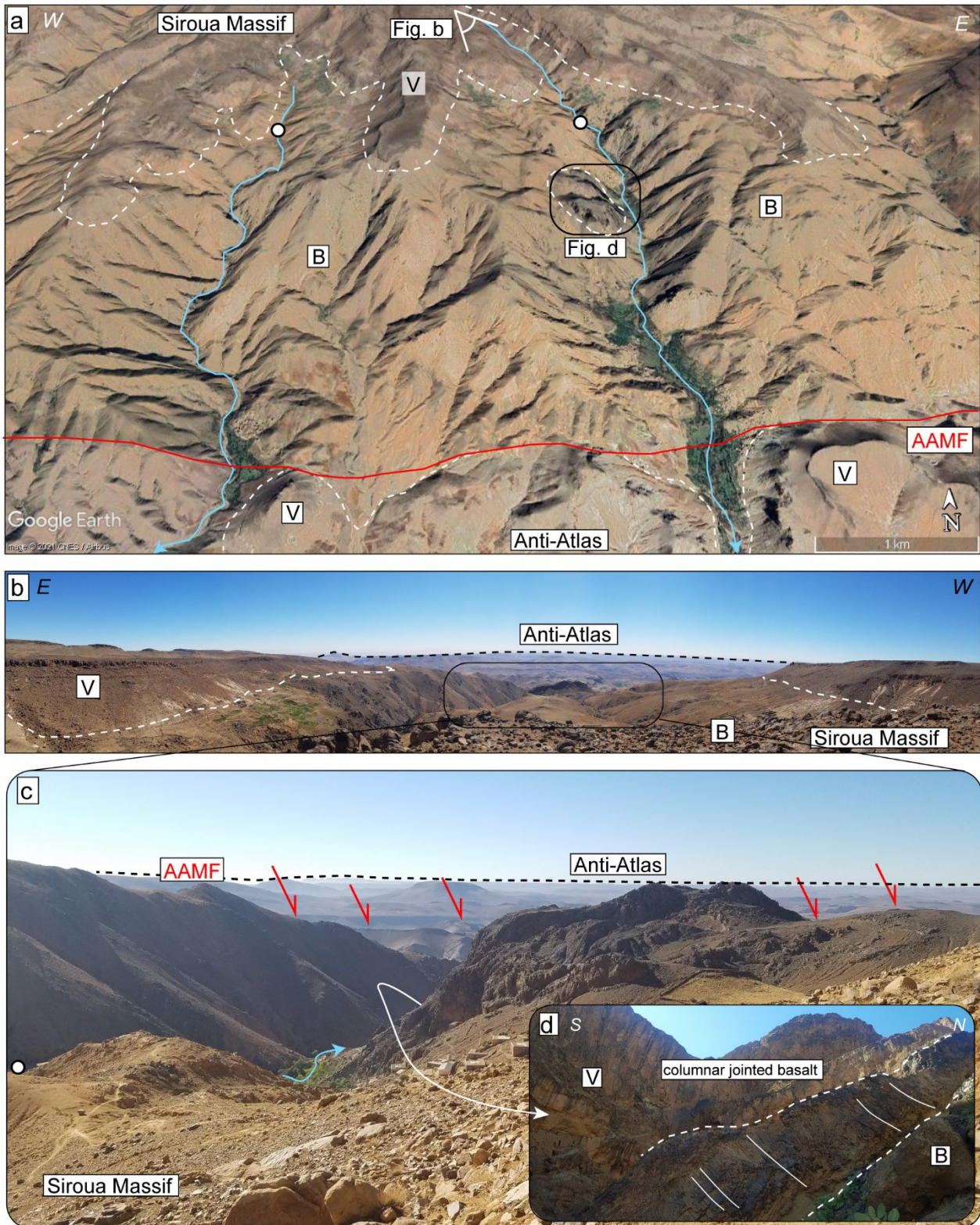
452 *Summary of k_{sn} and θ values upstream and downstream of non-lithological knickpoints obtained through a*
453 *linear regression of slope against drainage area on a log-log plot (see details in Table S1 of the supporting*
454 *information).*

| Sector | | Stream parameters | | | |
|-------------------|-----------------------|-------------------------------------|-----------------------|------------------------|-----------------|
| | | Downstream | | Upstream | |
| | | k_{sn} ($m^{0.9}$) ^f | θ ^g | k_{sn} ($m^{0.9}$) | θ |
| NCAA ^a | mean ^d | 102.6 ± 4.7 | 0.44 ± 0.02 | 33.2 ± 3.5 | 0.18 ± 0.04 |
| | σ ^e | 14.8 | 0.07 | 11.1 | 0.14 |
| SCAA ^b | mean | 82.2 ± 2.7 | 0.57 ± 0.03 | 22.7 ± 5.1 | 0.17 ± 0.07 |
| | σ | 7.2 | 0.09 | 13.5 | 0.2 |
| WAA ^c | mean | 65.07 ± 3.1 | 0.75 ± 0.1 | 33.23 ± 6.1 | 0.14 ± 0.07 |
| | σ | 9.45 | 0.3 | 18.5 | 0.22 |
| Tot | mean | 84.1 ± 3.8 | 0.58 ± 0.04 | 30.4 ± 2.9 | 0.17 ± 0.03 |
| | σ | 19.7 | 0.22 | 14.8 | 0.18 |

455
456 *Note.* ^a NCAA: rivers draining the northern flank of Anti-Atlas. ^b SCAA: rivers draining the southern flank of
457 Anti-Atlas. ^c WAA: rivers draining the western flank of Anti-Atlas. ^d Mean and standard error values. ^e
458 Standard deviation of the data. ^f Channel steepness calculated by using a reference concavity of 0.45. ^g
459 Concavity value calculated by the slope of the line regressed through a log-log plot of channel slope vs.
460 drainage area.

461
462

463 The k_{sn} and θ values obtained with the log S-log A method were compared with those estimated with
464 the χ method (Tables S1 and S2 in the supporting information). k_{sn} values lie mostly on the 1:1 trend
465 (Figure S3a), while the θ values show a higher dispersion (Figure S3b). Finally, the elevation and
466 distance from the river mouth show a positive correlation with the drainage area of catchments (Figure
467 S5b and S5c in the supporting information).



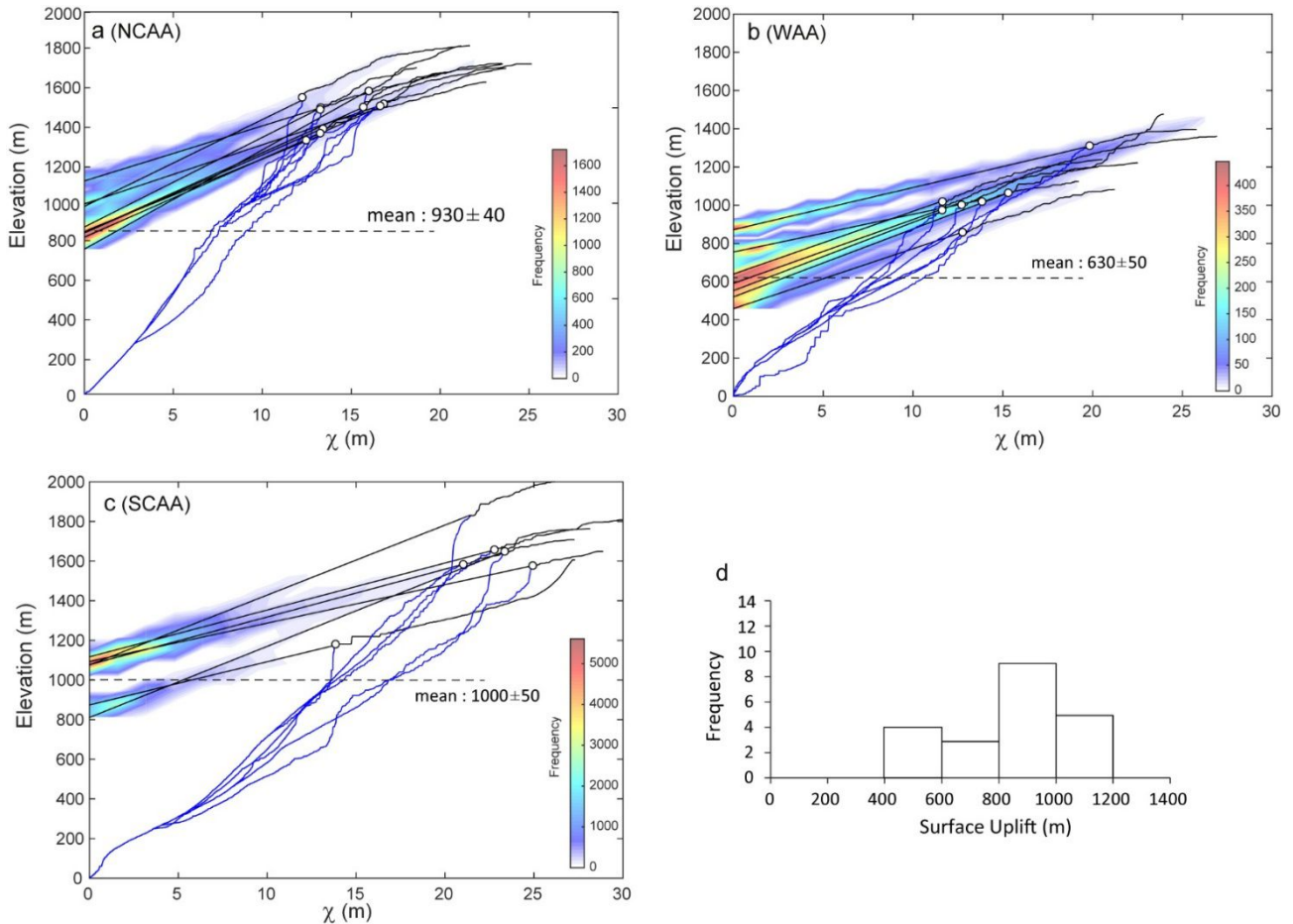
468

469 **Figure 6.** (a) Prospective Google Earth view of the southern flank of the Siroua Massif. Note the volcanic
 470 edifice composed of lava flows demarked by the dashed white line (see V symbol) resting unconformably over
 471 the Precambrian basement (see B symbol), the north-dipping Anti-Atlas Major Fault (AAMF), the non-
 472 lithological knickpoints shown with a white dot, the deep incisions and the remanence of the lava flows
 473 downstream of the non-lithological knickpoints. (b) Panoramic view from the Siroua Massif showing in the
 474 background the summit erosional surface (i.e., relict landscape) of the Anti-Atlas (see dashed black line) and
 475 in the foreground the contact between the basement (B) and the lava flows (V). (c) Detail of figure 6b showing
 476 the basalt flow preserved in the deeply incised valley of figure 6a. (d) Field picture of the basalt of figure 6c
 477 with high angle columnar joint (white lines) perpendicular to the valley bottom (dashed white lines).

478

479 4.3 Magnitude of fluvial incision

480 4.3.1 1D river projection



481

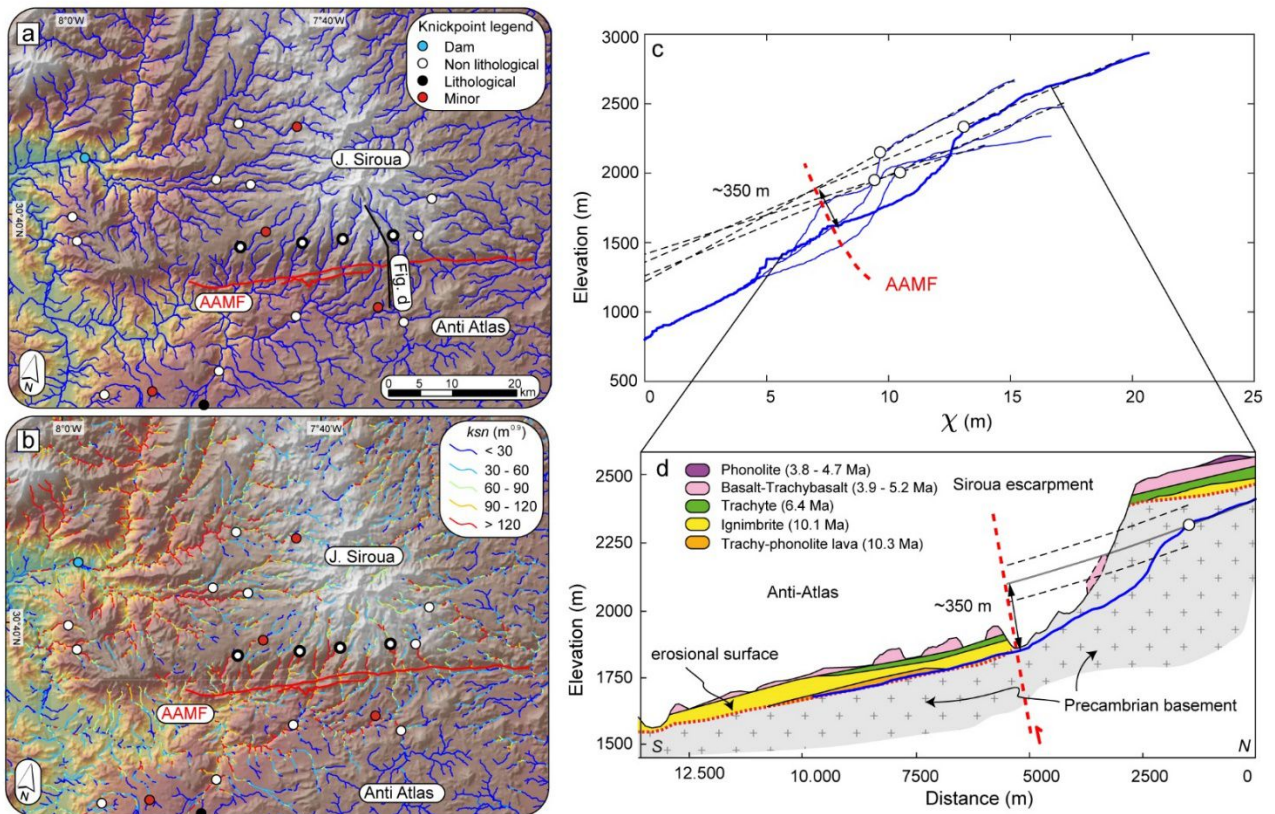
482 **Figure 7.** Modern longitudinal profiles and projections of the relict landscape (red segments) upstream of the
483 highest knickpoints using $\theta_{ref}=0.45$ and the k_{sn} of the relict portion for the (a) NCAA, (b) WAA, and (c) SCAA
484 sectors. Major knickpoints are marked with white circles. (d) Frequency diagrams of the magnitude of surface
485 uplift for the different sectors.

486

487 The Souss and Draa outlets represent the present-day base level for the NCAA and SCAA rivers,
488 respectively, whereas the WAA rivers flow directly into the Atlantic Ocean. To a first approximation
489 we can consider that all rivers of the Anti-Atlas drained toward the Atlantic before the relative base
490 level fall that initialized the most elevated non-lithological knickpoints (Figure 5). This because the
491 opening of the South Atlantic Ocean had already occurred in the early Cretaceous (Torsvik et al.,
492 2009) and hence much earlier than the Cenozoic topographic rejuvenation. Estimated values of fluvial
493 incision for the NCAA based on 10 stream profiles range from 822 ($2\sigma = 36$) to 1162 m ($2\sigma = 18$),

494 with an average of 930 ± 42 m ($\sigma = 133$) (Figure 7 and Table S4). In the WAA sectors, 7 transient
495 profiles yielded fluvial incision values ranging from 459 ($2\sigma = 9$) to 872 m ($2\sigma = 29$), with an average
496 of 630 ± 54 m ($\sigma = 143$) (Figure 7 and Table S4). Finally, in the SCAA where most of the rivers are
497 in equilibrium, the fluvial incision based on 6 stream profiles ranges from 811 ($2\sigma = 117$) to 1117 m
498 ($2\sigma = 18$), with an average of 1008 ± 53 m ($\sigma = 130$) (Figure 7 and Table S4). Furthermore, this
499 approach allows estimating the first-order paleo-topographic relief as the difference between the
500 modern mean drainage divide elevation and the elevation of the outlet of the reconstructed river
501 profiles. Our results indicate a mean elevation difference of 970, 890 and 750 m for the NCAA,
502 SCAA, and WAA sectors, respectively (Table S4 in the supporting information).

503 In the Siroua Massif, rivers are characterized by strong disequilibrium profiles with high-standing
504 non-lithological knickpoints lying at the margin of a low-topographic relief area and delimiting
505 streams with high k_{sn} from those with low k_{sn} values (Figures 6a, 8a, 8b). The magnitude of fluvial
506 incision along the Siroua escarpment was obtained by projecting the upstream portions of the rivers
507 up to the Anti-Atlas Major Fault (AAMF; Figure 8c, 8d). This value is ~ 350 m and represents only
508 the amount of incision between the Siroua Massif and the central sector of AA.



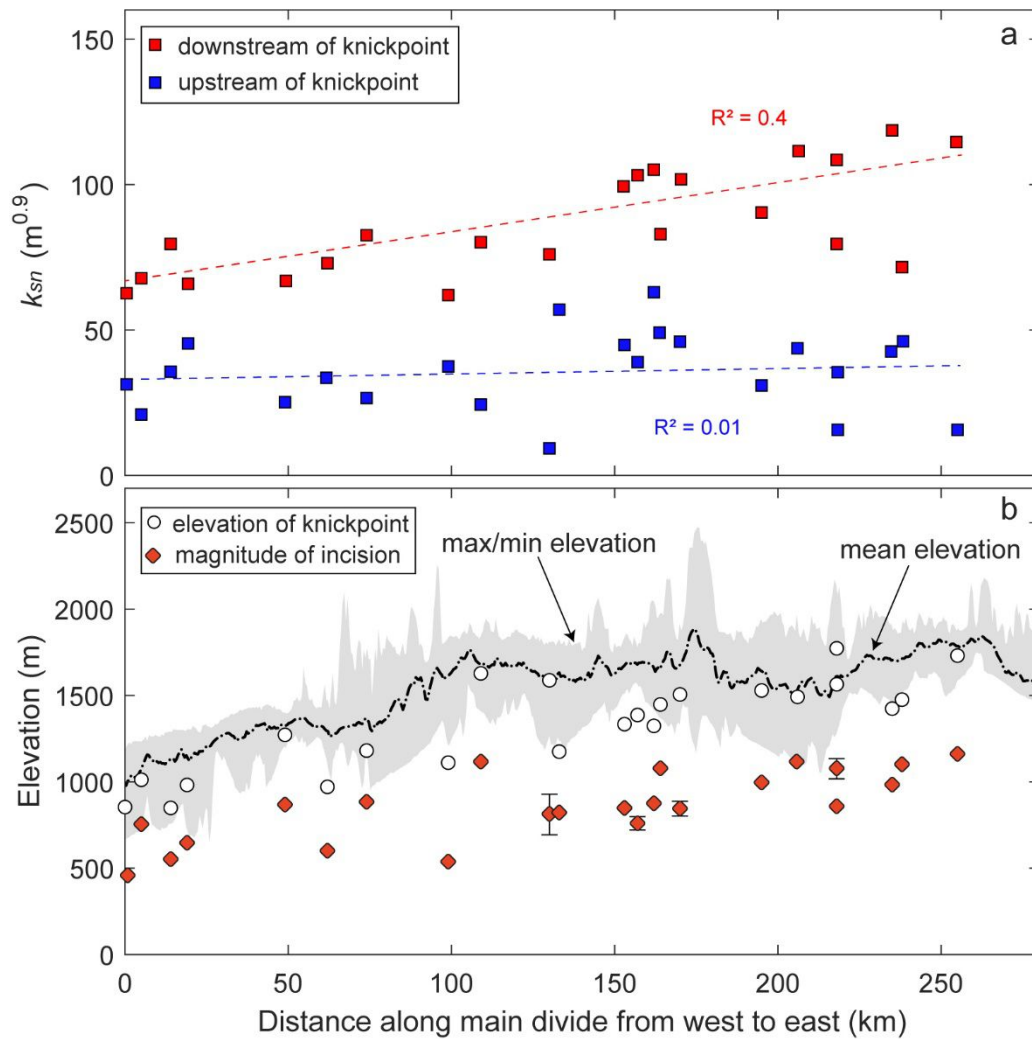
509

510 **Figure 8.** (a) Topographic map (SRTM DEM database) of study area with the main knickpoints. Bold blue
 511 rivers and associated knickpoints (thick black circles) are projected in figure 8c. (b) k_{sn} map of the stream
 512 network in the Siroua area. Location of the maps is shown in figure 3A. (c) Modern longitudinal profiles and
 513 projections from the Siroua relict landscape (dashed black segments) upstream of the highest knickpoints using
 514 $\theta_{ref} = 0.45$ and the k_{sn} of the relict portion. Note, the reconstructed river projections intersect each other between
 515 5 and 10 of χ . (d) Geological profile of the Siroua escarpment (after De Beer et al., 2000). The minimum value
 516 of incision is 350 m.

