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1   **Evolution of the continental margin of south to central Vietnam and its relationship**

2   **to opening of the South China Sea (East Vietnam Sea)**

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13

14   **Key Points:**

15   • Apatite thermochronometry record South Vietnam margin exhumation history

16   • Thermal models are used to test different causes of thermal history

17   • Regional enhanced cooling and rock uplift between 37-30 Ma record early stage rifting

18   **Abstract**

19   The continental margin of south to central Vietnam is notable for its high elevation plateaus

20   many of which are covered by late Cenozoic basalt flows. It forms the westernmost margin of a

21   wide continental rift of the South China Sea (East Vietnam Sea), and uplift has been considered a

22   result of either rifting or younger intraplate basalt magmatism. To investigate margin

23   development apatite thermochronometry was applied to a dense array of samples collected from

24   across and along the margin of south to central Vietnam. Results, including thermal history

25   models, identified a distinct regional episode of fast cooling between c. 37 and 30 Ma after

26   which cooling rates remained low. The fast cooling coincides with a period of fast extension

27   across the South China Sea (East Sea) region that preceded continental break-up recorded by

28   Oligocene grabens onshore. A thermal model is used test different processes that might influence

29   the inferred cooling including a distinct pulse of exhumation; a decrease in exhumation followed

30   by an associated transient decrease in geothermal gradients and, underplating coincident with

31   rifting. Thermal relaxation following Mesozoic arc magmatism is ruled out as geotherms

32   returned to background rates within 20 Myrs of emplacement, well before the onset of fast

33   cooling. Models support fast cooling attributed to accelerated erosion during early stages of

34   rifting. Some additional heating from either underplating, and/or hot mantle upwellings is also

35   possible. No evidence was found to support regional uplift associated with the intraplate

36   magmatism, enhanced monsoon-driven erosion or seafloor spreading dynamics.

37

38   **Plain Language Summary**

39   This study used apatite thermochronometry to examine the uplift history of South-Central

40   Vietnam to better understand how the elevated landscape formed, and to see if it is connected to

41   the eruption of the widespread basalt flows that cover many of the high elevation plateaus.

42   Results showed a match in the timing between a period of rapid cooling between 37-30 million

43   years ago that affected the entire region and a period of fast rifting and extension that ended with

44   continental breakup. To fully understand the cause of cooling thermal models tested different

45   scenarios. Thermal relaxation after magmatism was discounted at it took place well before the

46   period of rapid cooling. Models support fast cooling attributed to accelerated erosion during

47   early stages of rifting. Magmatic underplating associated with rifting might also have produced

48   additional heating but no evidence was found to support regional uplift associated with the basalt

49   magmatism.

50

51   **1. Introduction**

52   The continental margin of Indochina represents the westernmost margin of the South China Sea,

53   a wide continental rift that in the west is mostly submerged. In the region of south to central

54   Vietnam, the location of this study (**Fig. 1**), the continental margin extends offshore for > 100

55   km before oceanic crust is reached. The extension that led to continental break-up has yet to be

56   fully explained but is known to be related to the subduction of the paleo-Pacific plate, which at

57   the time formed a wide Cretaceous magmatic arc with widespread calc-alkaline magmatism

58   across south to central Vietnam and southern China. Subduction was directed to the northeast

59   beneath southern China and westwards below Vietnam, the eastern margin of the Indochina

60   block (Zhu et al., 2004; Hall, 2012; Hennig-Breitfeld et al., 2021).

61

62   Extension is imprinted on the landscape and geology of Vietnam but to what extent is not yet

63   clear. It also remains uncertain as to when surface uplift took place. Was uplift staggered over a

64   long time or did it occur within a short episode? Stratigraphic evidence provides some

65   constraints. In south-central Vietnam many of the high plateaus, with typical elevations between

66   500-1500m a.s.l. (Fig. 1), are covered by basalt flows (mainly tholeiites) the majority of which

67   were erupted between the late Miocene to Quaternary (Hoang et al., 2013; An et al., 2017).

68   Eruptions were often from extensional faults or small axial rifts that formed previously during

69   extrusion and opening of the South China Sea (Flower et al., 1998; Hoang and Flower, 1998).

70   Rangin et al., (1995) recognized that many of the normal faults associated with these basalts

71   were reactivated N 160°E to N-S right-lateral faults. However, the timing of extension across

72   this margin is not well defined. A common view is that extension involved uplift of the rift

73   shoulders and that this drove widespread erosion from the latest Cretaceous-early Paleocene

74   (Matthews et al., 1997; Roques et al., 1997; Lee and Watkins, 1998; Lee et al., 2001; Andersen

75   et al., 2005; Pubellier et al., 2005). However, onshore, there are no geological constraints at

76   outcrop to confirm this. Basalt flows can be seen to drape over Precambrian to Proterozoic

77   crystalline and metamorphic basement rocks and in some places Jurassic to Cretaceous and late

78   Neogene sedimentary and magmatic rocks, but the temporal gap between the youngest rocks is

79   large, e.g., exhumed Cretaceous granites, and their basalt cover spans 80-90 Myrs. Offshore

80   constrains are little better. In southern Vietnam the Cuu Long and Nam Con Som rift basins

81   record a c. 50 Myr gap between the Eocene clastic sediments that were deposited on exhumed

82   Cretaceous granites and older rocks (Fyhn et al., 2009a).

83

84   As the deformation and tectonic evolution of the continental margin of south-central Vietnam

85   and its relationship to the opening of the South China Sea is poorly resolved we conducted a

86   thermochronometry study based on apatite U-Th-He and fission track (FT) data using rock

87   samples collected from across the continental margin. The utility of this approach is

88   demonstrated by previous work on the Kontum region (**Fig. 1**) of central Vietnam (Carter et al.,

89   2000) that identified a late Miocene increase in exhumation rates across the central margin

90   contemporaneous with increased sedimentation in the adjacent Phu Khan marine basin. It was

91   unclear if this phase of exhumation was caused by surface uplift of the continental margin and

92   base level change associated with breakup of the southwest sub-basin of the South China Sea

93   (Fyhn et al., 2009; Savva et al., 2014), and/or switching between rifting and left-lateral

94   transtension along the East Vietnam Boundary Fault Zone (**Fig. 1**). An alternative explanation is

95   that late Cenozoic magmatism had a local influence on rock uplift (Carter et al., 2000). Finally,

96   work by Clift et al., (2015) proposed that basin subsidence histories may have been affected by

97   increased loading associated with monsoon intensification driving faster sediment delivery to

98   offshore basins and therefore enhanced exhumation could be due to erosional unloading related

99   to changes in monsoon intensity.

