1 **Thermochronology and Tectonics of the Mérida Andes and the Santander Massif, NW South** 2 **America**

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11

12 **Abstract**

13 New apatite U-Pb and multiphase ⁴⁰Ar/³⁹Ar data constrain the high to medium temperature (~500°C
14 - ~300°C) thermal histories of igneous and metamorphic rocks exposed in the Mérida Andes of
15 Venezuela, and new apatite and zircon fission track data constrain the ~500 °C - ~60°C thermal
16 histories of pre-Jurassic igneous and metamorphic rocks of the adjacent Santander Massif of
17 Colombia. Computed thermal history envelopes using apatite U-Pb dates and grain size information
18 from an Early Palaeozoic granodiorite in the Mérida Andes suggest that it cooled from >500°C to
19 <350°C between ~266 Ma and ~225 Ma. Late Permian to Triassic cooling is also recorded in Early
20 Palaeozoic granitoids and metasedimentary rocks in the Mérida Andes by numerous new muscovite
21 and biotite ⁴⁰Ar/³⁹Ar plateau dates spanning 257.1±1.0 Ma to 205.1±0.8 Ma. This episode of
22 cooling is not recognised in the Santander Massif, where ⁴⁰Ar/³⁹Ar data suggest that some Early
23 Palaeozoic rocks cooled below ~320 °C in the Early Palaeozoic. However, most data from pre -
24 Jurassic rocks reveal a regional heat pulse at ~200 Ma during the intrusion of numerous shallow
25 granitoids, resulting in temperatures in excess of ~520°C, obscuring late Palaeozoic histories.

26

27 The generally accepted timing of amalgamation of Pangaea along the Ouachita-Marathon suture
28 pre-dates Late Permian to Triassic cooling recorded in basement rocks of the Mérida Andes by >30
29 Ma, and its effect on rocks preserved in north-western South America is unknown. We interpret late
30 Permian to Triassic cooling in the Mérida Andes to be driven by exhumation. Previous studies have
31 suggested that a short phase of shortening and anatexis is recorded at ~253 Ma in the Maya Block,
32 which may have been adjacent to the basement rocks of the Mérida Andes in the Late Permian. The
33 coeval onset of exhumation in the Mérida Andes may be a result of increased coupling in the
34 magmatic arc, which was located along the western margin of Pangaea. Triassic extension is

documented in the Central Cordillera of Colombia and Ecuador between ~240 Ma and ~215 Ma, although extension at this time has not been clearly identified in the Mérida Andes or the Santander Massif. Permian to Triassic cooling is not recorded in the structurally isolated Caparo Block in the southern Mérida Andes, suggesting that it may have constituted a distinct fault block in the Triassic.

New fission track data from the Santander Massif suggest that it started exhuming at ~40 Ma during a period of accelerated convergence between the Nazca/Farallón Plate and the western margin of South America. Exhumation in the Santander Massif occurred diachronously since ~18 Ma in distinct fault blocks at rates of 0.5 – 1 km/Ma, and may have been driven by east-west compression as a result of the indentation of the Panama-Chocó terrane to western Colombia.

1. Introduction

The Mérida Andes of Venezuela and the Santander Massif of Colombia (Fig. 1a) may have been in the collision zone between Laurussia and Gondwana when Pangaea amalgamated during the Late Palaeozoic. The amalgamation of Pangaea is recorded by high-grade deformation and metamorphism between ~350 and ~270 Ma in the Variscan/Hercynian orogenic belt in Europe (e.g. Liew and Hofmann, 1988; Franke, 1989) and in the Alleghanian orogenic belt in Eastern North America (e.g. Hatcher, 2002 and references therein). In contrast, at the westernmost extent of the collision zone between north-western Gondwana and southern Laurussia, Pangaea amalgamation only resulted in low grade, sub-greenschist facies metamorphism in the Ouachita – Marathon fold-thrust belt, which started deforming at ~320 Ma (Ross, 1986; Thomas, 2004; Viele and Thomas, 1989). The cessation of deformation in the Ouachita – Marathon segment was diachronous from late Pennsylvanian in the east to Early Permian in the west (Poole et al., 2005). A paucity of magmatism in the Ouachita-Marathon segment has led to the suggestion that subduction of oceanic crust beneath north-western Gondwana accounted for convergence which led to collision between north-western Gondwana and Laurentia (e.g. Poole et al., 2005). This paper investigates the thermal histories of Palaeozoic granites and gneisses that formed within autochthonous crustal blocks along the north-western margin of Gondwana, and constrains the tectonic history of those blocks during

63 and subsequent to the amalgamation of Pangaea. We explore the crustal response of north-western
64 Gondwana to the collision between Laurentia and Gondwana, and its subsequent evolution.

65

66 Previous studies of Precambrian to Palaeozoic igneous and metamorphic rocks in the Mérida Andes
67 of Venezuela (Fig. 1) interpreted numerous K/Ar and Rb/Sr dates spanning between ~300 and ~200
68 Ma (e.g. Bass and Shagam, 1959; Martin-Bellizzia et al., 1968; Schubert, 1969) to reflect a
69 Carboniferous to Permian orogenic event in north-western Gondwana (Grauch, 1972; Maréchal,
70 1983; Bellizzia and Pimentel, 1994). However, i) Ar is lost from minerals at sub-solidus
71 temperatures by thermally activated, volume diffusion, alteration and metasomatism, ii) the
72 interpretation of Rb/Sr model dates is highly problematic because both Rb and Sr can be mobilised
73 by alteration (e.g. Rollinson, 1993); and iii) Rb/Sr model dates assume an initial composition of
74 $^{87}\text{Sr}/^{86}\text{Sr}$, which can be highly variable in the earth's crust (e.g. Rollinson, 1993). Consequently,
75 these data do not constitute a robust basis for interpreting periods of metamorphism.

76

77 Zircon U-Pb dates and Pb isotopic data from granitoids, gneisses and volcanic rocks in the Mérida
78 Andes and the Santander Massif suggest that they are autochthonous to north-western Gondwana
79 (van der Lelij et al., 2015), and crystallised in two main phases at ~499 Ma - ~415 Ma and ~250 Ma
80 - ~196 Ma, during i) subduction of the oceanic crust of the Iapetus Ocean beneath north-western
81 Gondwana, and ii) subduction of oceanic crust of the proto-Pacific ocean beneath western Pangaea,
82 respectively (van der Lelij et al., 2015). Remnants of a magmatic arc which started at ~290 Ma may
83 be preserved in the Central Cordillera of Colombia and in the Sierra Nevada de Santa Marta (Fig.
84 1a) (e.g. Cardona et al., 2010; Villagómez et al., 2011). However, the tectonic history of these
85 regions between ~415 Ma and ~290 Ma are not constrained by reliable geochronological data.

86

87 The Cainozoic tectonic history of the Mérida Andes is well constrained by several new detailed
88 apatite fission track studies (e.g. Bermúdez et al., 2010; Bermúdez et al., 2011; Erikson et al.,
89 2012). However, the timing of Cainozoic tectonic events in the Santander Massif is only

constrained by one fission track study (Shagam et al., 1984) utilising the population method, which may be inaccurate due to measurement bias during sample preparation and analysis (Gleadow, 1981). No track lengths were measured, and reliable age elevation profiles could not be constructed because of the numerous faults that intersect the basement (Fig. 2; Shagam et al., 1984). Consequently, the rate and timing of Cainozoic exhumation in the Santander Massif is largely unconstrained.

We present 30 new $^{40}\text{Ar}/^{39}\text{Ar}$ dates and apatite U-Pb data from basement rocks of the Mérida Andes (Fig. 1b), and 9 new $^{40}\text{Ar}/^{39}\text{Ar}$ dates and 23 new apatite and zircon fission track dates from Palaeozoic to Early Jurassic igneous and metamorphic rocks exposed in the Santander Massif (Fig. 2). These dates are combined with stratigraphic constraints to i) evaluate existing models for the Late Palaeozoic evolution of north-western Gondwana, and ii) constrain the Cainozoic tectonic history which gave rise to the high elevation, of up to 4500m, of the Santander Massif.