517

518 Overall, our projections document a progressive increase in the magnitude of fluvial incision and in
 519 the elevation of the highest knickpoints from the WAA to the central sector of AA with a culmination
 520 in the Siroua Massif (Figure 9b). This is also associated with a regional increase in k_{sn} values
 521 downstream of the major knickpoints (red dots in figure 9a).

522



523

524 **Figure 9.** (a) Distance along the main drainage divide versus k_{sn} down and upstream of the non-lithological
 525 knickpoints (data from Table S1). (b) Distance along the main drainage divide versus knickpoint elevation and
 526 estimated surface uplift values. The location of the swath profile is indicated in figure 2d.

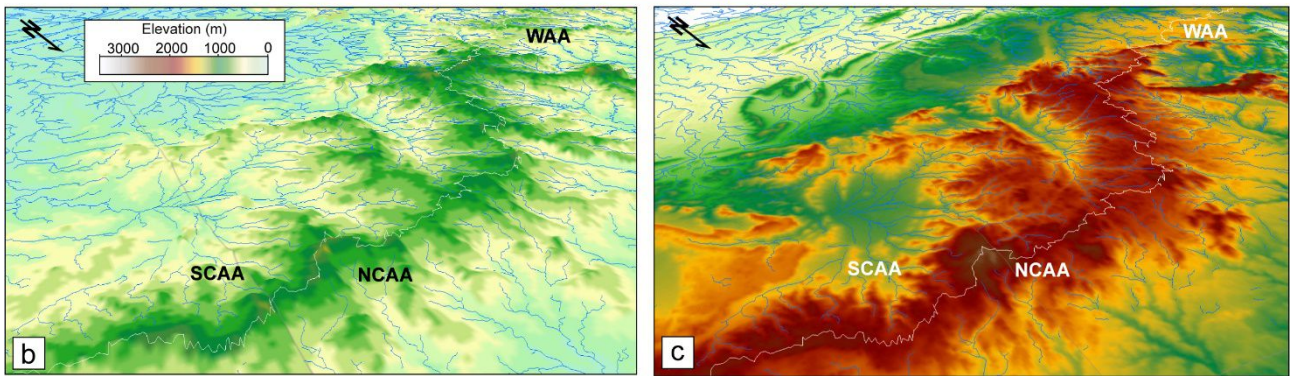
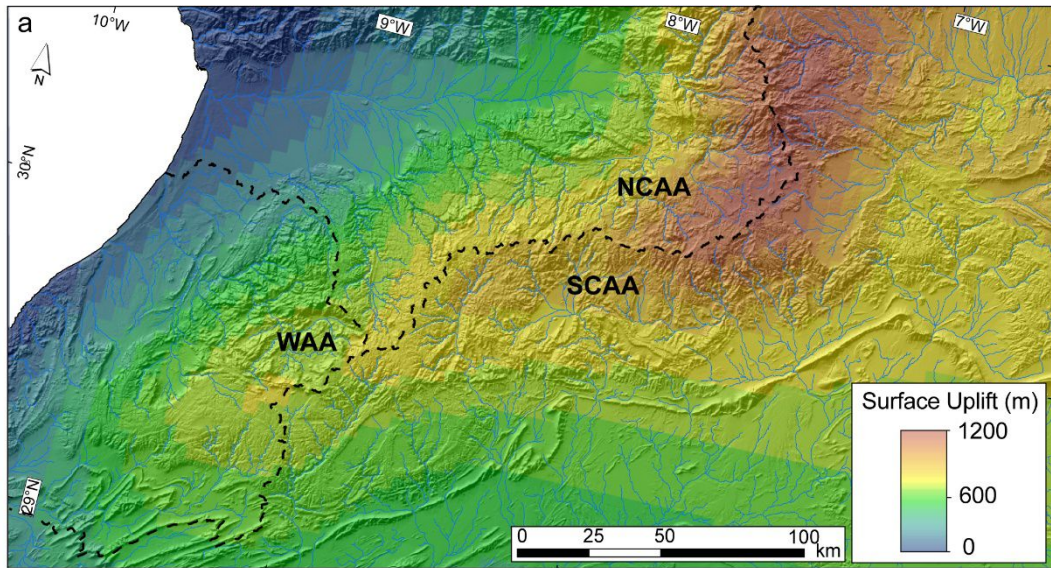
527

528 4.3.2 2D river projection

529 The 2D river projection method makes it possible to estimate the distribution of surface uplift (or
 530 base level fall) and normalized rock uplift rates, u^* proportional to k_{sn} , as a function of the elevations
 531 and the χ values from the relict river network (in the portion of landscape demarcated by black
 532 polygon in figure 3c). Due to the increased spatial variability provided by this approach, multiple
 533 solutions which pass through the χ and elevation data are expected, and results are non-unique.
 534 Damping forces smooth maps providing a means to choose a preferred result. Here damping
 535 parameters are chosen to ensure that results are consistent with the estimates obtained from projecting
 536 the information preserved in the upper reaches of the transient river profiles. Results suggest an

537 incision associated with a base level fall in the order of ~1200 m in the Siroua Massif, that decreases
538 gradually to ~500 m in the WAA (Figure 10a). The u^* values in the upstream relict landscape at the
539 maximum elevation are rather uniform with major variations coinciding with the highest mountain
540 peaks composed mostly of quartzites and metamorphosed conglomerates (Figures 1b and S6 in the
541 supporting information). Finally, the reconstructed paleo-topography indicates a maximum elevation
542 of ~1000 and ~500 m in the central sector of AA and the WAA, respectively before the onset of uplift
543 (Figure 10b). These results are consistent with those obtained from the river projections but with a
544 greater spatial resolution (Figure 10a, 10c). It should be noted that the fitted surface through the relict
545 topography above the incised landscape is less constrained in the Souss, Draa basins and in the Siroua
546 Massif. In the two basins, aggradation/deposition processes recorded by Plio-Quaternary sedimentary
547 sequences (Hssaine and Bridgland, 2009) are not taken into consideration. Moreover, the maximum
548 elevation obtained for the Siroua Massif represents a maximum estimate because the construction of
549 the volcanic edifice may be coeval or younger than the timing of the fluvial incision, as testified by
550 preserved lava flows along the Siroua escarpment (Figures 6d and 8d).

551



552

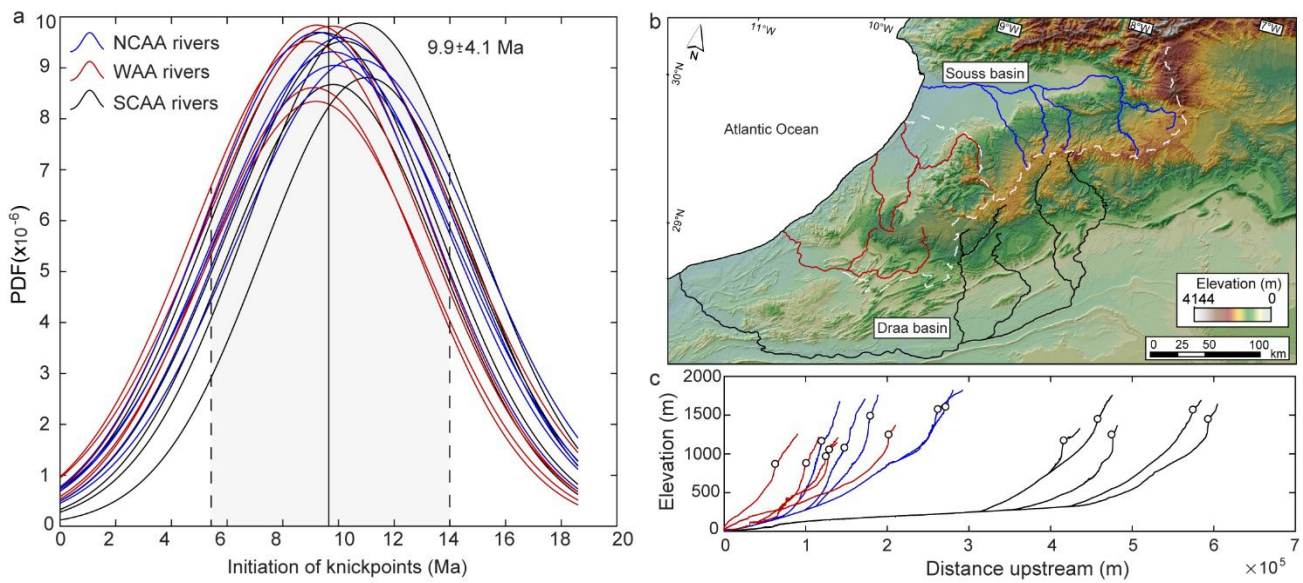
553 **Figure 10.** (a) Surface uplift (*S.U.*) map. The base level is defined at the relict landscape boundary; values of
 554 $m = 0.45$ and $A_0 = 1 \text{ m}^2$ were used to calculate χ and u^* (Equation 5). (b) Paleo-topographic reconstruction of
 555 the central and western Anti-Atlas. (c) Present-day topography of central and western Anti-Atlas.

556

557 4.4 Timescales of knickpoint migration

558 The time the highest non-lithological knickpoints took to travel from the base level to their present-
 559 day position is in the same range of that one required to erode the volume of rocks in a catchment
 560 between a high-standing preserved landscape and the minimum topography (*e.g.*, Gallen et al., 2013;
 561 Siame et al., 2015). This is called “excavation time” and is generally calculated assuming that erosion
 562 took place at constant rates. This volume of removed material includes the downstream (adjusted)
 563 and upstream portion (relict landscape) of the catchments, therefore the timing to remove it is higher
 564 than the time required to remove materials only in the rejuvenated portion. Furthermore, denudation
 565 rates derived from ^{10}Be concentrations for the Anti-Atlas Mountains are representative mainly for the
 566 relict landscape, as discussed in Clementucci et al. (2022). This because the quartz-bearing lithologies

567 are mostly exposed upstream of non-lithological knickpoints. Hence, denudation rates of 5.8 to 12.5
 568 m/Myr (Table S5 and Figure S7) provide a maximum age for the onset of topographic relief
 569 production in the relict landscape of the Anti-Atlas (8.8 to 18.6 Myr in Table S5 in the supporting
 570 information). Conversely, the highest value of denudation represents a mixed value (21.3 m/Myr in
 571 Table S5) from the Anti-Atlas relict and adjusted flanks, providing a minimum age \sim 3.8 Myr (Table
 572 S5).



573
 574 **Figure 11.** Initiation time of knickpoints propagation for the different sectors of the Anti-Atlas (NCAA, WAA,
 575 SCAA). (a) Probability Density Function of the timing of knickpoint initiation. (b) DEM showing the modelled
 576 rivers. (c) River longitudinal profiles and associated knickpoints of the modelled rivers. For further information
 577 see also table S3 in the supporting information.

578

579 The results from the selected river profiles are very consistent and indicate an onset for the upstream
 580 propagation of the base level fall at 9.9 ± 4.3 , 9.2 ± 4.1 and 10.4 ± 4.0 Myr for the NCAA, WAA and
 581 SCAA sectors, respectively (averaged value 9.9 ± 4.1 Myr; Figure 11 and Table S3).

582

583 5 Discussion

584 5.1 Significance of the transient topography

585 The strong variations in the normalized channel steepness indices, concavity values, topographic
586 slope, elevation, and fluvial incision between the portions of the landscape downstream and upstream
587 of highest non-lithological knickpoints (Figures 3 and 5) document the occurrence of a topographic
588 dichotomy, which is typical of a landscape in a transient state of development. Specifically, the
589 portion downstream of the highest knickpoints is eroding faster than the elevated relict landscape,
590 which is has been eroding at low and fairly uniform erosion rates at least since the late Cretaceous
591 (Clementucci et al., 2022). A transient topographic state could be attributed to several processes, such
592 as drainage reorganization, climate change, eustatic sea level fall and tectonic uplift (Hancock and
593 Kirwan, 2007; Kirby and Whipple, 2012; Miller et al., 2013, Ballato et al., 2015). Drainage
594 reorganization does not appear to be the cause because of the limited evidence of knickpoints, and
595 wind gaps formation related to river capture processes (Miller et al., 2013). An increase in erosion
596 rates due to climate variations (*i.e.*, increase in precipitation rates) is also unlikely, because such a
597 change would produce a decrease in the channel slopes rather than a steepening (Figure 5; Molnar et
598 al., 2004; Wobus et al., 2010). At the same time a decrease in precipitation rates would produce
599 sediment aggradation, in the downstream portions of non-lithological knickpoints and not stepper
600 flanks (Lanari et al., 2022).

601 Consequently, the abrupt break in the river longitudinal profiles and χ - z plots at the highest non-
602 lithological knickpoints and the position of the same knickpoints in the χ space (Figures 5b, 5d, 5f,
603 8a and S2) indicate that the two portions of the landscape erode at different rates (Schildgen et al.,
604 2012; Miller et al., 2012; 2013; Olivetti et al., 2016). This is consistent with an increase in erosion
605 and rock uplift rates propagating from the river outlet to the uppermost river segments (Miller et al.,
606 2013; Gallen and Wegmann, 2017; Racano et al., 2021). Consequently, the highest knickpoints
607 represent a response of the fluvial system to an increase in rock uplift rates and hence mark a phase
608 of topographic rejuvenation, while the uplifted relict landscape records the previous erosional
609 conditions predating such an increase. This implies that the magnitudes of fluvial incision represent
610 minimum estimates of the total amount of surface uplift (*e.g.*, Kirby and Whipple, 2012). A similar

611 scenario has been described in the Appalachian Mountains (Miller et al., 2013; Gallen et al., 2013)
612 and in other ancient orogens (Olivetti et al., 2012; Mandal et al., 2015; Calvet et al., 2015).
613 The minor and non-lithological knickpoints located in the lower segments of the fluvial network, at
614 700 to 1000 m and from 400 to 500 m in the NCAA, SCAA, and WAA sectors, respectively, can be
615 attributed to a Quaternary climate forcing, eustatic sea-level changes or both (Molnar and England,
616 1990; Crosby and Whipple, 2006; Hancock and Kirwan, 2007; Ballato et al., 2015). This pattern of
617 knickpoints is spatially consistent and does not show any channel steepness variation across
618 knickpoints in both longitudinal profiles and the χ - z plot (Figures 5b, 5d, 5f and S2) and hence cannot
619 reflect changes in rock uplift rates.

620

621 **5.2 Topographic evolution of the Anti-Atlas**

622 A peculiarity of the Anti-Atlas topography is the occurrence of a low-relief topography at its
623 maximum elevation plunging across and along strike over a wavelength of ~ 100 km (Figure 2b, 2c)
624 and ~ 600 km (Figure 2e), respectively, which can be described as a topographic swell. It follows,
625 that the high-standing relict landscape represents a warped geomorphic marker that can be used to
626 document the spatial distribution of the recent increase in surface uplift rates. The along strike pattern
627 of surface uplift ranges from ~ 500 to 1100 m from the western to the central sector of AA (Figure 9),
628 in agreement with the k_{sn} and elevation pattern of the most elevated non-lithological knickpoints. The
629 paleo-landscape must have been characterized by a low-topographic relief with localized ~ 1000 - to
630 700-m-high mountain peaks in the central Anti-Atlas (NCAA, SCAA) and the WAA, respectively
631 (Figure 10b and Table S4). These paleo-topographic heights correlate with high normalized channel
632 steepness values observed for low erodibility lithologies, such as quartzite and meta-conglomerates
633 (Figure S6; Clementucci et al., 2022).

634 The asymmetry of the AA flanks and the different locations of the highest knickpoints between the
635 southern (SCAA) and northern flanks (NCAA) can be attributed to a faster knickpoints migration in
636 the SCAA sector, as also suggested by the occurrence of quasi-equilibrium stream profiles (Figure

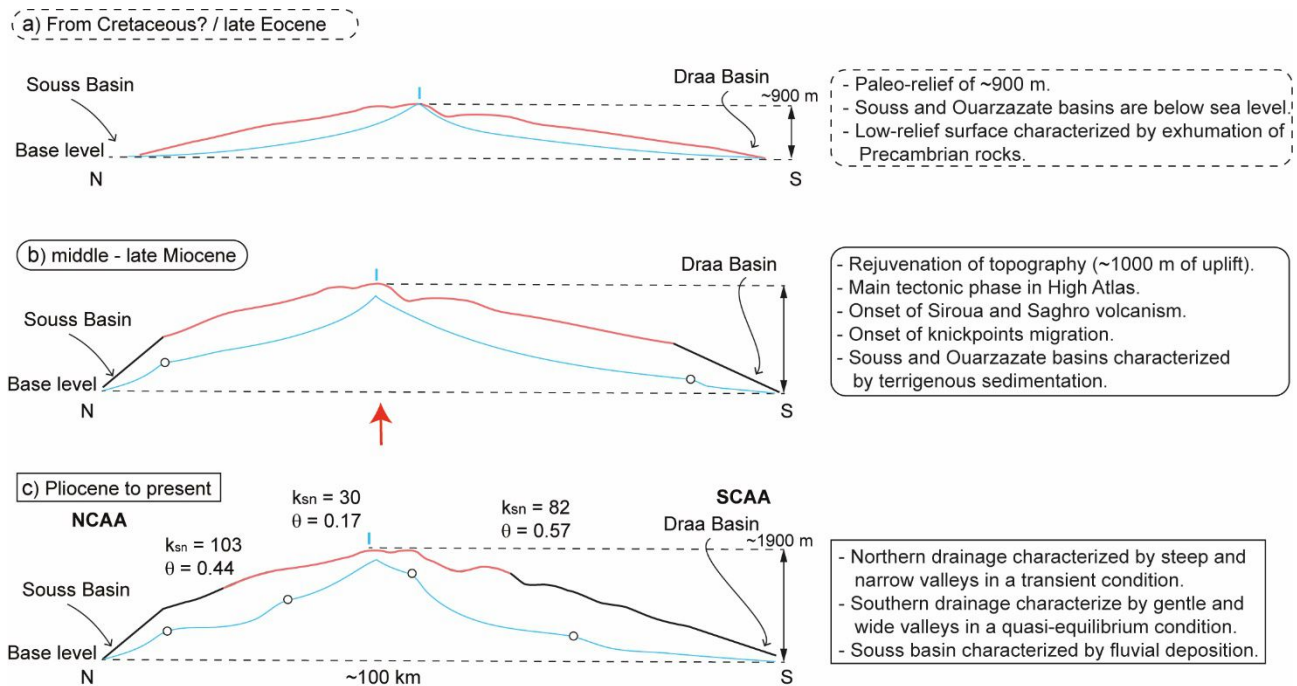
637 5e, 5f). Here, the presence of wide valleys sculptured on synclines within more erodible Paleozoic
638 rocks (limestones, siltstones and shists) may have promoted an increase in the contributing drainage
639 area and a faster knickpoint retreat (Figures 4b and S5b, S5c; Crosby and Whipple, 2006; Berlin and
640 Anderson, 2007; Schwanghart and Scherler, 2020). Conversely, in the northern flank, the knickpoint'
641 celerity is lower due to narrower and deeper valleys within less erodible late Precambrian carbonates
642 (Figures 4a and 13c). This highlights the key role that rock strength plays in controlling the valley
643 morphology, and in turn the location of genetically linked knickpoints (*e.g.*, Stokes et al., 2015).
644 Importantly, for the NCAA and SCAA rivers, the migrating wave of transient erosion propagated for
645 several kilometres within the Souss and Draa basins. These regions were likely characterized by
646 erosional processes, before the onset of regional uplift (*e.g.*, Hssaine and Bridgland, 2009). This
647 allows to dismiss a complex scenario with different erodibility parameters within the Neogene-
648 Quaternary sediments in the basins. However, this would not represent an issue because the
649 knickpoint would travel quickly in the mainstream of the Souss and Draa rivers due to the large
650 drainage areas. The knickpoints propagation rate would have decreased one reached the upstream
651 portions of the main mountain front (Crosby and Whipple, 2006; Schwanghart and Scherler, 2020).
652 Hence, the time of residence in the Draa and Souss basins would be negligible. Finally, it should be
653 noted that the occurrence of recent fluvial conglomerates, at high elevation in the major valleys of
654 the northern flank (~700 m) and in the Souss alluvial plain, has been interpreted as an indication that
655 uplift has decreased during the Quaternary (Lanari et al., 2022).

