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1 **Contrasting Thermal Evolution of the West African Equatorial and Central Atlantic**
2 **Continental Margins**

3

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15 **ABSTRACT**

16 **The landscape of the West African continental margins is the product of tectonic, thermal**
17 **and surface processes acting in concert during and following the breakup of Gondwana.**
18 **Central Atlantic opening was marked by the emplacement of the Central Atlantic**
19 **Magmatic Province (CAMP) and continental breakup proceeded through Late Jurassic**
20 **and Early Cretaceous divergent tectonics while opening of the Equatorial Atlantic was**
21 **dominated by Early and mid-Cretaceous transform movement. The onshore erosional**
22 **response to these events is poorly constrained yet is a crucial component of our**
23 **understanding of topographic evolution and sediment transfer across continental**
24 **margins. We present new apatite fission-track (AFT) data for 24 samples from Guinea**
25 **and 11 samples from Ivory Coast, and thermal histories from inverse modelling. Our data**
26 **and thermal histories show the following: the thermal effect of the CAMP across Guinea**
27 **and Ivory Coast; rapid cooling along the coast during the early to mid-Cretaceous related**
28 **to erosion of short-wavelength rift-shoulders; moderate cooling across longer**
29 **wavelengths reflecting a pattern of erosion across flexural margin upwarps; and low**
30 **cooling rates from the start of the Cenozoic to present day, consistent with low magnitudes**
31 **of erosion inferred by onshore geomorphological data. We present our results alongside**
32 **the published regional AFT dataset and draw inferences on the thermal and tectonic**
33 **evolution of the onshore margin.**

34

35 **Keywords: Thermochronology; Apatite fission-track; Transform Margin; Tectonics;**
36 **West Africa.**

37 1. Introduction

38

39 Divergent margins are first-order features of global plate tectonics. Constraining the
40 erosional history across the onshore domain of these margins is critical for understanding the
41 response of the Earth's surface to tectonic and climatic processes over million-year timescales
42 (Summerfield, 2000; Séranne and Anka, 2005). The Guinean-Ivory Coast sector of the West
43 African continental margins (Fig. 1) presents a complex case-study due to the multi-phase
44 history of rifting and interaction of a predominantly Jurassic extensional system in the Central
45 Atlantic along the northern Guinean margin (Biari et al., 2017) and an Early Cretaceous
46 transform dominated system in the Equatorial Atlantic along the Ivory Coast margin (Basile et
47 al., 2005).

48 Advances in understanding rifting chronology and the development of offshore
49 sedimentary basins along the margins has revealed links between the margins' tectonic and
50 geodynamic evolution and vertical motions (i.e., uplift and subsidence) across both the onshore
51 and offshore domains (Attoh et al., 2004; Burke and Gunnell, 2008; Grimaud et al., 2018; Ye
52 et al., 2019). Recent work involving mapping of dated lateritic palaeo-landforms (e.g.,
53 Beauvais and Chardon, 2013; Grimaud et al., 2014; Chardon et al., 2016) and inverting river
54 profiles (Lodhia et al., 2019) has revealed very low magnitude denudation and spatially
55 variable uplift from c. 45 Ma and has been interpreted within the context of basin-and-swell
56 mantle driven uplift, long-term sea-level fall, and climate changes during this time. This data
57 has been linked to offshore sediment accumulation data to constrain source-to-sink
58 relationships across the margin during the Cenozoic (Grimaud et al., 2018; Lodhia et al., 2019).
59 However, the paucity of Mesozoic geological markers means that the tectonic and geomorphic
60 evolution of West Africa from the onset of Central Atlantic rifting in the late Jurassic (Labails
61 et al., 2010) to the Cenozoic remains poorly understood. Moreover, there remains a lack of

62 thermochronological data that, in the absence of these markers, can provide insights into
63 periods of thermal or tectonic activity, enhanced erosion, and/or burial by constraining rock
64 thermal histories.

65 The apatite fission-track (AFT) thermochronometer has been used globally to constrain
66 denudation and burial along continental margins (see Wildman et al., 2020 for a review) by
67 providing data on the timing and rate of rock cooling. Apatite fission-track data in West Africa
68 has been limited to onshore studies in Burkina Faso (Gunnell, 2003), Ghana (Lisker et al., 2008;
69 Fernie et al., 2018), Benin (Wildman et al., 2019), and from offshore wells along a marginal
70 ridge in the Ivory Coast Basin (Clift et al., 1998; Bigot-Cormier et al., 2005). We present new
71 AFT data from Guinea and Ivory Coast (Fig. 1b, Fig. 2) and obtain thermal histories for these
72 regions using Bayesian inverse modelling to investigate the thermal response of the upper crust
73 along the Guinean and Ivory Coast margin segments in response to superimposed rifting. We
74 also discuss our data alongside previously published AFT datasets from West Africa to draw
75 more regional conclusions on the tectonic evolution of the region.

76

77 **2. West African geological setting**

78

79 2.1 Summary of rifting and break-up along the West African margin.

80

81 The West African continental margin is segmented by transform faults with the extensional
82 Central Atlantic Ocean segment North of the Guinea fracture zone (FZ) (Biari et al., 2017) (Fig.
83 1a). The Guinea-Liberia margin, Ivory Coast-Ghana margin and Togo-Benin sub-margins are
84 transtensional systems creating pull-apart oblique normal faults, splay faults, and step-over and
85 extensional duplex structures (Nemcok et al. 2016) bounded by the major St. Paul FZ,
86 Romanche FZ and Chain FZ transform faults, respectively (Ye et al., 2017, 2019; Basile et al.,

87 2005; Basile, 2015; Moulin et al., 2010; Heine et al., 2013). Transpressional structures have
88 also been mapped in offshore Ghana and Ivory Coast (Tari, 2006). The structural variations
89 across each of these segments and the onshore and offshore geological record are testament to
90 the prolonged and multi-phase history of rifting (Ye et al., 2017, 2019).

91 In the Central Atlantic, the emplacement of the extensive Central Atlantic Magmatic
92 Province (CAMP) in the early Jurassic is linked to thermal and tectonic processes preceding
93 and during continental rifting (Nomade et al., 2007; Greenroyd et al., 2008; Buitter and Torsvik,
94 2014). As observed along other margins (e.g., Clemson et al., 1997; Gibson et al., 2013; Peace
95 et al., 2018), reactivation of pre-existing structures likely focused the location of rifting (Attoh
96 et al 2005; Mercier de Lépinay et al., 2016). Rifting was predominantly symmetrical within the
97 Central Atlantic, involving normal faulting and the formation of rift basins (Withjack et al.,
98 2012; Biari et al., 2017) until seafloor spreading commenced around 180-200 Ma (Labails et
99 al., 2010; Mercier de Lépinay et al., 2016). While the CAMP coverage was extensive across
100 West Africa and Northeast America (Marzoli et al., 2004), intercontinental deformation at this
101 time was minor (Biari et al., 2017).

102 Along the Equatorial Atlantic are alternating zones of divergent and transform tectonics
103 (Basile, 2015). Oblique transform rifting via an eastward propagating system of strike-slip
104 faults initiated in the Early Cretaceous (c. 140 Ma) forming fault bounded pull-apart basins (Ye
105 et al., 2019). The Guinean Plateau, which sits at the transition between Central and Equatorial
106 Atlantic systems, exhibits evidence of crustal thinning related to the initial Jurassic extension
107 and deformation linked to later Cretaceous transform movement (Mercier de Lépinay et al.,
108 2016). This suggests the Guinea-Liberia margin initially began rifting during Central Atlantic
109 extension followed by the main transform rift phase during the Early Cretaceous (Bennett and
110 Rusk, 2002; Ye et al., 2019).

111 Movement along E-W to ENE-WSW intra-continental transform faults continued through
112 the Valanginian to Aptian (140–113 Ma) (Heine et al., 2013). The end of continental rifting of
113 the Equatorial Atlantic is marked by a Late Albian Breakup Unconformity (107 – 100 Ma)
114 mapped and correlated across diachronous oceanic crust from the Sierra Leone segment to
115 the Togo-Benin segment of the margin (Ye et al., 2019). At the time of break-up, South
116 American plate rotation shifted the tectonic regime of the Ivory Coast-Ghana margin segment
117 from transtension to transpression (Attoh et al., 2004) and an angular breakup unconformity
118 formed beneath Late Albian-Cenomanian Marine sediments (Ferne et al., 2018; Ye et al., 2019).
119 The transition to seafloor spreading is marked by a so-called ‘active-transform’ stage from the
120 Turonian to Santonian (c. 94–84 Ma) as oceanic spreading continues along transform faults
121 (Basile, et al., 2005; Nemcok et al., 2016; Ye et al., 2019; Ferne et al., 2018).

122 The tectonic processes that occurred during the formation of the West African Transform
123 Margin has made this sector the type-example for this form of continental break-up (Basile,
124 2015). The margin exhibits a classic geometry with an outer corner, where the continental
125 margin extends out towards the oceanic accretion axis, and inner corner, where the margin
126 extends inwards towards the continent, created by the divergent margin segments being
127 intersected and bound by major transform faults (Fig. 1a). There is also the presence of a
128 transform fault parallel marginal ridge along the western side of the Romanche FZ where
129 maximum strike-slip deformation has occurred (Fig. 1a, Basile, 2015) and transform marginal
130 plateaus in the Guinean, Liberian and Côte d'Ivoire-Ghana offshore segments (Loncke et al.,
131 2020).

132

133 2.2 Regional geology

134

135 The Guinea-Ivory Coast sector of West Africa is dominated by the Archean to
136 Palaeoproterozoic granitoids and greenstones of the Leo-Man Shield (Fig. 1b). In Guinea,
137 Archean gneisses in the west are separated from the younger Palaeoproterozoic rocks in the
138 east by a series of N-S and NE-SW trending shear zones (Rollinson, 2016). The trend of these
139 shear zones reflects the prevailing structural trend in Ivory Coast, with basement structures
140 accommodating transcurrent displacements during the Palaeoproterozoic orogeny (Lompo,
141 2010). A more complex structural fabric is observed in southeast Guinea where numerous faults
142 crosscut one another and are the product of pre-Cambrian and Pan-African deformation
143 accommodating mostly strike-slip movements (Guiraud et al., 2005) (Fig. 1a,b).

144 Across West Africa, Pan-African basin inversion and thrusting is manifested as two distinct
145 events (Villeneuve, 2005). The first Pan-African event occurred at c. 660 Ma and formed the
146 Mauritanides and Bassarides Belt, with the Mauritanides being reworked during Late
147 Palaeozoic Hercynian deformation (Guiraud et al., 2005). The second event at c. 550–500 Ma
148 produced the NNW-SSE trending Rokelides Belt, which extends through southwestern Guinea,
149 beneath the Bove Basin, and runs adjacent to the coastline through Sierra Leone and Liberia
150 (Villeneuve and Cornee, 1994; Guiraud et al., 2005; Deynoux et al., 2006) (Fig. 1a).