2. Geological Framework and previous work

The Andes of Venezuela and Colombia host numerous structurally isolated regions of prominent topography such as the Sierra Nevada de Santa Marta, the Mérida Andes, the Santander Massif and the Serranía del Perijá (Fig. 1a), which are fault-bounded against rapidly subsiding basins (e.g. the Llanos Basin, the Magdalena Valley and the Maracaibo Basin, Fig. 1a). These regions have been structurally controlled by the diffuse South Caribbean Plate Boundary Zone, which has accommodated dextral displacement between the Caribbean and South American plates since the Late Cretaceous (e.g. Dewey and Pindell, 1985; van der Lelij et al., 2010).

Mountain building in the South Caribbean Plate Boundary Zone is generally considered to be driven by coupling between the Caribbean and South American plates and the suturing of the Panama-Costa Rica arc (Fig. 1a) to South America at ~24 Ma (Farris et al., 2011). These interactions drive east-west oriented compression, leading to the northwards escape of the Maracaibo Triangular

117 Block (Fig. 1a; e.g. Kellogg, 1984; Mann and Burke, 1984). Although movement along the northern
118 margin of the Maracaibo Triangular Block is dominantly dextral strike-slip (Fig. 1a), transpression
119 along its south-eastern and western margins gave rise to the high topography of the 4978m
120 Venezuelan Mérida Andes (e.g. Hoorn et al., 1995; Bermúdez et al., 2011), and the ~4400 m
121 Colombian Santander Massif (e.g. Shagam et al., 1984) in the Cainozoic .

122

123 **2.1. Mérida Andes**

124 The Mérida Andes form a north-east trending orogenic belt parallel to the dextral Boconó Fault
125 (Fig. 1b). Contrasting interpretations on the deep fault structure beneath the Mérida Andes have led
126 to different orogenic models (e.g. Colletta et al., 1997; Audemard and Audemard, 2002; Monod et
127 al., 2010), although all models predict that most Cainozoic exhumation occurred in the centre of
128 the massif, close to the Boconó Fault. The basement of the Mérida Andes consists of amphibolite
129 facies metasedimentary gneisses, schists and phyllites, and orthogneisses of the Iglesias Complex
130 (Fig. 1b; Gonzales de Juana et al., 1980). These basement units were intruded by numerous
131 granitoids during two main pulses, from ~499 Ma to ~415 Ma, and from ~244 to ~211 Ma (van der
132 Lelij et al., 2015).

133

134 Published geochronological data from Precambrian to Triassic igneous and metamorphic rocks of
135 the Mérida Andes mainly utilised the K/Ar and Rb/Sr model age methods, and yield dates ranging
136 from 660 ± 30 Ma to 61 ± 15 Ma (see compilation in van der Lelij et al., 2015). The reliability of these
137 data cannot be assessed, and they cannot be used to accurately constrain the post-crystallisation
138 thermal histories of the basement rocks (e.g. Rollinson, 1993; McDougall and Harrison, 1999).
139 Apatite fission track studies on igneous and metamorphic rocks of the Mérida Andes show its rocks
140 were exhuming at ~27 Ma, and that exhumation rates increased at ~13 Ma (Bermúdez et al., 2010;
141 Bermúdez et al., 2011; Erikson et al., 2012). Exhumation was asymmetric across the Boconó Fault,
142 and discrete fault blocks exhumed at different times and rates (e.g. Bermúdez et al., 2011). The

143 south-western Mérida Andes are separated from the Santander Massif by a north-east to south-west
144 directed thrust system which became active in the Pliocene (Macellari, 1984).

145

146 **2.2. Santander Massif**

147 The basement of the Santander Massif is mainly composed of gneisses and schists that are intruded
148 by numerous granitoids and covered by Late Palaeozoic, Mesozoic and Cainozoic sedimentary
149 rocks (Fig. 2; Ward et al., 1974). The most striking structural feature in the Santander Massif is the
150 Santa-Marta Bucaramanga Fault (SMBF; Fig. 2), which has both sinistral and vertical displacement
151 components. The SMBF separates basement inliers to the east from Mesozoic and Cainozoic
152 sedimentary rocks of the Middle Magdalena Valley Basin to the west (Fig. 2) and defines the
153 western boundary of the Maracaibo Triangular Block (Fig. 1a). Displacement along the SMBF has
154 led to structural dissection of the Santander Massif, forming rhomboidal fault blocks that expose
155 Precambrian to Paleozoic basement inliers, which are intruded by numerous Mesozoic granitoids
156 (van der Lelij et al., 2015; Fig. 2).

157

158 The post-crystallisation thermal history of the Santander Massif is poorly constrained by published
159 thermochronological data. Although some K/Ar dates have been reported (Goldsmith et al., 1971),
160 the reliability of these dates cannot be assessed. More recent studies utilised the more reliable
161 $^{40}\text{Ar}/^{39}\text{Ar}$ method, although many of the studied samples yield disturbed age spectra (Restrepo-Pace,
162 1995; Restrepo-Pace et al., 1997; Cordani et al., 2005), precluding straightforward interpretation of
163 the data. All published $^{40}\text{Ar}/^{39}\text{Ar}$ plateau dates from Precambrian to Early Palaeozoic basement
164 rocks in the Santander Massif range between 184 ± 3 and 213 ± 3 Ma (Fig. 2; Restrepo-Pace, 1995). A
165 similar range of dates was obtained from the Bucaramanga gneiss (Fig. 2) sampled at unspecified
166 localities in the Santander Massif by Cordani et al. (2005). These late Triassic to early Jurassic
167 $^{40}\text{Ar}/^{39}\text{Ar}$ plateau dates are collectively considered to be a consequence of reheating during the
168 intrusion of numerous shallow plutons (Restrepo-Pace, 1995; Cordani et al., 2005) at ~ 200 Ma (van
169 der Lelij et al., 2015; Dörr et al., 1995).

170

171 Published fission track dates from the Santander Massif used the population method (Shagam et al.,
172 1984), and thus they may be inaccurate (Gleadow, 1981). Zircon fission track data collected from
173 Jurassic and older basement rocks exposed in the Santander Massif define two age groups at 50 ± 6
174 Ma to 67 ± 7 Ma and 91 ± 11 Ma to 109 ± 10 Ma (Shagam et al., 1984). The older age group was
175 considered to reflect partial annealing, and the younger age group was considered to reflect
176 complete annealing as a result of high heat flow during the waning stages of subsidence in the late
177 Cretaceous (Fig. 3; Shagam et al., 1984). Apatite fission track dates from Jurassic and older
178 basement rocks range between 3.8 ± 0.8 Ma and 18.9 ± 2.2 Ma and correlate poorly with elevation
179 (Shagam et al., 1984), suggesting that exhumation occurred at different times and rates in discrete
180 fault blocks. An earliest Oligocene (~ 34 Ma; Gradstein et al., 2012) onset of exhumation was
181 postulated, based on the interpretation that the sedimentary rocks in the Eastern Piedmont (Figs. 2
182 and 3) derive from erosion of the Santander Massif (Shagam et al., 1984).