656

657 **5.3 Surface uplift history of Anti-Atlas**

658 The AA and surrounding basins (*e.g.*, Souss and Ouarzazate) were the southern shoulders of the
659 Triassic-Jurassic rift (Lanari et al., 2020a) as documented by the absence of Mesozoic deposits older
660 than Cretaceous and sediment provenance data (Domènech et al., 2018). During the Cenozoic, the
661 Ouarzazate and Souss basins recorded marine sedimentation at least until the late Eocene, suggesting

662 that part of the study area must have been under sea level until 44 to 42 Ma (Figure 12a; El Harfi et
 663 al., 2001; Teson et al., 2010; Hssaine and Bridgland, 2009).



664

665

666 **Figure 12.** Conceptual model depicting the Anti-Atlas topographic evolution across the central sector. (a)
 667 Paleo-topography inferred until the late Eocene. (b) Topographic rejuvenation during the middle-late Miocene
 668 associated with the initiation of river incision and knickpoints upstream propagation. (c) Present-day
 669 configuration. At this stage, the first pattern of non-lithological knickpoints migrates upstream and a second
 670 pattern lies a lower elevation. Pink and black lines describe the preserved relict and the rejuvenated landscape,
 671 respectively. The blue line indicates the river profile.

672

673 In the AA, the presence of AHe cooling ages older than 70 Ma and consequently the low magnitude
 674 of exhumation (< 2-3 km) over the same time interval suggests that the paleo-topography has been
 675 developing at least since the late Cretaceous (Figure 12a; Gouiza et al., 2017; Lanari et al., 2020a;
 676 Charton et al., 2021). Moreover, our observations and celerity calculations indicate that surface uplift
 677 triggered the headward migration of the highest knickpoints starting from the middle-late Miocene
 678 (Figures 11 and 13b). These new time constraints suggest average surface uplift rates ranging from
 679 40 to 110 m/Myr (western sector) and from 70 to 180 m/Myr (central sector). These rates are based
 680 on the celerity model (*i.e.*, the age of uplift initiation of 9.9 ± 4.1 Ma) and hence are averaged over

681 time scales of 14 and 5.8 Ma, using the mean surface uplift values inferred from the river projections
682 (Table S4).

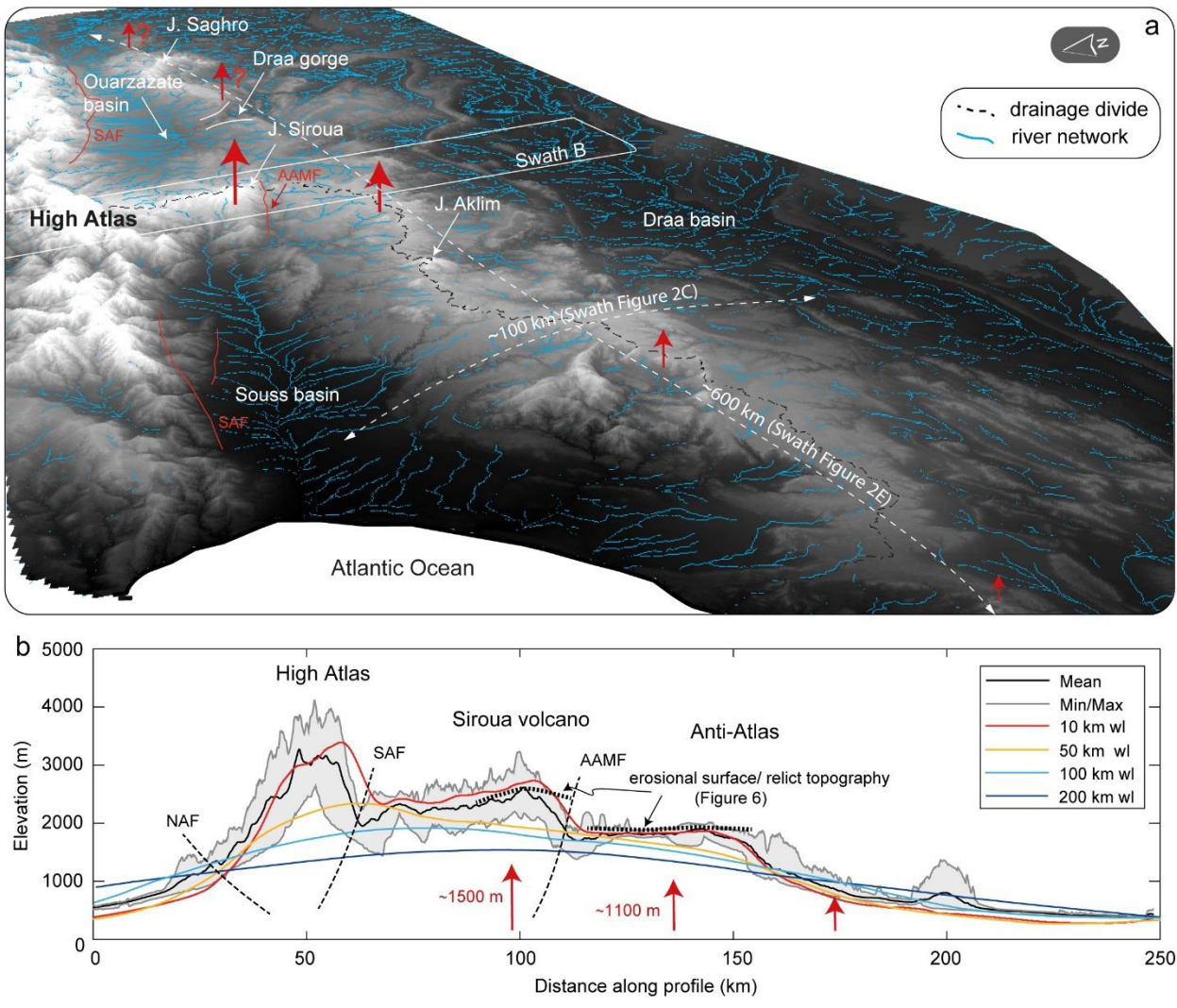
683 Downstream of non-lithological knickpoints, in the adjusted portion of the landscape, Clementucci et
684 al., (2022) estimated Quaternary denudation rates of up to 50 m/Myr. These rates are consistent with
685 the long-term rates inferred from the river projections averaged over the oldest age interval estimated
686 from the celerity model (14-10 Ma). Uplift initiation at 14-10 Ma, would agree with the onset of
687 volcanic activity in the Siroua Massif where more than 500 km² of crystalline basement are covered
688 by 11-to 3-Myr-old lava flows (Figure 12b; Berrahma and Delaoye, 1989; Missenard et al., 2008).
689 Moreover, this age estimates are consistent with the onset of clastic continental sedimentation in the
690 Ouarzazate Basin during the Langhian (16-14 Ma) following an Oligocene to early Miocene
691 sedimentary hiatus (El Harfi et al., 2001; Teson et al., 2010). Finally, the estimated time of
692 topographic rejuvenation in the Anti-Atlas agrees with the onset of contractional deformation in the
693 High Atlas Mountains, suggesting that there may a genetic link between the two events (Figure 12b;
694 Lanari et al., 2020a). Overall, these observations suggest that the topographic rejuvenation started
695 most likely at ~ 14-10 Ma.

696

697 **5.4 Topographic evolution and surface uplift history of Siroua Massif**

698 The Siroua Massif is also characterized by a transient topography as documented by the high-standing
699 basement representing the substratum of the Mio-Pliocene volcanic edifice (Figure 8; Missenard et
700 al., 2008). Along the southern margin of the massif, this transient topography is delimited by a pattern
701 of non-lithological knickpoints lying over 2000 m of elevation that on map view are subparallel to
702 the AAMF and cluster along similar χ values (Figure 8c). The projection of the stream profiles against
703 the AAMF documents about 350 m of surface uplift in the hanging wall of the fault in response to
704 fault activity (Figures 6a, 8c, 8d and 14). Furthermore, short- and long-term denudation rates,
705 estimated from ¹⁰Be cosmogenic concentration and eroded volcanic rocks from the Siroua Massif,

706 indicates averaged values of 40-50 m/Myr and 10-20 m/Myr in the adjusted topography downstream
 707 and upstream of non-lithological knickpoints, respectively (Clementucci et al., 2022).



708 **Figure 13.** (a) Tridimensional view of the Anti-Atlas region. (b) Swath profile from north to south, showing
 709 the local relief and maximum elevation of High Atlas, Siroua Massif and Anti-Atlas. wl: filtered topographic
 710 windows. Note that the erosional surface of the Siroua Massif and the Anti-Atlas is standing at more than 2000
 711 m and predated the onset of uplift in the Anti-Atlas and Siroua regions.
 712

713
 714 This suggests that the summit erosional surface of the AA must be the same surface underneath the
 715 Siroua volcanic edifice that has been uplifted by the AAMF (Figures 8d and 14b). These conclusions
 716 are also consistent with the occurrence of the same volcanic units at different elevations, in the
 717 hanging wall and the footwall of the AAMF. Some of these lava flows are also exposed along the
 718 steep landscape of the hanging wall of the AAMF downstream of the non-lithological knickpoints.

719 There, they present columnar jointed basalts orthogonal to the valley bottom, suggesting that the
720 valley must have been a reference cooling surface at ~ 4 Ma when the lava was emplaced (Figure 6d).
721 Consequently, the transient fluvial incision associated with the activity of the AAMF must have
722 started earlier than 4 Ma.

723 These data allow estimating the contribution of the different mechanisms that produced the modern
724 elevation of ~ 3300 m. Specifically the initial landscape could have been ~ 1000 m of elevation, like
725 the AA (see previous section), while the occurrence of basement rocks beneath the volcano at ~ 2500
726 m of elevation indicates that volcanic edifice contribute for ~ 800 m high. By subtracting to the
727 modern elevation, 1000 (paleo-topography) and 800 m (volcanic building), we can estimate a surface
728 uplift for the Siroua Massif of ~ 1500 m, where at least 350 meters result from the activity of the
729 AAMF (Figure 13b). A fraction of this ~ 350 m, however, may be associated with the injection of
730 magma at depth as suggested by the dome-like geometry ($\sim 50 \times 100$ km) of the basement beneath
731 the Siroua volcano (Missenard et al., 2008). In conclusion, if we exclude the ~ 350 m described above,
732 the regional surface uplift in the Siroua Massif will be ~ 1150 m in agreement with estimates from the
733 central sectors of the Anti-Atlas.

734

735 **5.5 Causes of surface uplift and topographic expression of the Anti-Atlas and Siroua Massif**

736 The intracontinental orographic system of the Atlas Mountains represents a natural laboratory for
737 studying the interaction between deep-seated, crustal, and surface processes. The amount of tectonic
738 (crustal) shortening and thickening is limited and in the order of 12 to 35% in the central HA
739 (Beauchamp et al., 1999; Gomez et al., 2000; Teixell et al., 2003; Domènech et al., 2016; Fekkak et
740 al., 2018; Lanari et al., 2020b) and less than 10% in the MA (Gomez et al., 1998; Arboleya et al.,
741 2004). Geophysical evidences, such as heat flow (Rimi et al., 1999; Teixell et al., 2005; Zeyen et al.,
742 2005), gravity anomalies (Ayarza et al., 2005; Missenard et al., 2006), seismological constraints
743 (Seber et al., 1996; Palomeras et al., 2013; Bezada et al., 2013; Miller and Becker, 2014; Spieker et
744 al., 2014) and seismic reflection data (Ayarza et al., 2014) point toward an insufficient thickness of

745 the crust to explain the observed high topography and the anomalously thin lithosphere (~65 km).
746 Geological and geomorphological constraints are consistent with these observations suggesting a
747 contribution of deep-seated processes to the recent regional uplift (Babault et al., 2008;
748 Benabdellouahed et al., 2017). Moreover, the late Miocene intraplate volcanism of the AA (Siroua
749 and Saghro Massif) and MA is almost coeval with the main phase of regional uplift inferred in this
750 study for the AA and from thermochronological data in the HA, indicating the occurrence of
751 magmatism during mountain building processes (Berrahma & Delaloye, 1989; El Azzouzi et al.,
752 1999; 2010).

753 The Atlas system displays short wavelength (10 – 50 km) topographic signals most likely linked to
754 processes operating at crustal scale such as shortening (Teixell et al., 2003; Missenard et al., 2006;
755 Faccenna and Becker, 2020) and long wavelength (> 100 km) signals associated with deeper
756 processes (Figure 13b). Examples of the former and the latter include the HA Mountains and the
757 Meseta, respectively, although multiple processes may concur to produce the observed topography
758 (Teixell et al., 2003; Lanari et al., 2020a). Our study documents that the AA Mountains consists of a
759 topographic swell that has an along- and across-strike wavelength of ~ 600 and 100 km, respectively
760 and a maximum amplitude of ~1100 m (Figures 9b, 10a and 13). This is well underlined by the high-
761 standing surface, which is warped along these two major directions (Figure 13a). The maximum
762 values of surface uplift are observed in the central sector of the AA and appear to be slightly lower
763 than those extracted from the adjacent Siroua Massif where surface uplift results from the sum of a
764 long wavelength, regional uplift, and multiple crustal-scale signals such as faulting along the AAMF
765 and possibly local magma injection. The occurrence of magma intrusions may be testified by the
766 occurrence of a localized topographic bulge on the somital erosional surface of the Siroua that stands
767 beneath the volcano. This would explain the short-wavelength topographic signal observed in figure
768 13b (Missenard et al., 2008). Similar geomorphic expressions have been described in different
769 settings where magma injection at variable depth in the crust generated a few hundred meters of uplift
770 over different timescales (Singer et al., 2018; Townsend, 2022). This process could have also

771 occurred in the Saghro mountain peak of the eastern Anti-Atlas where a localized maximum
772 topography associated with a few volcanic eruptions is observed (Figure 2e).

773 The long wavelength topographic swell (600 x 100 km) of the AA, however, still needs to be
774 explained (Figure 13a). Tectonic uplift associated with faulting along a 600-km-long (along strike)
775 crustal scale ramp appears to be unlikely because the AA domain has been tectonically quiescent
776 throughout the entire Cenozoic (Frizon de Lamotte et al., 2009; Lanari et al., 2020a; Clementucci et
777 al., 2022). Furthermore, to explain the ~100-km-long across strike wavelength, one would need a
778 fault-bend-fold mechanisms with a deep flat rooted into the basement, a ramp, and a shallower flat
779 accommodating at least 100 km of displacement. This appears to be quite unlikely, especially if one
780 considers that the AA is a Variscan orogen and not an undisturbed sedimentary multi-layer that may
781 have been detached along stratigraphic horizons. Furthermore, there are no evaporite layers in the pre
782 Variscan sedimentary sequence that could accommodate such an amount of displacement, as
783 observed in the adjacent HA (Lanari et al., 2020b). These observations suggest that the ~ 600 x 100
784 km AA topographic swell could be only explained through deeper, mantle driven processes.

785 A major deep-seated mechanism, which has been invoked as main contributor in the construction of
786 elevated orogenic plateaus is the delamination of the lithospheric mantle and possibly of the lower
787 crust (*e.g.*, Garzzone et al., 2006; Hatzfeld and Molnar, 2010). This would change the density structure
788 of lithosphere through the sinking of a relatively dense lithospheric mantle and the rise of a hotter and
789 less dense asthenosphere. This hypothesis requires the occurrence of a previously thickened
790 lithosphere, which does not appear to be the case for the AA where crustal shortening and thickening
791 processes following the Variscan orogeny have been very limited (Burkhard et al., 2006). Moreover,
792 the occurrence of a thickened lithosphere preceding mantle delamination would have produced
793 subsidence and sedimentation before the onset of uplift (*e.g.*, DeCelles et al., 2009), and this is not
794 supported by stratigraphic data, which suggest subaerial erosional conditions at least since the late
795 Cretaceous (Gouiza et al., 2017; Charton et al., 2021). In intracontinental areas, another possible
796 mechanism is the upwelling of asthenospheric mantle (Duggen et al., 2009; Faccenna et al., 2013;

797 Miller et al., 2013; Olivetti et al., 2016; Faccenna and Becker, 2020). Independently from the
798 geodynamic setting and the main reasons for such a rise, the upward movement of hot asthenospheric
799 mantle generates a deflection of the lithosphere in association with adiabatic melting and production
800 of a magma with an alkaline signature (*e.g.*, Wilson and Downes, 1991; Gibson et al., 2006). This
801 process will generate a strong positive free-air gravity anomaly (*e.g.*, McKenzie and Fairhead, 1997;
802 Watts, 2001; Faccenna and Becker, 2020). This appears to be the case for the AA, where all
803 observations are met (Berrahma and Delaloye, 1989; Spieker et al., 2014). Moreover, the wavelength
804 recorded in the topography of the AA is compatible with other examples from the oceanic floor,
805 where larger-scale swell signals have been observed (Hoggard et al., 2017; Cserepes et al., 2000).
806 Tomography analysis also suggests mantle upwelling in NW Africa, which may be associated with
807 plume activity recorded in the Atlantic Ocean in the Canary Islands (Duggen et al., 2009; Civiero et
808 al., 2019). In conclusion, our topographic analysis indicates that the topography of the Anti-Atlas
809 appears to be consistent with asthenospheric upwelling processes although the cause of these deep-
810 seated mechanism and the possible genetic link with the Canary Islands remain still poorly
811 understood.

812

813 **Conclusion**

814 Our study allows characterizing modern and past topography of the Anti-Atlas Mountains and
815 inferring its surface uplift history. Specifically, it provides new constrains for deciphering the
816 complex history of topographic growth resulting from the interplay between shallower and deep-
817 seated processes. Here, we show that:

818 1) The landscape is in a transient state and exhibits a main pattern of elevated non-lithological
819 knickpoints that mark a regional transition, from high to low values of topographic and channel
820 metrics. The topography upstream of the non-lithological knickpoints can be described as an erosional
821 surface (relict landscape) that plunges along and across strike. Our paleo-topographic reconstruction

822 suggests that such an erosional surface formed a subdued topography with a few local peaks in the
823 order of ~ 1000 m of elevation that started to develop from the late Cretaceous.

824 2) The magnitude of surface uplift increases along-strike from the western Anti-Atlas (500 m) to the
825 central Anti-Atlas (1100 m) and the Siroua Massif (1500 m), where the latter represents a main Mio-
826 Pliocene regional volcanic centre. Geometrically, the surface uplift can be described as $\sim 600 \times 100$
827 km swell. The initiation of knickpoints migration marks the onset of topographic rejuvenation and is
828 estimated to be middle-late Miocene (9.9 ± 4.1 Ma). This is consistent with the middle-late Miocene
829 initiation of a main tectonic phase in the High Atlas and with volcanism in the Siroua and the Saghro
830 Massifs. Our data indicate that surface uplift occurred at rates ranging from 40-110 m/Myr to 70-180
831 m/Myr (averaged over time scales of 14 and 5.8 Ma) from the western to the central Anti-Atlas. The
832 lower range of uplift rates seems to be more likely because it agrees with rates derived from available
833 ^{10}Be erosion rates. This observation, together with the aforementioned geological evidence, suggests
834 that the initiation of surface uplift (i.e., the topographic rejuvenation) in the Anti-Atlas and the Siroua
835 Massif occurred most likely at 14-10 Ma.

836 3) Surface uplift in the Siroua Massif is higher than in the Anti-Atlas because it results from the
837 contribution of different processes, such as the long-wavelength regional component (i.e., the
838 topographic swell; ~ 1150 m), the growth of a volcanic edifice (~ 800 m), faulting along the Anti-
839 Atlas Major Fault (~ 350 m) and possibly magma injection processes which cannot be easily
840 quantified.