151 The Neoproterozoic to Cambrian Madina-Kouta Basin, comprised of alternating
152 continental and shallow marine sandstones, siltstones and shales, overlies the Leo-Man shield
153 in northern Guinea and extends eastward through Mali and Burkina Faso along the southern
154 margin of the larger Taoudeni Basin (Villeneuve, 2008; Ennih and Liégeois, 2008) (Fig 1b).
155 The Bové Basin in northern Guinea (Fig. 1a) sits on top of the pre-Cambrian basement and
156 Pan-African belts and is comprised of Cambrian–Devonian fluvial-deltaic sandstones and
157 conglomerates, Silurian marine shales with interbedded sandstones, and alternating sequences
158 of Devonian marine sandstones and shales (Villeneuve, 2005; Deynoux et al., 2006). Northwest
159 of Guinea, in Guinea-Bissau, Upper Cretaceous sediments are present and belong to the larger

160 Senegal Basin (Fig. 1a,b), which extends north along the Central Atlantic margin and west into
161 the offshore domain (Brownfield and Charpentier, 2003). These rocks are primarily marine
162 shales, siltstones and sandstones, which form the primary hydrocarbon reservoirs and seals in
163 the Senegal Basin (Brownfield and Charpentier, 2003; Davison, 2005).

164 Emplacement of the CAMP (Fig. 1) is believed to have occurred during a short-lived (c. <
165 1 Myr) period of peak magmatism at c. 201 Ma followed by ongoing activity until c. 192 Ma
166 based on $^{40}\text{Ar}/^{39}\text{Ar}$ and zircon $^{206}\text{Pb}/^{238}\text{U}$ ages (see Marzoli et al. 2018 for a review). Large
167 volumes of mafic intrusive material were emplaced during this magmatism in the form of long,
168 dense dyke swarms and voluminous sills (Deckart et al., 2005). Lava flows can be found and
169 traced across sedimentary basins; however, they are thin (<500 m) and rarely preserved
170 (Marzoli et al., 2018). Outcrop in Guinea is predominantly in the form of layered intrusions
171 forming caps on elevated regions. Estimated pre-erosional volumes of the CAMP have been
172 given in the range of $2\text{--}3 \times 10^6 \text{ km}^3$ (McHone, 2003; Svensen et al. 2018) (Fig. 1b).

173

174 2.3 Topography and geomorphology

175

176 The topography in Guinea is dominated by the Guinea Rise topographic massif (Chardon
177 et al., 2016) that trends NW-SE through Guinea, reaching elevations >1000 m capped by low-
178 relief plateaus (Fig. 2). Inward from the coast, a series of stepped escarpments mark rapid
179 increases in elevation (up to 700 m) in northern Guinea. Across Sierra Leone and Liberia, a
180 100-200 km wide low-relief, coastal plain with a gentle elevation increase up to c. 200 m is
181 observed. Along the Guinea sector of the Central Atlantic margin, the coastal plain is much
182 narrower < 100 km, with a region of high relief and elevations up to c. 600 m, following the
183 structural trend of the Rokelide Belt, separating the coastal region from the Guinea Rise. Four
184 hundred kilometres inland, the topography drops to a low-relief, gently inland sloping region

185 with elevations of c. 400 m (Fig. 2). In the southern part of the Guinea Rise, the elevated plateau
186 is fragmented with the eastern crest of the plateau dissected in a series of N-S trending ridges
187 following the cratonic shear zone. Further south, the relief, comprised of peaks and scattered
188 plateaus, is controlled by the NE trending structural grain with maximum elevations reaching
189 1948m on Mont Loma in NE Sierra Leone. Maximum elevations across Ivory Coast are lower
190 than along the Central Atlantic margin. Elevation gently increases from sea-level at the coast
191 to c. 400–600 m at 500 km inland.

192 The Guinean Rise is one of three major topographic features (including the Hoggar swell
193 in the north and Jos Plateau to the east), that control the large-scale regional drainage of West
194 Africa (Chardon et al., 2016). These features form part of the continental-scale basin-and-swell
195 topography (Burke, 1996), which is proposed to be a product of long-term surface processes in
196 response to the break-up of Gondwana, post-rift tectonics and intraplate mantle driven dynamic
197 uplift (Burke, 1996; Chardon et al., 2016). The drainage is divided by Grimaud et al. (2018)
198 into 4 main domains: (i) Senegambia (ii) Short Atlantic drainages, (iii) Long Atlantic drainages,
199 and (iv) the Niger catchment (Fig. 2). The existing drainage network is suggested to have been
200 relatively stable since the early Oligocene (29 Ma) or possibly earlier at the end of the Eocene
201 (34 Ma) (Chardon et al., 2016).

202 Reconstruction of the Cenozoic drainage evolution has been achieved through a series of
203 studies using ^{40}Ar - ^{39}Ar dating of K-rich Mn oxides and regionally correlated remnant lateritic
204 palaeo-landforms across West Africa (e.g., Vasconcelos et al., 1994; Beauvais et al., 2008;
205 Beauvais and Chardon, 2013; Grimaud et al., 2014; Chardon et al., 2016). Several weathering
206 sequences were identified and include the bauxitic “African Surface”, which developed under
207 humid conditions from the Late Cretaceous to c. 45 Ma, the ferricrete capped “Intermediate”
208 surface, which developed between 29–24 Ma, and three Neogene pediment systems abandoned
209 at c. 11, 6 and 3 Ma, respectively, known as the High, Middle and Low glacia systems

210 (Beauvais and Chardon, 2013; Chardon et al., 2016, 2018; Grimaud et al., 2014, 2018).
211 Dissection of these landforms have enabled facilitated estimates of dissection driven
212 denudation to be made with rates being extremely low since the Eocene (c. 2–10 m/Myr, mean
213 values of 5–7 m/Myr) (Beauvais and Chardon, 2013; Grimaud et al., 2014, 2018).

214

215 2.4 Previous thermochronology studies.

216

217 The new AFT data we present here compliments and adds to existing thermochronology
218 data across the West African Equatorial margin (Fig. 1a). Gunnel (2003) presented central AFT
219 ages from Burkina Faso, ranging from 175 ± 10 to 218 ± 7 Ma, and thermal histories implying
220 slow monotonic cooling since the Mesozoic across the West African craton. In Ghana, Lisker
221 et al. (2008) and Fernie et al. (2018) present AFT data from samples along the coastline and
222 continental interior, respectively. AFT ages from basement samples along the southern Ghanian
223 coastline range from 130 to 415 Ma (Lisker et al., 2008). Thermal history modelling of this
224 data infers two stages of cooling during the Palaeozoic, following the Pan-African Orogeny,
225 and Cretaceous, related to exhumation following continental rifting. Inland, central AFT ages
226 range from 65 ± 11 to 298 ± 18 Ma (Fernie et al., 2018) with thermal histories inferring Late
227 Triassic - Early Jurassic cooling attributed to post-CAMP thermal relaxation, early to mid-
228 Cretaceous cooling (145 – 90 Ma) due to rifting related exhumation, and Late Cretaceous-
229 Cenozoic cooling due to exhumation driven by post-rift tectonic reactivation along the Bole-
230 Nangodi shear zone (Fig. 1a). Further east, along the Benin sector of the margin Wildman et al.
231 (2019) present AFT ages ranging from 106 ± 5 to 401 ± 45 Ma with single-grain apatite (U-
232 Th)/He ages (corrected for alpha-ejection) ranging from 26 ± 4 to 500 ± 95 . These data and
233 associated thermal history modelling are used to advocate for relatively rapid exhumation
234 driven cooling in the Early Cretaceous along the coastal margin and the interior Iullemeden

235 Basin (Fig. 1a) and very slow cooling of the region between the coast Iullemmeden Basin
236 through the Palaeozoic. The thermochronometry study in Benin does not preclude heating
237 related to burial along the coastal basins and interior basins during the mid-Cretaceous (110 –
238 90 Ma) and Late Cretaceous to Early Cenozoic, respectively.

239 Offshore, Clift et al. (1998) and Bigot-Cormier et al. (2005) present AFT data from
240 sedimentary samples collected from Ocean Drilling Program borehole sites through and surface
241 samples along the marginal ridge (Fig. 1a). Clift et al. (1998) present central AFT ages from
242 the borehole sites ranging from 88 ± 4 Ma to 113 ± 4 Ma. Central ages from samples from the
243 southern continental slope of the ridge, reported by Bigot-Cormier et al. (2005) have a
244 comparable but slightly larger range from 64.2 ± 4.7 to 125.3 ± 10.6 Ma, with the majority of
245 AFT ages less than 100 Ma. Thermal history modelling of the data presented by Clift et al.
246 (1998) infers pre-depositional cooling, which represents evolution of the source region (i.e.,
247 the evolving continental margin), between 130 and 110 Ma in three samples and between 110
248 and 100 Ma in another. Minor post-depositional heating and fission-track annealing is inferred
249 following deposition. Clift et al. (1998) suggest the first cooling event was driven by uplift and
250 erosion during transform deformation and a later cooling event by a post-transform period of
251 inversion, respectively. As the majority of the AFT ages presented by Bigot-Cromier et al.
252 (2005) are younger than the samples Lower Cretaceous stratigraphic ages, they advocate
253 significant sedimentary burial of these samples followed by erosion during the ‘active-
254 transform’ stage.

255

256 **3. Apatite fission track thermochronology**

257

258 **3.1 AFT Methods**

259

260 Apatite fission-track ages were obtained from 24 samples from Guinea and 11 from the
261 Ivory Coast using the zeta-calibration external detector method (Hurford and Green, 1983).
262 Apatite fission-track ages were combined with horizontal confined track length measurements
263 to constrain thermal annealing of tracks through the AFT partial annealing zone (PAZ, 60–110
264 $\pm 10^\circ\text{C}$) (Gleadow et al., 1986). The angle of measured track lengths to the crystallographic c-
265 axis was measured to correct for anisotropic annealing (Ketcham et al., 2007) and Dpar (length
266 of the etch pit formed by tracks intersecting the polished grain surface) was measured as a
267 proxy for the bulk compositional influence on track annealing (Carlson et al. 1999; Donelick
268 et al. 2005). Radial plots and track length distributions for all samples can be found in
269 Supplementary Data, Fig. S1 and Fig. S2, respectively.

270

271 3.2 Apatite Fission-Track Results

272

273 3.2.1 Guinea

274

275 Twenty-four samples were collected from Guinea (Table 1; Supplementary Data, Table S1)
276 and can be divided into two regional groups (Fig. 2). In the northwest, 15 samples (Guinea
277 North) form a coast-perpendicular transect that extends from the coast, through the elevated
278 Guinean Rise to c. 400 km inland. Nine samples (Guinea South) cover the south-eastern region
279 of the Guinean Rise at c. 800–1000 m elevation. Central AFT ages across the entire Guinea
280 dataset range from 99.1 ± 13.0 to 204.2 ± 34.3 Ma. Nine samples fail the $P(\chi^2)$ age homogeneity
281 test (i.e., $P(\chi^2) < 0.05$) with four of these samples showing excess single grain age dispersion
282 ($>20\%$) than that expected for a single population (see radial plots in Supplementary Data, Fig.
283 S1). In most of the Guinea samples, Dpar measurements were collected across the entire sample
284 (i.e., not on individual dated grains) and so the relationship between single grain age and Dpar

285 cannot be assessed. In those samples where Dpar was measured on dated grains
286 (Supplementary Data, Fig. S1) no relationship between Dpar and age is apparent suggesting
287 excess scatter cannot be solely attributed to apatite composition influence annealing. Given the
288 high number of tracks for the samples in the Guinean dataset, it is possible that the chi-sq test
289 is penalising the data for being too precise (e.g., Vermeesch, 2021). Mean track lengths (MTL),
290 corrected for their c-axis orientation (c-MTL) range from 13.57 ± 0.12 to 14.78 ± 0.08 with
291 standard deviations of 0.78 to 1.47.