183

184 **3. Analytical methods**

185 **3.1. Apatite U-Pb analysis**

186 Granodiorite 08VDL44 from the Mérida Andes was selected for U-Pb apatite thermochronology
187 because it yields large amounts of inclusion free, euhedral hexagonal apatite crystals with grain
188 radii varying between ~ 20 μm and ~ 250 μm . The sample was crushed to <500 μm and apatites were
189 extracted using heavy liquids and magnetic separation. Inclusion free, euhedral apatites with a range
190 of grain radii were selected for analysis. Grain length and diameter were measured parallel and
191 perpendicular to the crystallographic c-axis using calibrated photomicrographs of the crystals, and
192 individual grains were rinsed and washed by ultrasound cleaning in ultrapure H_2O and ultrapure
193 acetone. After cleaning, single apatite crystals were loaded in 300 μl Savillex microcapsules, spiked
194 with 5-10 mg of the ^{205}Pb - ^{233}U - ^{235}U EARTHTIME tracer solution (Condon et al., 2007) and
195 dissolved in 6.2N HCl for 48h at 180°C . After dissolution, the solutions were dried down and re-
196 dissolved in 3.1N HCl. U and Pb were separated using single-column anion exchange chemistry. U

197 and Pb fractions were loaded separately on outgassed, zone-refined single Re filaments with a silica
 198 gel/HPO₃ emitter solution (Gerstenberger and Haase, 1997).
 199
 200 U and Pb isotopic ratios were measured on a Thermo Triton thermal ionization mass spectrometer
 201 (TIMS) in the Radiogenic Isotope Laboratory at the University of Geneva. Pb was measured using a
 202 dynamic peak jumping routine on a MasCom secondary electron multiplier (SEM). Mass
 203 fractionation on the electron multiplier was determined to be 0.13±0.03 ‰/amu based on daily
 204 analyses of the NBS-981 common Pb standard. U was measured as U-oxide in static mode on
 205 Faraday cups equipped with 10¹² Ω resistors, or using a dynamic peak jumping routine on the SEM.
 206 Measured isotopic ratios were corrected for interferences of ²³³U¹⁸O¹⁶O on ²³⁵U¹⁶O₂ using an ¹⁸O/¹⁶O
 207 of 0.00205, measured on large U500 loads, and mass fractionation was controlled using a double
 208 spike with a ²³³U/²³⁵U ratio of 0.99506. A laboratory blank value of 0.9±0.5 pg was subtracted from
 209 the total common Pb composition and the remaining common Pb was corrected using the measured
 210 whole rock initial Pb composition (²⁰⁶Pb/²⁰⁴Pb=18.387±0.184, ²⁰⁷Pb/²⁰⁴Pb=15.636±0.078,
 211 ²⁰⁸Pb/²⁰⁴Pb=38.154±0.5723; van der Lelij et al., 2015). U-Pb ratios and dates were calculated relative
 212 to a ²³⁵U/²⁰⁵Pb ratio of 100.23±0.023% (1σ) for the tracer and raw data were reduced using Tripoli
 213 and U-Pb_Redux software (Bowring et al., 2011; McLean et al., 2011). All uncertainties are reported
 214 at the 2σ confidence level and exclude systematic uncertainties associated with tracer calibration and
 215 decay constants. Isotopic dates were calculated using λ(²³⁸U)=1.55125x10⁻¹⁰ and λ(²³⁵U)=9.8485x10⁻
 216 ¹⁰ (Jaffey et al., 1971).
 217

218 **3.2. ⁴⁰Ar/³⁹Ar analysis**

219 Igneous and metamorphic rocks from the Mérida Andes and the Santander Massif were crushed and
 220 mineral separates were recovered using a Wilfley table, heavy liquids and magnetic separation
 221 techniques. Unaltered hornblende, biotite, muscovite and alkali feldspar were hand-picked, and the
 222 fractions were rinsed in de-ionized water in an ultrasonic bath. The mineral separates were
 223 subsequently packed in copper foil and irradiated in the CLICIT facility of the TRIGA reactor at

Oregon State University. J values were determined by measurement of simultaneously irradiated Fish Canyon Tuff sanidine, using an age of 28.02 ± 0.16 Ma (Renne et al., 1998). Samples were degassed by CO₂-IR laser step heating, and the extracted gas was passed through a cold trap chilled to ~150 Kelvin and purified in a stainless steel extraction line fitted with AP10 and GP50 (ST101) getters. Argon isotopes were analysed using a GV Instruments Argus mass spectrometer fitted with four high-gain 1×10^{12} Ω faraday detectors and one 1×10^{11} Ω faraday detector (⁴⁰Ar). Dates were calculated using the ⁴⁰K decay constant of Steiger and Jaeger (1977). Data reduction was performed using the ArArCALC 2.4 software (Koppers, 2002), plateau dates were determined using the criteria of Dalrymple and Lanphere (1974), and calculated dates are presented in Table 2 and on Figs. 1b and 2. Analytical data are reported in the appendix.

3.3. Fission track analysis

Zircon and apatite fractions were extracted from crushed igneous and metamorphic rock samples from the Santander Massif using conventional heavy liquid and magnetic separation techniques. Fission track analyses were performed by Apatite to Zircon Inc.. Apatite grains were mounted in epoxy, ground and polished to reveal an interior surface. Polished grain mounts were etched in 5.5 M HNO₃ at 21°C for 20 seconds, and spontaneous track densities were measured. Subsequently, grain mounts were exposed to ²⁵²Cf fission fragments to increase the etching efficiency of confined tracks (Donelick and Miller, 1991), and re-etched in the same conditions. Track lengths and their angle to the crystallographic c-axis were measured and calibrated by track length measurements in the standards used to construct the annealing model (Carlson et al., 1999; Ketcham et al., 2007).

Zircons were mounted in Teflon disks, ground and polished, and grain mounts were etched to reveal spontaneous fission tracks by immersion in a eutectic melt of NaOH + KOH at 220°C for 12-72 hours. The etched grain mounts were then cleaned in 48% HF for 15 min at room temperature. Spontaneous track densities were determined from multiple grains within each sample.

251 The uranium concentrations of apatite and zircon grains, from which track densities were previously
252 determined, were measured by laser ablation inductively coupled plasma mass spectrometry
253 (Hasebe et al., 2004) in the Geoanalytical Laboratory at Washington State University. Grain dates
254 were calculated following the procedure of Donelick et al. (2005) using zeta values of
255 8.2727 ± 0.1407 (apatite) and 0.0401 ± 0.0005 (zircon), determined by the measurement of multiple
256 standards by Paul O'Sullivan (Apatite to Zircon Inc.). The obtained fission track dates are shown in
257 Tables 2 and 3, and on Fig. 2.

258

259 **4. Results**

260 **4.1. Apatite U-Pb results**

261 All nine analysed apatites from granodiorite 08VDL44 in the Mérida Block (zircon concordant U-
262 Pb age 466.8 ± 2.5 Ma; van der Lelij, 2015; north-east in Fig. 1), with radii ranging from 34 μm to
263 250 μm yield concordant U-Pb dates, with $^{238}\text{U}/^{206}\text{Pb}$ dates ranging between 251.9 ± 7.7 Ma and
264 343.4 ± 2.8 Ma (Fig. 4a, Table 1). Most $^{238}\text{U}/^{206}\text{Pb}$ dates are coeval regardless of their corresponding
265 grain radii (Fig. 4b), although grains with radii of 100 μm and 206 μm yield slightly older dates of
266 272.9 ± 5.7 Ma and 343.4 ± 2.8 Ma. The poor correlation between U-Pb date and grain radius (Fig.
267 4b) suggests that either i) not all apatite grains underwent Pb loss by volume diffusion, ii) there
268 were unidentified crystal lattice defects or inclusions in some of the apatites measured, which
269 altered the average diffusion length relative to the grain size, iii) diffusion kinetics may not have
270 been identical for all apatite grains, or iv) there were influences of variable U zonation inside the
271 apatites.

272

273 The diffusivity of Pb in apatite may be a function of apatite composition, similar to the relationship
274 found between fission track annealing kinetics and halogen content in apatite (e.g. Carlson et al.,
275 1999). The chemical composition of the analysed apatites is unknown with the exception of their
276 Th/U content (Table 1), and the youngest 7 apatites yield a tight cluster of Th/U ratios that range
277 between 0.07 and 0.35 (Fig. 4C), whereas the two oldest grains yield very different Th/U ratios of

278 2.31 and 4.08 (Fig. 4C). Low Th/U ratios <0.35 are typical of apatites which formed in S- Type
279 granites, whereas high Th/U ratios >1 are typical of apatites which formed in I type granites (Chu et
280 al., 2009). Quartz O isotope data, radiogenic isotope data and whole rock geochemistry (van der
281 Lelij et al., 2015) suggest that Granodiorite 08VDL44 is a peraluminous hybrid I-Type – S-Type
282 granitoid, and therefore the high Th/U apatites may be xenocrystic (Cochrane et al., 2014a). The
283 Th/U ratios may co-vary with other trace and rare earth elements in apatite (Chu et al., 2009), and
284 we hypothesize that these compositional effects may have a significant influence on the diffusion
285 kinetics of Pb. Even though the effect of apatite chemistry on Pb diffusion is hypothetical, we
286 assume that such effects are likely, and we choose to ignore data from the two grains that have
287 higher Th/U ratios.