841 4) The long wavelength ($\sim 600 \times 100$ km) topographic swell documented through the topographic
842 analysis can be explained by upwelling of hot asthenosphere, which is also responsible for the
843 generation of magmatism in the Siroua and Saghro Massif. Although the genesis of such a rise is
844 unknown, it appears to be the only feasible mechanisms.

845

846 **Acknowledgments**

847 This study is part of the PhD thesis of RC at the University of Roma Tre (PhD Cycle XXXIV). It was
848 supported by the PhD School of Roma Tre and grant “Vinci 2020” awarded to RC (Number: C2-
849 1403). PB was supported by the MIUR (Ministry of Education University and Research), Excellence
850 Department Initiative, Art. 1, com. 314-337, Law 232/2016. The ASTER (CEREGE, Aix-en-
851 Provence) AMS national facility, is supported by the INSU/CNRS, and the IRD. We thank P-H Blard,
852 S. Willett and T. Schildgen for revising this manuscript as a chapter of the PhD Thesis of the RC, to
853 L. Benedetti, M. Della Seta, P. Molin and O. Bellier for providing valuable comments during the PhD
854 defence of RC. The Supporting Information is available online (<https://osf.io/fnvke/>).

855

856 References

- 857 Ait Malek, H. A., Gasquet, D., Bertrand, J. M., & Leterrier, J. (1998). Géochronologie U-Pb sur zircon de
858 granitoïdes éburnéens et panafricains dans les boutonnières protérozoïques d'Igherm, du Kerdous et
859 du Bas Drâa (Anti-Atlas occidental, Maroc). *Comptes Rendus de l'Académie des Sciences-Series IIA-*
860 *Earth and Planetary Science*, 327(12), 819-826. [https://doi.org/10.1016/S1251-8050\(99\)80056-1](https://doi.org/10.1016/S1251-8050(99)80056-1)
- 861 Arboleya, M. L., Teixell, A., Charroud, M., & Julivert, M. (2004). A structural transect through the High and
862 Middle Atlas of Morocco. *Journal of African Earth Sciences*, 39(3-5), 319-327.
863 <https://doi.org/10.1016/j.jafrearsci.2004.07.036>
- 864 Arboleya, M.-L., Babault, J., Owen, L. A., Teixell, A., & Finkel, R. C. (2008). Timing and nature of
865 Quaternary fluvial incision in the Ouarzazate foreland basin, Morocco. *Journal of the Geological*
866 *Society*, 165(6), 1059-1073. <https://doi.org/10.1144/0016-76492007-151>
- 867 Ayarza, P., Alvarez-Lobato, F., Teixell, A., Arboleya, M. L., Teson, E., Julivert, M., & Charroud, M. (2005).
868 Crustal structure under the central High Atlas Mountains (Morocco) from geological and gravity
869 data. *Tectonophysics*, 400(1-4), 67-84. <https://doi.org/10.1016/j.tecto.2005.02.009>
- 870 Azizi Samir, M. R., Ferrandini, J., & Tane, J. L. (1990). Tectonique et volcanisme tardi-Pan Africains (580-
871 560 Ma) dans l'Anti-Atlas Central (Maroc): interprétation géodynamique à l'échelle du NW de
872 l'Afrique. *Journal of African Earth Sciences*, 10(3), 549-563. [https://doi.org/10.1016/0899-](https://doi.org/10.1016/0899-5362(90)90105-N)
873 [5362\(90\)90105-N](https://doi.org/10.1016/0899-5362(90)90105-N)

- 874 Babault, J., Teixell, A., Arboleya, M. L., & Charroud, M. (2008). A Late Cenozoic age for long-wavelength
875 surface uplift of the Atlas Mountains of Morocco. *Terra nova*, 20(2), 102-107.
876 <https://doi.org/10.1111/j.1365-3121.2008.00794.x>
- 877 Baldwin, J. A., Whipple, K. X., & Tucker, G. E. (2003). Implications of the shear stress river incision model
878 for the timescale of postorogenic decay of topography. *Journal of Geophysical Research: Solid*
879 *Earth*, 108(B3). <https://doi.org/10.1029/2001JB000550>.
- 880 Balestrieri, M. L., Moratti, G., Bigazzi, G., & Algouti, A. (2009). Neogene exhumation of the Marrakech
881 High Atlas (Morocco) recorded by apatite fission - track analysis. *Terra Nova*, 21(2), 75-82.
882 <https://doi.org/10.1111/j.1365-3121.2008.00857.x>
- 883 Ballato, P., Landgraf, A., Schildgen, T. F., Stockli, D. F., Fox, M., Ghassemi, M. R., Kirby, E., & Strecker,
884 M. R. (2015). The growth of a mountain belt forced by base-level fall: Tectonics and surface
885 processes during the evolution of the Alborz Mountains, N Iran. *Earth and Planetary Science*
886 *Letters*, 425, 204-218. <https://doi.org/10.1016/j.epsl.2015.05.051>
- 887 Barbero, L., Teixell, A., Arboleya, M. L., del Río, P., Reiners, P. W., & Bougadir, B. (2007). Jurassic - to -
888 present thermal history of the central High Atlas (Morocco) assessed by low - temperature
889 thermochronology. *Terra Nova*, 19(1), 58-64. <https://doi.org/10.1111/j.1365-3121.2006.00715.x>
- 890 Baudon, C., Fabuel - Perez, I., & Redfern, J. (2009). Structural style and evolution of a Late Triassic rift
891 basin in the central High Atlas, Morocco: Controls on sediment deposition. *Geological Journal*,
892 44(6), 677-691. <https://doi.org/10.1002/gj.1195>
- 893 Beauchamp, W., Allmendinger, R. W., Barazangi, M., Demnati, A., El Alji, M., & Dahmani, M. (1999).
894 Inversion tectonics and the evolution of the High Atlas Mountains, Morocco, based on a geological -
895 geophysical transect. *Tectonics*, 18(2), 163-184. <https://doi.org/10.1029/1998TC900015>
- 896 Benabdellouahed, M., Klingelhoefer, F., Gutscher, M. A., Rabineau, M., Biari, Y., Hafid, M., Duarte J. C.,
897 Schnabel M., Baltzer A., Pedoja K., Le Roy P., Reichert C., & Sahabi, M. (2017). Recent uplift of
898 the Atlantic Atlas (offshore West Morocco): Tectonic arch and submarine
899 terraces. *Tectonophysics*, 706, 46-58. <https://doi.org/10.1016/j.tecto.2017.03.024>

900 Berlin, M. M., & Anderson, R. S. (2007). Modeling of knickpoint retreat on the Roan Plateau, western
901 Colorado. *Journal of Geophysical Research: Earth Surface*, 112(F3).
902 <https://doi.org/10.1029/2006JF000553>

903 Berrahma, M., & Delaloye, M. (1989). Données géochronologiques nouvelles sur le massif volcanique du
904 Siroua (Anti-Atlas, Maroc). *Journal of African Earth Sciences (and the Middle East)*, 9(3-4), 651-
905 656. [https://doi.org/10.1016/0899-5362\(89\)90049-3](https://doi.org/10.1016/0899-5362(89)90049-3).

906 Berrahma, M. (1995). Etudes pétrologiques des laves récentes du massif du Siroua (Anti-Atlas, Maroc). Éd.
907 du Service Géologique du Maroc.

908 Bezada, M. J., Humphreys, E. D., Davila, J. M., Carbonell, R., Harnafi, M., Palomeras, I., & Levander, A.
909 (2014). Piecewise delamination of Moroccan lithosphere from beneath the Atlas
910 Mountains. *Geochemistry, Geophysics, Geosystems*, 15(4), 975-985.
911 <https://doi.org/10.1002/2013GC005059>

912 Burkhard, M., Caritg, S., Helg, U., Robert-Charrue, C., & Soulimani, A. (2006). Tectonics of the anti-Atlas
913 of Morocco. *Comptes Rendus Geoscience*, 338(1-2), 11-24.
914 <https://doi.org/10.1016/j.crte.2005.11.012>.

915 Calvet, M., Gunnell, Y., & Farines, B. (2015). Flat-topped mountain ranges: Their global distribution and
916 value for understanding the evolution of mountain topography. *Geomorphology*, 241, 255-291.
917 <https://doi.org/10.1016/j.geomorph.2015.04.015>

918 Charton, R., Bertotti, G., Arnould, A. D., Storms, J. E., & Redfern, J. (2021). Low-temperature
919 thermochronology as a control on vertical movements for semi-quantitative source-to-sink analysis:
920 A case study for the Permian to Neogene of Morocco and surroundings. *Basin Research*, 33(2),
921 1337-1383. <https://doi.org/10.1111/bre.12517>

922 Choubert, G. (1952). Histoire géologique du domaine de l'Anti-Atlas. *Notes et Mém. Serv. Géol. Maroc*,
923 (100), 77-172.

924 Civiero, C., Custódio, S., Rawlinson, N., Strak, V., Silveira, G., Arroucau, P., & Corela, C. (2019). Thermal
925 nature of mantle upwellings below the Ibero-western Maghreb region inferred from teleseismic
926 tomography. *Journal of Geophysical Research: Solid Earth*, 124(2), 1781-1801.
927 <https://doi.org/10.1029/2018JB016531>

928 Clementucci, R., Ballato, P., Siame, L. L., Faccenna, C., Yaaqoub, A., Essaifi, A., Leanni, L., & Guillou, V.
929 (2022). Lithological control on topographic relief evolution in a slow tectonic setting (Anti-Atlas,
930 Morocco). *Earth and Planetary Science Letters*, 596, 117788.
931 <https://doi.org/10.1016/j.epsl.2022.117788>.

932 Crosby, B. T., & Whipple, K. X. (2006). Knickpoint initiation and distribution within fluvial networks: 236
933 waterfalls in the Waipaoa River, North Island, New Zealand. *Geomorphology*, 82(1-2), 16-38.
934 <https://doi.org/10.1016/j.geomorph.2005.08.023>

935 Cserepes, L., Christensen, U. R., & Ribe, N. M. (2000). Geoid height versus topography for a plume model
936 of the Hawaiian swell. *Earth and Planetary Science Letters*, 178(1-2), 29-38.
937 [https://doi.org/10.1016/S0012-821X\(00\)00065-0](https://doi.org/10.1016/S0012-821X(00)00065-0)

938 De Beer, C. H., Chevallier, L. P., De Kock, G. S., Gresse, P. G., & Thomas, R. J. (2000). Mémoire explicatif
939 de la carte géologique du Maroc au 1/50 000, Feuille Sirwa. *Notes Mem. Serv. Geol. Maroc*, 395, 86.

940 DeCelles, P. G., Ducea, M. N., Kapp, P., & Zandt, G. (2009). Cyclicality in Cordilleran orogenic
941 systems. *Nature Geoscience*, 2(4), 251-257. <https://doi.org/10.1038/ngeo469>

942 DiBiase, R. A., & Whipple, K. X. (2011). The influence of erosion thresholds and runoff variability on the
943 relationships among topography, climate, and erosion rate. *Journal of Geophysical Research: Earth*
944 *Surface*, 116(F4). <http://dx.doi.org/10.1029/2011JF002095>.

945 Domènech, M., Teixell, A., Babault, J., & Arboleya, M. L. (2015). The inverted Triassic rift of the
946 Marrakech High Atlas: A reappraisal of basin geometries and faulting histories. *Tectonophysics*, 663,
947 177-191. <https://doi.org/10.1016/j.tecto.2015.03.017>

948 Domènech, M., Teixell, A., & Stockli, D. F. (2016). Magnitude of rift - related burial and orogenic
949 contraction in the Marrakech High Atlas revealed by zircon (U - Th)/He thermochronology and
950 thermal modeling. *Tectonics*, 35(11), 2609-2635. <https://doi.org/10.1002/2016TC004283>

951 Domènech, M., Stockli, D. F., & Teixell, A. (2018). Detrital zircon U-Pb provenance and palaeogeography
952 of Triassic rift basins in the Marrakech High Atlas. *Terra Nova*, 30(4), 310-318.
953 <https://doi.org/10.1111/ter.12340>

954 Duggen, S., Hoernle, K. A., Hauff, F., Kluegel, A., Bouabdellah, M., & Thirlwall, M. F. (2009). Flow of
955 Canary mantle plume material through a subcontinental lithospheric corridor beneath Africa to the
956 Mediterranean. *Geology*, 37(3), 283-286. <https://doi.org/10.1130/G25426A.1>

957 Duvall, A., Kirby, E., & Burbank, D. (2004). Tectonic and lithologic controls on bedrock channel profiles
958 and processes in coastal California. *Journal of Geophysical Research: Earth Surface*, 109(F3).
959 <https://doi.org/10.1029/2003JF000086>

960 El Azzouzi, M., Bellon, H., Maury, R. C., Pique, A., Cotten, J., Griffiths, J. B., ... & Hernandez, J. (1999).
961 Evolution of the sources of Moroccan volcanism during the Neogene. *Comptes Rendus de l'Academie*
962 *des Sciences. Serie 2, Sciences de la Terre et des Planetes*, 95-102.

963 El Azzouzi, M. H., Maury, R. C., Bellon, H., Youbi, N., Cotten, J., & Kharbouch, F. (2010). Petrology and
964 K-Ar chronology of the Neogene-quadernary Middle Atlas basaltic province, Morocco. *Bulletin de la*
965 *société géologique de France*, 181(3), 243-257. <https://doi.org/10.2113/gssgfbull.181.3.243>

966 El Harfi, A., Lang, J., Salomon, J., & Chellai, E. H. (2001). Cenozoic sedimentary dynamics of the
967 ouarazate foreland basin (Central High Atlas Mountains, Morocco). *International Journal of Earth*
968 *Sciences*, 90, 393-411. <https://doi.org/10.1007/s005310000115>

969 England, P., & Molnar, P. (1990). Surface uplift, uplift of rocks, and exhumation of rocks. *Geology*, 18(12),
970 1173-1177. [https://doi.org/10.1130/0091-7613\(1990\)018<1173:SUUORA>2.3.CO;2](https://doi.org/10.1130/0091-7613(1990)018<1173:SUUORA>2.3.CO;2)

971 Faccenna, C., Becker, T. W., Conrad, C. P., & Husson, L. (2013). Mountain building and mantle
972 dynamics. *Tectonics*, 32(1), 80-93. <https://doi.org/10.1029/2012TC003176>

973 Faccenna, C., & Becker, T. W. (2020). Topographic expressions of mantle dynamics in the
974 Mediterranean. *Earth-Science Reviews*, 103327. <https://doi.org/10.1016/j.earscirev.2020.103327>

975 Fekkak, A., Ouanaïmi, H., Michard, A., Soulaïmani, A., Ettachfini, E. M., Berrada, I., et al. (2018). Thick -
976 skinned tectonics in a Late Cretaceous - Neogene intracontinental belt (High Atlas Mountains,
977 Morocco): The flat - ramp fault control on basement shortening and cover folding. *Journal of African*
978 *Earth Sciences*, 140, 169-188. <https://doi.org/10.1016/j.jafrearsci.2018.01.008>

979 Flint, J. J. (1974). Stream gradient as a function of order, magnitude, and discharge. *Water Resources*
980 *Research*, 10, 969-973.

981 Forte, A. M., & Whipple, K. X. (2019). Short communication: The Topographic Analysis Kit (TAK) for
982 TopoToolbox, *Earth Surface Dynamics*, 7, 87–95. <https://doi.org/10.5194/esurf-7-87-2019>.

983 Fox, M. (2019). A linear inverse method to reconstruct paleo-topography. *Geomorphology*, 337, 151-164.
984 <https://doi.org/10.1016/j.geomorph.2019.03.034>

985 Fox, M., & Carter, A. (2020). How continuous are the “Relict” landscapes of Southeastern Tibet?. *Frontiers*
986 *in Earth Science*, 8(587597). <https://doi.org/10.3389/feart.2020.587597>

987 Fraissinet, C., El Zouine, M., Morel, J. L., Poisson, A., Andrieux, J., & Faure-Muret, A. (1988). Structural
988 evolution of the southern and northern central High Atlas in Paleogene and Mio-Pliocene times.
989 In *The Atlas System of Morocco* (pp. 273-291). Springer, Berlin, Heidelberg.

990 Frizon de Lamotte, D., Saint Bezar, B., Bracène, R., & Mercier, E. (2000). The two main steps of the Atlas
991 building and geodynamics of the western Mediterranean. *Tectonics*, 19, 740-761.
992 <https://doi.org/10.1029/2000TC900003>

993 Frizon de Lamotte, D., Leturmy, P., Missenard, Y., Khomsi, S., Ruiz, G., Saddiqi, O., Guillocheau F., &
994 Michard A. (2009). Mesozoic and Cenozoic vertical movements in the Atlas system (Algeria,
995 Morocco, Tunisia). An overview: *Tectonophysics*. 475, 9-28.
996 <https://doi.org/10.1016/j.tecto.2008.10.024>.

997 Froitzheim, N., Stets, J., & Wurster, P. (1988). Aspects of western High Atlas tectonics. In *The Atlas System*
998 *of Morocco*, (pp. 219-244). Berlin: Springer.

999 Gallen, S. F., Wegmann, K. W., & Bohnenstiehl, D. R. (2013). Miocene rejuvenation of topographic relief in
1000 the southern Appalachians. *GSA Today*, 23(2), 4-10. <https://doi.org/10.1130/GSATG163A.1>.

1001 Gallen, S. F., & Wegmann, K. W. (2017). River profile response to normal fault growth and linkage: An
1002 example from the Hellenic forearc of south-central Crete, Greece. *Earth Surface Dynamics*, 5(1),
1003 161-186. <https://doi.org/10.5194/esurf-5-161-2017>

1004 Gallen, S. F. (2018). Lithologic controls on landscape dynamics and aquatic species evolution in post-
1005 orogenic mountains. *Earth and Planetary Science Letters*, 493, 150-160.
1006 <https://doi.org/10.1016/j.epsl.2018.04.029>.