292 The Ivory Coast dataset contains 11 samples (Fig. 2) (Table 1; Supplementary Data, Table
293 S1) forms a broadly coast perpendicular transect from the coast to c. 500 km inland with a
294 gradual elevation gain from sea-level to 500 m. Approximately 350 km from the coast, two
295 samples (YOR-1 and YAL-1) fall to the west of the main transect along an outer ridge of the
296 Guinean rise. Central AFT ages in Ivory Coast range from 114.7 ± 13.1 to 188.9 ± 22.5 Ma, c-
297 MTLs range from 13.99 ± 0.14 to 14.39 ± 0.07 μm with unimodal track length distributions
298 and MTL-SD of 0.74–1.03. Five samples fail the $P(\chi^2)$, however, only one of these samples
299 show excessive dispersion (>20%). As suggested above for the Guinean samples, apatite
300 composition variability may have an influence on creating excess single grain age dispersion
301 but the high track counts in these grains may be a contributing factor causing the $P(\chi^2)$ to fail.

302 Our West African data set yields a relatively narrow range in MTLs 13.57 ± 0.12 to 14.78
303 ± 0.08 over a reasonably large age range 99.1 ± 13.0 to 204.2 ± 34.3 Ma, which does not show
304 any clear correlation (Fig. 3a). The negative correlation between MTL and MTL St. Dev. (Fig.
305 3b) reflects shorter MTLs having a broader track length distribution.

306 Inland from the coast, AFT ages in the Ivory Coast and the northern Guinea increase
307 gradually to a distance of c. 200 km (Fig. 3c). In northern Guinea, this distance corresponds to
308 the region of increasing topography from sea-level at the coast to the highest sampled
309 elevations in the Guinean Rise (i.e., samples GN-06 to GN-08) (Fig. 2). The relationship to

310 topography is less clear in Ivory Coast due to the modest elevation range across the entire
311 country (Fig. 2). However, in general this c. 200 km region corresponds to the low-lying (<
312 200 m) coastal strip.

313 Beyond 200 km inland, AFT ages gradually decrease again (Fig. 3c). This is best observed
314 in the Guinea North dataset, with three samples from Ivory Coast also plotting on this trend. In
315 northern Guinea, the decreasing trend is coincident with a general decrease in elevation from
316 peak elevation >1000m in the Guinean rise to c. 400 m inland (Fig. 2). Although some of the
317 Ivory Coast data agree with the lateral trend in AFT age, the elevation beyond 200 km inland
318 gradually increases to c. 400 m. Four samples from the Ivory Coast profile also appear to form
319 a negative correlation from c. 350 to 500 km inland that is offset from the main data trend (Fig.
320 3c). This type of offset may be attributable to fault offsetting with down-to-the-northeast
321 displacement and differential denudation across the fault. However, more samples would be
322 required to better relate the spatial relationship of the AFT data to fault displacement. Samples
323 from the Guinea South dataset, situated between 300 and 400 km inland, are fairly scattered.
324 Five samples appear to have AFT ages consistent with the main AFT age-distance from coast
325 trend, while two seem more consistent with the offset trend defined by the Ivory Coast samples.

326 The Guinea North dataset shows the strongest AFT age-elevation correlation (Fig. 3d),
327 which connects increasing and decreasing elevations across the Guinean Rise to the AFT age-
328 distance from the coast relationship. The Guinea South data are more scattered but while they
329 do not show a strong correlation, they appear to be consistent with the Guinea North AFT age-
330 elevation trend. Several of the Ivory Coast samples fall on the trend defined by the Guinea
331 North data but at 350–400 m there is more scatter in the Ivory Coast AFT ages. The offset
332 samples here are also the samples that form the offset part of the AFT age-distance from the
333 coast trend for Ivory Coast.

334

335 4. Thermal History Modelling

336

337 4.1 Modelling approach

338

339 Thermal histories for all samples were acquired using QTQt, which incorporates a
340 Bayesian transdimensional Markov Chain Monte Carlo (MCMC) approach to data inversion
341 (Gallagher, 2012). Apatite fission-track data were modelled using the multi-kinetic fission-
342 track annealing model of Ketcham et al. (2007) with anisotropic annealing and compositional
343 influences on the annealing rate taken into account using c-axis projected track lengths and
344 sample average Dpar measurements, respectively. We present the expected thermal history
345 (ExTH) and 95% credible intervals in figure 4 and summarise the modelling results below. We
346 provide a more detailed discussion of the modelling strategy, preliminary testing, alternative
347 model scenarios and plots of the observed data vs. model predicted data in Supplementary Data,
348 Data S1, Table S2, Fig. S3–S9. All QTQt input files are provided in Supplementary Data,
349 Dataset S1.

350 Our modelling strategy involved initially modelling all samples individually and without
351 any additional constraints to generate thermal histories with complexity solely driven by the
352 data of one sample (Supplementary Data, Fig. S3). We then look to combine samples with the
353 aim of reinforcing mutually consistent features observed in individual sample thermal histories
354 and making these parts of the thermal history more robust. We take this step because due to the
355 relatively low precision of AFT thermochronology, the data for a single sample may not
356 robustly constrain detail in the thermal history the model (i.e., timing of maximum heating or
357 inflection points in the thermal history path). Moreover, thermal histories from single samples
358 taken from locations near to one another may result in apparent regional variability in thermal
359 histories between adjacent samples. These variations in thermal histories could be interpreted

360 as having a geological meaning (e.g., migrating patterns of erosion or fault movement causing
361 differential denudation). Before this conclusion can be drawn, it should be assessed whether
362 data from several samples within a region can be explained by a single thermal history that is
363 consistent with all the data and their uncertainties. A multi-sample thermal history may also
364 identify a sample or samples that are inconsistent with data from adjacent samples, and this
365 may indicate a more complex spatial-thermal history relationship.

366 We take a simplistic approach to grouping samples, grouping first into 100 km bins
367 (starting from the coastline), and assessing whether the joint model suitably reproduces all of
368 the grouped data. If some samples are not reproduced, we group the data into 50 km bins, and
369 then, if necessary to fit the data, resort to the individually modelled sample (see Supplementary
370 Data, Data S1, Fig. S4 for further details). We assign a temperature offset based on an assumed
371 geothermal gradient of $25\pm 5^\circ\text{C}/\text{km}$. However, it should be appreciated that these samples do
372 not form a pseudo-vertical profile or are not taken from a borehole profile where the definition
373 of a thermal offset would have more relevance. The grouped samples are typically at similar
374 elevations, have the same present-day surface temperatures and assumed to have had similar
375 palaeo-temperatures. The specified geothermal gradient and the range on this value is used to
376 build some uncertainty into this assumption.

377 We do not explicitly define an initial starting constraint for the models and in some cases
378 the output thermal history appears to begin at temperatures somewhere much colder than the
379 base of the PAZ. In these circumstances it is inferred that rapid cooling from temperatures
380 hotter than the PAZ has occurred immediately before the starting point time.

381

382 4.2 Thermal History Modelling Results

383

384 4.2.1 Guinea North

385

386 Samples GM01, GM02, GN1, GN3 and GN4 reside on the low elevation coastal margin
387 zone that extends 150–200 km inland from the coastline and are modelled together (Fig. 4a).
388 The bauxite capped plateaus and pervasive Middle glaciae relics in that area suggest that this
389 part of the landscape experienced approximately 400 and 1000 m of denudation between 45
390 and 6 Ma, followed by negligible denudation to present-day. Assuming a geothermal gradient
391 of $25\pm 5^\circ\text{C}/\text{km}$ from 45 Ma, and a surface temperature of 20°C , this equates to cooling from a
392 temperature in the range of 28 to 50°C .

393 The geomorphological information can be used as additional constraints to force cooling
394 from 45 and 6 Ma. To prevent additional time-temperature points being proposed after 45 Ma,
395 the lower time range of the prior was reduced to 45 Ma. The early portion of the ExTH (>70
396 Ma) shows the main cooling episode observed at c. 110 Ma. Heating of c. 20°C is then inferred
397 between 70 and 45 Ma, before finally cooling to surface temperatures at present-day as
398 constrained by the geomorphology (Fig. 4b). In supplementary information (Supplementary
399 Data, Fig. S5) we explore the necessity for the Early Cenozoic cooling event to be forbidding
400 reheating and observe that the data can still be reproduced without reheating. However, as this
401 is a very restrictive constraint to impose, and the inferred reheating is fairly minor we continue
402 our discussion with model presented in Fig. 4a.

403 While some ambiguity remains in the ExTH model, we can have confidence in the
404 consistent and recurrent parts of the history. Specifically, this includes an initial period of
405 protracted cooling through the Late Jurassic and Early Cretaceous followed by a phase of rapid
406 cooling in the mid-Cretaceous (110 – 90 Ma). Some burial in the Late Cretaceous to Early
407 Cenozoic prior to the samples cooling to the surface through the Late Cenozoic is possible, but
408 not unequivocally required by the observed data. Regardless of the complexity of the

409 Cretaceous and Early Cenozoic history, the samples remained at temperatures $<50^{\circ}\text{C}$ since the
410 mid-Late Cretaceous (c. 90 Ma).

411 The thermal history for samples *GN6*, *GN7* and *GN8* shows cooling from 100°C at 190
412 Ma to 35°C at 100 Ma. From 100 Ma to present the expected model predicts around 10°C of
413 cooling to surface temperatures, which is consistent with the low levels of denudation
414 throughout the Cenozoic predicted by geomorphological studies (Beauvais and Chardon, 2013;
415 Grimaud et al., 2018).

416 The *GN09*, *GN10*, *GM12* thermal history reproduces the data well and shows cooling at a
417 moderate rate between 100 and 40°C from 170 Ma to 120 Ma before low levels of cooling from
418 120 Ma to present-day. The *GMI3* single sample (Fig. 4a) also shows low amounts of cooling
419 through low temperatures (i.e., $<40^{\circ}\text{C}$) from c. 120 Ma but has experienced more rapid cooling
420 at c. 140 Ma.

421 Furthest from the coast, *GMI4*, *GN11* and *GN12* yield a model where the top sample in
422 the profile cools from 100°C at 170 Ma to 60°C at 135 Ma (Fig. 4a). The ExTH then predicts
423 protracted cooling to bring the samples to present-day surface temperatures. The ExTH
424 reproduces the data reasonably well and the path is consistent with the geomorphological
425 constraints in the region.