288

289 The 7 remaining apatite dates are similar, although the grain with the largest radius yields a
290 distinguishably older U-Pb date (Fig. 4b). These dates, combined with the range in grain sizes
291 suggests that Pb loss could have been controlled by thermally activated, volume diffusion, and that
292 granodiorite 08VDL44 cooled rapidly through the apatite partial Pb retention zone of the different
293 measured grains.

294

295 **4.2. $^{40}\text{Ar}/^{39}\text{Ar}$ results - Mérida Andes**

296 Hornblende, muscovite, biotite and alkali feldspar extracted from 21 igneous and metamorphic
297 rocks exposed in the Mérida Andes yield 30 $^{40}\text{Ar}/^{39}\text{Ar}$ dates ranging from 464.3 ± 2.4 Ma to
298 116.4 ± 1.1 Ma (Table 2; Fig. 1b; Fig. 5). Of these, thirteen aliquots yield disturbed $^{40}\text{Ar}/^{39}\text{Ar}$ spectra,
299 which suggests that K/Ar distributions in the analysed mineral phases were affected by alteration.
300 These yield weighted mean $^{40}\text{Ar}/^{39}\text{Ar}$ dates either of i) statistically non-age equivalent steps, or ii)
301 age equivalent steps representing less than 50% of the total ^{39}Ar , which are less reliable than plateau
302 dates (Table 2) and were interpreted with caution.

303

304 The oldest $^{40}\text{Ar}/^{39}\text{Ar}$ dates from hornblende, muscovite and biotite in the Mérida Andes were
305 obtained from granitoids exposed in the south-eastern foothills (Caparo Block, Fig. 1b), and range
306 from 464.3 ± 2.4 (Hornblende; 08VDL55; Table 2; Fig. 5) to 404.7 ± 2.1 Ma (Biotite; 08VDL57). In
307 contrast, granitoids, schists and gneisses from the Mérida Block yield younger $^{40}\text{Ar}/^{39}\text{Ar}$ dates (Fig.
308 5), with only four $^{40}\text{Ar}/^{39}\text{Ar}$ dates ranging from 379.6 ± 1.5 Ma to 323.8 ± 1.0 Ma (Fig.1b, Fig. 5).
309 These Devonian to Carboniferous dates were obtained from i) granitic gneiss 08VDL27 (e.g. Fig. 6;
310 Table 2) and granodiorite 08VDL32, located close to the Caparo Block (Fig. 1b), and ii) a
311 muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ weighted mean date of 379.6 ± 1.5 Ma was obtained from granite 08VDL41,
312 which is exposed in the central part of the Mérida Andes.

313
314 Most granitoids and gneisses from the Mérida Block with zircon U-Pb crystallisation ages of ~ 489
315 $- 211$ Ma yield muscovite and biotite $^{40}\text{Ar}/^{39}\text{Ar}$ dates which cluster between ~ 257 Ma and ~ 200 Ma
316 (Fig. 5). Biotite-muscovite-chlorite schist 08VDL16 from the Mucuchachi Fm (Fig. 1a), which is
317 interfingering with volcanic tuffs yielding a zircon U-Pb age of 452.6 ± 2.7 Ma (Van der Lelij et al.,
318 2015) has biotite porphyroblasts yielding a $^{40}\text{Ar}/^{39}\text{Ar}$ plateau date of 254.7 ± 1.0 Ma (Fig.6, Table 2).

319
320 Five mineral separates yield muscovite and biotite disturbed $^{40}\text{Ar}/^{39}\text{Ar}$ spectra with weighted mean
321 dates ranging between 182.3 ± 1.1 Ma (08VDL21; Table 2) and 116.4 ± 1.1 Ma (08VDL23; Table 2),
322 although the disturbed $^{40}\text{Ar}/^{39}\text{Ar}$ spectra (Appendix Fig. 1) suggest that these dates may not be
323 reliable.

324

325 **4.3. $^{40}\text{Ar}/^{39}\text{Ar}$ results - Santander Massif**

326 Palaeozoic to Jurassic igneous and metamorphic rocks in the Santander Massif yield nine new
327 hornblende, muscovite and biotite $^{40}\text{Ar}/^{39}\text{Ar}$ dates, ranging between 450.2 ± 1.6 Ma and 196.6 ± 1.1
328 Ma (Table 2, Fig. 2, Fig. 5). Four of the analysed aliquots do not yield age plateaux as defined by
329 Dalrymple and Lanphere (1974), suggesting that the K-Ar systematics were disturbed (Table 2).

330 Three gneisses and granodiorites yield Early Palaeozoic $^{40}\text{Ar}/^{39}\text{Ar}$ dates. Muscovite from migmatite
331 10VDL45 in Bucaramanga Gneiss yields a $^{40}\text{Ar}/^{39}\text{Ar}$ plateau date of 450.2 ± 1.6 Ma. Anatectite
332 10VDL48 crosscuts Silgará Gneiss sample 10VDL49 (U/Pb age of 479.8 ± 3.1 Ma; van der Lelij et
333 al., 2015) and yields a weighted mean muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ date of 407.4 ± 1.4 , although this was
334 calculated from only 44% of the age spectrum (Table 2). Biotite from granodiorite 10VDL55 (U/Pb
335 age: 443.4 ± 3.2 Ma) yields a weighted mean $^{40}\text{Ar}/^{39}\text{Ar}$ date of 428.7 ± 1.6 Ma from the two highest
336 temperature degassing steps (Table 2).

337

338 Three samples with Early Palaeozoic zircon U/Pb crystallisation ages (10VDL23, 10VDL37 and
339 10VDL51; Table 2; van der Lelij et al., 2015) yield Late Triassic to Early Jurassic $^{40}\text{Ar}/^{39}\text{Ar}$ dates.
340 Banded gneiss 10VDL23 (U/Pb age of 461.0 ± 2.1 Ma; van der Lelij et al., 2015) may be a meta-
341 igneous rock which underwent post-magmatic deformation at lower amphibolite facies (see van der
342 Lelij et al., 2015), and yields a weighted mean hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ date of $204.1.4 \pm 3.5$ Ma.
343 Hornblende from Orthogneiss 10VDL37, which formed at 451.5 ± 1.3 Ma and locally underwent
344 anatexis at $208.8 \text{ Ma} \pm 1.2$ Ma (see Figure 9 in van der Lelij et al., 2015), yields a plateau $^{40}\text{Ar}/^{39}\text{Ar}$
345 date of 205.7 ± 0.9 Ma (Table 2, Fig. 2). Finally, orthogneiss 10VDL51 (U/Pb age of 472.5 ± 3.4 Ma;
346 van der Lelij et al., 2015) yields a biotite plateau $^{40}\text{Ar}/^{39}\text{Ar}$ date of 198.8 ± 0.8 Ma.

347

348 Biotite from granodiorite 10VDL32 (zircon U/Pb age of 198.0 ± 0.8 Ma; van der Lelij et al., 2015)
349 yields a weighted mean $^{40}\text{Ar}/^{39}\text{Ar}$ date of 196.6 ± 1.1 Ma from a slightly disturbed spectrum
350 (MSWD=4.1; Table 2). Finally, $^{40}\text{Ar}/^{39}\text{Ar}$ data were obtained from two samples from the Silgará
351 Formation, which did not yield U-Pb ages: muscovite from anatectite 10VDL20 yielded a plateau
352 age of 210.4 ± 1.0 Ma, and muscovite from garnet metatuff 10VDL36 yields a plateau date of
353 199.3 ± 0.7 Ma.