- 1007 Garzione, C. N., Molnar, P., Libarkin, J. C., & MacFadden, B. J. (2006). Rapid late Miocene rise of the
1008 Bolivian Altiplano: Evidence for removal of mantle lithosphere. *Earth and Planetary Science*
1009 *Letters*, 241(3-4), 543-556. <https://doi.org/10.1016/j.epsl.2005.11.026>
- 1010 Gasquet, D., Ennih, N., Liégeois, J. P., Soullaimani, A., & Michard, A. (2008). The pan-african belt.
1011 In *Continental evolution: the geology of Morocco* (pp. 33-64). Springer, Berlin, Heidelberg.
1012 https://doi.org/10.1007/978-3-540-77076-3_2
- 1013 Ghorbal, B., Bertotti, G., Foeken, J., & Andriessen, P. (2008). Unexpected Jurassic to Neogene vertical
1014 movements in 'stable' parts of NW Africa revealed by low temperature geochronology. *Terra*
1015 *Nova*, 20(5), 355-363. <https://doi.org/10.1111/j.1365-3121.2008.00828.x>
- 1016 Gibson, S. A., Thompson, R. N., & Day, J. A. (2006). Timescales and mechanisms of plume–lithosphere
1017 interactions: $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology and geochemistry of alkaline igneous rocks from the Paraná–
1018 Etendeka large igneous province. *Earth and Planetary Science Letters*, 251(1-2), 1-17.
1019 <https://doi.org/10.1016/j.epsl.2006.08.004>
- 1020 Gomez, F., Allmendinger, R., Barazangi, M., Er-Raji, A., & Dahmani, M. (1998). Crustal shortening and
1021 vertical strain partitioning in the Middle Atlas Mountains of Morocco. *Tectonics*, 17(4), 520-533.
1022 <https://doi.org/10.1029/98TC01439>
- 1023 Gomez, F., Beauchamp, W., & Barazangi, M. (2000). Role of the Atlas Mountains (northwest Africa) within
1024 the African - Eurasian plate-boundary zone. *Geology*, 28(9), 775-778. [https://doi.org/10.1130/0091-
1025 7613\(2000\)28<775:ROTAMN>2.0.CO;2](https://doi.org/10.1130/0091-7613(2000)28<775:ROTAMN>2.0.CO;2)
- 1026 Goren, L., Fox, M., & Willett, S. D. (2014). Tectonics from fluvial topography using formal linear inversion:
1027 Theory and applications to the Inyo Mountains, California. *Journal of Geophysical Research: Earth*
1028 *Surface*, 119(8), 1651-1681. <https://doi.org/10.1002/2014JF003079>
- 1029 Görler, K., Helmdach, F.-F., Gaemers, P., Heiig, K., Hinsch, W., Mdler, K., et al. (1988). The uplift of the
1030 central High Atlas as deduced from Neogene continental sediments of the Ouarzazate province,
1031 Morocco. In *The Atlas System of Morocco*, (pp. 359-404). Berlin: Springer.
1032 <https://doi.org/10.1007/s00531-016-1325-0>

- 1033 Gouiza, M., Charton, R., Bertotti, G., Andriessen, P., & Storms, J. E. A. (2017). Post-Variscan evolution of
1034 the Anti-Atlas belt of Morocco constrained from low-temperature geochronology. *International*
1035 *Journal of Earth Sciences*, 106(2), 593-616.
- 1036 Guimerà, J., Arboleya, M. L., & Teixell, A. (2011). Structural control on present-day topography of a
1037 basement massif: the Central and Eastern Anti-Atlas (Morocco). *Geologica Acta: an international*
1038 *earth science journal*, 9(1), 55-65. <https://doi.org/10.1344/105.00.0001643>.
- 1039 Hack, J. T. (1957). *Studies of longitudinal stream profiles in Virginia and Maryland* (Vol. 294). US
1040 Government Printing Office.
- 1041 Hafid, M. (2000). Triassic - early Liassic extensional systems and their Tertiary inversion, Essaouira Basin
1042 (Morocco). *Marine and Petroleum Geology*, 17, 409-429. [https://doi.org/10.1016/S0264 -](https://doi.org/10.1016/S0264-8172(98)00081-6)
1043 [8172\(98\)00081 - 6](https://doi.org/10.1016/S0264-8172(98)00081-6)
- 1044 Hancock, G., & Kirwan, M. (2007). Summit erosion rates deduced from ¹⁰Be: Implications for relief
1045 production in the central Appalachians. *Geology*, 35(1), 89-92. <https://doi.org/10.1130/G23147A.1>
- 1046 Harkins, N., Kirby, E., Heimsath, A., Robinson, R., & Reiser, U. (2007). Transient fluvial incision in the
1047 headwaters of the Yellow River, northeastern Tibet, China. *Journal of Geophysical Research: Earth*
1048 *Surface*, 112(F3). <https://doi.org/10.1029/2006JF000570>
- 1049 Hatzfeld, D., & Molnar, P. (2010). Comparisons of the kinematics and deep structures of the Zagros and
1050 Himalaya and of the Iranian and Tibetan plateaus and geodynamic implications. *Reviews of*
1051 *Geophysics*, 48(2). <https://doi.org/10.1029/2009RG000304>
- 1052 Hefferan, K. P., Admou, H., Karson, J. A., & Saquaque, A. (2000). Anti-atlas (Morocco) role in
1053 neoproterozoic western Gondwana reconstruction. *Precambrian Research*, 103(1-2), 89-96.
1054 [https://doi.org/10.1016/S0301-9268\(00\)00078-4](https://doi.org/10.1016/S0301-9268(00)00078-4)
- 1055 Heidarzadeh, G., Ballato, P., Hassanzadeh, J., Ghassemi, M. R., & Strecker, M. R. (2017). Lake overflow
1056 and onset of fluvial incision in the Iranian Plateau: Insights from the Mianeh Basin. *Earth and*
1057 *Planetary Science Letters*, 469, 135-147. <https://doi.org/10.1016/j.epsl.2017.04.019>
- 1058 Helg, U., Burkhard, M., Caritg, S., & Robert-Charrue, C. (2004). Folding and inversion tectonics in the anti-
1059 Atlas of Morocco. *Tectonics*, 23(4). <https://doi.org/10.1029/2003TC001576>

1060 Hoggard, M. J., Winterbourne, J., Czarnota, K., & White, N. (2017). Oceanic residual depth measurements,
1061 the plate cooling model, and global dynamic topography. *Journal of Geophysical Research: Solid*
1062 *Earth*, 122(3), 2328-2372. <https://doi.org/10.1002/2016JB013457>.

1063 Hollard, H., Choubert, G., Bronner, G., Marchand, J., Sougy, J. (1985). Carte géologique du Maroc, scale 1:
1064 1,000,000. *Serv. Carte géol. Maroc*. 260(2).

1065 Howard, A. D., & Kerby, G. (1983). Channel changes in badlands. *Geological Society of America*
1066 *Bulletin*, 94(6), 739-752. [http://dx.doi.org/10.1130/0016-7606\(1983\)94<739:CCIB>2.0.CO;2](http://dx.doi.org/10.1130/0016-7606(1983)94<739:CCIB>2.0.CO;2).

1067 Howard, A. D. (1994). A detachment-limited model of drainage basin evolution. *Water resources*
1068 *research*, 30(7), 2261-2285. <https://doi.org/10.1029/94WR00757>

1069 Hssaine, A. A., & Bridgland, D. (2009). Pliocene-Quaternary fluvial and aeolian records in the Souss Basin,
1070 southwest Morocco: A geomorphological model. *Global and Planetary Change*, 68, 288-296.
1071 <https://doi.org/10.1016/j.gloplacha.2009.03.002>

1072 Jansen, J. D., Codilean, A. T., Bishop, P., & Hoey, T. B. (2010). Scale dependence of lithological control on
1073 topography: Bedrock channel geometry and catchment morphometry in western Scotland. *The*
1074 *Journal of geology*, 118(3), 223-246. <https://doi.org/10.1086/651273>.

1075 Kirby, E., & Whipple, K. (2001). Quantifying differential rock-uplift rates via stream profile
1076 analysis. *Geology*, 29(5), 415-418. [https://doi.org/10.1130/0091-](https://doi.org/10.1130/0091-7613(2001)029<0415:QDRURV>2.0.CO;2)
1077 [7613\(2001\)029<0415:QDRURV>2.0.CO;2](https://doi.org/10.1130/0091-7613(2001)029<0415:QDRURV>2.0.CO;2).

1078 Kirby, E., & Whipple, K. X. (2012). Expression of active tectonics in erosional landscapes. *Journal of*
1079 *Structural Geology*, 44, 54-75. <https://doi.org/10.1016/j.jsg.2012.07.009>.

1080 Lanari, R., Fellin, M. G., Faccenna, C., Balestrieri, M. L., Pazzaglia, F. J., Youbi, N., & Maden, C. (2020a).
1081 Exhumation and surface evolution of the western high atlas and surrounding regions as constrained
1082 by low-temperature thermochronology. *Tectonics*, 39(3), e2019TC005562.
1083 <https://doi.org/10.1029/2019TC005562>

1084 Lanari, R., Faccenna, C., Fellin, M. G., Essaifi, A., Nahid, A., Medina, F., & Youbi, N. (2020b). Tectonic
1085 evolution of the western high Atlas of Morocco: oblique convergence, reactivation, and
1086 transpression. *Tectonics*, 39(3), e2019TC005563. <https://doi.org/10.1029/2019TC005563>

- 1087 Lanari, R., Reitano, R., Giachetta, E., Pazzaglia, F. J., Clementucci, R., Faccenna, C., & Fellin, M. G.
1088 (2022). Is the Anti-Atlas of Morocco still uplifting?. *J. Afr. Earth. Sci.*, 188, 104481.
1089 <https://doi.org/10.1016/j.jafrearsci.2022.104481>
- 1090 Leblanc, M. (1975). Ophiolites précambriennes et gites arseniés de cobalt: Bou Azzer (Maroc). *These*
1091 *Doctorat d'Etat, Faculte des Science Paris VI, Memoires Centre Geologique et Geophysique.*
- 1092 Leprêtre, R., Missenard, Y., Barbarand, J., Gautheron, C., Jouvie, I., & Saddiqi, O. (2018). Polyphased
1093 inversions of an intracontinental rift: Case study of the Marrakech High Atlas, Morocco. *Tectonics*,
1094 37(3), 818-841. <https://doi.org/10.1002/2017TC004693>
- 1095 Malusa, M. G., Polino, R., Feroni, A. C., Ellero, A., Ottria, G., Baidder, L., & Musumeci, G. (2007). Post-
1096 Variscan tectonics in eastern anti-atlas (Morocco). *Terra Nova*, 19(6), 481-489.
1097 <https://doi.org/10.1111/j.1365-3121.2007.00775.x>
- 1098 Mandal, S. K., Lupker, M., Burg, J. P., Valla, P. G., Haghypour, N., & Christl, M. (2015). Spatial variability
1099 of 10Be-derived erosion rates across the southern Peninsular Indian escarpment: A key to landscape
1100 evolution across passive margins. *Earth and Planetary Science Letters*, 425, 154-167.
1101 <https://doi.org/10.1016/j.epsl.2015.05.050>.
- 1102 McKenzie, D., & Fairhead, D. (1997). Estimates of the effective elastic thickness of the continental
1103 lithosphere from Bouguer and free air gravity anomalies. *Journal of Geophysical Research: Solid*
1104 *Earth*, 102(B12), 27523-27552. <https://doi.org/10.1029/97JB02481>.
- 1105 Miller, S. R., Baldwin, S. L., & Fitzgerald, P. G. (2012). Transient fluvial incision and active surface uplift in
1106 the Woodlark Rift of eastern Papua New Guinea. *Lithosphere*, 4(2), 131-149.
1107 <https://doi.org/10.1130/L135.1>
- 1108 Miller, S. R., Sak, P. B., Kirby, E., & Bierman, P. R. (2013). Neogene rejuvenation of central Appalachian
1109 topography: Evidence for differential rock uplift from stream profiles and erosion rates. *Earth and*
1110 *Planetary Science Letters*, 369, 1-12. <https://doi.org/10.1016/j.epsl.2013.04.007>
- 1111 Miller, M. S., & Becker, T. W. (2014). Reactivated lithospheric-scale discontinuities localize dynamic uplift
1112 of the Moroccan Atlas Mountains. *Geology*, 42(1), 35-38. <https://doi.org/10.1130/G34959.1>.

- 1113 Miller, M. S., O'Driscoll, L. J., Butcher, A. J., & Thomas, C. (2015). Imaging Canary Island hotspot material
1114 beneath the lithosphere of Morocco and southern Spain. *Earth and Planetary Science Letters*, 431,
1115 186-194. <https://doi.org/10.1016/j.epsl.2015.09.026>
- 1116 Missenard, Y., Zeyen, H., Frizon de Lamotte, D., Leturmy, P., Petit, C., Sébrier, M., & Saddiqi, O. (2006).
1117 Crustal versus asthenospheric origin of relief of the Atlas Mountains of Morocco. *Journal of*
1118 *Geophysical Research: Solid Earth*, 111(B3). <https://doi.org/10.1029/2005JB003708>
- 1119 Missenard, Y., Saddiqi, O., Barbarand, J., Leturmy, P., Ruiz, G., El Haimer, F. Z., & Frizon de Lamotte, D.
1120 (2008). Cenozoic denudation in the Marrakech High Atlas, Morocco: insight from apatite fission-
1121 track thermochronology. *Terra Nova*, 20(3), 221-228. [https://doi.org/10.1111/j.1365-](https://doi.org/10.1111/j.1365-3121.2008.00810.x)
1122 [3121.2008.00810.x](https://doi.org/10.1111/j.1365-3121.2008.00810.x).
- 1123 Missenard, Y., & Cadoux, A. (2012). Can Moroccan Atlas lithospheric thinning and volcanism be induced
1124 by Edge-Driven Convection?. *Terra Nova*, 24(1), 27-33. [https://doi.org/10.1111/j.1365-](https://doi.org/10.1111/j.1365-3121.2011.01033.x)
1125 [3121.2011.01033.x](https://doi.org/10.1111/j.1365-3121.2011.01033.x)
- 1126 Molnar, P., & England, P. (1990). Late Cenozoic uplift of mountain ranges and global climate change:
1127 chicken or egg?. *Nature*, 346(6279), 29-34. <https://doi.org/10.1038/346029a0>
- 1128 Molnar, P. (2004). Late Cenozoic increase in accumulation rates of terrestrial sediment: How might climate
1129 change have affected erosion rates?. *Annu. Rev. Earth Planet. Sci.*, 32, 67-89.
1130 <https://doi.org/10.1146/annurev.earth.32.091003.143456>
- 1131 Montgomery, D. R., & Foufoula-Georgiou, E. (1993). Channel network source representation using digital
1132 elevation models. *Water Resources Research*, 29(12), 3925-3934.
1133 <https://doi.org/10.1029/93WR02463>
- 1134 Mudd, S. M., Attal, M., Milodowski, D. T., Grieve, S. W., & Valters, D. A. (2014). A statistical framework
1135 to quantify spatial variation in channel gradients using the integral method of channel profile
1136 analysis. *Journal of Geophysical Research: Earth Surface*, 119(2), 138-152.
1137 <https://doi.org/10.1002/2013JF002981>
- 1138 Norton, K. P., von Blanckenburg, F., Schlunegger, F., Schwab, M., & Kubik, P. W. (2008). Cosmogenic
1139 nuclide-based investigation of spatial erosion and hillslope channel coupling in the transient foreland

1140 of the Swiss Alps. *Geomorphology*, 95(3-4), 474-486.
1141 <https://doi.org/10.1016/j.geomorph.2007.07.013>

1142 Olivetti, V., Cyr, A. J., Molin, P., Faccenna, C., & Granger, D. E. (2012). Uplift history of the Sila Massif,
1143 southern Italy, deciphered from cosmogenic ¹⁰Be erosion rates and river longitudinal profile
1144 analysis. *Tectonics*, 31(3). <https://doi.org/10.1029/2011TC003037>

1145 Olivetti, V., Godard, V., Bellier, O., & ASTER team. (2016). Cenozoic rejuvenation events of Massif
1146 Central topography (France): Insights from cosmogenic denudation rates and river profiles. *Earth
1147 and Planetary Science Letters*, 444, 179-191. <https://doi.org/10.1016/j.epsl.2016.03.049>

1148 Palomeras, I., Thurner, S., Levander, A., Liu, K., Villaseñor, A., Carbonell, R., & Harnafi, M. (2014). Finite-
1149 frequency Rayleigh wave tomography of the western Mediterranean: Mapping its lithospheric
1150 structure. *Geochemistry, Geophysics, Geosystems*, 15(1), 140-160.
1151 <https://doi.org/10.1002/2013GC004861>

1152 Pastor, A., Babault, J., Owen, L. A., Teixell, A., & Arboleya, M. L. (2015). Extracting dynamic topography
1153 from river profiles and cosmogenic nuclide geochronology in the Middle Atlas and the High Plateaus
1154 of Morocco. *Tectonophysics*, 663, 95-109. <https://doi.org/10.1016/j.tecto.2015.06.007>

1155 Perron, J. T., & Royden, L. (2013). An integral approach to bedrock river profile analysis. *Earth Surface
1156 Processes and Landforms*, 38(6), 570-576. <https://doi.org/10.1002/esp.3302>.

1157 Piqué, A., & Michard, A. (1989). Moroccan Hercynides; a synopsis; the Paleozoic sedimentary and tectonic
1158 evolution at the northern margin of West Africa. *American Journal of science*, 289(3), 286-330.
1159 <https://doi.org/10.2475/ajs.289.3.286>

1160 Piqué, A., Bouabdelli, M., Soulaïmani, A., Youbi, N., & Illiani, M. (1999). Les conglomérats du PIII
1161 (Protérozoïque terminal) de l'Anti-Atlas (Sud du Maroc): Molasses tardi-Panafricaines, ou marq n
1162 rifting fini-Protérozoïque. *Comptes Rendus Académie Sciences Paris*, 328, 409-414.

1163 Racano, S., Schildgen, T. F., Cosentino, D., & Miller, S. R. (2021). Temporal and spatial variations in rock
1164 uplift from river-profile inversions at the Central Anatolian Plateau southern margin. *Journal of
1165 Geophysical Research: Earth Surface*, 126(8), e2020JF006027.
1166 <https://doi.org/10.1029/2020JF006027>

- 1167 Rimi, A. (1999). Mantle heat flow and geotherms for the main geologic domains in Morocco. *International*
1168 *Journal of Earth Sciences*, 88(3), 458-466. <https://doi.org/10.1007/s005310050278>
- 1169 Roberts, G. G., & White, N. (2010). Estimating uplift rate histories from river profiles using African
1170 examples. *Journal of Geophysical Research: Solid Earth*, 115(B2).
1171 <https://doi.org/10.1029/2009JB006692>
- 1172 Royden, L., & Taylor Perron, J. (2013). Solutions of the stream power equation and application to the
1173 evolution of river longitudinal profiles. *Journal of Geophysical Research: Earth Surface*, 118(2),
1174 497-518. <https://doi.org/10.1002/jgrf.20031>
- 1175 Safran, E. B., Bierman, P. R., Aalto, R., Dunne, T., Whipple, K. X., & Caffee, M. (2005). Erosion rates
1176 driven by channel network incision in the Bolivian Andes. *Earth Surface Processes and Landforms:*
1177 *The Journal of the British Geomorphological Research Group*, 30(8), 1007-1024.
1178 <https://doi.org/10.1002/esp.1259>.
- 1179 Scharf, T. E., Codilean, A. T., De Wit, M., Jansen, J. D., & Kubik, P. W. (2013). Strong rocks sustain ancient
1180 postorogenic topography in southern Africa. *Geology*, 41(3), 331-334.
1181 <https://doi.org/10.1130/G33806.1>.
- 1182 Schildgen, T. F., Cosentino, D., Bookhagen, B., Niedermann, S., Yıldırım, C., Echtler, H., ... & Strecker, M.
1183 R. (2012). Multi-phased uplift of the southern margin of the Central Anatolian plateau, Turkey: A
1184 record of tectonic and upper mantle processes. *Earth and Planetary Science Letters*, 317, 85-95.
1185 <https://doi.org/10.1016/j.epsl.2011.12.003>
- 1186 Schwanghart, W., & Scherler, D. (2014). TopoToolbox 2—MATLAB-based software for topographic
1187 analysis and modeling in Earth surface sciences. *Earth Surface Dynamics*, 2(1), 1-7.
1188 <https://doi.org/10.5194/esurf-2-1-2014>.
- 1189 Schwanghart, W., & Scherler, D. (2020). Divide mobility controls knickpoint migration on the Roan Plateau
1190 (Colorado, USA). *Geology*, 48(7), 698-702. <https://doi.org/10.1130/G47054.1>
- 1191 Seber, D., Barazangi, M., Tadili, B. A., Ramdani, M., Ibenbrahim, A., & Ben Sari, D. (1996). Three-
1192 dimensional upper mantle structure beneath the intraplate Atlas and interplate Rif mountains of
1193 Morocco. *Journal of Geophysical Research: Solid Earth*, 101(B2), 3125-3138.
1194 <https://doi.org/10.1029/95JB03112>