426

427 4.2.2 *Guinea South*

428

429 All models across the Guinea South dataset (Fig. 2) show a similarly protracted cooling
430 history (Fig. 4b). All samples cool through the base of the AFT PAZ during the Late Triassic
431 to Late Jurassic, with the exception of *GN18*, which cooled through this boundary during the
432 mid-Permian. Although slow cooling is predominant, there is some variation in the cooling rate
433 through the Cretaceous (c. 150–70 Ma) with some samples showing slower cooling and cooling

434 through lower temperatures than others. For example, *GN18* experiences cooling from 61 to
435 42°C at a rate of 0.24°C/Myr between 150 and 100 Ma, compared to the joint model of *GN31*
436 and *GN33* cooling from 90 to 42°C at a rate of 0.6°C/Myr.

437

438 4.2.3 Ivory Coast

439

440 Two samples, *SA14* and *SA3*, were collected from the coastline and when modelled
441 together produce an ExTH that shows rapid cooling in the Early Cretaceous between 140 and
442 110 Ma (Fig. 4c), followed by minimal cooling until present day. The wide 95% credible
443 intervals in the late Cretaceous imply that the initial cooling may have been greater than the
444 expected model shows, but it is also possible that these samples experienced some minor burial
445 related reheating (<60°C). The *DL03BisA* and *SOK1A* ExTH model shows cooling through the
446 base of the PAZ at c. 200 Ma, followed by a moderate cooling rate of c. 0.6–0.7°C/Myr through
447 the Late Jurassic and Early Cretaceous, before the cooling rate drops quickly over the last 110
448 Ma (Fig. 4c).

449 Further inland, *BEZ-VAA* yields an ExTH model showing relatively moderate cooling
450 through the PAZ during the Cretaceous with only minor, slow cooling through the Cenozoic
451 (Fig. 4c). The *MK64-SEN* ExTH model shows cooling through the PAZ during the Late Triassic
452 and Jurassic (c. 215–150 Ma), and residence at near surface temperatures from the Cretaceous
453 to the present-day (Fig. 4c). The ExTH model for the furthest sample from the coast, *KG33*,
454 shows rapid cooling between 150 and 140 Ma followed by low-temperature (<40°C) protracted
455 cooling from the Cretaceous to present-day.

456 Two samples (*YAL* and *YOR*) sit west of the main transect (Fig. 2) on the older palaeo-
457 to meso-Archean basement. *YOR* resided at the foot of an escarpment, which marks a jump in
458 elevation from c. 300 m to up to 1000 m. *YAL* is sampled inland of these maximum elevations,
459 c. 30 km northwest of *YOR*, from an elevation of 511 m. The individual models for these

460 samples show an initially rapid phase of cooling at two distinct times. For *YAL*, cooling through
461 the PAZ is inferred to occur between 170 and 155 Ma, while *YOR* infers rapid cooling from
462 temperatures hotter from the base of the PAZ to c. 70°C at 120 Ma (Fig. 4c). As these samples
463 are collected from a faulted region marking the boundary between the older Archean basement
464 and the younger Archean and Proterozoic basement (Fig. 1), fault reactivation may have caused
465 the distinct thermal histories.

466

467 **5. Discussion**

468

469 5.1 Spatial patterns of palaeo-temperature and denudation

470

471 Using the ExTH for the models presented above, we can infer the palaeotemperature
472 pattern across the West Africa margin during specific time periods (Fig. 5). Assuming a
473 geothermal gradient over time then allows us to estimate the amount of denudation or burial
474 that may have occurred over time (Fig. 6). In Figure 6 we highlight our denudation/burial
475 estimates assuming a standard geotherm of 25°C/km and use values derived from this geotherm
476 in our discussion. However, we acknowledge that the geothermal gradient was likely elevated
477 during continental breakup along passive and transform margins (Balázs et al., 2022) and the
478 time of CAMP emplacement. We also acknowledge that the geothermal gradient more
479 generally can vary over time and space due to deep thermal processes and due to the addition
480 and removal of material with different thermal properties (e.g., Łuszczak et al., 2017) with
481 implications for the temporal and spatial pattern of denudation. To reflect this, in Figure 6 we
482 show a range for denudation/burial estimates for geothermal gradients between 20-70° C/km.
483 The main observation that can be made using this range of values is that if geothermal gradients
484 were higher and decayed over time, the total amount of material removed in the late Jurassic

485 - Cretaceous would have been significantly less than that predicted using a gradient of 25°
486 C/km. However, it is not clear how far an elevated geotherm would have reached in the context
487 of our sample locations relative to the main rift zone, and how this would have evolved in
488 tandem with post-rift surface, tectonic and thermal processes. Acquiring thermochronometric
489 data from different depths from borehole profiles would potentially yield greater insights into
490 the evolution of the geotherm.

491 At 200 Ma, *MK64* and *SEN* reside at temperatures cooler than 120°C (Fig. 5). All other
492 samples were either residing at temperatures hotter than 100 - 120°C at 200 Ma or were brought
493 to temperatures hotter than 120°C after 200 Ma but prior to cooling through the PAZ during
494 the Cretaceous (Fig. 5). If these palaeotemperatures were solely due to burial, then this would
495 equate to a total of 4 to 5 km of overburden that has been removed since the late Jurassic (for
496 a geothermal gradient of (25°C/km)). Most samples show a progressive decrease in the amount
497 of cooling and consequently in denudation from the Early Cretaceous to the Cenozoic (Fig. 6).
498 Some samples show slight increases in cooling in the Late Cenozoic, however, due to the low
499 temperature nature of this cooling it is uncertain.

500 The emplacement of CAMP rocks may have caused the temperature of upper crustal rocks
501 to increase by direct heating due to their proximity to intrusions, associated regional
502 hydrothermal activity, or burial of under thick layers of extrusive rocks, and/or changes in the
503 geothermal gradient during magmatism and changes in thermal conductivity of the rock
504 blanketing (i.e., the extrusive volcanic rocks) the previously exposed basement. CAMP lavas
505 would have covered the 'pre-rift' landscape, which likely had some Palaeozoic cover (Hubbard,
506 1983). Given the location of the Guinean samples within the CAMP it is possible that this event
507 reset the AFT data at c. 200 Ma.

508 With the exception of samples *YAL* and *YOR*, the Ivory Coast samples lie out with the
509 preserved CAMP intrusion field (Fig. 1b). If these samples were reset at c. 200 Ma with the

510 Guinean samples, then it is likely it was due to burial under CAMP extrusive rocks or more
511 regional hydrothermal activity at the time. *MK64* and *SEN* have not been completely reset at
512 this time, which may be due to their residence at a higher elevation and consequently not being
513 covered by as thick a cover of volcanic rocks or less thick cover of Palaeozoic sediments.
514 However, as the AFT data do not record when the data were reset, only when they cooled below
515 the closure temperature, burial of the samples entirely beneath a sedimentary cover during the
516 Palaeozoic cannot be excluded as a possibility.

517 For both transects, the samples nearest to the coast reside at the highest temperatures at
518 150 Ma (Ivory Coast samples are hotter than the base of the PAZ) and show episodes of rapid
519 cooling (Fig. 5). In Guinea North, this rapid cooling occurs during the 110–90 Ma interval,
520 after an initially moderate phase of cooling, whereas in Ivory Coast rapid cooling occurs during
521 130–110 Ma. Both of these events involve cooling of c. 30–40°C at a rate of 1.5–2°C/Myr. For
522 a geothermal gradient of 25°C/km, this cooling equates to a denudation thickness of 1.2–1.6
523 km (Fig. 6) at a rate of 60 to 80 m/Myr.

524 In Guinea North, the interior samples cool at a low to moderate rate through the Early to
525 mid-Cretaceous (~0.5 – 0.6°C/Myr over 150–90 Ma) before the cooling rate and the total
526 amount of cooling rapidly decreases through the Late Cretaceous and into the Cenozoic (~0.1
527 – 0.2°C/Myr over 70–0 Ma). Similar trends in cooling rate and total amount of cooling through
528 the Cretaceous and the Cenozoic are also observed along the interior of the Ivory Coast transect.
529 Some of the initial cooling in the Early Cretaceous may be attributable to thermal relaxation
530 following the CAMP event but it is also possible that a significant proportion of the cooling
531 through the Early to mid-Cretaceous is attributable to exhumation driven by moderate erosion
532 rates in response to continental break-up. This style of landscape evolution is comparable to
533 that observed along passive continental margins (e.g., Gallagher and Brown, 1999; Persano et

534 al., 2002; Spotila et al., 2004; Balestrieri et al., 2005; Campanile et al., 2008; Peulvast et al.,
535 2008).

536 Due to the temperature sensitivity of the AFT thermochronometer, we are not able to
537 resolve any detail on thermal events colder than 60°C. However, our data remain informative
538 as they require denudation to be less than c. 1–2 km (Fig. 6). In the absence of data or any
539 additional information driving inflections (e.g., episodes of rapid cooling or reheating) in the
540 low-temperature thermal history during the Cenozoic our models infer simple monotonic
541 cooling. Our data and models are therefore consistent with observations by Beauvais and
542 Chardon (2013) and Grimaud et al. (2014, 2018) that denudation was limited to c. 2–10 m/Myr
543 across this part of West Africa. Application of a lower temperature thermochronometer such as
544 apatite (U-Th)/He would potentially provide additional insights on the timing and rate of
545 Cenozoic cooling.

546

547 5.2 Thermal and tectonic processes along the West African margin.

548

549 Our new AFT data is an important contribution to a growing thermochronological record
550 with coverage from the Guinean margin at along the Central Atlantic to the Benin margin along
551 the Equatorial Atlantic (Fig. 7, Fig. 8). This regional dataset helps to better understand the
552 spatial and temporal record of exhumation and the driving geological processes. All data from
553 the West African transform margin are presented in an AFT age-MTL plot (Fig. 7a) and show
554 a partial ‘boomerang’-style relationship. ‘Boomerang’ relationships in AFT datasets are formed
555 by a population of ‘young’ AFT ages and one of ‘old’ AFT ages, both with long MTLs, and a
556 zone between these age populations where the AFT-MTL relationship follows a U-shaped trend.
557 The data can be interpreted as preserving an early (old) rapid cooling event and then recording
558 a later (young) rapid cooling event, with intervening samples exhibiting the effects of partial

559 resetting (Green, 1986). In the West African data set the younger peak is evident corresponding
560 to long MTLs ($>12.5 \mu\text{m}$) with Late Jurassic and Early- to mid-Cretaceous (145 – 90 Ma) AFT
561 ages, signifying the importance of this period for the exhumation and cooling of West Africa.
562 However, the older peak is absent. This suggests that all samples across this region of West
563 Africa experienced some thermal annealing after the Cambrian/Early Ordovician.