354

355 4.4 Fission Track results

Eleven igneous and metamorphic rocks from the Santander Massif yield pooled zircon fission track (ZFT) dates ranging between 227.2 ± 22.1 Ma and 37.4 ± 4.0 Ma (Table 3; Fig 2). All zircon fission track dates have $P(\chi^2)$ greater than 17.9 % (Table 3), suggesting that the zircon fission track dates of each rock define single date populations. The oldest ZFT date of 227.2 ± 22.1 Ma was obtained from Tonalite 10VDL52, which has a zircon U-Pb crystallisation age of 199.8 ± 1.2 Ma (van der Lelij et al., 2015), suggesting that there may have been a slight counting bias during analysis of this sample, and that it cooled rapidly to temperatures lower than $\sim 200^\circ\text{C}$ after crystallisation. All other ZFT dates are younger than the zircon U-Pb crystallisation ages obtained from the same rocks, and increase with linear distance from the Santa Marta - Bucaramanga Fault (Fig. 10).

Twelve igneous and metamorphic rocks from the Santander Massif yield pooled apatite fission track (AFT) dates ranging between 20.0 ± 3.3 Ma and 4.6 ± 0.4 Ma ($\pm 2\sigma$; Table 4; Fig. 2). The $P(\chi^2)$ value of all apatite samples is $>24\%$, suggesting that all samples are composed of homogeneous age populations. Mean c-axis projected track lengths for all samples range between 14.47 ± 0.09 μm and 15.37 ± 0.06 μm , suggesting that all samples cooled rapidly through the apatite partial annealing zone (60°C - 110°C ; Laslett et al., 1987). Granitoids located to the west of the SMBF yield increasing AFT dates with distance from the fault (Fig. 10), while this trend is absent for samples to the east of the SMBF. There are no clear date v elevation relationships. The youngest AFT date of 4.6 ± 0.4 Ma is from gneiss 10VDL49, which crops out in the easternmost Santander Massif (Fig. 2).

5. Recovery of thermal histories

5.1. Apatite U-Pb data

Most published studies interpret apatite U-Pb data either by i) assuming certain closure temperatures for analysed grains (e.g. Chamberlain and Bowring, 2001; Flowers et al., 2005), ii) generating forward thermal history models and qualitatively comparing them with measured data (Schoene and Bowring, 2007), or iii) estimating Pb loss by the degree of discordance of several measured grains (Blackburn et al., 2011). However, i) the effective closure temperature of an apatite

383 grain depends on multiple variables including cooling rate and grain size; ii) forward modelling
384 generally prohibits testing a large range of t-T possibilities, and iii) the analytical uncertainties of
385 measured $^{235}\text{U}/^{207}\text{Pb}$ dates in apatite are usually too high to resolve Pb-loss in Phanerozoic apatites.
386 Cooling rates of 100 - 1°C/Ma yield bulk closure temperatures (Dodson, 1973) that range between
387 590 - 487°C (for our largest grain with a radius of 250 μm) and 499 – 415°C (for our smallest grain
388 with a radius of 34 μm) using the diffusion coefficients of Cherniak et al. (1991).

389

390 We follow the approach of Cochrane et al. (2014a) and utilise a Monte Carlo simulation on forward
391 models of thermal histories, using the HeFTy software (Ketcham, 2005). Input parameters were the
392 kinetic parameters determined by Pb diffusion experiments in apatite ($E_A=54.6$ kcal/mol and D_0
393 $=0.000127$ cm/s; Cherniak et al., 1991) and measured apatite $^{238}\text{U}/^{206}\text{Pb}$ dates and radii of seven
394 grains with similar Th/U ratios. Approximately 200,000 random theoretical time(t)-temperature(T)
395 paths were generated, and model dates were generated for each measured grain radius by
396 determining the net summation of in-growth and diffusive loss of ^{206}Pb over the entire time span of
397 the simulation. The model dates were compared to the measured $^{238}\text{U}/^{206}\text{Pb}$ dates, and goodness-of-
398 fit values were assigned to each t-T path as a function of the deviations of modelled dates from
399 measured dates and their assigned uncertainties (see also Ketcham et al., 2000; Ketcham, 2005).
400 Theoretical t-T paths with a goodness of fit of >0.05 were considered to be acceptable fits and paths
401 with a goodness of fit of >0.5 were considered to be good fits. The acceptable and good fit
402 envelopes, and the best fit thermal history solution are presented in Fig. 7b. Time-temperature paths
403 that undergo rapid cooling from $>500^\circ\text{C}$ at ~ 260 Ma to below 360°C at ~ 230 Ma, at minimum rates
404 of $4.4^\circ\text{C}/\text{Ma}$, yield theoretical U-Pb dates that match the measured dates of the low Th/U grains
405 (Fig. 7a).

406

407 **5.2. $^{40}\text{Ar}/^{39}\text{Ar}$ data**

408 Daughter isotopes are lost from minerals by volume diffusion within a temperature range that
409 depends on intrinsic diffusion properties, diffusion length and cooling rate (e.g. Dodson, 1973).

410 Assuming a range of cooling rates of 100-1°C/Ma and a diffusion length of 100 µm, calculated Tc
411 ranges during cooling are: i) 520±40°C for hornblende (spherical geometry; Harrison, 1982), ii)
412 440±50°C for muscovite (planar sheet geometry; Harrison et al., 2009) and iii) 320±30°C for biotite
413 (planar sheet geometry; McDougall and Harrison, 1999). Alkali feldspar has more complicated
414 diffusion properties related to its mineral structure (e.g. Lovera et al., 1989), and we assume that the
415 high temperature degassing steps have an approximate Tc of ~300°C (McDougall and Harrison,
416 1999).

417

418 **5.3. Fission track data**

419 Numerous studies of fission track annealing kinetics in zircon have led to contrasting estimates of
420 the temperature range of partial annealing (e.g. Yamada et al., 1995; Tagami et al., 1996; Tagami
421 and O'Sullivan, 2005; Bernet, 2009). Our zircon fission track data is not modelled to generate t-T
422 solutions. Rather, we use a partial annealing zone of 180-260°C (Bernet, 2009) to calculate
423 minimum temperatures in the Santander Massif.

424

425 Variations in fission track annealing kinetics in apatite are estimated in our models by utilising the
426 kinetic parameter r_{mr0} , which was calibrated against in-situ measurements of Cl, F, Fe, Mn, OH and
427 the sum of other cations (Ketcham et al., 2007). Forward modelling was performed using the
428 HeFTy software (Ketcham, 2005) using the annealing model of Ketcham (2007), which accounts
429 for crystallographic orientation effects. Input constraints were identical for all modelled samples,
430 allowing thermal histories to commence between 100°C and 200°C, in or above the apatite partial
431 annealing zone (60°C - 110°C; Laslett et al., 1987) and between 30 and 20 Ma. Between 10^4 and
432 10^5 random thermal history paths were generated using a controlled random search procedure, and
433 model fission track dates and track length distributions were calculated for each random path. The
434 modelled data were compared to the measured data using the Kolmogorov-Smirnov test, and
435 goodness of fit values were attributed accordingly. Paths with a goodness of fit of >0.05 were
436 considered to be acceptable fits and paths with a goodness of fit of >0.5 were considered to be good

fits. Minimum and maximum cooling rates were determined by determining the minimum and maximum slopes of lines in the envelope defined by the good fit thermal history solutions of a given sample.

Modelled thermal history solutions for the AFT data yield good fits for all samples except for granodiorite 10VDL22, which yields only acceptable fits (Appendix Fig. 2). AFT data from tonalite 10VDL52 only yields acceptable or good fits when the model is allowed to cool to $<10^{\circ}\text{C}$ at the present day (Appendix Fig. 2) because its long mean track length of $15.37 \pm 0.06 \mu\text{m}$ (Table 4) is incompatible with recent annealing. This granodiorite was collected at an altitude of 3731m (Table 4) where mean surface temperatures of $\sim 5^{\circ}\text{C}$ may be typical in Colombia (Sarmiento, 1986). However, the palaeo-elevation of sample 10VDL52 is unknown, and these temperatures lie well outside the range of temperature sensitivity of apatite fission track data (APAZ: $60^{\circ}\text{C} - 110^{\circ}\text{C}$; Laslett et al., 1987), and the thermal histories below 60° are not constrained by our data.