- 1195 Sébrier, M., Siame, L., Zouine, E. M., Winter, T., Missenard, Y., & Leturmy, P. (2006). Active tectonics in
1196 the moroccan high atlas. *Comptes Rendus Geoscience*, 338(1-2), 65-79.
1197 <https://doi.org/10.1016/j.crte.2005.12.001>.
- 1198 Sebti, S., Saddiqi, O., El Haimer, F. Z., Michard, A., Ruiz, G., Bousquet, R., Baidder, L., & de Lamotte, D.
1199 F. (2009). Vertical movements at the fringe of the West African Craton: First zircon fission track
1200 datings from the Anti-Atlas Precambrian basement, Morocco. *Comptes Rendus Geoscience*, 341(1),
1201 71-77. <https://doi.org/10.1016/j.crte.2008.11.006>
- 1202 Sehart, M., Glasmacher, U. A., Stockli, D. F., Jabour, H., & Kluth, O. (2018). The southern Moroccan passive
1203 continental margin: An example of differentiated long-term landscape evolution in
1204 Gondwana. *Gondwana Research*, 53, 129-144. <https://doi.org/10.1016/j.gr.2017.03.013>
- 1205 Sembroni, A., Molin, P., Pazzaglia, F. J., Faccenna, C., & Abebe, B. (2016). Evolution of continental-scale
1206 drainage in response to mantle dynamics and surface processes: An example from the Ethiopian
1207 Highlands. *Geomorphology*, 261, 12-29. <https://doi.org/10.1016/j.geomorph.2016.02.022>
- 1208 Sembroni, A., Molin, P., Soligo, M., Tuccimei, P., Anzalone, E., Billi, A., Franchini, S., Rinaldi, M., &
1209 Tarchini, L. (2020). The uplift of the Adriatic flank of the Apennines since the Middle Pleistocene:
1210 New insights from the Tronto River basin and the Acquasanta Terme Travertine (central Italy).
1211 *Geomorphology*, 352, 106990. <https://doi.org/10.1016/j.geomorph.2019.106990>.
- 1212 Siame, L. L., Sébrier, M., Bellier, O., Bourlès, D., Costa, C., Ahumada, E. A., Gardini, C.E., & Cisneros, H.
1213 (2015). Active basement uplift of Sierra Pie de Palo (Northwestern Argentina): rates and inception
1214 from ^{10}Be cosmogenic nuclide concentrations. *Tectonics*, 34(6), 1129-1153.
1215 <https://doi.org/10.1002/2014TC003771>
- 1216 Singer, B. S., Le Mével, H., Licciardi, J. M., Córdova, L., Tikoff, B., Garibaldi, N., ... & Feigl, K. L. (2018).
1217 Geomorphic expression of rapid Holocene silicic magma reservoir growth beneath Laguna del
1218 Maule, Chile. *Science advances*, 4(6), eaat1513. <https://doi.org/10.1126/sciadv.aat1513>
- 1219 Siravo, G., Molin, P., Sembroni, A., Fellin, M. G., & Faccenna, C. (2021). Tectonically driven drainage
1220 reorganization in the Eastern Cordillera, Colombia. *Geomorphology*, 389, 107847.
1221 <https://doi.org/10.1016/j.geomorph.2021.107847>

- 1222 Smith, A. G., Fox, M., Schwanghart, W., & Carter, A. (2022). Comparing methods for calculating channel
1223 steepness index. *Earth-Science Reviews*, 103970. <https://doi.org/10.1016/j.earscirev.2022.103970>
- 1224 Snyder, N. P., Whipple, K. X., Tucker, G. E., & Merritts, D. J. (2000). Landscape response to tectonic
1225 forcing: Digital elevation model analysis of stream profiles in the Mendocino triple junction region,
1226 northern California. *Geological Society of America Bulletin*, 112(8), 1250-1263.
1227 [https://doi.org/10.1130/0016-7606\(2000\)112<1250:LRTTFD>2.0.CO;2](https://doi.org/10.1130/0016-7606(2000)112<1250:LRTTFD>2.0.CO;2)
- 1228 Soulaïmani, A., Bouabdelli, M., & Piqué, A. (2003). L'extension continentale au Néo-Protérozoïque
1229 supérieur-Cambrien inférieur dans l'Anti-Atlas (Maroc). *Bulletin de la Société géologique de France*.
- 1230 Soulaïmani, A., & Burkhard, M. (2008). The Anti-Atlas chain (Morocco): the southern margin of the
1231 Variscan belt along the edge of the West African Craton. *Geological Society, London, Special
1232 Publications*, 297(1), 433-452. <https://doi.org/10.1144/SP297.20>
- 1233 Soulaïmani, A., Michard, A., Ouanaïmi, H., Baidder, L., Raddi, Y., Saddiqi, O., & Rjimati, E. C. (2014).
1234 Late Ediacaran–Cambrian structures and their reactivation during the Variscan and Alpine cycles in
1235 the Anti-Atlas (Morocco). *Journal of African Earth Sciences*, 98, 94-112.
1236 <https://doi.org/10.1016/j.jafrearsci.2014.04.025>.
- 1237 Spieker, K., Wölbern, I., Thomas, C., Harnafi, M., & El Moudnib, L. (2014). Crustal and upper-mantle
1238 structure beneath the western Atlas Mountains in SW Morocco derived from receiver
1239 functions. *Geophysical Journal International*, 198(3), 1474-1485. <https://doi.org/10.1093/gji/ggu216>
- 1240 Stäblein, G. (1988). Geomorphological aspects of the Quaternary evolution of the Ouarzazate Basin,
1241 southern Morocco. In *The Atlas System of Morocco* (pp. 433-444). Springer, Berlin, Heidelberg.
1242 DOI: <https://doi.org/10.1007/BFb0011603>
- 1243 Stock, J. D., & Montgomery, D. R. (1999). Geologic constraints on bedrock river incision using the stream
1244 power law. *Journal of Geophysical Research: Solid Earth*, 104(B3), 4983-4993.
1245 <https://doi.org/10.1029/98JB02139>.
- 1246 Stock, J., & Dietrich, W. E. (2003). Valley incision by debris flows: Evidence of a topographic
1247 signature. *Water Resources Research*, 39(4). <https://doi.org/10.1029/2001WR001057>
- 1248 Teixell, A., Arboleya, M. L., Julivert, M., & Charroud, M. (2003). Tectonic shortening and topography in the
1249 central High Atlas (Morocco). *Tectonics*, 22(5). <https://doi.org/10.1029/2002TC001460>

1250 Teson, E., & Teixell, A. (2006). Sequence of thrusting and syntectonic sedimentation in the eastern Sub-
1251 Atlas thrust belt (Dades and Mgoun valleys, Morocco). *International Journal of Earth*
1252 *Sciences*, 97(1), 103-113. <https://doi.org/10.1007/s00531-006-0151-1>

1253 Tesón, E., Pueyo, E. L., Teixell, A., Barnolas, A., Agustí, J., & Furió, M. (2010). Magnetostratigraphy of the
1254 Ouarzazate Basin: Implications for the timing of deformation and mountain building in the High
1255 Atlas Mountains of Morocco. *Geodinamica Acta*, 23(4), 151-165. [https://doi.org/10.3166/ga.23.151-](https://doi.org/10.3166/ga.23.151-165)
1256 [165](https://doi.org/10.3166/ga.23.151-165)

1257 Thomas, R. J., Chevallier, L. P., Gresse, P. G., Harmer, R. E., Eglington, B. M., Armstrong, R. A., de Beer,
1258 C.H., Martini, J. E. J., Kock, G. S., Macey, P. H., & Ingram, B. A. (2002). Precambrian evolution of
1259 the Sirwa window, Anti-Atlas orogen, Morocco. *Precambrian Research*, 118(1-2), 1-57.
1260 [https://doi.org/10.1016/S0301-9268\(02\)00075-X](https://doi.org/10.1016/S0301-9268(02)00075-X)

1261 Torsvik, T. H., Rouse, S., Labails, C., & Smethurst, M. A. (2009). A new scheme for the opening of the
1262 South Atlantic Ocean and the dissection of an Aptian salt basin. *Geophysical Journal*
1263 *International*, 177(3), 1315-1333. <https://doi.org/10.1111/j.1365-246X.2009.04137.x>

1264 Townsend, M. (2022). Linking surface deformation to thermal and mechanical magma chamber
1265 processes. *Earth and Planetary Science Letters*, 577, 117272.
1266 <https://doi.org/10.1016/j.epsl.2021.117272>

1267 Tucker, G. E., & van Der Beek, P. (2013). A model for post-orogenic development of a mountain range and
1268 its foreland. *Basin Research*, 25(3), 241-259. <https://doi.org/10.1111/j.1365-2117.2012.00559.x>

1269 Walsh, G. J., Aleinikoff, J. N., Benziane, F., Yazidi, A., & Armstrong, T. R. (2002). U–Pb zircon
1270 geochronology of the Paleoproterozoic Tagragra de Tata inlier and its Neoproterozoic cover, western
1271 Anti-Atlas, Morocco. *Precambrian Research*, 117(1-2), 1-20. [https://doi.org/10.1016/S0301-](https://doi.org/10.1016/S0301-9268(02)00044-X)
1272 [9268\(02\)00044-X](https://doi.org/10.1016/S0301-9268(02)00044-X)

1273 [Watts, A. B. \(2001\). *Isostasy and Flexure of the Lithosphere*. Cambridge University Press.](https://doi.org/10.1016/S0301-9268(02)00044-X)

1274 West, A. J., Fox, M., Walker, R. T., Carter, A., Harris, T., Watts, A. B., & Gantulga, B. (2013). Links
1275 between climate, erosion, uplift, and topography during intracontinental mountain building of the
1276 Hangay Dome, Mongolia. *Geochemistry, Geophysics, Geosystems*, 14(12), 5171-5193.
1277 <https://doi.org/10.1002/2013GC004859>

- 1278 Whipple, K. X., & Tucker, G. E. (1999). Dynamics of the stream-power river incision model: Implications
1279 for height limits of mountain ranges, landscape response timescales, and research needs. *Journal of*
1280 *Geophysical Research: Solid Earth*, 104(B8), 17661-17674. <https://doi.org/10.1029/1999JB900120>.
- 1281 Whipple, K. X., DiBiase, R. A., Ouimet, W. B., & Forte, A. M. (2017). Preservation or piracy: Diagnosing
1282 low-relief, high-elevation surface formation mechanisms. *Geology*, 45(1), 91-94.
1283 <https://doi.org/10.1130/G38490.1>
- 1284 Whittaker, A. C. (2012). How do landscapes record tectonics and climate?. *Lithosphere*, 4(2), 160-164.
1285 <https://doi.org/10.1130/RF.L003.1>
- 1286 Willett, S. D., & Brandon, M. T. (2002). On steady states in mountain belts. *Geology*, 30(2), 175-178.
1287 [https://doi.org/10.1130/0091-7613\(2002\)030<0175:OSSIMB>2.0.CO;2](https://doi.org/10.1130/0091-7613(2002)030<0175:OSSIMB>2.0.CO;2)
- 1288 Wilson, M., & Downes, H. (1991). Tertiary—Quaternary extension-related alkaline magmatism in western
1289 and central Europe. *Journal of Petrology*, 32(4), 811-849. <https://doi.org/10.1093/petrology/32.4.811>
- 1290 Wobus, C., Whipple, K. X., Kirby, E., Snyder, N., Johnson, J., Spyropolou, K., Crosby, B., & Sheehan, D.
1291 (2006). Tectonics from topography: Procedures, promise, and pitfalls. *Special papers-Geological*
1292 *Society of America*, 398, 55. [https://doi.org/10.1130/2006.2398\(04\)](https://doi.org/10.1130/2006.2398(04))
- 1293 Wobus, C. W., Tucker, G. E., & Anderson, R. S. (2010). Does climate change create distinctive patterns of
1294 landscape incision?. *Journal of Geophysical Research: Earth Surface*, 115(F4).
1295 <https://doi.org/10.1029/2009JF001562>
- 1296 Yang, R., Willett, S. D., & Goren, L. (2015). In situ low-relief landscape formation as a result of river
1297 network disruption. *Nature*, 520(7548), 526-529. <https://doi.org/10.1038/nature14354>
- 1298 Zeyen, H., Ayarza, P., Fernández, M., & Rimi, A. (2005). Lithospheric structure under the western African-
1299 European plate boundary: A transect across the Atlas Mountains and the Gulf of Cadiz. *Tectonics*,
1300 24(2). <https://doi.org/10.1029/2004TC001639>

Tectonics1#2 Supporting Information for**Surface uplift and topographic rejuvenation of a tectonically inactive range: Insights from Anti-Atlas and Siroua Massif (Morocco)**

Clementucci R.^{1,2}, Ballato P.¹, Siame L.², Fox M.³, Lanari R.^{1,4}, Sembroni A.¹, Faccenna C.^{1,5}, Yaaqoub A.⁶, Essaifi A.⁶

¹ Dipartimento di Scienze, Università Roma Tre, Largo San Leonardo Murialdo 1, 00146 Rome, Italy.

² Aix-Marseille Univ., CNRS, IRD, INRAE, Collège de France, CEREGE, Aix-en Provence, France.

³ Department of Earth Sciences, University College London, Gower Street, London, WC1E 6BT, United Kingdom.

⁴ Dipartimento di scienze, Università di Firenze, Italia.

⁵ GFZ-German Research Centre for Geosciences, Potsdam, Germany.

⁶ Département de Géologie, FSSM, B.P. 2390, Université Cadi Ayyad, Marrakech, Morocco.

Contents of this file

Figures S1 to S7

Introduction

The supporting information provided in this file (named 1#2) are complemented by additional tables reported in another supporting file (named 2#2). The data reported here comprise:

- Figures S1 and S2 provide a conceptual framework for the distribution of knickpoints in the Anti-Atlas and for the interpretation of the knickpoints in the river longitudinal profiles and the χ versus elevation plot.
- Figure S3 and S5 report additional data concerning the knickpoint and river profile analysis presented in the main text.
- Figure S4 provides estimates of the erodibility parameters, used to constraint the knickpoints celerity model.

- Figure S6 provide the normalized channel steepness map to constraints the linear inverse model (2D river projection).
- Figure S7 provide the information about the cosmogenic nuclides data from Clementucci et al., 2022.

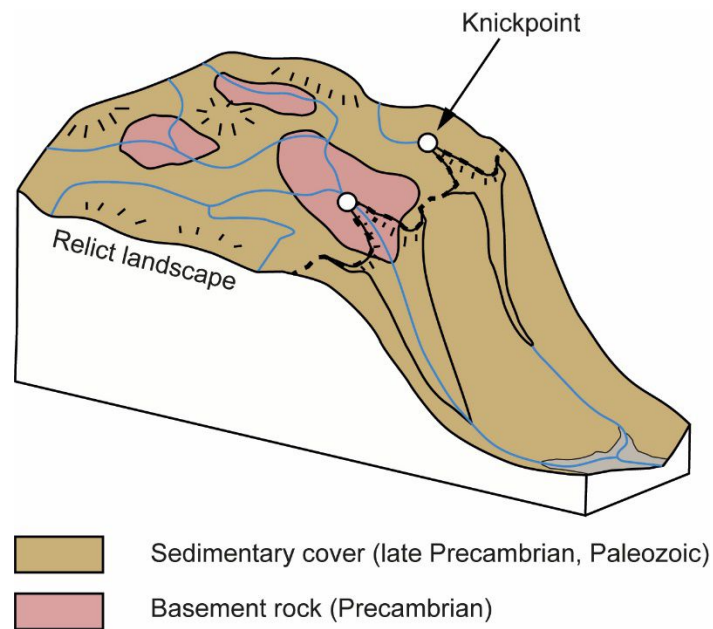


Figure S1. Cartoon of the Anti-Atlas topography. The major lithological contrast is between crystalline/ metamorphic Precambrian rock and late Precambrian/ Paleozoic sedimentary cover.

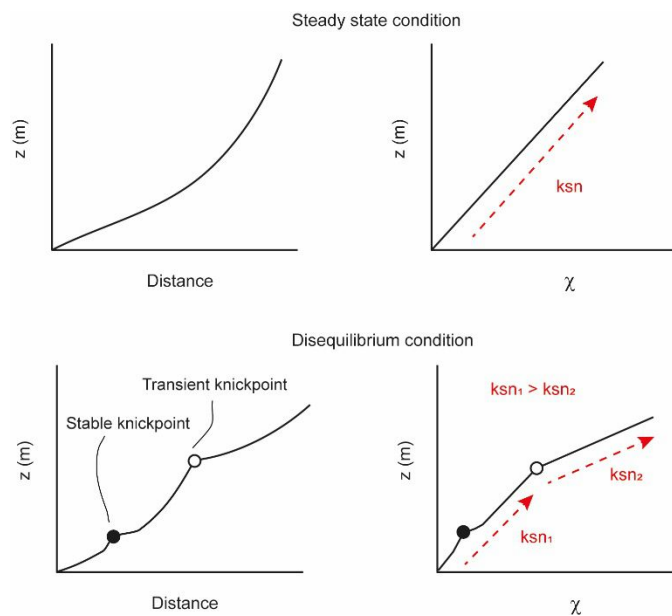


Figure S2. The simplified sketch of longitudinal river profiles (to the left) and χ -plot (to the right) in a steady-state condition and disequilibrium state. In a steady-state condition, the river can be well explained by a single value of k_{sn} , thus a straight line in a χ space (top). In a disequilibrium condition, the river is characterized by knickpoints. The transient knickpoints show an abrupt break in the χ -plot reflected the variation of rock uplift. Whereas the stable knickpoints (lithological, dam, landslide, local heterogeneities, or climate related) is not characterized by k_{sn} variation in along the profile.

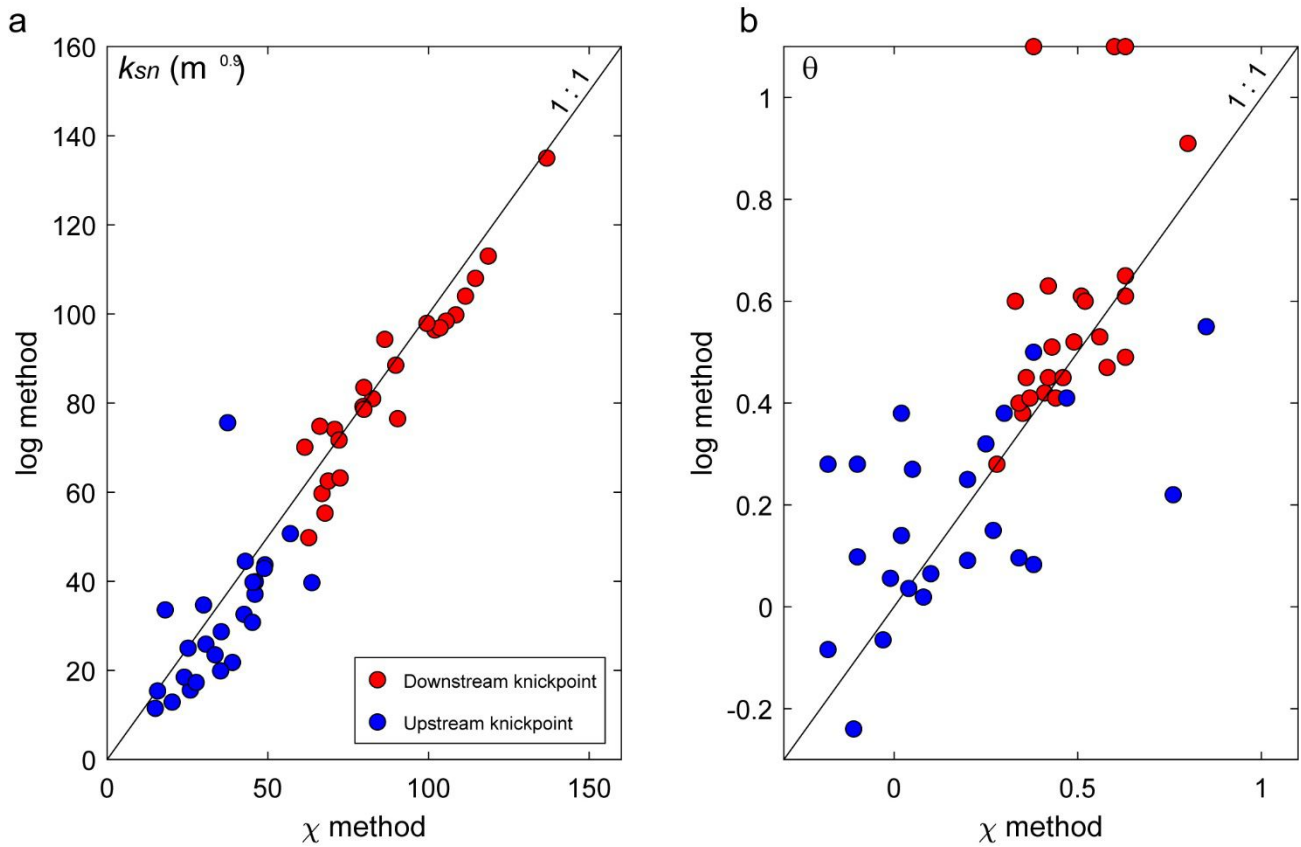


Figure S3. Comparison normalized channel steepness (k_{sn}) and concavity values (θ) using the logS-logA method and integration method (χ). Note the k_{sn} values approximately follow the 1:1 line (plot to left), whereas the concavity values are more scattered (plot to right).