564 The data defining the ‘younger’ peak are scattered due to variation in the AFT ages and
565 MTLs. The long period of time over which samples yield long MTLs (Fig. 7a) may be
566 indicative of temporal variations for the onset of cooling and variations in cooling rate across
567 West Africa in the Mesozoic. The variation of the MTLs may be directly related to cooling, but
568 it should be noted that Fig. 7a presents lengths that have not been corrected for their orientation
569 to the crystallographic c-axis (e.g., Ketcham et al., 2007), which may improve consistency in
570 the length measurements (Ketcham et al., 2018). Gunnell (2003), Clift et al. (1998), Bigot-
571 Cormier (2005) and Fernie et al. (2018) do not state that a c-axis correction was made to the
572 length measurements. The absence of this additional data is common in older studies as
573 annealing anisotropy and appropriate numerical corrections were not typically considered.
574 Some studies (e.g., de Grave et al., 2011; van Ranst et al., 2020) question the extent of
575 anisotropy in fossil fission tracks and whether it is appropriate to make a c-axis correction, and
576 this consideration may have been adopted by Fernie et al. (2018). Clift et al. (1998) and Bigot-
577 Cormier et al. (2005) also do not include compositional information (e.g., Dpar or Cl wt. %),
578 which is now commonplace in AFT studies. The absence of these data may help to explain
579 some of the conflicts in timing in the thermal histories observed along the offshore marginal
580 ridge by Clift et al. (1998) and Bigot-Cormier et al. (2005).

581 An AFT age-elevation relationship is most apparent in the Guinean samples, where a clear
582 positive correlation is observed with increasing AFT age over c. 1200 m of elevation gain (Fig.
583 7b). Despite AFT ages reaching up to c. 400 Ma in Benin, no age-elevation relationship is

584 apparent, with samples residing in a fairly narrow range of elevations c. 200–400 m.,
585 suggesting prevailing lateral variations in erosion. It is also apparent that AFT ages greater than
586 200 Ma (i.e., CAMP emplacement) are only found in datasets in the eastern part of the West
587 African margin (e.g., Benin and Ghana). This lends further support to the thermal influence of
588 the CAMP on the Guinean and Ivory Coast datasets and subsequent cooling driven by
589 progressive exhumation of the Guinea and Ivory Coast regions of the West African margin (Fig.
590 9a,b).

591 The absence of the CAMP influence on the eastern datasets has made it possible to better
592 observe the spatial variation in response of surface processes to major tectonic processes (Fig.
593 9). The lack of CAMP influence is inferred due to the preservation of relatively older (i.e., pre-
594 CAMP) AFT ages in Benin (Wildman et al., 2019), Burkina Faso (Gunnell, 2003) and Ghana
595 (Lisker et al., 2008; Fernie et al., 2018) (Fig. 1; Fig. 8). Samples with older (> 200 Ma) AFT
596 ages and relatively short MTLs in these studies reflect regions of the intracontinental craton
597 that have experienced protracted cooling and slow rates and/or low magnitudes of erosion
598 throughout the Palaeozoic and Mesozoic. However, Fernie et al. (2018) do present thermal
599 histories that cool through the PAZ between 200 and 150 Ma, which they attribute to post-
600 CAMP thermal relaxation.

601 The jump in AFT ages in the range 110–130 Ma along a coastal strip in Benin to ages
602 greater than 300 Ma further inland coincident with the location of the present-day continental
603 drainage-divide at c. 375 km inland (Fig. 8), is comparable to models of passive margin
604 landscape evolution, where erosion during escarpment retreat or plateau downwearing drives
605 exhumation (Wildman et al., 2019). Due to the flexural isostatic response to erosion,
606 denudation magnitudes can be several kilometres and is greatest in the region between the coast
607 and the continental divide (Fig. 9) (Gallagher and Brown, 1999; Braun, 2018; Wildman et al.,
608 2019). While the increase in AFT age is not as dramatic along the Guinean and Ivory Coast

609 margins a similar scenario may be invoked. Along the Guinea North transect, the trend of
610 increasing AFT age with distance from the coast, and with increasing elevation, may also mark
611 the influence of a drainage divide at c. 200 km inland, where the oldest AFT ages are observed.
612 In Ivory Coast, AFT ages progressively increase from the coast to c. 350 km inland, where
613 *MK64* and *SEN-1* preserve older AFT ages.

614 The patterns of long wavelength deformation, driving the exhumation patterns across the
615 Central and Equatorial Atlantic, are consistent with the model of transform margin evolution
616 described by Ye et al. (2017) involving the formation and persistence of flexural margin up-
617 warps (e.g., Gilchrist and Summerfield, 1990; Gallagher and Brown, 1997) (Fig. 9). Within
618 this model, the divergent Central Atlantic further North than our study area exhibits a longer
619 wavelength, coast parallel, up-warp compared to the transform Equatorial Atlantic. The
620 difference in up-warp wavelength is attributed to the age of the respective margins, with
621 wavelength increasing for the older central Atlantic margin due to greater thermal relaxation
622 and lithospheric strengthening (Ye et al., 2017). The main pattern of denudation following
623 rifting in the Equatorial Atlantic is extends c. 350 km inland from the coast from Guinea along
624 to Benin. However, in Guinea we have no data >350 km from the coast along the Guinea North
625 transect.

626 Across Guinea, Ivory Coast and Benin, samples inland of the divide infer Mesozoic
627 cooling suggesting erosion of interior flanks of the marginal up-warp beyond the drainage
628 divide (Fig. 9) that may have fed sediment into interior basins (see Ye et al., 2017). Ye et al.
629 (2017) attribute the long wavelength of the Ivory Coast upwarp to processes in the
630 asthenosphere and highlight the emplacement timing (150 and 135 Ma) of Leo-Man shield
631 kimberlitic province in support for chemical and physical changes in the lithospheric mantle,
632 which drives vertical motions (e.g., Ault et al., 2013; Stanley et al, 2015). However, the
633 inference from our thermal histories that crustal cooling across the Ivory Coast was already

634 underway in the late Jurassic-Early Cretaceous suggests a closer link to the tectonic and or
635 thermal processes forming the up-warp along the divergent Guinean margin.

636 Thermal histories from samples from the coastline along Guinea North and Ivory Coast
637 also support more rapid cooling events (Fig. 4a, Fig. 4c). These rapid cooling events may be
638 attributed to more rapid exhumation due to erosion of short-wavelength uplift rift flanks. Along
639 the Ivory Coast equatorial margin, the timing of this rapid exhumation is coeval with the main
640 phase of transform faulting in the Equatorial Atlantic suggesting a causal link (Fig. 9c). Data
641 from Benin (Wildman et al., 2019) and Ghana (Ferne et al., 2018; Lisker et al. 2008) suggest
642 cooling driven by exhumation related to rifting and associated processes occurred along the
643 entire Equatorial Atlantic margin and further into the continental interior (e.g., Benue Trough
644 in Ghana, Gao rift in Benin). The abrupt transition in AFT ages in Ghana across the Bole-
645 Nangodi shear zone (Fig. 1a, Fig 8) invoke intracontinental reactivation, during the rift period,
646 causing differential exhumation across the structure (Fig. 9) (Ferne et al. 2018). Possibly fault
647 reactivation or hydrothermal circulation can explain the offset ages along the Ghanian coast
648 (Fig. 8) (Lisker et al., 2008).

649 The rapid exhumation along the Guinea coastline is approximately 80 Ma after the onset
650 of seafloor spreading in the Central Atlantic and overlaps with the end of transform faulting in
651 the Equatorial Atlantic (Fig. 9d), possibly suggesting some regional tectonic uplift in response
652 to continental break-up in the Equatorial Atlantic. Ferne et al. (2018) present central AFT ages
653 as young as 65 ± 11 Ma and tentatively propose a Late-Cretaceous – early Cenozoic, minor
654 cooling event, attributed to inversion during NNW-SSE shortening within western Africa
655 driving a phase of enhanced exhumation. This timing would also be consistent with the timing
656 of exhumation of the offshore marginal ridge proposed by Bigot-Cormier et al. (2005) (Fig.
657 9d,e). Reactivation of brittle faults during the syn- and post-rift phase has also been proposed
658 along several type-example passive continental margins (e.g., Ksienzyk et al., 2014; Cogné et

659 al., 2011; Wildman et al., 2016). Moreover, inversion during the Late Albian has been observed
660 in structures along the Guinea plateau (Ye et al., 2017) and so fault reactivation along the
661 Guinean coast, driving the rapid cooling between 110 and 90 Ma cannot be ruled out. Uplift
662 and tectonic reactivation across the Central and Equatorial margins following continental
663 break-up in the Equatorial Atlantic and through the active-transform phase is therefore possible
664 but still poorly resolved by the AFT datasets.

665 The limitations on the temperature sensitivity of AFT thermochronology and the low
666 magnitude of Mesozoic cooling make resolving the thermal history and the chronology of
667 denudation and burial through the Cenozoic challenging. Along the coastal regions of the
668 Central and Equatorial Atlantic there are remnants of Late Cretaceous sedimentary basins (i.e.,
669 embayments). However, the past extent of these basins, in terms of both their regional coverage
670 and thickness, is uncertain. Our data from Guinea and Ivory Coast and that presented by
671 Wildman et al., (2019) from Benin do not exclude burial of the coast (e.g., Fig. 9e) and along
672 the fringes of interior basins (e.g., Fig. 9d) as a possibility but they do infer that any burial was
673 not sufficient to cause significant track annealing and so the total amount of heating related to
674 this burial was $<60^{\circ}\text{C}$.

675 More generally across West Africa, the chronology of landscape evolution has been
676 resolved using dated relict lateritic landforms (Beauvais and Chardon, 2013; Grimaud et al.,
677 2014; Chardon et al., 2016). These datasets all appear to converge on the conclusion that in
678 West Africa there was greater tectonic stability and lower magnitudes of erosion during the
679 Cenozoic and certainly since the Eocene. Surface uplift during the Cenozoic, driven by mantle
680 upwellings, has been proposed as a mechanism to form the ‘basin-and-swell’ topography
681 described across Africa (e.g., Burke, 1996). The growth of the Hoggar Swell, to the northeast
682 of our study area, since the mid-Eocene caused significant denudation and increase in sediment
683 flux from the Niger-Benue catchment. Elsewhere in West Africa any minor surface uplift

684 contribution of mantle driven upwelling did not trigger km-scale erosion. Instead, erosion
685 would have been limited to enhanced dissection of successive weathering profiles.

686 The regional drainage network response to tectonic and isostatically driven surface
687 deformation will determine how the eroded sediments are partitioned into offshore and interior
688 basins (e.g., Ye et al., 2017, 2019). A source-to-sink analysis by Ye et al. (2016) observe that in
689 the Ivory Coast–Ghana basin the largest accumulation of siliciclastic sediments occurs during
690 the Campanian (85–72 Ma). Over the mid-Late Cretaceous (104 – 66 Ma), accumulation rates
691 increase from 4.3 to $13.3 \times 10^3 \text{ km}^3/\text{Myr}$. At this time, most samples across the Guinea and
692 Ivory Coast margins, except for the coastal samples in Guinea, show very low cooling rates, or
693 rapidly decreasing cooling rates, reflecting slow, low-magnitude exhumation. These low
694 cooling rates continue throughout the Cenozoic, while offshore accumulation rates first
695 decrease in the Eocene to $3.3 \times 10^3 \text{ km}^3/\text{Myr}$ before increasing to $12.3 \times 10^3 \text{ km}^3/\text{Myr}$ in the
696 Plio-Pleistocene, with a short-lived peak of $18.3 \times 10^3 \text{ km}^3/\text{Myr}$ in the Early Oligocene (34–31
697 Ma). The general trend observed in the Ivory Coast - Ghana basin of high Late Cretaceous
698 accumulation rates, low rates in the early Cenozoic and high rates in the Late Cenozoic is also
699 observed in other basins in the Equatorial Atlantic suggesting a discordance between the timing
700 of uplift driven erosion and peaks in offshore accumulation (Ye et al., 2016). Ye et al. (2016)
701 conclude that the disagreement between the timing and magnitude of onshore denudation and
702 the record of accumulation can be attributed to drainage reorganization and changes in size of
703 paleo-catchments and/or post-depositional sedimentary redistribution on sediment supply.