100

101   **2. Geological Background**

102   During the Cretaceous eastern Indochina was part of a wide Andean type continental margin.

103   Calc-alkaline magmatism dates to between 118 – 87 Ma (Thuy et al., 2004; Shellnutt et al., 2013;

104   Hennig-Breitfield et al., 2021). The end of magmatism coincided with docking of the Luconia-

105   Dangerous Ground block with Indochina, and South China moving to the southeast to become

106   part of the South Asia margin (Fyhn et al., 2009a; Hall, 2012). These changes likely drove the

107   early rifting of the Proto-South China Sea in either a back-arc setting on the East Asian margin

108   or slab-pull induced microcontinent detachment from subduction along Northern Borneo (Doust

109   and Sumner, 2007).

110

111   Spreading in the South China Sea, that started at c. 32 Ma (Barckhausen et al., 2014) affected the

112   continental margins of both Indochina and SE China. Pre-spreading extension of the continental

113   lithosphere produced more or less E-W oriented normal faults across SE China and South

114   Vietnam margins although in the Cu Long basin the orientation is closer to SW-NE (Schmidt et

115   al., 2019). With the onset of ocean spreading, the orientation of new faults moved to NE-SW

116   reflecting the progression of spreading to the southwest (Savva et al., 2014). A ridge jump at 25

117   Ma relocated spreading to a southwestern sub-basin until spreading ended at 15 Ma (Li et al.,

118   2014), when the subduction zone became blocked by collision with continental fragments

119   (northern part of Palawan and/or a part of the Dangerous Grounds).

120

121   It is widely assumed that the Late Cretaceous to Early Paleocene early extension drove uplift of

122   the rift shoulders followed by widespread erosion and peneplanation (Taylor and Hayes, 1983),

123   but this is based on offshore seismic interpretations of poorly dated early syn-rift sediments. The

124   imprint of extension and rifting on the south-central Vietnam continental margin is different

125   from the SE China margin due to a combination of re-activation of inherited structures (mainly

126   Indosinian) and the local influence of tectonic extrusion driven by India-Eurasia collision (Fyhn

127   et al, 2009a). Although the amount of left-lateral extrusion is debated (Searle et al., 2010) the

128   position and orientation of Indochina must have changed throughout the period of South China

129   Sea opening, and any change would have affected the regional stress field. Four phases of late

130   Mesozoic–Cenozoic deformation are recognised in Vietnam; 1) E-W compression, N-S

131   extension, 2) NNW-SSE compression, ENE-WSW Extension, 3) N-S compression, E-W

132   extension and 4) NNE-SSW compression, ESE-WNW extension (Nguyen and Hoai, 2019).

133

134   **2.1. Records from offshore basins**

135   Extension and rifting led to the formation of the Phu Khanh and Cuu Long basins (**Fig. 1**). The

136   Phu Khanh Basin was controlled by the N-S striking East Vietnam Boundary Fault, considered a

137   continuation of the Red River Shear Zone, as well as NE-SW directed normal faults (Fyhn et al.,

138   2009a). Further south the Cuu Long Basin is bounded by NE-SW to E-W striking normal faults

139   (Schmidt et al., 2019). The sedimentary archives of these basins reflect onshore erosion.

140

141   **2.2. Phu Khanh Basin**

142   This mainly deep-water basin is situated at the base of the continental slope off central Vietnam

143   and is separated from the mainland by the East Vietnam Boundary Fault (**Fig. 1**) (Fyhn et al.,

144   2009b). The basin history records two major rifting events (Lee, and Watkins, 1998; Savva et al.,

145   2014). Syn-Rift I in the Palaeogene was accompanied by the deposition of clastic and lacustrine

146   sediments. Syn-Rift II is associated with the opening of the South China Sea and from the end of

147   the Oligocene involved mainly marine sedimentation. A distinct unconformity marks the

148   Paleogene–Neogene boundary (Fyhn et al., 2009b) across which rifting decreased significantly.

149   Post-rift sediments include upper Miocene–Pliocene turbidite fans interpreted as the product of

150   high rates of onshore erosion (**Fig. 2**).

151

152   **2.3. Cuu Long Basin**

153   This basin is located offshore southeast Vietnam (**Fig. 1**) across the continental shelf and was

154   formed by rifting during the Late Eocene–Early Oligocene (Schmidt et al., 2019). The basin

155   contains up to 8 km of mainly clastic sediments deposited between the Eocene to Quaternary

156   (Morley et al., 2019). A narrow valley developed during the first (syn-rift) phase of extension,

157   and in the basin axial zone rapid subsidence took place in the late Oligocene. Whilst extensional

158   deformation extended into the Miocene, a distinct unconformity at ~ 27 Ma marks inversion and

159   a transition to post-rift sag. The main phase of post-rift thermal sag took place after 23 Ma

160   (Morley et al., 2019).

161

162   In the Early Miocene, a new phase of seafloor spreading was accompanied by a period of sea

163   level rise. This caused a marine transgression in all of the basins along the western edge of the

164   South China Sea (Lee, and Watkins, 2001) leading to the formation of carbonate and coral reefs

165   (Fyhn et al., 2009a). Later post rift subsidence formed a broader shallow sag basin with clastic

166   sedimentation. Our low-temperature thermochronometry results should show which parts of the

167   continental margin experienced the enhanced erosion that drove the clastic sedimentation.

168

169   **3. Methodology**

170   The partial retention zone (c. 80–40 °C) of the (U-Th-Sm)/He (AHe) and partial annealing zone

171   (c. 130–60 °C) of the apatite fission track (AFT) methods are well placed to provide constraints

172   on timing, rate, and magnitude of bedrock exhumation in the uppermost crust (Lisker et al.,

173   2009). For this study a primary goal is to use apatite thermochronometry to detect when rock

174   uplift and exhumation accelerated in response to either surface uplift and/or denudation.

175   Mapping the pattern of exhumation in relation to the extended continental margin will allow

176   assessment of the impact of extension on the continental margin of southern Vietnam. This aim

177   guided our bedrock sampling strategy that comprised a series of north to south, coast to interior

178   transects across the study area, spanning the local relief. Studies of rift and passive margin

179   erosion patterns based on apatite fission track data (Wildman et al., 2019) have shown that the

180   amount of erosion is generally greatest along the coastal areas and decreases inland. This is

181   mainly due to isostatic unloading: As a rift escarpment migrates inland, rock uplift rates remain

182   high close to the margin leading to higher exhumation rates and younger ages. Variations in

183   erosion caused by local geomorphic conditions and geology such as reactivation of inherited

184   faults and magmatic underplating add noise to this dominant trend.