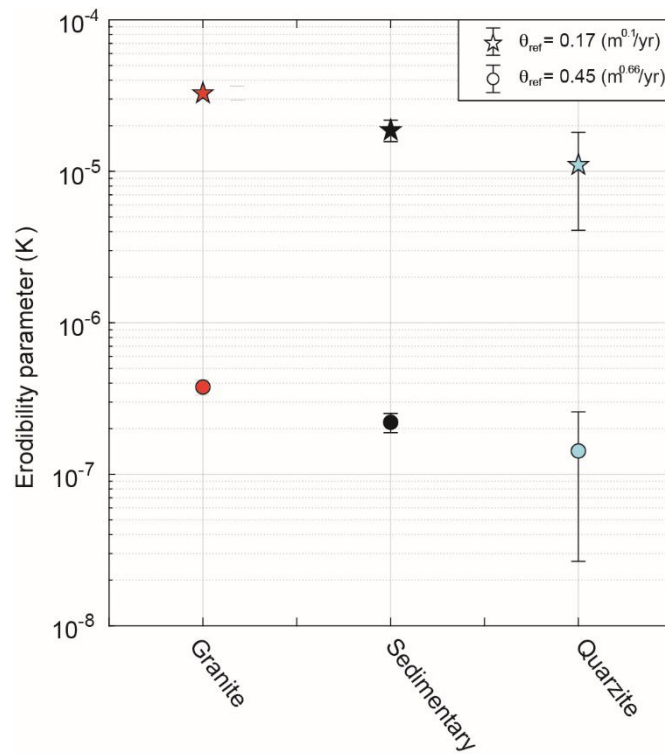


Figure S4. Rock-type versus erodibility parameter. Erodibility values were estimate from the forced linear regression between basin-wide denudation rates and basin-averaged k_{sn} using θ_{ref} of 0.45 and 0.17 (mean value of concavity relative to uplifted relict landscape, Table S1), assuming $n = 1$.

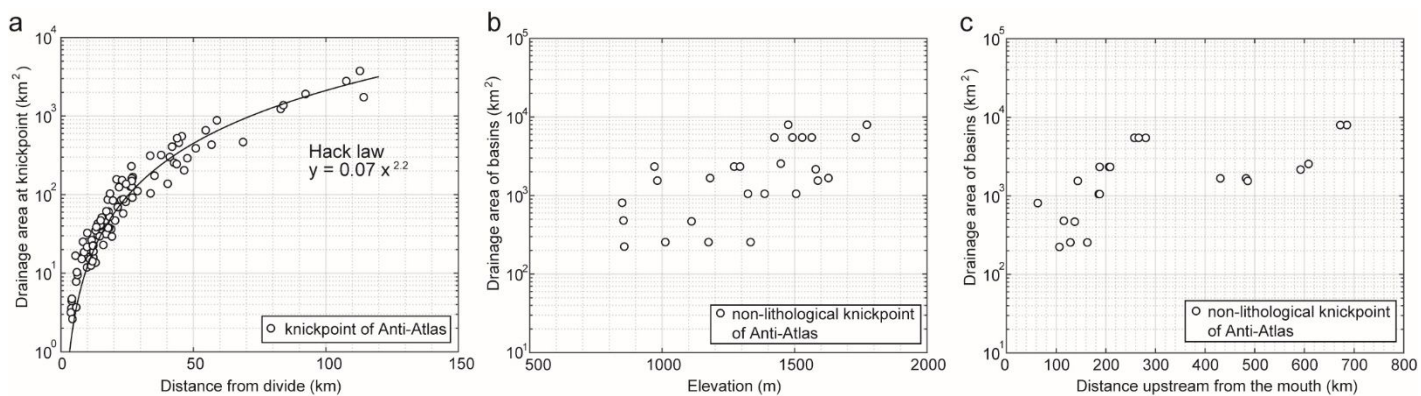


Figure S5. (a) Distance from divide versus upstream area at knickpoints location. Best-fit regression confirm the power law relation of Hack law for knickpoints in the study area. (b) Elevation of knickpoints versus drainage area of basins calculating at 500 m of elevation. (c) Distance from mouth of knickpoints versus drainage area of basins calculating at 500 m of elevation.

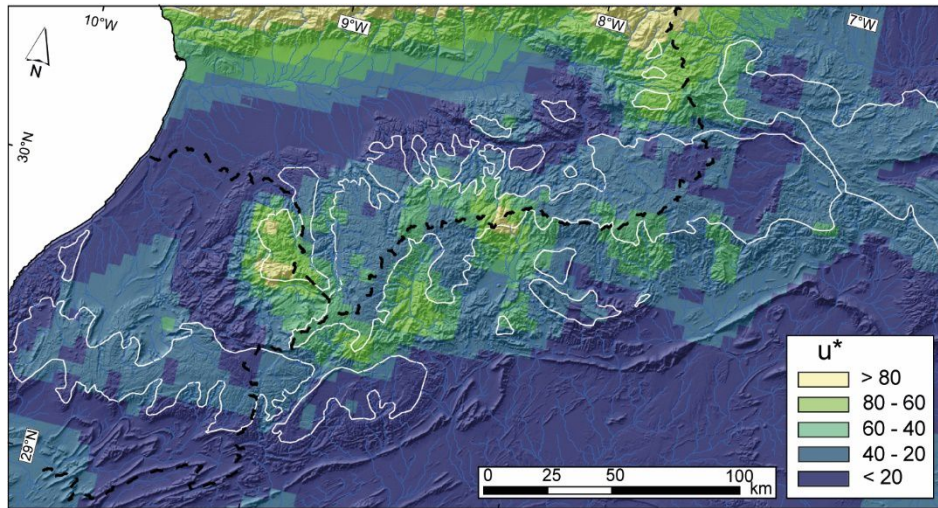


Figure S6. Normalized rock uplift rate (u^* or k_{sn} , using $\theta_{\text{ref}} = 0.45$) map relative to the high-standing relict landscape (white polygons).

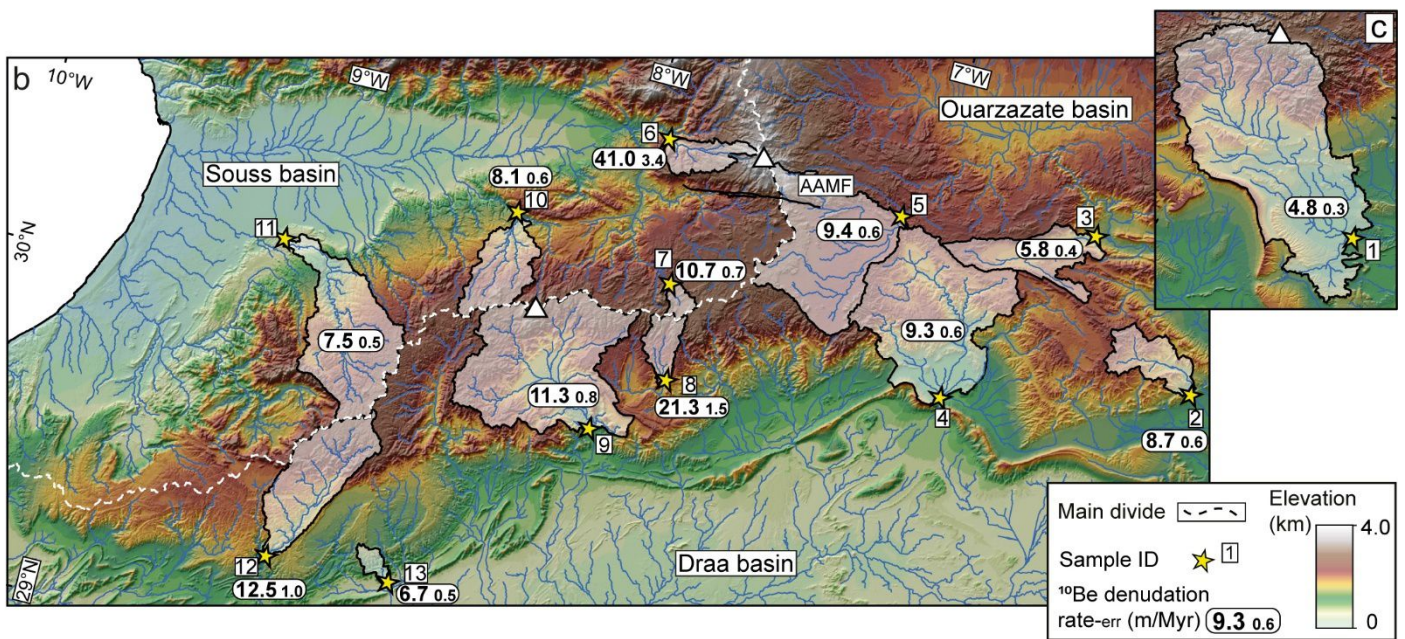
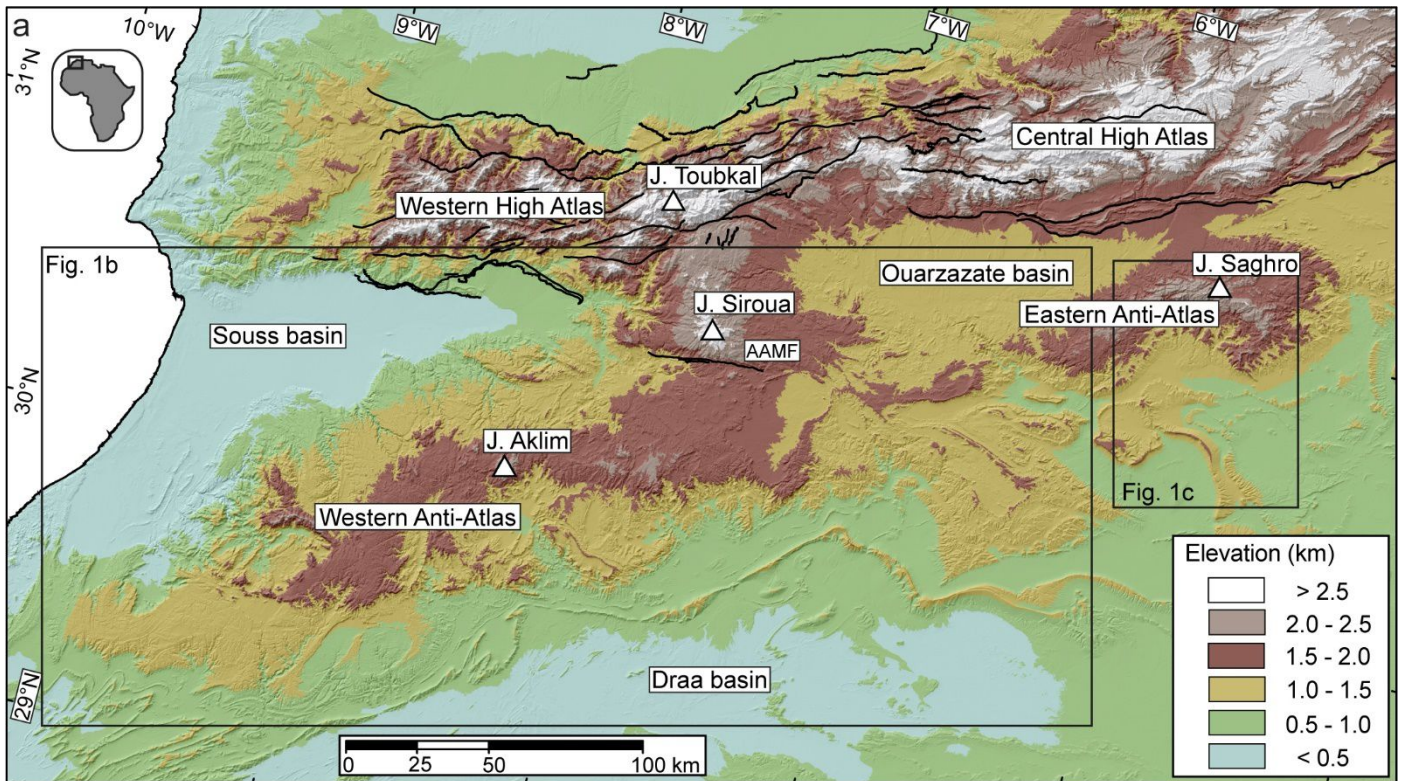


Figure S7. (a) Digital elevation model of the High Atlas and the Anti-Atlas Mountains (SRTM DEM, pixel size 90 m). The black lines are the major tectonic lineaments. (b-c) Topographic map of the Anti-

Atlas and the Siroua Massif, with ^{10}Be sampling locations. The white polygons are the limits of the sampled catchments (Clementucci et al., 2022). The sample ID is used in Table S5 of the supplementary material. AAMF: Anti-Atlas Major Fault.

References

- Clementucci, R., Ballato, P., Siame, L. L., Faccenna, C., Yaaqoub, A., Essaifi, A., Leanni, L., & Guillou, V. (2022). Lithological control on topographic relief evolution in a slow tectonic setting (Anti-Atlas, Morocco). *Earth and Planetary Science Letters*, 596, 117788. <https://doi.org/10.1016/j.epsl.2022.117788>

Tectonics2#2 Supporting Information for**Surface uplift and topographic rejuvenation of a tectonically inactive range: Insights from Anti-Atlas and Siroua Massif (Morocco)**

Clementucci R.^{1,2}, Ballato P.¹, Siame L.², Fox M.³, Lanari R.^{1,4}, Sembroni A.¹, Faccenna C.^{1,5}, Yaaqoub A.⁶, Essaifi A.⁶

¹ Dipartimento di Scienze, Università Roma Tre, Largo San Leonardo Murialdo 1, 00146 Rome, Italy.

² Aix-Marseille Univ., CNRS, IRD, INRAE, Collège de France, CEREGE, Aix-en Provence, France.

³ Department of Earth Sciences, University College London, Gower Street, London, WC1E 6BT, United Kingdom.

⁴ Dipartimento di scienze, Università di Firenze, Italia.

⁵ GFZ-German Research Centre for Geosciences, Potsdam, Germany.

⁶ Département de Géologie, FSSM, B.P. 2390, Université Cadi Ayyad, Marrakech, Morocco.

Contents of this file

Tables S1 to S5

Introduction

The supporting information provided in this file (named 2#2) are complemented by additional figures reported in another supporting file (named 1#2). The data reported here comprise:

- Tables S1 and S2 contain the results of the knickpoint and river profiles of the Anti-Atlas.
- Table S3 summarizes the results of the knickpoints celerity model.
- Table S4 summarizes the results of the 1D river projections.

- Table S5 provides the maximum excavation time estimated from basin-wide denudation rates from Clementucci et al., 2022.

Table S1. Non-lithological knickpoint and river profile data from log-log of channel slope vs. drainage area.

| Stream ^a | Long. | Lat. | Elevation | DFD ^b | DFM ^c | Drainage area | Stream parameters | | | |
|------------------------|----------|-----------|-----------|------------------|------------------|-------------------|---|--------------|------------------------------|---------------|
| | | | | | | | Downstream | | Upstream | |
| | °W | °N | (m) | (m) | (m) | (m ²) | k_{sn} (m ^{0.9}) ^f | θ^g | k_{sn} (m ^{0.9}) | θ |
| NCAA1 | 637465.1 | 3379132.5 | 1731 | 20354 | 280487 | 1.58E+08 | 108 ± 2 | 0.45 ± 0.15 | 11.5 ± 2.4 | 0.27 ± 0.09 |
| NCAA2 | 613556.8 | 3372613.8 | 1424 | 56831 | 257482 | 4.33E+08 | 113 ± 3 | 0.47 ± 0.18 | 32.6 ± 2.9 | 0.056 ± 0.14 |
| NCAA3 | 608426.8 | 3354013.8 | 1565 | 23537 | 266134 | 1.42E+08 | 99.8 ± 2.8 | 0.51 ± 0.13 | 28.7 ± 1.5 | 0.32 ± 0.084 |
| NCAA4 | 596666.8 | 3351313.8 | 1492 | 18508 | 257691 | 1.03E+08 | 104 ± 3 | 0.42 ± 0.11 | 44.5 ± 1.2 | 0.38 ± 0.17 |
| NCAA5 | 589316.8 | 3345523.8 | 1529 | 11566 | 266423 | 2.64E+07 | 76.5 ± 8 | 0.28 ± 0.085 | 34.7 ± 2.6 | 0.096 ± 0.13 |
| NCAA6 | 561056.8 | 3341233.8 | 1506 | 13041 | 185763 | 3.52E+07 | 96.4 ± 2.4 | 0.52 ± 0.092 | 37.1 ± 2.5 | 0.25 ± 0.13 |
| NCAA7 | 555176.8 | 3340753.8 | 1324 | 16987 | 185931 | 4.73E+07 | 98.4 ± 1.6 | 0.38 ± 0.093 | 39.7 ± 4.3 | 0.083 ± 0.11 |
| NCAA8 | 550076.8 | 3341713.8 | 1387 | 20427 | 187603 | 4.69E+07 | 96.9 ± .7 | 0.4 ± 0.089 | 21.8 ± 2.2 | 0.28 ± 0.11 |
| NCAA9 | 543836.8 | 3339913.8 | 1334 | 21396 | 162843 | 69549300 | 97.9 ± 2.8 | 0.53 ± 0.39 | 30.8 ± 4.8 | -0.084 ± 0.14 |
| NCAA10 | 525536.8 | 3328513.8 | 1175 | 24446 | 128875 | 80991900 | 135 ± 5 | 0.45 ± 0.34 | 50.7 ± 3.7 | 0.15 ± 0.11 |
| Mean NCAA ^d | | | | | | | 102.6 ± 4.7 | 0.44 ± 0.02 | 33.2 ± 3.5 | 0.18 ± 0.04 |
| σ^e | | | | | | | 14.8 | 0.07 | 11.1 | 0.14 |
| WAA1 | 497197.4 | 3309532.6 | 1111 | 15696 | 137453 | 54819423 | 70.1 ± 3.8 | 1.1 ± 0.3 | 75.6 ± 3.2 | 0.55 ± 0.077 |
| WAA2 | 443907.4 | 3255074.6 | 857 | 25669 | 106609 | 1.49E+08 | 71.7 ± 6.4 | 0.91 ± 0.3 | 42.9 ± 4.6 | -0.065 ± 0.19 |
| WAA3 | 433249.4 | 3252884.6 | 849 | 29224 | 62967 | 1.11E+08 | 78.6 ± 2.3 | 1.1 ± 0.78 | 19.9 ± 2.8 | 0.098 ± 0.094 |
| WAA4 | 422883.4 | 3241277.6 | 854 | 11251 | 115444 | 12464531 | 49.8 ± 1.4 | 0.6 ± 0.079 | 25.9 ± 1.2 | 0.38 ± 0.31 |
| WAA5 | 428504.4 | 3236605.6 | 1012 | 12228 | 128546 | 22472393 | 55.3 ± 4.1 | 1.1 ± 0.17 | 12.9 ± 1.3 | 0.28 ± 0.2 |
| WAA6 | 445513.4 | 3233028.6 | 982 | 14909 | 143614 | 29352132 | 74.8 ± 1.1 | 0.49 ± 0.087 | 39.8 ± 7.9 | 0.055 ± 0.13 |
| WAA7 | 471939.4 | 3246752.6 | 1272 | 10221 | 206751 | 30897542 | 59.7 ± 0.7 | 0.41 ± 0.037 | 25 ± 5.6 | -0.24 ± 0.41 |
| WAA8 | 482670.4 | 3251424.6 | 1294 | 11699 | 208687 | 33460791 | 62.5 ± 0.7 | 0.45 ± 0.057 | 33.6 ± 5 | 0.019 ± 0.15 |
| WAA9 | 483765.4 | 3234050.6 | 971 | 43222 | 187564 | 2.55E+08 | 63.2 ± 0.8 | 0.63 ± 0.18 | 23.5 ± 1.8 | 0.14 ± 0.11 |
| Mean WAA | | | | | | | 65.07 ± 3.1 | 0.75 ± 0.1 | 33.23 ± 6.1 | 0.14 ± 0.07 |
| σ | | | | | | | 9.45 | 0.3 | 18.5 | 0.22 |
| SCAA1 | 497927.4 | 3246825.6 | 1181 | 27564 | 430725 | 1.62E+08 | 81 ± 1.6 | 0.61 ± 0.19 | 15.6 ± 2.4 | 0.22 ± 0.091 |

| | | | | | | | | | | |
|-----------|----------|-----------|------|-------|--------|----------|------------|--------------|------------|--------------|
| SCAA2 | 520922.4 | 3281573.6 | 1628 | 14831 | 482538 | 39338678 | 79.2 ± 0.8 | 0.5 ± 0.066 | 18.5 ± 2.9 | 0.036 ± 0.19 |
| SCAA4 | 529390.4 | 3272156.6 | 1588 | 6291 | 485819 | 9485620 | 75.3 ± 2.2 | 0.66 ± 0.12 | 8.44 ± 3.8 | 0.041 ± 0.4 |
| SCAA5 | 549465.4 | 3322015.6 | 1580 | 12837 | 592340 | 32309727 | 94.3 ± 0.9 | 0.41 ± 0.065 | 17.3 ± 1.2 | 0.41 ± 0.073 |
| SCAA6 | 564284.4 | 3324643.6 | 1448 | 17885 | 608379 | 1.51E+08 | 83.5 ± 2.2 | 0.6 ± 0.13 | 43.7 ± 1.5 | 0.5 ± 0.1 |
| SCAA7 | 618961.4 | 3336542.6 | 1773 | 13358 | 685609 | 21443896 | 88.5 ± 2 | 0.61 ± 0.089 | 15.4 ± 3.8 | 0.065 ± 0.17 |
| SCAA8 | 636992.4 | 3343550.6 | 1476 | 22860 | 672273 | 86074008 | 74.1 ± 2 | 0.65 ± 0.13 | 39.9 ± 3.8 | 0.091 ± 0.11 |
| Mean SCAA | | | | | | | 82.2 ± 2.7 | 0.57 ± 0.03 | 22.7 ± 5.1 | 0.17 ± 0.07 |
| σ | | | | | | | 7.2 | 0.09 | 13.5 | 0.2 |
| Mean tot | | | | | | | 84.1 ± 3.8 | 0.58 ± 0.04 | 30.4 ± 2.9 | 0.17 ± 0.03 |
| σ | | | | | | | 19.7 | 0.22 | 14.8 | 0.18 |

^a **NCAA**: rivers draining the northern flank of Anti-Atlas; **WAA**: rivers draining the western flank of Anti-Atlas; **SCAA**: rivers draining the southern flank of Anti-Atlas.