6. Periods of heating and cooling

6.1. Mérida Andes

The new $^{40}\text{Ar}/^{39}\text{Ar}$ data from the Mérida Andes can be divided into three main groups according to locality and date (Fig. 5): i) $^{40}\text{Ar}/^{39}\text{Ar}$ dates from the Caparo Block (Fig. 1b), which range between $464.3 \pm 2.4 \text{ Ma}$ (08VDL55, hornblende; Fig. 5) and $404.7 \pm 2.1 \text{ Ma}$ (08VDL57, biotite); ii) Devonian to Carboniferous $^{40}\text{Ar}/^{39}\text{Ar}$ dates from 3 granitoids in a transect across the Mérida Block ranging between $379.6 \pm 1.5 \text{ Ma}$ (08VDL41, muscovite) and $323.8 \pm 1.0 \text{ Ma}$ (08VDL32, biotite), although sample 08VDL41 also yields a biotite $^{40}\text{Ar}/^{39}\text{Ar}$ date of $212.8 \pm 0.9 \text{ Ma}$, and iii) Late Permian to Mesozoic $^{40}\text{Ar}/^{39}\text{Ar}$ dates that range between $257.1 \pm 1.0 \text{ Ma}$ (08VDL04, muscovite) and $168.7 \pm 1.8 \text{ Ma}$ (08VDL05, biotite) from scattered locations with no geographic age trends.

6.1.1. Caparo Block

463 The Caparo Block is the only area in the Mérida Andes where non-metamorphosed, Early
464 Palaeozoic sedimentary rocks (the Caparo and El Horno Fms.; Fig. 3) are preserved. Tonalite
465 08VDL55 is unconformably overlain by rocks of the Llandovery – Early Wenlock El Horno Fm.
466 (444-426 Ma; Gonzales de Juana et al., 1980), and it must have exhumed before ~444 Ma. The
467 hornblende $^{40}\text{Ar}/^{39}\text{Ar}$ plateau date of 464.3 ± 2.4 Ma may relate to exhumation after crystallisation at
468 ~473.1 $\pm$ 3.5 Ma (Fig. 5). Muscovite and biotite $^{40}\text{Ar}/^{39}\text{Ar}$ dates of 425.3 ± 1.9 and 404.7 ± 2.1 Ma,
469 respectively, from granite 08VDL57 (U-Pb zircon age: 492.3 ± 5.3 Ma; Fig. 1b) suggest that it
470 cooled during this interval. A magmatic and sedimentary hiatus starts at ~415 Ma in the Mérida
471 Andes (Fig. 3; Van der Lelij et al., 2015), which was coeval with a magmatic hiatus along most of
472 the western margin of Gondwana (Bahlburg et al., 2009). Coeval sedimentary formations are not
473 known in the Mérida Andes and the Santander Massif, and perhaps these regions were above sea
474 level during the Early Devonian.

475

476 **6.1.2. Mérida Block**

477 The oldest recognised cover sequence in the Mérida Block is the more than 5000m thick
478 Mucuchachi Fm. (Compañía Shell de Venezuela and Creole Petroleum Corporation, 1964), which
479 unconformably overlies the upper amphibolite facies basement of the Iglesias Complex throughout
480 the mountain range (Fig. 1b; Fig. 3). New zircon U-Pb data from volcanic tuffs in the Mucuchachi
481 Fm. constrain the age of its basal layers to 452.6 ± 2.7 Ma (van der Lelij et al., 2015). This Late
482 Ordovician age conflicts with palaeontological determinations made from non-metamorphic rocks
483 attributed to the Mucuchachi Fm. near the Caparo Fault (Fig. 1b), which suggest a Pennsylvanian
484 age (~318 - 299 Ma; Odremán and Useche, 1997). However, it was previously suggested that these
485 ages are not representative of the entire Mucuchachi Fm. (Compañía Shell de Venezuela and
486 Creole Petroleum Corporation, 1964), and the unit mapped as the Mucuchachi Fm. may therefore
487 include Ordovician to Carboniferous sedimentary rocks. Chlorite schist 08VDL16 (Fig. 1b) was
488 sampled from the Mucuchachi Fm. where it is interfingered with Ordovician volcanic tuffs
489 (Staargaard and Carlson, 2000). Biotite crystals in this rock are randomly oriented and surrounded

490 by aligned chlorite crystals. The largest chlorite crystals are located in the strain shadows of the
491 biotite, suggesting that the biotite crystals are porphyroclastic. The poorly defined Mucuchachi Fm.
492 hosts both low grade and non-metamorphosed sedimentary sequences of different ages, which may
493 be separated by at least one unconformity (Figs. 3 and 9).

494

495 The total age range during which the Mucuchachi Fm. may have been deposited brackets the period
496 of cooling recorded by muscovite and biotite $^{40}\text{Ar}/^{39}\text{Ar}$ dates of $\sim 379.6 \pm 1.5$ to 323.8 ± 1.0 Ma in
497 three granitoids in the north-eastern Mérida Andes (08VDL27, 08VDL32, 08VDL41; Fig. 1b).
498 These granitoids are located close to each other although they are separated by faults. The $^{40}\text{Ar}/^{39}\text{Ar}$
499 dates from these rocks are older than the ~ 257 -200 Ma dates obtained from other samples in the
500 Mérida Andes, suggesting that during the Late Permian to Triassic, these granitoids may already
501 have been below the closure temperature of muscovite ($440 \pm 50^\circ\text{C}$) and at least one granitoid may
502 already have been below the closure temperature of biotite ($320 \pm 30^\circ\text{C}$) by ~ 324 Ma. Alternatively,
503 Ar may have been partially lost from the analysed mineral phases before ~ 257 Ma, although our
504 data does not discriminate between these possibilities. Granite 08VDL41 (zircon U-Pb age
505 454.5 ± 3.1 Ma, Fig. 1b; Van der Lelij et al., 2015) also yields a biotite $^{40}\text{Ar}/^{39}\text{Ar}$ date of 212.8 ± 0.9 ,
506 which we interpret to be reset by the intrusion of granodiorite 08VDL14 at $\sim 211.2 \pm 1.2$ Ma, which
507 is located 9 km from the sample. A muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ plateau date of 234.8 ± 1.3 Ma from
508 granodiorite 08VDL44 falls outside of the acceptable fit envelope defined by forward modelled
509 apatite U-Pb data (Fig. 7b). Therefore, either the utilised diffusion kinetics i) for Pb in apatite
510 (Cherniak et al., 1991), or ii) for Ar in muscovite (Harrison et al., 2009) may not be applicable to
511 the mineral phases in our samples, suggesting that either i) the closure temperature for apatites from
512 granodiorite 08VDL44 are underestimated, or ii) the closure temperature of muscovite from
513 granodiorite 08VDL44 is overestimated. Unfortunately, our data are not sufficient to distinguish
514 between these possibilities.

515

516 A summary t-T envelope of combined apatite U-Pb thermal history solutions and $^{40}\text{Ar}/^{39}\text{Ar}$ data
517 shows that most Early Palaeozoic igneous and metamorphic rocks in the Mérida Block cooled
518 rapidly from $>500^\circ\text{C}$ to $<320\pm 30^\circ\text{C}$ between 257 Ma and 200 Ma (Fig. 9). The apatite U-Pb
519 thermal history model (Figs. 7 and 9) suggests that granodiorite 08VDL44 was at $\geq 500^\circ\text{C}$ before
520 ~ 260 Ma. High grade metamorphism resulting in anatexis is not recorded in the Mérida Andes
521 during this period (Van der Lelij et al., 2015), suggesting that peak temperatures of Early
522 Palaeozoic igneous and metamorphic rocks were below $\sim 600^\circ\text{C}$ (Thompson, 1999), which provides
523 an upper temperature range for our samples between Early Palaeozoic crystallisation and cooling at
524 ~ 257 Ma. The numerous muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ dates that range between ~ 251 Ma and ~ 200 Ma from
525 Early Palaeozoic granitoids (Fig. 5) also suggest that these were at minimum temperatures of
526 $>440\pm 50^\circ\text{C}$ before ~ 251 Ma. Three granitoids from the Mérida Andes yield Triassic zircon U-Pb
527 ages that range between 243.5 ± 3.4 Ma and 211.2 ± 1.2 Ma (Fig. 5), and one biotite $^{40}\text{Ar}/^{39}\text{Ar}$ date
528 (08VDL41) may have been reset by the intrusion of the largest granodiorite body in the Mérida
529 Andes (08VDL11; Fig. 1b) at ~ 211 Ma. In contrast, contact metamorphism does not seem plausible
530 to explain the spatial distribution of late Permian to Triassic $^{40}\text{Ar}/^{39}\text{Ar}$ dates, and we suggest that
531 cooling was driven by exhumation.