185

186   In total 67 samples were collected of which 42 (**Fig. 3**) produced good quality apatite suitable for

187   analysis. Each sample weighed between 2–5 kg and were mainly granites as early sampling

188   found that other rock types contained little, if any, apatite. Supplementary table S1 details

189   sample locations. All thermochronometry analyses were conducted at the London

190   Geochronology Centre. Fission track analyses used conventional neutron irradiation procedures

191   based on the external detector method and zeta calibration approach (Hurford and Green, 1983).

192   Apatites were mounted, polished and etched in 5N nitric acid 20 °C for 20 seconds. Track length

193   measurements were used to constrain sample cooling rate, and grain bulk composition was

194   monitored using etch pit length (DPAR) measured parallel to the crystallographic c-axis

195   (Donelick 1993). (U-Th-Sm)/He analyses typically involved between 4 and 6 replicates. Apatite

196   grains were placed in platinum tubes and outgassed using a 25 W, 808 nm diode laser and ^{4}He

197   measured on a Balzers quadrupole mass spectrometer. Following apatite dissolution and spiking

198   U, Th, and Sm concentrations were measured on an Agilent 7700x ICP-MS. Further details on

199   the analytical protocols can be found in the supplementary section.

200

201   **4. Results**

202   **4.1. Low-Temperature Thermochronometry**

203   Apatite (U-Th-Sm)/He analysis was performed on 18 representative samples from across the

204   elevation range (Table S2). These typically show age dispersion between individual replicates,

205   some of which can be explained by variations in grain size, accumulated radiation damage, and

206   apatite chemical composition (Fig. S1). Such factors can affect helium diffusion kinetics and

207   produce a range of (U-Th)/He dates from a single rock. However, spurious ages can also arise

208   from alpha particle implantation from external sources, U-Th rich inclusions, inaccurate grain

209   size measurement and alpha ejection correction, problematic outgassing, incomplete grain

210   dissolution and sample handling. Outlier ages were only rejected where there was clear evidence

211   of experimental issues.

212   AFT analysis was carried out on 47 samples of which 44 yielded good quality age data (Table

213   S3. By contract, track length data was affected by low spontaneous track densities and only 11

214   samples yielded sufficient track lengths for modelling. Sample central ages range from 70±5 to

215   22± 3 Ma, with a weighted average of 41 Ma. Apatite etch pit lengths measured parallel to the

216   grain c-axis (DPAR), a proxy for grain bulk composition, ranges from 1.7 to 3.8 μm. If cooling

217   rates were low DPAR and AFT age would be expected to correlate, due to a compositional

218   influence on FT annealing. As there is no correlation this implies cooling rates were not slow,

219   borne out by FT track length data (samples with > 50 measurements) that show unimodal

220   distributions and long mean lengths between 13.5μm and 14.2μm (n = 16). Furthermore, a

221   comparison between AHe ages and effective uranium (eU) values and grain radius, based on a

222   spherical geometry, revealed no trends to support slow cooling (Fig. S1). If this were the case

223   single AHe ages should correlate with eU (proxy for radiation da,mage) (Flowers et al., 2009)

224   Likewise, larger grains, with a slightly higher closure temperature, would be appreciably older

225   than smaller grains if cooling was slow.

226

227   Figure 4 plots all replicate AHe ages and AFT central ages against sample elevation. Most ages

228   fall between 20-60 Ma, irrespective of sample elevation, that points to a phase of regional uplift.

229   There is no well-developed age trend between coastal areas and furthest points inland as would

230   be expected from a conventional passive margin whereby the youngest ages would be confined

231   to the coastal areas and oldest ages retricted to the highest elevations inland. Within the dataset,

232   AHe ages tend to increase with elevation but this is less well developed in the AFT data. Clear

233   evidence is seen in **Fig. 5** that shows ages plotted for a suite of samples collected across a 1421m

234   elevation range within a 92 Ma granite, located close to Nha Trang on the coast. Proximity to the

235   coast means that the data are more likely to record a primary signal of rock uplift and erosion:

236   points further from the coast will be less sensitive to a geomorphic response to rock uplift. In

237   other words, any delay between uplift and erosion will depend on geomorphic processes, which

238   may take longer to respond to changes in tectonics further from the baselevel (sea-level). The

239   plot shows exhumation rates accelerated after 40-50 Ma based on the changing slopes of the age-
240   elevation relationship.

241

242   **4.2. Thermal history models.**

243   To constrain rock uplift history, the thermal histories of representative samples, with sufficient
244   track length measurements, were obtained using the QTQt software of Gallagher (2012). This is
245   based on a Bayesian transdimensional approach to data inversion to extract probable thermal
246   histories. Model outputs are accepted thermal history models that can be combined to give a
247   mean thermal history model weighted by the posterior probability of each individual thermal
248   history with 95% credible intervals that provide a measure of uncertainty. Model runs allowed
249   the temperature offset to vary over time and data were predicted using the annealing and
250   diffusion models Ketcham et al (2007) and Gautheron et al., (2009). Granite emplacement age
251   was the only time-temperature constraint used in the inversion based on our own zircon U-Pb
252   analyses or published ages (Shellnutt et al., 2013; Hennig-Breitfield 2021) .