704

705 **7. Conclusions**

706

707 The landscape of the West African continental margin has largely evolved in response to
708 thermal, tectonic, and surface processes since the emplacement of the Central Atlantic

709 Magmatic Province in the Late Jurassic. This event marked the beginning of a series of tectonic
710 events including rifting and continental break-up in the Central Atlantic followed by oblique
711 movement and break-up in the Equatorial Atlantic. We infer the following from our new AFT
712 data, thermal histories and their association with previously published AFT data.

713 (i) The thermal effect of the CAMP, either by direct heating or by burial under thick
714 magmatic rocks, was significant across Guinea and Ivory Coast and sufficient to
715 heat the presently exposed crust to temperatures $>110^{\circ}\text{C}$. Further east, the absence
716 of the CAMP facilitates preservation of older pre-rift AFT ages.

717 (ii) Patterns of AFT data with distance from the coast are comparable to those expected
718 for traditional models of passive margin evolution, where most erosion occurs
719 coastward of a continental divide during the main phase of rifting. Coastward limbs
720 of long-wavelength flexural upwarps (as described by Ye et al., 2017) in Guinea
721 and Ivory Coast were eroded, at moderate rates, in response to the opening of the
722 Central Atlantic in the Late Jurassic-Early Cretaceous and in response to the onset
723 of transform movement in the Equatorial Atlantic during the early to mid-
724 Cretaceous (145 – 90 Ma). Landward limbs inland of the upwarp crest also
725 experienced suggest significant interior erosion and fed interior basins.

726 (iii) Fast cooling along the Ivory Coast coastline between c. 130 and 110 Ma, is
727 attributed to rapid erosion of a short-wavelength rift shoulder formed during the
728 onset of transform faulting in the Equatorial Atlantic. The onset of fast cooling
729 along the Guinean coast at 110–90 Ma is coeval with the timing of continental
730 break-up in the Equatorial Atlantic. While some of the regional West African AFT
731 data, particularly in Ghana, suggest fault reactivation of basement structures during
732 the Cretaceous and potentially in the Early Cenozoic, it is poorly resolved elsewhere
733 but remains possible

734 (iv) The AFT data and thermal histories infer low cooling rates and total amount of
735 cooling through the Late Cretaceous and Cenozoic suggesting low erosion rates
736 prevailed, which is consistent with Cenozoic denudation estimates of 2-10 m/Myr
737 observed from Geomorphology studies. However, due to the lower temperature
738 limit of the AFT thermochronometer, sub-km scale denudation cannot be
739 confidently resolved. Similarly, burial of the coast and along the fringes of interior
740 basins during the Late Cretaceous is a possibility but the total amount of heating
741 related to this burial was $<60^{\circ}\text{C}$.

742

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744

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751

752 **Figure Captions**

753 Figure 1: (a) Tectonic map of West Africa showing major sedimentary basins, deformation belts,
754 faults and Leo-Man cratonic shield onshore and major fracture zones (FZ) and oceanic crust
755 age offshore. BN SZ – Bole Nangodi shear zone. Location of apatite fission-track data, colour
756 coded for their publication are shown. (b) Geology of part of the map area of Fig. 1a showing
757 the location of the samples from Guinea and Ivory Coast presented in this study.

758 Figure 2: Topographic map of the Guinea-Ivory Coast region of West Africa showing major
759 rivers and drainage divides, sample locations and lines of transects shown in Fig. 5 and Fig. 6 .
760 Inset figure shows West African drainage patterns after Grimaud et al. (2018) and major
761 elevated topographic features (G. R. – Guinean Rise, Tag. – Tagant, Jos. – Jos Plateau, Hog. –
762 Hoggar Swell, Ad. – Adamaoua massif).

763 Figure 3: AFT data plots: (a) c-axis projected mean track length vs. Central AFT age, (b) c-axis
764 projected mean track length vs. c-axis projected mean track length standard deviation, (c)
765 Central AFT age vs. Distance from the coastline, (d) Elevation vs. Central AFT Age. All
766 uncertainties are 1σ .

767 Figure 4: Thermal history models for samples from (a) Guinea North (b) Guinea South, and (c)
768 Guinea South. Thermal histories for each sample or sample grouping in each area are overlain
769 on each other. Further details on the modelling methodology and sample groupings are
770 provided in Supplementary Data, Data S1, Table S2, Fig. S3 – S9. Thick line coloured line
771 with black trim is the Expected Thermal History, coloured semi-transparent shading is the 95%
772 credible intervals. Events correspond to 1: End of the CAMP activity, 2: Onset of sea floor
773 spreading in the Central Atlantic, 3: Oblique transform faulting in the Equatorial Atlantic, 4:
774 Continental break-up in the Equatorial Atlantic, 5: ‘Active-transform’ phase (Ye et al., 2019)
775 in Equatorial Atlantic.

776 Figure 5: Topographic profiles for the Guinea North and Ivory Coast transects with sample
777 palaeotemperature at different time periods estimated from the expected thermal history model.
778 Lines of transect are shown on Fig. 2. For clarity, palaeotemperature uncertainties are not
779 shown but can be estimated using the 95% credible intervals on the expected thermal history.

780 Figure 6: Topographic profiles for the Guinea North and Ivory Coast transects with estimates
781 of denudation for time intervals: 150 – 130, 130 – 110, 110 – 90, 90 – 70, 70 – 50, 50 – 30, 30
782 – 10 and 10 – 0 Ma. Yellow line highlights the estimates when a geothermal gradient of
783 25°C/km is used. Bars show the range of possible denudation magnitudes for geothermal
784 gradients between 20 and 70°C/km. Lines of transect are shown on Fig. 2.

785 Figure 7: West African AFT data plots: (a) Mean track length vs. Central AFT age. Measured
786 mean track lengths (i.e., not c-axis projected lengths) are used because c-axis lengths are not
787 presented in all publications) and (b) elevation vs. Central AFT age. All uncertainties are 1σ .

788 Fig. 8: Maps of West African published AFT data coloured to indicate AFT age. Sample source
789 shown on Figure 1a.

790 Fig. 9: Maps of West Africa summarising tectonic, thermal, and surface evolution over the Late
791 Jurassic and Cretaceous. Palaeogeography and tectonics structures adopted after Ye et al.
792 (2017). (a) During Late Jurassic the emplacement of the Central Atlantic Magmatic Province
793 caused thermal resetting of samples across Guinea, Ivory Coast and potentially in Ghana.
794 Samples in Benin are not reset and preserve a record of Palaeozoic denudation. (b) Denudation
795 across the length of continental margin in response to rifting in the Equatorial Atlantic.
796 Denudation inland due to erosion of marginal upwarps, inland of drainage divides or erosion
797 of uplifted blocks due to intraplate fault reactivation along the Bole-Nangodi Shear Zone
798 (Fernie et al., 2018). (c) Denudation along coastal margin and across some inland regions
799 continues through late Early Cretaceous at moderate rates, faster denudation observed along

800 the Ivory Coast coastline and the offshore marginal ridge in response to the main phase of
801 rifting and onset of transform faulting in the Equatorial Atlantic. (d) Denudation is generally
802 lower along continental margin during the mid-Cretaceous but enhanced denudation inferred
803 along the Guinean coastline and the southward slope of the offshore Ivory Coast marginal ridge
804 (Bigot-Cromier et al., 2005). Burial inferred inland in Benin along the margins of the
805 Iullemmenden Basin (Wildman et al., 2019) and possibly the northern slopes of the offshore
806 marginal ridge (Clift et al., 1998). (e) Denudation magnitudes are inferred to be low from the
807 Late Cretaceous until present-day. Some interior regions continue to have low-moderate rates
808 of denudation and another phase of intracontinental fault inversion may have occurred.

809 Table 1: Apatite Fission Track data. ρ_s , ρ_i , ρ_d are track density of induced, spontaneous,
810 dosimeter tracks. $P(\chi^2)$ is p-value of the chi-sq age homogeneity test (Galbraith, 2005). AFT
811 ages are central ages calculated with 1σ standard error. 'GN' samples were etched using 5M
812 HNO₃ after Gleadow and Lovering, (1978) and ages were calculated using a $\zeta = 338\pm 38$ with
813 a standard CN-5 glass. All other samples were etched using 5.5M HNO₃ after Donelick et al.
814 (2005) and ages calculated with $\zeta = 303.1\pm 9.6$ using a standard IRMM540 glass. Dispersion is
815 the standard deviation of the single-grain ages as a percentage of their central age. Mean track
816 lengths are corrected for their orientation to the c-axis after Ketcham et al. (2007). Sample
817 location and rock type information are found in Supplementary Data, Table S1.

818 **Supplementary Data**

819 Data S1: Further details on the thermal history modelling methodology

820 Fig. S1: Radial plots for single-grain AFT ages.

821 Fig. S2: Apatite fission-track length distributions

822 Fig. S3: Individual sample models with no additional constraints

823 Fig. S4: Map showing sample groupings

824 Fig. S5: Model testing and data predictions for the Guinean North coastal samples.

825 Fig. S6: Detailed models and data predictions for Guinea North inland samples.

826 Fig. S7: Detailed models and data predictions for Guinea South samples.

827 Fig. S8: Detailed models and data predictions for Ivory Coast samples.

828 Fig. S9: Alternative sample groupings

829 Table S1: Location and rock type information for samples

830 Table S2: Summary of QTQt model set up

831

832 **References**

833

834 Attoh, K., Brown, L., & Haenlein, J. (2005). The role of Pan-African structures in intraplate
835 seismicity near the termination of the Romanche fracture zone, West Africa. *J. Afr. Earth.*
836 *Sci.*, 43(5), 549-555.

837 Attoh, K., Brown, L., Guo, J., & Heanlein, J. (2004). Seismic stratigraphic record of
838 transpression and uplift on the Romanche transform margin, offshore Ghana.
839 *Tectonophysics*, 378(1-2), 1-16.

840 Ault, A. K., Flowers, R. M., & Bowring, S. A. (2013). Phanerozoic surface history of the Slave
841 craton. *Tectonics*, 32(5), 1066-1083.