532

533 Biotite from metasedimentary chlorite schist 08VDL16 (Mucuchachi Fm; Fig. 1b1b) yields a
534 $^{40}\text{Ar}/^{39}\text{Ar}$ plateau date of 254.7 ± 1.0 Ma, which suggests that it was at temperatures of $> 320\pm 30^\circ\text{C}$
535 after deposition. This is consistent with the mineral assemblage of this rock, which suggests that it
536 was metamorphosed at low temperatures (Spear, 1993). The schist is inter-fingered with volcanic
537 tuffs of the Mucuchachi Fm., which erupted at $\sim 452.6\pm 2.7$ Ma (van der Lelij et al., 2015), and
538 hence it must have been metamorphosed between ~ 450 Ma and ~ 255 Ma, and cooled through
539 $320\pm 30^\circ\text{C}$ at ~ 255 Ma.

540

541 Cooling between ~ 260 and ~ 200 Ma post-dates the deposition of the Sabaneta and Palmarito Fms.
542 and pre-dates the deposition of the La Quinta Fm. (Fig. 9), which crops out throughout the Mérida

Andes (Fig. 1b). The basal tuffs of the La Quinta Fm. yield a biotite $^{40}\text{Ar}/^{39}\text{Ar}$ plateau date of 200.4 ± 0.9 Ma (Table 2), which is indistinguishable from their zircon U-Pb age of $\sim 202.0 \pm 1.6$ Ma (Table 2; Van der Lelij et al., 2015), suggesting that the base of the La Quinta Fm. was not exposed to temperatures in excess of $320 \pm 30^\circ\text{C}$ (Fig. 9).

Paragneiss 08VDL05 from the Sierra Nevada Fm. (Fig. 1b) is the only sample which yields a muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ plateau date younger than the La Quinta Fm., of 170.1 ± 1.0 Ma, and its significance is unclear, although it is equivalent to a disturbed biotite date of 168.7 ± 1.8 Ma from the same rock. All other $^{40}\text{Ar}/^{39}\text{Ar}$ dates younger than ~ 200 Ma were weighted mean dates calculated from disturbed age spectra, suggesting that most of the presently exposed basement in the Mérida Andes may have been at temperatures $< 320 \pm 30^\circ\text{C}$ by ~ 200 Ma.

Late Cainozoic rapid cooling is documented by apatite fission track data, which show that discrete fault blocks in the Mérida Andes cooled diachronously from $\sim 120^\circ\text{C}$ at rates of $16\text{--}40^\circ\text{C}/\text{Ma}$ since ~ 12 Ma (Bermúdez et al., 2011; Erikson et al., 2012).

6.2. Santander Massif

Only the three easternmost and northernmost samples yield Early Palaeozoic $^{40}\text{Ar}/^{39}\text{Ar}$ dates, of which only one is a plateau date. Muscovite from a migmatite in Bucaramanga gneiss (10VDL45; Table 2, Fig. 2) yields a plateau date of 450.2 ± 1.6 Ma, which likely reflects cooling below $440 \pm 50^\circ\text{C}$ following regional anatexis during peak metamorphism at ~ 472 Ma (van der Lelij et al., 2015). The two disturbed $^{40}\text{Ar}/^{39}\text{Ar}$ dates from granodiorite 10VDL55 (428.7 ± 1.6 Ma; biotite; Table 2) and anatectite 10VDL48 (407.4 ± 1.4 Ma; muscovite, Table 2) may loosely constrain cooling below ~ 320 and $\sim 440^\circ\text{C}$ respectively, but these dates may also result from partial loss of radiogenic ^{40}Ar during the intense magmatic activity in the late Triassic to early Jurassic. Zircon fission track data from the easternmost and northernmost samples indicate that the samples were hotter than $180\text{--}260^\circ\text{C}$ (Bernet, 2009) at ~ 195 Ma (Fig. 2), but were not sufficiently hot for extended periods of time

570 to fully reset the K-Ar systematics of biotite ($320\pm 30^{\circ}\text{C}$; 10VDL55) or muscovite ($440\pm 50^{\circ}\text{C}$
571 10VDL45; 10VDL48).

572

573 A vast majority of new and previously reported $^{40}\text{Ar}/^{39}\text{Ar}$ plateau dates from Precambrian,
574 Ordovician and Mesozoic granitoids and gneisses from the Santander Massif cluster between ~ 215
575 Ma and ~ 185 Ma (Restrepo-Pace, 1995; Cordani et al., 2005; Fig. 2), and most of these dates are
576 interpreted to reflect high heat flow during the intrusion of magma that formed the Santander
577 Plutonic Group between ~ 205 Ma and ~ 196 Ma (Fig. 2; Dörr et al., 1995; Restrepo-Pace, 1995;
578 Cordani et al., 2005; van der Lelij et al., 2015). These data also suggest that a majority of the ~ 200
579 Ma granitoids in the Santander Massif rapidly cooled below the biotite closure temperature
580 ($\sim 320\pm 30^{\circ}\text{C}$), which is consistent with their porphyritic texture (van der Lelij et al., 2015),
581 suggesting intrusion at shallow depth.

582

583 Zircon fission track dates from the Santander Massif span between ~ 227 Ma and ~ 37 Ma, and the
584 oldest dates are coeval with the intrusion of numerous granitoids at ~ 200 Ma (Fig. 2; van der Lelij
585 et al., 2015). We follow the hypothesis of previous authors (e.g. Restrepo-Pace, 1995; Cordani et
586 al., 2005) that the crystallisation of large volumes of magma at ~ 200 Ma gave rise to very high
587 geothermal gradients and complete annealing of fission tracks in zircon at ~ 200 Ma in our samples
588 which yield Early Palaeozoic zircon U-Pb ages. Therefore, the oldest zircon fission track dates of
589 ~ 200 Ma reflect cooling through $\sim 260^{\circ}\text{C}$ - 180°C (Bernet, 2009) during thermal relaxation after the
590 intrusion events. The youngest zircon fission track dates of 37.4 ± 4.0 Ma, 38.3 ± 4.3 Ma and 44.1 ± 4.6
591 Ma were obtained from samples located close to the Santa-Marta Bucaramanga Fault (Figs. 2 and
592 10) and the intermediate zircon fission track dates that range between 72.9 ± 7.4 Ma and 172.0 ± 16.4
593 Ma increase with distance from the fault, on both its eastern and western walls, regardless of the
594 zircon U-Pb ages of the samples (Fig. 10). Sedimentation over basement rocks of the Santander
595 Massif appears to have been discontinuous in the Jurassic, and continuous throughout the
596 Cretaceous to Palaeocene (Fig. 3), resulting in a cumulative cover thickness of ~ 2500 - ~ 3500 m

597 (Shagam et al., 1984). Burial was insufficient to heat samples yielding zircon fission track dates of
598 ~200 Ma to within the partial annealing zone (~180°C to ~260°C; Bernet, 2009), perhaps because
599 they were not sufficiently deep below the unconformity with the Mesozoic cover sequences. Zircon
600 fission track dates between 72.9 ± 7.4 Ma and 172.0 ± 16.4 Ma are coeval with periods of subsidence
601 rather than exhumation (Fig. 3), and therefore may reflect prolonged residence in the zircon partial
602 annealing zone. However, the three youngest dates between ~37 and ~44 Ma are statistically
603 indistinguishable, post-date the youngest cover sequences exposed in the Santander Massif, and are
604 roughly coeval with the onset of sedimentation in the northern Middle Magdalena Valley Basin and
605 in the Eastern Piedmont (Figs. 2 and 3), and therefore may record cooling through the zircon partial
606 annealing zone at ~40 Ma.