253

254   Results from the multi-sample vertical profile inversion shown in **Fig. 5** are presented in **Fig. 6**.
255   Models produce a reasonable fit to the data although some AHe ages are underpredicted. For the
256   low elevation data underprediction, could be due to a recent increase in cooling such that the
257   amount of rock uplift and exhumation is insufficient to provide a clear signal in the AHe data.
258   However, thermal history models of low elevation samples with the youngest AHe ages from
259   across the study area (S2) show no significant departure from the post 30 Ma cooling path in
260   figure 6. Likewise models of high elevation samples from across the study area are consistent
261   (residence at crustal temperatures < 60°C since 50-60 Ma) with the thermal history model in **Fig.**
262   **6**. The models suggest the region experienced a multi-stage thermal history that involved
263   accelerated cooling between 60-50 Ma, minor reheating between 50-35 Ma, and accelerated
264   cooling between ~ 37-30 Ma followed by constant cooling to the present. The early phases of
265   cooling are not well constrained as oldest tracks do not extend to the middle Cretaceous hence
266   the thermal history between 90 Ma and about 50 Ma are largely driven by the requirement of
267   granite emplacement. Resolution increases through time as shown by the difference between the
268   credible intervals for specific samples. The reheating stage of the history from 50 to 37 Ma is
269   relatively well resolved. From 37 Ma, however, resolution increases and the accelerated cooling

270   between 37-30 Ma is well constrained and consistent with the regional age-elevation data plotted

271   in **Fig. 5**. These two stages require average offset temperatures (differences between lowest and

272   highest elevation samples) to rise to up to 80°C equivalent to a geothermal gradient of

273   ~57°C/km, but then temperature offsets decrease through time. This decrease in geothermal

274   gradient through time is partly required by the fact that both high and low elevation samples

275   must reach similar temperatures at the surface. However, the ages are sensitive to times as recent

276   as 20 Ma and therefore, the change in geothermal gradient is likely to be robust and not an

277   artifact.

278

279   **5. Discussion**

280   Dense sampling across the study area, extending from the coast up to 75 km inland yielded 42

281   AFT ages and 77 AHe replicate single grain ages distributed across an elevation range of 1524

282   m. Ages do not increase from the coast to inland as would be expected from a typical rifted

283   margin where there is limited erosion landward of an escarpment (Gallagher and Brown, 1997).

284   The majority of sample ages cluster between 20-50 Ma, independent of elevation, consistent with

285   regional block uplift. Although some of the highest elevation data show a relationship between

286   age and elevation reflecting an earlier exhumation history, most of the data below 800 m record

287   similar ages indicative of faster cooling. This contrast in cooling rates is best seen in data from a

288   suite of samples collected across a 1.4 km vertical section of a 92 Ma granite (**Fig. 5**) that

289   captures the regional thermal history. QTQt thermal history models (**Fig. 6**) show a multi-stage

290   history with reheating between ~ 50-37 Ma, and well-resolved accelerated cooling between 37-

291   30 Ma followed by uninterrupted slower cooling to the present. Differences in temperature

292   between the lowermost and uppermost samples between 60-30 Ma require a high geothermal

293   gradient (c. 57°C/km) that when applied to the accelerated cooling between ~37-30 Ma would

294   crudely translate to about 1500 m of rock uplift and exhumation. This regional thermal history

295   raises several questions: 1) What caused the interval of reheating between 50-37 Ma? 2) What

296   caused the interval of accelerated cooling between 37-30 Ma and was this period related to the

297   generation of present-day topography? 3) Are there any connections between changes in cooling

298   rate, the opening of the South China Sea (East Vietnam Sea) and sedimentation in the marine

299   basins?

300

301   In order to address these questions, we require a thermal model that relates cooling histories of

302   rocks to changes in exhumation rate and the thermal structure of the crust. Using this thermal

303   model we can test different factors that might influence the inferred cooling. In the next section,

304   we first highlight that the elevated heating from arc magmatism and emplacement of the sampled

305   granites is a transient feature that does not persist beyond about 20 Myrs. Therefore our resolved

306   changes in cooling since about 50 Ma are related to other processes. We highlight three different

307   processes that predict similar cooling histories: a distinct pulse of exhumation; a decrease in

308   exhumation followed by an associated transient decrease in geothermal gradients; and, recent

309   underplating coincident with rifting but unrelated to arc magmatism and recent basaltic

310   magmatism.

311

312   We use a simple 1D transient thermal model to explore these processes. The model solves the

313   heat transport equation with Dirichlet boundary conditions at surface and lower crustal

314   temperatures, tracking material points through the evolving thermal field to provide time-

315   temperature paths. Numerically, the model uses finite differences in space and the Crank-

316   Nicolson method for temporal integration (see Fox and Carter, 2020). The thermal model is 35

317   km thick, used boundary conditions set at 15 $^{\circ}$C at the surface and 800 $^{\circ}$C at the base to give an

318   unperturbed geothermal gradient of 22 $^{\circ}$C/km as this is about the average geothermal gradient

319   predicted by the QTQt models over the last 40 Ma. A thermal diffusivity of 25 km^2/Ma is used

320   for all models. The timestep length is set to 1×10^{-3} Myr and the verical resolution is 0.2318 km.

321   The intial condition is modified to reflect the emplacement of magmas at different depths.

322   Exhumation rate histories are also modified to explore the different scenarios.

323

324   **5.1. Relationship between cooling and exhumation**

325

326   <u>**The influence of magmatic heating during emplacement**</u>

327   As a former magmatic arc the study area would have experienced increased heating and elevated

328   geothermal gradients followed by post-magmatic thermal relaxation. A fundamental question is

329   whether this impacted on the low-temperature thermochronology data. Murray et al., (2018)

330   investigated this question using 1D and 3D models in which heat diffused from a magmatic

331   emplacement and rocks were advected towards the surface. They were able to show that large

332   midcrustal plutons, emplaced at depths between 10–15 km, can reset low-temperature

333   thermochronometers in the upper crust. This is not surprising, but of importance to this study is

334   the question of how long such thermal effects persisted following magma emplacement? There is

335   at least a 40 Myr gap between granite ages and the earliest cooling constrained by the apatite

336   data. Could this early stage cooling be a consequence of post magmatic thermal relaxation? To

337   explore this we used the same approach as Murray et al., (2018) to assess the effects of heating at

338   depth on the time-temperature paths experienced by thermochronometry samples. The cartoon in

339   Figure 7 shows the basic evolution of the model. Heating can lead to increased geothermal

340   gradients at all depths. Samples that are exhuming will experience this heating and subsequent

341   cooling. The exact heating and cooling rates will depend on the duration of heating, the

342   exhumation rate, the depth of heating and the samples and the thermal diffusivity of the crust

343   (Murray et al., 2018).

344

345   Figure 8 shows the results of a numerical model designed to simulate high temperature at the

346   start of the model. Here, temperatures are high during the first 2 Myrs of the model run from 8-

347   12 km depth. This heats a large section of the crust. After this initialization period, heat is lost

348   through diffusion and temperatures rapidly return to a steady state solution. It is important to

349   note that the steady state solution shows slightly higher geothermal gradients than the initial

350   condition due to the advection of heat driven by erosion which is constant at 0.1 km/Myr. The

351   rocks that are at the surface at the end of the model are tracked along the red curve. This is

352   highlighted Figure 8B and shows the rapid cooling following emplacement. For reference, a

353   thermal history is shown in which there is no heating at the start of the model. Only the high

354   temperature $^{40}Ar/^{39}Ar$ in hornblende and muscovite thermochronometric systems would be

355   influenced by this heating event. The other thermochronometers would not be sensitive to this

356   thermal event and therefore, we can begin the subsequent models at 40 Ma.