842 Balázs, A., Gerya, T., May, D., & Tari, G. (2022). Contrasting transform and passive margin
843 subsidence history and heat flow evolution: insights from 3D thermo-mechanical modelling.
844 *Geological Society, London, Special Publications*, 524, doi: 10.1144/SP524-2021-94

845 Balestrieri, M. L., Stuart, F. M., Persano, C., Abbate, E., & Bigazzi, G. (2005). Geomorphic
846 development of the escarpment of the Eritrean margin, southern Red Sea from combined
847 apatite fission-track and (U–Th)/He thermochronometry. *Earth Planet. Sci. Lett.*, 231(1-2),
848 97-110.

849 Basile, C. (2015). Transform continental margins—Part 1: Concepts and models.
850 Tectonophysics, 661, 1-10.

851 Basile, C., Mascle, J., & Guiraud, R. (2005). Phanerozoic geological evolution of the
852 Equatorial Atlantic domain. *J. Afr. Earth. Sci.*, 43(1-3), 275-282.

853 Beauvais, A., & Chardon, D. (2013). Modes, tempo, and spatial variability of Cenozoic
854 cratonic denudation: The West African example. *Geochem., Geophys., Geosy.*, 14(5), 1590-
855 1608.

856 Beauvais, A., Ruffet, G., Hénocque, O., & Colin, F. (2008). Chemical and physical erosion
857 rhythms of the West African Cenozoic morphogenesis: the ³⁹Ar-⁴⁰Ar dating of supergene
858 K-Mn oxides. *J. Geophys. Res.-Earth*, 113(F4).

859 Bennett, K. C., & Rusk, D. (2002). Regional 2D seismic interpretation and exploration
860 potential of offshore deepwater Sierra Leone and Liberia, West Africa. *The Leading Edge*,
861 21(11), 1118-1124.

862 Biari, Y., Klingelhoefer, F., Sahabi, M., Funck, T., Benabdellouahed, M., Schnabel, M.,
863 Reichert, C., Gutscherm M. A., Bronner, A., & Austin, J. A. (2017). Opening of the central
864 Atlantic Ocean: implications for geometric rifting and asymmetric initial seafloor spreading
865 after continental breakup. *Tectonics*, 36(6), 1129-1150.

866 Bigot-Cormier, F., Basile, C., Poupeau, G., Bouillin, J. P., & Labrin, E. (2005). Denudation of
867 the Côte d'Ivoire-Ghana transform continental margin from apatite fission tracks. *Terra*
868 *Nova*, 17(2), 189-195. doi:10.1111/j.1365-3121.2005.00605.x.

869 Braun, J. (2018). A review of numerical modeling studies of passive margin escarpments
870 leading to a new analytical expression for the rate of escarpment migration velocity.
871 *Gondwana Res.*, 53, 209-224.

872 Brownfield, M. E., & Charpentier, R. R. (2003). Assessment of the undiscovered oil and gas of
873 the Senegal Province, Mauritania, Senegal, the Gambia, and Guinea-Bissau, Northwest
874 Africa., U.S. Geological Survey Bulletin 2207–A (2003), p. 28

875 Buitter, S. J., & Torsvik, T. H. (2014). A review of Wilson Cycle plate margins: A role for mantle
876 plumes in continental break-up along sutures?. *Gondwana Res.*, 26(2), 627-653.

877 Burke, K. (1996). The African plate. *S. Afr. J. Geol.*, 99(4), 341-409.

878 Burke, K., & Gunnell, Y. (2008). The African erosion surface: a continental-scale synthesis of
879 geomorphology, tectonics, and environmental change over the past 180 million years (Vol.
880 201). *Geol. Soc. Am. Mem.*

881 Campanile, D., Nambiar, C. G., Bishop, P., Widdowson, M., & Brown, R. (2008).
882 Sedimentation record in the Konkan–Kerala Basin: implications for the evolution of the
883 Western Ghats and the Western Indian passive margin. *Basin Res.*, 20(1), 3-22.

884 Carlson, W. D., Donelick, R. A., & Ketcham, R. A. (1999). Variability of apatite fission-track
885 annealing kinetics: I. Experimental results. *Am. Mineral.*, 84(9), 1213-1223.

886 Chardon, D., Grimaud, J. L., Rouby, D., Beauvais, A., & Christophoul, F. (2016). Stabilization
887 of large drainage basins over geological time scales: Cenozoic West Africa, hot spot swell
888 growth, and the Niger River. *Geochem., Geophy., Geosy.*, 17(3), 1164-1181.

889 Chardon, D., Grimaud, J.-L., Beauvais, A., Bamba, O., (2018). West African lateritic pediments:
890 landform-regolith evolution processes and mineral exploration pitfalls. *Earth-Sci. Rev.*, 179,
891 124-146.

892 Clemson, J., Cartwright, J., & Booth, J. (1997). Structural segmentation and the influence of
893 basement structure on the Namibian passive margin. *J Geol. Soc. London*, 154(3), 477-482.

894 Clift, P. D., Carter, A., & Hurford, A. J. (1998). Apatite fission track analysis of Sites 959 and
895 960 on the transform continental margin of Ghana, West Africa, *Proceedings of the Ocean*
896 *Drilling Program, Scientific Results*, Vol. 159.

897 Cogné, N., Gallagher, K., & Cobbold, P. R. (2011). Post-rift reactivation of the onshore margin
898 of southeast Brazil: Evidence from apatite (U–Th)/He and fission-track data. *Earth Planet.*
899 *Sci. Lett.*, 309(1-2), 118-130.

900 Davison, I. (2005). Central Atlantic margin basins of North West Africa: geology and
901 hydrocarbon potential (Morocco to Guinea). *J. Afr. Earth. Sci.*, 43(1-3), 254-274.

902 Deckart, K., Bertrand, H., & Liégeois, J. P. (2005). Geochemistry and Sr, Nd, Pb isotopic
903 composition of the Central Atlantic Magmatic Province (CAMP) in Guyana and Guinea.
904 *Lithos*, 82(3-4), 289-314.

905 de Grave, J., Glorie, S., Buslov, M. M., Izmer, A., Fournier-Carrie, A., Batalev, V. Y., van
906 haecke, F., Elburg, M., & van den Haute, P., (2011). The thermo-tectonic history of the Song-
907 Kul plateau, Kyrgyz Tien Shan: Constraints by apatite and titanite thermochronometry and
908 zircon U/Pb dating. *Gondwana Research*, 20(4), 745-763. Doi: 10.1016/j.gr.2011.03.011

909 Deynoux, M., Affaton, P., Trompette, R., & Villeneuve, M. (2006). Pan-African tectonic
910 evolution and glacial events registered in Neoproterozoic to Cambrian cratonic and foreland
911 basins of West Africa. *J. Afr. Earth. Sci.*, 46(5), 397-426.

912 Donelick, R. A., O’Sullivan, P. B., & Ketcham, R. A. (2005). Apatite fission-track analysis.
913 *Rev. Mineral. Geochem.*, 58(1), 49-94.

914 Ennih, N., & Liégeois, J. P. (2008). The boundaries of the West African craton, with special
915 reference to the basement of the Moroccan metacratonic Anti-Atlas belt. *Geol. Soc. Spec.*
916 *Publ.*, 297(1), 1-17.

917 Fernie, N., Glorie, S., Jessell, M. W., & Collins, A. S. (2018). Thermochronological insights
918 into reactivation of a continental shear zone in response to Equatorial Atlantic rifting
919 (northern Ghana). *Scientific reports*, 8(1), 1-14.

920 Gallagher, K. (2012). Transdimensional inverse thermal history modeling for quantitative
921 thermochronology. *J. Geophys. Res.-Sol. Ea.*, 117(B2).

- 922 Gallagher, K., & Brown, R. (1997). The onshore record of passive margin evolution. *J Geol.*
923 *Soc. London*, 154(3), 451-457.
- 924 Gallagher, K., & Brown, R. (1999). Denudation and uplift at passive margins: the record on
925 the Atlantic Margin of southern Africa. *Philos. Trans. R. Soc. Series A: Mathematical,*
926 *Physical and Engineering Sciences*, 357(1753), 835-859.
- 927 Gibson, G. M., Totterdell, J. M., White, L. T., Mitchell, C. H., Stacey, A. R., Morse, M. P., &
928 Whitaker, A. (2013). Pre-existing basement structure and its influence on continental rifting
929 and fracture zone development along Australia's southern rifted margin. *J Geol. Soc.*
930 *London*, 170(2), 365-377.
- 931 Gilchrist, A. R., & Summerfield, M. A. (1990). Differential denudation and flexural isostasy in
932 formation of rifted-margin upwarps. *Nature*, 346(6286), 739-742.
- 933 Gleadow, A. J. W., & Lovering, J. F. (1978). Thermal history of granitic rocks from western
934 Victoria: A fission-track dating study. *J Geol. Soc. Aust.*, 25(5-6), 323-340.
- 935 Gleadow, A. J., Duddy, I. R., Green, P. F., & Hegarty, K. A. (1986). Fission track lengths in the
936 apatite annealing zone and the interpretation of mixed ages. *Earth and planetary science*
937 *letters*, 78(2-3), 245-254.
- 938 Green, P. F. (1986). On the thermo-tectonic evolution of Northern England: Evidence from
939 fission track analysis. *Geol. Mag.*, 123, 493-506.
940 <https://doi.org/10.1017/S0016756800035081>
- 941 Greenroyd, C. J., Peirce, C., Rodger, M., Watts, A. B., & Hobbs, R. W. (2008). Demerara
942 plateau—The structure and evolution of a transform passive margin. *Geophys. J. Int.*, 172(2),
943 549-564.
- 944 Grimaud, J. L., Chardon, D., & Beauvais, A. (2014). Very long-term incision dynamics of big
945 rivers. *Earth Planet. Sci. Lett.*, 405, 74-84.

946 Grimaud, J. L., Rouby, D., Chardon, D., & Beauvais, A. (2018). Cenozoic sediment budget of
947 West Africa and the Niger delta. *Basin Res.*, 30(2), 169-186.

948 Guiraud, R., Bosworth, W., Thierry, J., & Delplanque, A. (2005). Phanerozoic geological
949 evolution of Northern and Central Africa: an overview. *J. Afr. Earth. Sci.*, 43(1-3), 83-143.

950 Gunnell, Y. (2003). Radiometric ages of laterites and constraints on long-term denudation rates
951 in West Africa. *Geology*, 31(2), 131-134.

952 Heine, C., Zoethout, J., & Müller, R. D. (2013). Kinematics of the South Atlantic rift. *Solid*
953 *Earth*, 4, 215–253.

954 Hubbard, F. H. (1983). The Phanerozoic cover sequences preserved as xenoliths in the
955 kimberlite of eastern Sierra Leone. *Geol. Mag.*, 120(1), 67-71.

956 Hurford, A. J., & Green, P. F. (1983). The zeta age calibration of fission-track dating. *Chem.*
957 *Geol.*, 41, 285-317.

958 Ketcham, R. A., Carter, A., & Hurford, A. J. (2015). Inter-laboratory comparison of fission
959 track confined length and etch figure measurements in apatite. *Am. Mineral.*, 100(7), 1452-
960 1468.

961 Ketcham, R. A., Carter, A., Donelick, R. A., Barbarand, J., & Hurford, A. J. (2007). Improved
962 modeling of fission-track annealing in apatite. *Am. Mineral.*, 92(5-6), 799-810.