^b Distance from divide.

^c Distance from river mouth.

^d Mean and standard error values.

^e Standard deviation of the data.

^f Channel steepness of selected river segments, by using a reference concavity of 0.45.

^g Concavity value calculated by the slope of the line regressed through a log–log plot of channel slope vs. drainage area.

Table S2. Non-lithological knickpoint and river profile data from χ -plot (data from Clementucci et al., 2022).

| Stream ^a | Long. | Lat. | Elevation | DFD ^b | DFM ^c | Drainage area | Stream parameters | | | |
|------------------------|----------|-----------|-----------|------------------|------------------|-------------------|---|-----------------------|------------------------------|-------------|
| | | | | | | | Downstream | | Upstream | |
| | °W | °N | (m) | (m) | (m) | (m ²) | k_{sn} (m ^{0.9}) ^f | θ ^g | k_{sn} (m ^{0.9}) | θ |
| NCAA1 | 637465.1 | 3379132.5 | 1731 | 20354 | 280487 | 1.58E+08 | 114.6 ± 0.2 | 0.42 | 15 ± 0.3 | 0 |
| NCAA2 | 613556.8 | 3372613.8 | 1424 | 56831 | 257482 | 4.33E+08 | 118.5 ± 0.4 | 0.58 | 42.6 ± 0.78 | 0 |
| NCAA3 | 608426.8 | 3354013.8 | 1565 | 23537 | 266134 | 1.42E+08 | 108.5 ± 0.4 | 0.43 | 35.5 ± 0.5 | 0.25 |
| NCAA4 | 596666.8 | 3351313.8 | 1492 | 18508 | 257691 | 1.03E+08 | 111.5 ± 0.43 | 0.41 | 43 ± 0.4 | 0.3 |
| NCAA5 | 589316.8 | 3345523.8 | 1529 | 11566 | 266423 | 2.64E+07 | 90.4 ± 0.5 | 0.28 | 30 ± 0.3 | 0.34 |
| NCAA6 | 561056.8 | 3341233.8 | 1506 | 13041 | 185763 | 3.52E+07 | 102 ± 0.55 | 0.49 | 46 ± 0.7 | 0.2 |
| NCAA7 | 555176.8 | 3340753.8 | 1324 | 16987 | 185931 | 4.73E+07 | 105 ± 0.42 | 0.35 | 63.7 ± 1 | 0.38 |
| NCAA8 | 550076.8 | 3341713.8 | 1387 | 20427 | 187603 | 4.69E+07 | 103 ± 0.43 | 0.34 | 39 ± 0.9 | -0.1 |
| NCAA9 | 543836.8 | 3339913.8 | 1334 | 21396 | 162843 | 69549300 | 99.5 ± 0.47 | 0.56 | 45.2 ± 1.2 | -0.18 |
| NCAA10 | 525536.8 | 3328513.8 | 1175 | 24446 | 128875 | 80991900 | 136.8 ± 0.44 | 0.46 | 57 ± 0.56 | 0.27 |
| Mean NCAA ^d | | | | | | | 109 ± 3.9 | 0.43 ± 0.03 | 41.7 ± 4.2 | 0.15 ± 0.06 |
| σ^e | | | | | | | 12.6 | 0.09 | 13.5 | 0.19 |
| WAA1 | 497197.4 | 3309532.6 | 1111 | 15696 | 137453 | 54819423 | 61.5 ± 0.2 | 0.6 | 37.5 ± 1.17 | 0.86 |
| WAA2 | 443907.4 | 3255074.6 | 857 | 25669 | 106609 | 1.49E+08 | 72.16 ± 0.6 | 0.8 | 49 ± 1.1 | 0.1 |
| WAA3 | 433249.4 | 3252884.6 | 849 | 29224 | 62967 | 1.11E+08 | 79.9 ± 0.4 | 0.63 | 35.3 ± 0.75 | -0.1 |
| WAA4 | 422883.4 | 3241277.6 | 854 | 11251 | 115444 | 12464531 | 62.7 ± 0.3 | 0.33 | 30.7 ± 0.43 | 0.02 |
| WAA5 | 428504.4 | 3236605.6 | 1012 | 12228 | 128546 | 22472393 | 67.8 ± 0.37 | 0.39 | 20.24 ± 0.6 | -0.18 |
| WAA6 | 445513.4 | 3233028.6 | 982 | 14909 | 143614 | 29352132 | 66.2 ± 0.37 | 0.63 | 45.4 ± 3.44 | -0.90 |
| WAA7 | 471939.4 | 3246752.6 | 1272 | 10221 | 206751 | 30897542 | 66.86 ± 0.17 | 0.37 | 25.26 ± 0.85 | -0.10 |
| WAA8 | 482670.4 | 3251424.6 | 1294 | 11699 | 208687 | 33460791 | 68.8 ± 0.14 | 0.36 | 18.1 ± 0.47 | 0.08 |
| WAA9 | 483765.4 | 3234050.6 | 971 | 43222 | 187564 | 2.55E+08 | 72.5 ± 0.21 | 0.42 | 33.6 ± 0.43 | 0.02 |
| Mean WAA | | | | | | | 68.7 ± 1.8 | 0.5 ± 0.05 | 32.7 ± 3.5 | 0 ± 0.14 |
| σ | | | | | | | 5.6 | 0.16 | 10.5 | 0.44 |
| SCAA1 | 497927.4 | 3246825.6 | 1181 | 27564 | 430725 | 1.62E+08 | 82.56 ± 0.21 | 0.51 | 25.9 ± 0.57 | 0.76 |
| SCAA2 | 520922.4 | 3281573.6 | 1628 | 14831 | 482538 | 39338678 | 79.58 ± 0.1 | 0.38 | 24 ± 0.38 | 0.04 |
| SCAA4 | - | - | - | - | - | - | - | - | - | - |
| SCAA5 | 549465.4 | 3322015.6 | 1580 | 12837 | 592340 | 32309727 | 86.4 ± 0.1 | 0.43 | 27.86 ± 0.23 | 0.47 |
| SCAA6 | 564284.4 | 3324643.6 | 1448 | 17885 | 608379 | 1.51E+08 | 79.85 ± 0.26 | 0.52 | 49.15 ± 0.35 | 0.37 |
| SCAA7 | 618961.4 | 3336542.6 | 1773 | 13358 | 685609 | 21443896 | 89.76 ± 0.67 | 0.63 | 15.7 ± 0.5 | -0.03 |
| SCAA8 | 636992.4 | 3343550.6 | 1476 | 22860 | 672273 | 86074008 | 70.85 ± 0.3 | 0.63 | 46.17 ± 0.6 | 0.20 |
| Mean SCAA | | | | | | | 81.5 ± 2.4 | 0.51 ± 0.03 | 31.4 ± 5 | 0.30 ± 0.1 |

| σ | 6.5 | 0.1 | 13.2 | 0.29 |
|----------|-------------|-------------|-------------|-------------|
| Mean tot | 87.9 ± 4.07 | 0.47 ± 0.02 | 36.01 ± 2.5 | 0.12 ± 0.07 |
| σ | 20.35 | 0.12 | 12.3 | 0.34 |

^a **NCAA**: rivers draining the northern flank of Anti-Atlas; **WAA**: rivers draining the western flank of Anti-Atlas; **SCAA**: rivers draining the southern flank of Anti-Atlas.

^b Distance from divide.

^c Distance from river mouth.

^d Mean and standard error values.

^e Standard deviation of the data.

^f Channel steepness of selected river segments, by using a reference concavity of 0.45.

^g Concavity value calculated by finding the value that minimizes the misfit to the linear relationship between χ and elevation.

Table S3. Celerity model of non-lithological knickpoints in the NCAA, WAA and SCAA sectors.

| Stream ^a | Elevation (m) | DFD ^b (m) | DFM ^c (m) | Drainage area (m ²) | Model constrains | | | | | | Model results | | | | | | |
|---------------------|------------------|-------------------------|-------------------------|---------------------------------------|------------------|------------------|-----|------|-----------|------|---------------|-------|-----------|-------|---------------------------|--------|-------------------|
| | | | | | K | | m | | Time (Ma) | | K | m | Time (Ma) | | Model DFM ^d | Misfit | RMSE ^e |
| | | | | | Min | Max | Min | Max | Min | Max | Value | Value | Value | Error | (m) | (m) | |
| NCAA1 | 1731 | 20354 | 280487 | 1.58E+08 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 5.29E-05 | 0.66 | 10.1 | 4.2 | 280488 | 0,392 | 0.198 |
| NCAA3 | 1565 | 23537 | 266134 | 1.42E+08 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 6.14E-05 | 0.69 | 9.9 | 4.4 | 266254 | 0,043 | 0.065 |
| NCAA6 | 1506 | 13041 | 185763 | 3.52E+07 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 6.33E-05 | 0.69 | 9.8 | 4.3 | 185763 | 0,009 | 0.031 |
| NCAA9 | 1334 | 21396 | 162843 | 6.95E+07 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 5.84E-05 | 0.69 | 10.7 | 4.3 | 163048 | 0,001 | 0.012 |
| NCAA10 | 1175 | 24446 | 128875 | 8.10E+07 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 6.04E-05 | 0.67 | 9.3 | 4.1 | 129128 | 0,044 | 0.066 |
| Mean | | | | | | | | | | | 5.93E-05 | 0.68 | 9.9 | 4.3 | | | |
| WAA1 | 1111 | 15696 | 137453 | 5.48E+07 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 5.91E-05 | 0.64 | 9.2 | 4.1 | 137453 | 0,026 | 0.051 |
| WAA2 | 857 | 25669 | 106609 | 1.49E+08 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 5.22E-05 | 0.61 | 9.2 | 4.0 | 106609 | 0,190 | 0.138 |
| WAA3 | 849 | 29224 | 62967 | 1.11E+08 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 5.48E-05 | 0.58 | 9.8 | 4.5 | 62967 | 0,082 | 0.090 |
| WAA5 | 1012 | 12228 | 128546 | 2.25E+07 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 5.53E-05 | 0.65 | 9.1 | 3.9 | 128547 | 0,413 | 0.203 |
| WAA7 | 1272 | 10221 | 206751 | 3.09E+07 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 5.34E-05 | 0.68 | 9.0 | 4.2 | 206752 | 0,348 | 0.186 |
| Mean | | | | | | | | | | | 5.50E-05 | 0.63 | 9.2 | 4.1 | | | |
| SCAA1 | 1181 | 27564 | 430725 | 1.62E+08 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 5.47E-05 | 0.68 | 10.2 | 4.2 | 430725 | 0,166 | 0.128 |
| SCAA2 | 1628 | 14831 | 482538 | 3.93E+07 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 6.69E-05 | 0.71 | 10.8 | 4.0 | 482538 | 0,136 | 0.116 |
| SCAA4 | 1588 | 6291 | 485819 | 9.49E+06 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 7.00E-05 | 0.72 | 11.0 | 3.8 | 485819 | 0,005 | 0.022 |
| SCAA5 | 1580 | 12837 | 592340 | 3.23E+07 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 6.63E-05 | 0.71 | 9.8 | 3.8 | 592340 | 0,216 | 0.147 |
| SCAA6 | 1448 | 17885 | 608379 | 1.51E+08 | 10 ⁻⁷ | 10 ⁻⁴ | 0 | 0.75 | 3.8 | 18.6 | 6.42E-05 | 0.70 | 10.2 | 4.3 | 608379 | 0,119 | 0.109 |
| Mean | | | | | | | | | | | 6.44E-05 | 0.70 | 10.4 | 4.0 | | | |
| Mean tot | | | | | | | | | | | 5.96E-05 | 0.67 | 9.9 | 4.1 | | | |

^a **NCAA**: rivers draining the northern flank of Anti-Atlas; **WAA**: rivers draining the western flank of Anti-Atlas; **SCAA**: rivers draining the southern flank of Anti-Atlas.

^b Distance from divide.

^c Distance from river mouth.

^d Modelled distance from river mouth.

^e **Root** mean square error.

Table S4. Reconstructed relict stream profile and 95% confidence interval.

| Stream ^a | m/n | Drainage divide elevation ^b | Paleo-relief ^c | Elevation knickpoint | ΔZ ^d | Error (2 σ) |
|---------------------|------|--|---------------------------|----------------------|-------------------------|---------------------|
| | | (m) | (m) | (m) | (m) | (m) |
| NCAA1 | 0.45 | 1900 | 738.2 | 1731 | 1161.8 | 18.3 |
| NCAA2 | 0.45 | 1900 | 916.1 | 1424 | 983.9 | 22.3 |
| NCAA3 | 0.45 | 1900 | 1047.1 | 1565 | 852.9 | 17.4 |
| NCAA4 | 0.45 | 1900 | 782.7 | 1492 | 1117.3 | 8.4 |
| NCAA5 | 0.45 | 1900 | 903.2 | 1529 | 996.8 | 11.1 |
| NCAA6 | 0.45 | 1900 | 1055.1 | 1506 | 844.9 | 42.5 |
| NCAA7 | 0.45 | 1900 | 1024.1 | 1324 | 875.9 | 28.3 |
| NCAA8 | 0.45 | 1900 | 1139.8 | 1387 | 760.2 | 38.4 |
| NCAA9 | 0.45 | 1900 | 1054.2 | 1334 | 845.8 | 29.1 |
| NCAA11 | 0.45 | 1900 | 1077.4 | 1175 | 822.6 | 36.5 |
| Mean ^e | | | 973.8 | | 926.2 ± 42.1 | |
| σ^f | | | | | 133.15 | |
| WAA1 | 0.45 | 1400 | 879.5 | 1111 | 520.5 | 31.8 |
| WAA3 | 0.45 | 1400 | 846.8 | 849 | 553.2 | 15.1 |
| WAA4 | 0.45 | 1400 | 941.2 | 854 | 458.8 | 9.1 |
| WAA5 | 0.45 | 1400 | 644.9 | 1012 | 755.1 | 15.7 |
| WAA6 | 0.45 | 1400 | 760.1 | 982 | 639.9 | 13.9 |
| WAA7 | 0.45 | 1400 | 527.9 | 1272 | 872.1 | 28.8 |
| WAA9 | 0.45 | 1400 | 808.6 | 971 | 591.4 | 10.0 |
| Mean | | | 772.7 | | 627.3 ± 54.1 | |
| σ | | | | | 143.17 | |
| SCAA1 | 0.45 | 1900 | 1025.8 | 1181 | 874.2 | 5.9 |
| SCAA2 | 0.45 | 1900 | 783.3 | 1628 | 1116.7 | 18.3 |
| SCAA4 | 0.45 | 1900 | 1089.1 | 1588 | 810.9 | 117.7 |
| SCAA6 | 0.45 | 1900 | 822.9 | 1448 | 1077.1 | 12.7 |
| SCAA7 | 0.45 | 1900 | 824.3 | 1773 | 1075.7 | 58.4 |
| SCAA8 | 0.45 | 1900 | 806.9 | 1476 | 1093.1 | 20.9 |
| Mean | | | 892 | | 1008.0 ± 53.3 | |
| σ | | | | | 130.5 | |

^a **NCAA**: rivers draining the northern flank of Anti-Atlas; **WAA**: rivers draining the western flank of Anti-Atlas. **SCAA**: rivers draining the southern flank of Anti-Atlas.

^b Mean drainage divide elevation.

^c Difference between mean drainage divide elevation and surface elevation (ΔZ).

^d Estimated surface uplift from reconstructed relict stream profiles.

^e Mean and standard error values.

^f Standard deviation of the data.

Table S5. Maximum excavation time estimated from basin-wide denudation rates of Anti-Atlas watersheds in Clementucci et al. (2022).

| Stream | Basin area ^a | Eroded material ^b | Ratio V/A ^c | Denudation rate ^d (m/Myr) | | Excavation time ^e (Myr) | |
|--------|-------------------------|------------------------------|------------------------|---|-------|---------------------------------------|-------|
| ID | (km ²) | (km ³) | (km) | Value | Error | Value | Error |
| 11 | 1200.6 | 148.1 | 0.12 | 7.53 | 0.51 | 16.37 | 0.64 |
| 10 | 480.1 | 58.6 | 0.12 | 8.10 | 0.55 | 15.07 | 0.59 |
| 12 | 842.7 | 117.4 | 0.14 | 12.46 | 0.95 | 11.18 | 0.50 |
| 9 | 1905.1 | 260.0 | 0.14 | 11.34 | 0.77 | 12.04 | 0.47 |
| 7 | 74.1 | 7.0 | 0.09 | 10.72 | 0.71 | 8.79 | 0.34 |
| 8 | 195.8 | 15.9 | 0.08 | 21.32 | 1.47 | 3.81 | 0.15 |
| 4 | 1828.1 | 314.7 | 0.17 | 9.25 | 0.65 | 18.61 | 0.76 |
| 3 | 558.3 | 56.0 | 0.10 | 5.78 | 0.39 | 17.36 | 0.68 |
| 5 | 1438.9 | 229.5 | 0.16 | 9.38 | 0.64 | 17.00 | 0.67 |

Note. ^a Basin area extracted from ArcGIS. ^b Estimated missing volume of eroded material between a smooth surface fitting the preserved summit and present-day topography (e.g., Siame et al., 2015). ^c Ratio between volume of eroded material and basin area. ^d Denudation rate from Clementucci, et al. 2022 (¹⁰Be - derived denudation rate). ^e Ratio between eroded material from river basin and denudation rate. The values are maximum time using the denudation relative only to the relict portion of the landscape, which providing quartz grains

References

- Clementucci, R., Ballato, P., Siame, L. L., Faccenna, C., Yaaqoub, A., Essaifi, A., Leanni, L., & Guillou, V. (2022). Lithological control on topographic relief evolution in a slow tectonic setting (Anti-Atlas, Morocco). *Earth and Planetary Science Letters*, 596, 117788. <https://doi.org/10.1016/j.epsl.2022.117788>
- Siame, L. L., Sébrier, M., Bellier, O., Bourlès, D., Costa, C., Ahumada, E. A., Gardini C. E., & Cisneros, H. (2015). Active basement uplift of Sierra Pie de Palo (Northwestern Argentina): rates and inception from ^{10}Be cosmogenic nuclide concentrations. *Tectonics*, 34(6), 1129-1153. <https://doi.org/10.1002/2014TC003771>