357

358   Next, we consider two other processes that may produce similar, high rates of cooling over the

359   same sort of time interval: 1) relaxation of geothermal gradient following the cessation of

360   exhumation and, 2) cooling following a pulse of heating associated with emplacement of

361   underplating material below the lower crust associated with the rifting.

362

363 **Changes in exhumation rate**

364 We use a reference scenario in which the cooling is interpreted in terms of exhumation (Fig. 9).

365 In this scenario, a pulse of exhumation with rates of 0.6 km/Ma is simulated between 37 and 30

366 Ma and the background rate is set to 0.02 km/Ma (consistent with Quaternary cosmogenic

367 erosion rates measured in a river catchment to the north of the study area by Jonell et al., 2017).

368 This leads to relatively fast cooling over this time interval recreating the accelerated cooling

369 between 37 and 30 Ma. However, the peak in cooling rate is at 30 Ma and there is just a small

370 change in the geothermal gradient. This is in contrast with the results of the QTQt model that

371 highlights large changes in the geothermal gradient through time. It is important to note that we

372 are highlighting the geothermal gradient at the changing depth of the synthethic sample as it is

373 exhumed. This is to be consistent with the temperature offset resolved by QTQt. In addition, it is

374 not clear how exhumation rates might change so abruptly given the gradual changes in surface

375 processes expected following changes in rock uplift.

376

377 A second scenario is simulated in which a pulse of exhumation that began at 40 Ma at rates of

378 0.75 km/Ma and then decreased exponentially from 37 Ma to the present to a background rate of

379 0.01 km/Ma (**Fig. 9**). This exhumation rate history approximates a geomorphic response to rock

380 uplift in which erosion rates are high during rock uplift and then decay after the cessation of

381 active rock uplift. High rock uplift rates elevate the geothermal gradient due to the advection of

382 heat. As exhumation rates decrease, the geothermal gradients relax but rocks continue to

383 exhume. Importantly, the peak in cooling rate actually postdates the phase of active rock uplift

384 and therefore the phase of rapid cooling resolved by the thermochronometric data may be the

385 result of an earlier phase of active rock uplift. Although this scenario modifies the geothermal

386 gradient due to heat advection, the magnitude of the change in geothermal gradient is not as large

387 as inferred using QTQt. As with the first scenario, geothermal gradients decrease following the

388 phase of fast exhumation.

389

390 **Changes in boundary conditions**

391 A third scenario simulates the emplacement of underplating material below the lower crust (**Fig**

392 **9**). In seismic sections further to the north in the South China Sea, lower-crust high-velocity

393 anomalies have been observed and related to magmatic underplating at depths of about 25-30 km

394   (Li et al., 2020). In this model a pulse of heating is set to 1500 $^\circ$C between depths of 25 and 35

395   km from 40-38 Ma and this simulates a change in temperature at the base of the crust associated

396   with rifting. Such depths and temperatures are not unrealistic as the geochemistry of local

397   Cenozoic calc alkali magmatism indicates a hotter than normal shallow asthenosphere (Hoang et

398   al., 2013) and the lithosphere in this study area is some 20-30 km thinner than northernVietnam

399   (Vu et al., 2021), presumably in part due to extension and rifting. Following this heating event,

400   the lower boundary is returned to 800 $^\circ$C and temperatures diffuse back to the normal conditions

401   over time. We acknowledge that it is unrealistic that temperature would return to a fixed value

402   immediately after this heating event, however, this is sufficient to highlight this process. The

403   background exhumation rate is set to 0.075 km/Ma for the duration of the model. This value is

404   slightly higher than the 0.02 km/Ma inferred using cosmogenic nuclide concentration

405   measurements (Jonell et al., 2017) but we hope to highlight that a constant exhumation rate

406   through time can produce the observed cooling history, and 0.02 km/Ma for the duration of the

407   model would require impossibly high geothermal gradients to exhume rocks from high enough

408   temperatures. This scenario produces a reheating signal as temperatures increase, then as

409   temperatures decrease and rocks continue to exhume, a pulse of cooling is predicted. There is a

410   hint of reheating in the QTQt models. However it is not very strong and is poorly resolved.

411   Cooling rates during the cooling pulse are lower than in the other two scenarios. However, the

412   geothermal gradient is much higher and decreases through time, as required by the data.

413

414   These three scenarios highlight the range of geological models that produce similar time-

415   temperature paths. Other geological models could be envisioned including changing thermal

416   diffusivities through time, different boundary conditions or many other factors. The large

417   changes in geothermal gradient through time predicted by QTQt are hard to reproduce with a

418   simple advection-diffusion equation and require changes in boundary conditions through time or

419   the emplacement of hot material at the base of the model, as simulated in scenario 3. However,

420   the rate of cooling following this is too slow in scenario 3 and this suggests that there is

421   additional exhumation during rifting. The data are, therefore, likely providing a record of

422   accelerated erosion during rifting and additional heating possibly related to magmatic

423   underplating in the lower crust.

424

425 **5.2. Enhanced denudation associated with monsoon climate?**

426 Climate is a key driver of denudation. The study area is affected by the East Asian monsoon and

427 therefore the changes in cooling rate detected by the thermochronometry might reflect enhanced

428 rates of erosion and exhumation driven by monsoon climate change. Reconstructions of the East

429 Asian monsoon show a relatively stable wet environment throughout the late Eocene and

430 Oligocene that strengthened after 23 Ma, reaching a maximum between 18-10 Ma (Clift et al.,

431 2014, Farnsworth et al., 2019). These timings do not coincide with the changes seen in the

432 apatite thermal histories. Notably, cooling rates declined at a time when the monsoon intensified.

433 Hence there is no obvious connection between changes in monsoon intensity and regional

434 thermal history.