963 Ketcham, R. A., van der Beek, P., Barbarand, J., Bernet, M., and Gautheron, C. (2018)
964 Reproducibility of thermal history reconstruction from apatite fission-track and (U-Th)/He
965 data, *Geochem., Geophys., Geosy.* 19(8), 2411-2436.

966 Ksienzyk, A. K., Dunkl, I., Jacobs, J., Fossen, H., & Kohlmann, F. (2014). From orogen to
967 passive margin: constraints from fission track and (U–Th)/He analyses on Mesozoic uplift
968 and fault reactivation in SW Norway. *Geol. Soc. Spec. Publ.*, 390(1), 679-702.

969 Labails, C., Olivet, J. L., Aslanian, D., & Roest, W. R. (2010). An alternative early opening
970 scenario for the Central Atlantic Ocean. *Earth Planet. Sci. Lett.*, 297(3-4), 355-368.

971 Lisker, F., T. John, and B. Ventura. (2008). Denudation and uplift across the Ghana transform
972 margin as indicated by new apatite fission track data, In. Katlenburg-Lindau: Katlenburg-
973 Lindau, Germany: Copernicus GmbH on behalf of the European Geosciences Union (EGU).

974 Lodhia, B. H., Roberts, G. G., Fraser, A. J., Jarvis, J., Newton, R., & Cowan, R. J. (2019).
975 Observation and Simulation of Solid Sedimentary Flux: Examples From Northwest Africa.
976 *Geochem., Geophy., Geosy.*, 20(11), 4613-4634.

977 Lompo, M. (2010). Paleoproterozoic structural evolution of the Man-Leo Shield (West Africa).
978 Key structures for vertical to transcurrent tectonics. *J. Afr. Earth. Sci.*, 58(1), 19-36.

979 Loncke, L., W. R. Roest, Frauke Klingelhoefer, C. Basile, David Graindorge, A. Heuret, Boris
980 Marcaillou, Thomas Museur, Anne-Sophie Fanget, and M. Mercier de Lépinay. "Transform
981 marginal plateaus." *Earth-Science Reviews* 203 (2020): 102940.

982 Łuszczak, K., Persano, C., Braun, J., & Stuart, F. M. (2017). How local crustal thermal
983 properties influence the amount of denudation derived from low-temperature
984 thermochronometry. *Geology*, 45(9), 779-782.

985 Marzoli, A., Bertrand, H., Knight, K. B., Cirilli, S., Buratti, N., Vérati, C., Nomade, S., Renne,
986 P. R., Youbi, N., Martini, R., Allenbach, K., Neuwerth, R., Rapaille, C., Zaninetti, L., &
987 Bellieni, G. (2004). Synchrony of the Central Atlantic magmatic province and the Triassic-
988 Jurassic boundary climatic and biotic crisis. *Geology*, 32(11), 973-976.

989 Marzoli, A., Callegaro, S., Dal Corso, J., Davies, J. H., Chiaradia, M., Youbi, N., Bertrand, H.,
990 Reisberg, L., Merle, R., & Jourdan, F. (2018). The Central Atlantic magmatic province
991 (CAMP): a review. In *The Late Triassic World* (pp. 91-125). Springer, Cham.

992 McHone, J. G. (2003). Volatile emissions from Central Atlantic Magmatic Province basalts:
993 Mass assumptions and environmental consequences. *Geoph. Monog. Series*, 136, 241-254.

994 Mercier de Lépinay, M., Loncke, L., Basile, C., Roest, W. R., Patriat, M., Maillard, A., & De
995 Clarens, P. (2016). Transform continental margins—Part 2: A worldwide review.
996 *Tectonophysics*, 693, 96-115.

997 Moulin, M., Aslanian, D., & Unternehr, P. (2010). A new starting point for the South and
998 Equatorial Atlantic Ocean. *Earth Sci. Rev.*, 98(1-2), 1-37.

999 Nemcok, M., Rybar, S., Sinha, S. T., Hermeston, S. A., & Ledvenyiova, L. (Eds.). (2016,
1000 September). *Transform Margins: Development, Controls and Petroleum Systems*.
1001 Geological Society of London.

1002 Nomade, S., Knight, K. B., Beutel, E., Renne, P. R., Verati, C., Féraud, G., Marzoli, A., Youbi,
1003 N., & Bertrand, H. (2007). Chronology of the Central Atlantic Magmatic Province:
1004 implications for the Central Atlantic rifting processes and the Triassic–Jurassic biotic crisis.
1005 *Palaeogeogr. Palaeoclimatol. Palaeoecol.*, 244(1-4), 326-344.

1006 Peace, A., McCaffrey, K., Imber, J., van Hunen, J., Hobbs, R., & Wilson, R. (2018). The role
1007 of pre-existing structures during rifting, continental breakup and transform system
1008 development, offshore West Greenland. *Basin Res.*, 30(3), 373-394.

1009 Persano, C., Stuart, F. M., Bishop, P., & Barfod, D. N. (2002). Apatite (U–Th)/He age
1010 constraints on the development of the Great Escarpment on the southeastern Australian
1011 passive margin. *Earth Planet. Sci. Lett.*, 200(1-2), 79-90.

1012 Peulvast, J. P., Sales, V. C., Bétard, F., & Gunnell, Y. (2008). Low post-Cenomanian denudation
1013 depths across the Brazilian Northeast: implications for long-term landscape evolution at a
1014 transform continental margin. *Global Planet. Change*, 62(1-2), 39-60.

1015 Rollinson, H. (2016). Archaean crustal evolution in West Africa: A new synthesis of the
1016 Archaean geology in Sierra Leone, Liberia, Guinea and Ivory Coast. *Precambrian Res.*, 281,
1017 1-12.

1018 Séranne, M., & Anka, Z. (2005). South Atlantic continental margins of Africa: a comparison of
1019 the tectonic vs climate interplay on the evolution of equatorial west Africa and SW Africa
1020 margins. *J. Afr. Earth. Sci.*, 43(1-3), 283-300.

1021 Spotila, J. A., Bank, G. C., Reiners, P. W., Naeser, C. W., Naeser, N. D., & Henika, B. S. (2004).
1022 Origin of the Blue Ridge escarpment along the passive margin of Eastern North America.
1023 *Basin Res.*, 16(1), 41-63.

1024 Stanley, J. R., Flowers, R. M., & Bell, D. R. (2015). Erosion patterns and mantle sources of
1025 topographic change across the southern African Plateau derived from the shallow and deep
1026 records of kimberlites. *Geochem., Geophy., Geosy.*, 16(9), 3235-3256.

1027 Summerfield, M.A., 2000. *Geomorphology and Global Tectonics*. Wiley, Chichester (386
1028 pp.).

1029 Svensen, H. H., Torsvik, T. H., Callegaro, S., Augland, L., Heimdal, T. H., Jerram, D. A.,
1030 Planke, S., & Pereira, E. (2018). Gondwana Large Igneous Provinces: plate reconstructions,
1031 volcanic basins and sill volumes. *Geol. Soc. Spec. Publ.*, 463(1), 17-40.

1032 Tari, G. (2006). Traditional and new play types of the offshore Tano Basin of Cote D'Ivoire
1033 and Ghana, West Africa., *Houston Geological Society Newsletter*, January, 48, 27 – 34.

1034 Van Ranst, G., Pedrosa-Soares, A. C., Novo, T., Vermeesch, P., & De Grave, J. (2020). New
1035 insights from low-temperature thermochronology into the tectonic and geomorphologic
1036 evolution of the south-eastern Brazilian highlands and passive margin. *Geosci. Front.*, 11(1),
1037 303-324.

1038 Vasconcelos, P. M., Brimhall, G. H., Becker, T. A., & Renne, P. R. (1994). $^{40}\text{Ar}/^{39}\text{Ar}$ analysis
1039 of supergene jarosite and alunite: Implications to the paleoweathering history of the western
1040 USA and West Africa. *Geochim. Cosmochim. Ac.*, 58(1), 401-420.

1041 Vermeesch, P. (2021). On the treatment of discordant detrital zircon U–Pb data.
1042 *Geochronology*, 3(1), 247-257.

- 1043 Villeneuve, M. (2005). Paleozoic basins in West Africa and the Mauritanide thrust belt. *J. Afr.*
1044 *Earth. Sci.*, 43(1-3), 166-195.
- 1045 Villeneuve, M. (2008). Review of the orogenic belts on the western side of the West African
1046 craton: the Bassarides, Rokelides and Mauritanides. *Geol. Soc. Spec. Publ.*, 297(1), 169-
1047 201.
- 1048 Villeneuve, M., & Cornée, J. J. (1994). Structure, evolution and palaeogeography of the West
1049 African craton and bordering belts during the Neoproterozoic. *Precambrian Res.*, 69(1-4),
1050 307-326.
- 1051 Wildman, M., Brown, R., Beucher, R., Persano, C., Stuart, F., Gallagher, K., Schwanethal, J.,
1052 & Carter, A. (2016). The chronology and tectonic style of landscape evolution along the
1053 elevated Atlantic continental margin of South Africa resolved by joint apatite fission track
1054 and (U-Th-Sm)/He thermochronology. *Tectonics*, 35(3), 511-545.
- 1055 Wildman, M., Cogné, N., & Beucher, R. (2020). Fission-track thermochronology applied to the
1056 evolution of passive continental margins. In *Fission-Track Thermochronology and its*
1057 *Application to Geology* (pp. 351-371). Springer, Cham.
- 1058 Wildman, M., Webster, D., Brown, R., Chardon, D., Rouby, D., Ye, J., Huyghe, D., & Dall'Asta,
1059 M. (2019). Long-term evolution of the West African transform margin: estimates of
1060 denudation from Benin using apatite thermochronology. *J Geol. Soc. London*, 176(1), 97-
1061 114.
- 1062 Withjack, M. O., R. W. Schlische, and P. E. Olsen (2012), Development of the passive margin
1063 of eastern North America: Mesozoic rifting, igneous activity, and breakup, in *Regional*
1064 *Geology and Tectonics*, edited by D. G. Roberts and A. W. Bally, pp. 301–335, *Phanerozoic*
1065 *Rift Systems and Sedimentary Basins*, Elsevier, New York.

1066 Ye, J., Chardon, D., Rouby, D., Guillocheau, F., Dall'asta, M., Ferry, J. N., & Broucke, O.
1067 (2017). Paleogeographic and structural evolution of northwestern Africa and its Atlantic
1068 margins since the early Mesozoic. *Geosphere*, 13(4), 1254-1284.

1069 Ye, J., Rouby, D., Chardon, D., Dall'asta, M., Guillocheau, F., Robin, C., & Ferry, J. N. (2019).
1070 Post-rift stratigraphic architectures along the African margin of the Equatorial Atlantic: Part
1071 I the influence of extension obliquity. *Tectonophysics*, 753, 49-62.

1072 Ye, J. (2016). Evolution topographique, tectonique et sédimentaire syn- à post-rift de la marge
1073 transformante ouest Africaine. GET Toulouse, PhD Thesis, 273 p.