435

436 In the offshore basins lacustrine/fluvial environments dominated the Eocene, and sediment

437 accumulation rates do not appear to change, although this could be due to limited

438 accommodation space and poor development of regional drainage systems. Between the

439 Oligocene and early Miocene there was a switch from siliciclastic sedimentation to carbonates

440 associated with shallow marine lagoonal and reefal environments. Only in the latest part of the

441 Miocene do offshore basins record a marked rise in siliciclastic sedimentation, largely associated

442 with shoreface facies. At this time accumulation rates were sufficiently high to cause

443 progradation, especially in the Phu Khanh Basin (Fyhn et al., 2009; Lee and Watkins, 1998).

444 Paradoxically, this occurred at a time when the East Asian monsoon began weakening (Clift et

445 al., 2014). Likewise, [10]Be Cosmogenic derived denudation rates for the Song Gianh monsoon-

446 dominated river in northern central Vietnam (Jonell et al., 2017), show lower rates for the strong

447 monsoon of the Early Holocene compared to longer-term erosion rates based on local apatite

448 thermochronometry data. Based on this evidence past changes in monsoon climate intensity do

449 not explain the thermochronometry data.

450

451 **5.3. Timing of surface uplift**

452 By the early Cenozoic, it is widely held that a peneplain or low relief surface extended across

453 much of Indochina (Tran et al., 2011) hence surface uplift must post-date this. Tighter

454 constraints are provided by the onshore geology. Onshore South Vietnam, the c. 300 km long

455 Song Ba Rift formed within a major NW–SE-trending strike-slip fault zone (**Fig 1**) that

456 continues offshore and separates the Phu Khanh Basin from the Cuu Long Basin. Structural

457 observations show the faults cross-cut Cretaceous rocks and were therefore active in the

458 Cenozoic when deformation reactivated the strike-slip faults as extensional faults (Nielsen et al.,

459 2007). The Song Ba rift contains up to 500 m of syn-rift Oligocene sediments but their thermal

460 maturity (vitrinite reflectance % Ro c. 0.4) indicates that the grabens were originally deeper and

461 contained more fill, between 1-2 km. This means that substantial uplift and denudation occurred

462 after the Oligocene but before eruption of the Miocene basalts that drape the uplift unconformity

463 (Nielson et al., 2007). Some basalt flows are also seen to infill pre-existing valleys. Hoang and

464 Flowers (2013) provide examples of 6–8 Ma lava flows (their Fig. 2) infilling over 300m of local

465 relief. Such evidence favours a close relationship between topographic growth and rifting

466 although it is possible surface uplift continued after active rifting had ended.

467

468 **5.4. Paleogene rifting**

469 The 37–30 Ma interval of accelerated cooling recorded by the thermochronometry data overlaps

470 active rifting in both onshore and offshore basins across central and south Vietnam. Onshore

471 evidence for active erosion during this time can be found within Oligocene clastic deposits of the

472 Di Linh Formation of the Da Lat zone. Detrital zircon U-Pb data from these rocks are dominated

473 by Cretaceous magmatic ages, consistent with erosion of the Cretaceous igneous rocks on which

474 they sit (Hennig et al., 2018). The Di Linh Formation is possibly an extension of the offshore

475 Cuu Long Basin where sediments were first deposited during the latest middle Eocene. Similar to

476 onshore, Paleogene sediments rest on an older basement that includes exhumed Cretaceous

477 granites. Active rifting and inversion took place during the late Eocene to early Oligocene and

478 had almost ceased by the end of the Lower Oligocene, recorded by a basin-wide unconformity

479 dated to c. 28 Ma (Morley, et al., 2019). After this the basin entered a short period of

480 compression due to the Mekong Delta Fault Zone switching from left-lateral to right-lateral

481 strike-slip (Schmidt et al., 2019). This regime ended circa 25 Ma when strike-slip motion on the

482 Mekong Delta Fault Zone stopped, and the basin then transitioned into post rift thermal

483 subsidence.

484

485 The influence of a large fault zone is also seen in the Phu Khan Basin where left-lateral motion

486 along the East Vietnam Boundary Fault Zone (**Fig. 1**) affected early rifting (Fyhn et al., 2009a).

487   Easternmost parts of the basin also show evidence for contemporaneous NW to SE extension

488   suggesting some of the extension was due to other factors, such as slab pull from subduction of a

489   proto-South China Sea (Vu et al., 2017). An unconformity at the Oligocene-Miocene boundary

490   marks the end of rifting in this basin, and a switch to regional post-rift subsidence led to a

491   transgression.

492

493   The 37–30 Ma interval of accelerated cooling recorded by the thermochronometry data also

494   spanned regional changes in the stress regime associated with rifting and the onset of ocean

495   spreading c. 32 Ma (**Fig. 10**). Ocean Discovery Program expeditions 367/368, at the northern

496   South China Sea margin, identified a major period of fast extension during the late Eocene to

497   early Oligocene (Larsen et al, 2018), roughly 40-30 Ma. This relatively short rifting event

498   spanned the rift-to-igneous crustal accretion transition and was associated with rapid upwelling

499   of the asthenosphere, seeding the seafloor spreading that took place soon after. This would have

500   affected all surrounding margins, and it culminated in the formation of a regional breakup

501   unconformity at circa 33–28 Ma after which there was a regional reduction in extensional

502   activity (Morley et al., 2016; 2019). This timing is shared by basins offshore the study area.

503   During this time Indochina also underwent extrusion along the Ailao Shan-Red River Fault Zone

504   (ARRFZ). Left-lateral shearing along the Day Nui Con Voi shear zone, the southern extension of

505   the ARRFZ, took place between ~35 Ma to ~20 Ma (Jolivet et al., 2001; Leloup et al., 2001 Liu

506   et al., 2020; Searle et al., 2010). However, it seems unlikely that this had a major influence on

507   deformation across the study area since the duration of extrusion extended well beyond the

508   episode of fast cooling.

509

510   **5.5. Relationship to ocean spreading.**

511   Onset of seafloor spreading at magnetic anomaly 11, c. 32 Ma (Barckhausen et al., 2014; Sibouet

512   et al., 2016) began in the East South China Sea sub-basin opening in a roughly N-S direction.

513   But, at 23 Ma (anomaly 6a) a ridge jump changed the position and direction of spreading by

514   rotating about 15–20° anticlockwise to form the South West sub-basin (Sibouet et al., 2016).

515   This shift in orientation caused a change in regional stress and coincides with unconformities in

516   the Cuu Long and Phu Khanh basins (**Figs. 1&8**) (Fyhn et al., 2009a; Schmidt et al., 2019).

517   Changes in the regional stress regime are also indicated by changes in rates of the propagation of

518   spreading. Between 32–23 Ma rates of propagation in the East South China Sea sub-basin were

519   slow but increased in both sub-basins after the ridge jump (Le Pourhiet et al., 2017).

520

521   Were these changes sufficient to explain the fast cooling? During spreading, as the asthenosphere

522   rises and spreads along the rift axis there will be accompanying changes in the force balance

523   between far-field stresses, resulting from margin composition, structure and topography, and

524   local buoyancy from thinning of the lithosphere. Margin uplift would be expected where the

525   rising asthenosphere induces gradients of gravitational potential energy by juxtaposing denser

526   asthenospheric material against thicker crustal material (Rey, 2001; Mondy et al., 2018)

527   however, a large topographic load from the surrounding continental margin can transmit

528   significant compressive stresses that oppose buoyancy forces. In the case of the South China Sea

529   (East Vietnam Sea) 3D numerical simulations have demonstrated how a small amount of

530   compression, and/or extension, acting normal to the direction of propagation can influence the

531   rate of breakup propagation (Le Pourhiet et al., 2017). The slow propagation stage of rifting

532   within the East South China Sea sub-basin can be explained by resistive, compressive stresses

533   from the west-to-east topographic load of Indochina. The change following the ridge jump at 23

534   Ma, caused the direction of spreading propagation to become oblique to Indochina and reduced

535   much of the resisting out-of plane compression allowing the propagation of spreading to the SW

536   to accelerate. Whilst the slow propagation stage overlaps with the fast cooling seen in the QTQt

537   models (**Fig. 6**), this is not the case for the timing of the switch to fast rift propagation since it

538   significantly post-dates the slowdown in cooling rates. Thus, whilst it may be appealing to link

539   the onshore thermal history to changes in spreading and regional stresses there is no strong

540   evidence to support this.

541

542   **5.6. Relationships to mantle structure**

543   Between ~17–0.2 Ma widespread intraplate volcanism took place across south-central Vietnam

544   (Hoang et al., 2013) and is effectively ongoing. A common view is that the late Miocene

545   increased siliciclastic sedimentation in basins along the margins of south-central Vietnam was a

546   response to margin surface uplift (Fyhn et al., 2009) and that this may have been linked to the

547   widespread basaltic magmatism (Carter et al., 2000). Hoang et al., (2013) observed that the large

548   magmatic centers in Vietnam formed in pull apart basins and that the pattern of volcanism had

549   evolved from SW–NE transtension to E–W extension linked to changes in the lithospheric stress

550   field due to either collision-related lithospheric response or an effect of asthenospheric flow.

551   Using a 1-D shear velocity model for the Indochina block, Yang et al. (2015) identified a low-Vs

552   anomaly below depths of 100 km along South-Central Vietnam and considered it likely reflects

553   active mantle flow in the asthenosphere beneath the region due to the stretching and thinning of

554   the lithosphere. The nature of flow is unclear, however, but could be related to edge-driven

555   convection as this has been used to explain intraplate volcanism and high topography in areas

556   where there is a significant gradient in lithosphere thickness (Kaislaniemi and Hunen, 2014).

557   Whilst the patterns and consequences of edge-driven convection on the continental side of a

558   passive margin has yet to be fully investigated, it is noted that local conditions for such flow are

559   suitable. The study area has an average crustal thickness between 29–33km (Bai et al., 2010; Vu

560   et al., 2021), thinner than global average values (~36–41 km; Szwillus et al., 2019) and the study

561   area lithospheric thickness is lower, between 110–120 km, compared to > 130km for northern

562   Vietnam (Vu et al., 2021). In relation to the apatite thermochronometry results early flow

563   associated with initial thinning and breakup would explain the interval of rapid cooling and

564   account for some of the margin uplift.

565

566   An alternative explanation to the above is related to low velocity anomalies detected in the lower

567   mantle below the study area (spanning depths between ∼700–2889 km). Using multiscale global

568   tomography Zhao et al., (2021) interpreted the low velocity anomalies below southern Indochina

569   as hot mantle upwellings associated with a cluster of dying or already dead plumes that existed in

570   the Cenozoic. Basalt compositions across the study area are Ocean Island Basalt (OIB) mantle or

571   Enriched Mantle types (EM). An et al., (2016) argued that whilst there is large-ion lithophile

572   element enrichment without high field strength element depletion, typical of OIB, NiO -olivine

573   plots show compositions that fall within the range of Hawaiian and Hainan basalt olivines. As

574   these may be produced by partial melting of silica-poor eclogite and peridotite An et al., (2016)

575   favoured a Hainan plume source from recycled eclogitic oceanic crust. An alternative

576   explanation for basalt compositions is provided by Hoang et al., (2003) who highlighted the role

577   of active mantle flow in the asthenosphere by arguing that the late Cenozoic alkalic magmatism

578   in East Asia originated in the shallow ductile asthenosphere from adiabatic decompression

579   caused by pronounced changes in stress regimes linked to India - Eurasia collision and

580   subduction of the Pacific Plate. India - Eurasia collision and subduction clearly impacted on

581   mantle structure beneath Indochina as shown by a regional-scale receiver function study of the

582   mantle transition zone discontinuities (Yu et al., 2017). This work highlighted the potential role

583   of the subducted Indian slab and how broken segments sinking into the mantle transition zone

584   could have influenced mantle structure and dynamics beneath the Indochina Peninsula. Clearly

585   further work is needed to fully understand the nature of the magmatism and its relationship to

586   mantle structure.

587

588

589   **6. Conclusions**

590   The dense array of apatite thermochronometry results from across and along the margin of south-

591   central Vietnam revealed a regional episode of fast cooling between 37-30 Ma. This signal is

592   present within the coastal margins as well as inland. Consideration of the multiple factors that

593   affected the regional stress field at this time highlighted the coincidence between the episode of

594   fast cooling and the period of fast extension across the South China Sea (East Sea) region. This

595   shows a close relationship to rifting, however the episode of cooling might also be due to

596   transient changes in geothermal gradient. Thermal models explored the relationship between

597   cooling and exhumation. Models rule out thermal relaxation following Mesozoic arc magmatism

598   as geotherms returned to background rates 40 to 30 Myrs before the onset of fast cooling.

599   Instead, models suggest fast cooling could be attributed to accelerated erosion during early stages

600   of rifting, possibly with some additional heating from either underplating, and/or hot mantle

601   upwellings although the timing does not coincide with known periods of magmatism. No

602   evidence was found to connect regional uplift with the Miocene to Quaternary intraplate

603   magmatism. If this did occur the magnitude of associated surface uplift is beyond the resolution

604   of this study.

605

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613
614 **Open Research**
615 **Data Availability Statement**
616 Data sets for this research are provided in Supporting Information Tables S1 to S3. Data are also
617 available in Earth and Space Science Open Archive doi.org/10.1002/essoar.10507473.1
618
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831

832 **Figure Captions**

833

834 **Fig. 1.** Location of the study area that compliments an earlier study of the Kontum region (Carter

835 et al., 2000).

836

837 **Fig. 2.** Seismic profile across the continental margin adjacent to the study area (Vietnam

838 Petroleum Institute). A distinct Oligocene unconformity records basin inversion and a mid

839 Miocene unconformity records the end of rifting after which sediment supply increased leading

840 to progradation and slope mass transport deposits.

841

842 **Fig. 3.** Sample locations and regional geology modified from Tran, et al., (1988).

843

844 **Fig. 4.** Plot comparing AFT and AHe (corrected) replicate ages against sample elevation.

845 Because the samples are from a range of locations with different exhumation rate histories, the

846 slope of this relationship will not approximate the exhumation rate. In addition, elevation varies

847 with distance from the coast and so this variability also masks obvious trends. Error bars are 1s.

848

849 **Fig. 5.** AFT and AHe (corrected) ages for a suite of samples collected across a 1421m elevation

850 range within a 92 Ma granite. Ages above 800m record slower exhumation than samples at lower

851 elevation. Key to geological units given in **Fig. 3**.

852

Fig. 6. Thermal history model of the vertical profile shown in Fig. 5. Dark column denotes the circa 37-30 Ma onset of fast cooling recorded by the best-fit models. Solid lines are best fit models. The credible intervals are shown (dashed lines) are shown only for the lowermost and uppermost samples. Plot 6b shows how well the models fit the measured data by plotting the predicted and observed values as a function of elevation. Error bars include the mean 95% credible range for the predictions from all thermal history models accepted during the post-burnin MCMC sampling.

Fig. 7. A simple 1D model is used to assess the effects of heating at depth on the time-temperature paths experienced by thermochronometry samples. In this cartoon, the initial condition is shown by the dotted line. Here temperature increases linearly with depth and then there is an increase in temperature associated with the magma (1). Temperatures remain high to simulate an active magma chamber. This leads to heating above the magma chamber (2). After a set amount of time, temperatures at the base of the model return to a temperature defined by the initial geothermal gradient (3). Temperatures remain high across the upper part of the model (4). Over time temperature return to a steady state thermal model determined by the boundary conditions and exhumation rate. The inset shows the evolution of temperature a material point experiences as it is exhumed towards the surface. Initially there is heating to the elevation geotherm (EG) and then temperatures decrease to the steady state geotherm (SSG).

Fig. 8. The duration of heating associated with magma intrusion. A magma intrusion is simulated between 8-12 km depth. Temperatures remain high for 2 million years before the basal boundary condition at 35 km is forced to 800C. The exhumation rate remains constant at 0.1 km/Ma. The grey curves show the evolving thermal model over the first 10 million years of the simulated history. After approximately 20 million years the steady state solution is reached and temperature as a function of depth remains constant. The red line shows the temperature-depth paths rocks that reach the surface took. The circles show the closure temperatures and depths evaluated using Dodson's approximation. B) This same path can be plotted as a function of temperature and age. This highlights the rapid cooling from temperatures of 1500°C to temperatures of about 250°C over the first 20 million years of the simulated history. For

883   reference, a time-temperature path from a thermal model with no heating is shown by the blue

884   dashed line.

885

886   **Fig. 9:** Thermal models for three different thermokinematic scenarios used to predict the

887   evolving thermal field: Dotted line (scenario 1) is where the cooling was interpreted in terms of

888   exhumation, with rates of 0.6 km/Ma between 37.5 and 30 Ma and a background rate of 0.02

889   km/Ma. The solid line (scenario 2) approximates a geomorphic response to rock uplift where

890   exhumation began at 40 Ma at rates of 0.75 km/Ma but then decreased exponentially from 37 Ma

891   to the present background rate of 0.01 km/Ma. The dashed line (scenario 3) represents a pulse of

892   heating associated with emplacement of underplating material below the lower crust and

893   highlights how a pulse of heating within the lower crust could also produce a pulse of cooling.

894   Blue shaded areas show the interval of rapid cooling indicated by QTQt models (**Fig. 6**). All

895   curves show increased cooling rates between about 35 and 25 Ma. Scenario 3 also shows a

896   reheating event.

897

898   **Fig. 10.** Principal features of the marine basins adjacent to the study area (Fyhn et al., 2009;

899   Schmidt et al., 2019; Vu et al, 2017; Morely et al., 2019) compared to the South China Sea

900   spreading history (Larsen et al., 2018; Le Pourhiet et al., 2017), exhumation data and volcanism

901   across the Indochina margin (Hoang et al., 2013).

902

Mid-Miocene unconformity Slope incised valley Carbonate build-up

Turbidite

Top of Oligocene

E

Two-way travel time (s)

0.5
1.0
1.5
2.0
2.5
3.0
3.5
4.0

Pre-Cenozoic basement

9 km

108°45'   109°00'   109°15'

12°30'

12°20'

Nha Trang

12°10'

SVN8
60.2±5.2

U-Pb
92.0±0.3

SVN4
31.6±2.7

SVN9
58.3±4.3

SVN7
40.8.3±3.7

SVN5
33.8±2.8

U-Pb
92.9±0.7

10 km

Inset plot:
- Elevation (m) vs Age (Ma)
- AHe, AFT

A)

Temperature (°C)

Granite emplacement

80   60   40   20   0
Time (Ma)

37-30 Ma

■	1474
■	1187
■	836
■	439
■	141
■	53
○	FT Age
△	AHe Age

Average temperature offset

Temperature offset (°C)

80   60   40   20   0
Time (Ma)

B)

Age (Ma)

Elevation (m)

Elevation (m)

MTL microns

▣	Kin in			
●	FT Age Obs	– –	FT Age Pred.	── Sampled FTA Pred.	⋯ MTL Pred.
✦	Sampled Pred.	▽	AHe Age Pred.	▲ AHe Age Obs.	── Sampled MTL Pred.
				◆ MTL Obs.

A)

B)