607

608 Detrital zircon studies (Nie et al., 2012) reveal a change in sedimentary provenance at ~40 Ma in
609 the Middle Magdalena Valley Basin where the source regions changed from mainly 100 Ma - 300
610 Ma rocks for the La Paz Fm. to mainly Precambrian sources for the Esmeraldas and younger
611 formations (Fig. 3). This switch is interpreted to reflect the onset of exhumation of the Santander
612 Massif and the Eastern Cordillera (Nie et al., 2012). However, the Santander Massif has numerous
613 igneous and metamorphic rocks with zircon U-Pb ages of ~483 - ~439 Ma and ~209 – 196 Ma (van
614 der Lelij et al., 2015), which are not documented in the Middle Magdalena Valley Basin until the
615 Oligocene Mugrosa Fm. (~30 Ma; Fig. 3; Nie et al., 2012). Cretaceous rocks in the adjacent
616 Floresta Massif (Fig. 1a) mainly host Precambrian zircons (Horton et al., 2010), and we hypothesize
617 that Cretaceous cover sequences of the Santander Massif may have hosted similar detritus as the
618 adjacent Floresta Massif during that time, and that they were the main source of Precambrian
619 zircons of the Middle Magdalena Valley Basin between ~40 Ma and ~30 Ma, during exhumation of
620 the Santander Massif. The strong correlation between zircon fission track date and distance from the
621 Santa-Marta-Bucaramanga Fault (Fig. 10), and the change in sedimentary provenance of the Middle
622 Magdalena Valley Basin (Fig. 2) strongly suggests that the Santa-Marta – Bucaramanga Fault was
623 active at ~40 Ma. This was coeval with the onset of deformation and exhumation in the Eastern

624 Cordillera (e.g. Horton et al., 2010; Moreno et al., 2011; Nie et al., 2012), and therefore reactivation
 625 of the Santander Massif and the Eastern Cordillera was probably synchronous. If we assume that the
 626 youngest zircon fission track dates record cooling through the zircon partial annealing zone (180°C
 627 - 260°C; Bernet, 2009) at ~40 Ma, then the total cooling recorded by samples close to the Santa
 628 Marta – Bucaramanga Fault since ~40 Ma may be >~250°C. Eocene magmatic rocks are not
 629 documented in the Santander Massif, and due to the spatial relationships between cooling patterns
 630 and the Santa Marta-Bucaramanga Fault, we assume that cooling was driven by variable
 631 exhumation across the fault. If we assume a geothermal gradient of 25°C/km, the surface near the
 632 Santa-Marta – Bucaramanga Fault may record >10km of exhumation since ~40 Ma. Although we
 633 calculate a minimum average rate of ~0.25 km/Ma, apatite fission track data show that exhumation
 634 rates were highly variable through time.

635 The highest cooling rates were determined from gneiss 10VDL49, which cooled through the APAZ
 636 at ~4.6 Ma at rates of >33°C/Ma (Fig.8a, Table 4). Most of the granitoids and gneisses from the
 637 western and central Santander Massif cooled rapidly through the APAZ between 12 and 3.5 Ma at
 638 rates of >9°C/Ma to >25°C/Ma (Fig. 8b), whereas granodiorite 10VDL05 cooled through the APAZ
 639 between 16.8 Ma and 13.6 Ma at minimum rates of 25°C/Ma. Granodiorites from the southern
 640 Santander Massif and to the west of the Santa Marta- Bucaramanga fault yield the lowest cooling
 641 rates, which range between 6 and 12°C/Ma (Table 4; Fig. 8c).

642

643 Apatite fission track data from the Santander Massif show that the current surface cooled through
 644 the apatite partial annealing zone at different times and rates (Fig. 8), and that the timing of cooling
 645 was controlled by displacement along the Santa-Marta Bucaramanga Fault (Fig. 10). Granodiorites
 646 from the west of the Santa-Marta Bucaramanga Fault cooled either earlier (10VDL22) or at lower
 647 rates (10VDL28, 10VDL32) than igneous and metamorphic rocks to the east. Exceptions include: i)
 648 gneiss 10VDL49 (Fig. 2) cooled rapidly through the apatite partial annealing zone at ~5 Ma (Fig.
 649 8), and ii) granodiorite 10VDL05 (Fig. 2) cooled rapidly at ~17 Ma (Fig. 8). Gneiss 10VDL49 is
 650 located close to the faulted boundary between the Santander Massif and the Mérida Andes, where

NE-SW oriented folds and faults developed in the last ~5 Ma as a result of convergence between the two massifs (Macellari, 1984). Granodiorite 10VDL05 suggest that it cooled earlier than other faulted blocks from the east of the Santa-Marta Bucaramanga Fault. This rock forms part of the same fault block as orthogneiss 10VDL37, which yields a younger apatite fission track date of 8.5 ± 1.5 Ma (Fig. 2), suggesting that this fault block cooled diachronously.

Due to a paucity of coeval magmatic rocks which could have disturbed the isotherms, we assume that our AFT data record cooling driven by exhumation through a geothermal gradient of $25^{\circ}\text{C}/\text{km}$. Our AFT data suggest that exhumation in the Santander Massif occurred diachronously between ~12 Ma and ~3.5 Ma, at minimum rates of 0.5-1 km/Ma (Fig. 8b). The fault block hosting granodiorite 10VDL05 and orthogneiss 10VDL37 (Fig. 2) may have exhumed diachronously during tilting, with its south-western boundary exhuming first. Granodiorites to the west of the Santa-Marta Bucaramanga Fault may have exhumed at rates of 0.25-0.5 km/Ma between 18 Ma and 6 Ma (Fig. 8c), and the higher exhumation rates recorded to the east of the fault suggest that exhumation along the SMBF may have been ongoing since ~40 Ma. The structural evolution of the Santander Massif has not been studied in detail, although the diachronous rapid exhumation of discrete fault blocks between 12 Ma and ~3.5 Ma can be explained by displacement along secondary faults that were active during sinistral displacement of the Santa Marta – Bucaramanga Fault.

7. Relationships with tectonic setting

7.1. 425 Ma to 405 Ma

Cooling between ~425 Ma and ~405 Ma recorded in granite 08VDL57 in the Caparo Block (Fig. 1b) was roughly coeval with a cessation of arc magmatism in the Mérida Andes and along most of the western margin of Gondwana (van der Lelij et al., 2015; Bahlburg et al., 2009). During this period, Gondwana was sufficiently close to Laurentia to permit faunal exchange between these two continents, although this predates the collision between Laurentia and Gondwana (e.g. Torsvik and Cocks, 2004). Devonian sedimentary rocks may be absent in the Mérida Andes, but Early to Middle

Devonian (~407- 385 Ma; Gradstein et al., 2012) marine sedimentary rocks are exposed in the Santander Massif (Fig. 3) and the Serranía del Perijá (Fig. 1a). This suggests that the mean elevation of these regions was low during this period. We propose that the magmatic hiatus spanning between ~415 Ma and ~290 Ma along the north-western margin of Gondwana may have resulted from a prolonged period of oblique subduction of oceanic crust of the Iapetus Ocean. Changing convergence vectors may have driven extension in the upper crust, localised exhumation of basement blocks and the formation of basins that hosted sedimentary units such as the Floresta Fm. (Fig. 3).

7.2. ~370 Ma to ~260 Ma

Two $^{40}\text{Ar}/^{39}\text{Ar}$ plateau dates and two weighted mean $^{40}\text{Ar}/^{39}\text{Ar}$ dates obtained from disturbed age spectra from the Mérida Andes span between ~370 Ma and ~320 Ma (Fig. 5). However, the interpretation of these dates as reflecting a distinct period of cooling is ambiguous because of the large spread in dates and the paucity of plateau dates. No evidence supports the presence an active margin during this time outboard of the basement of the Mérida Andes and the Santander Massif, and detrital zircons of ~400 Ma – ~300 Ma are scarce in autochthonous rocks of Colombia (e.g. Horton et al., 2010). This period was previously considered to pre-date the amalgamation of Pangaea along the Ouachita-Marathon suture (Fig. 11) between ~325 Ma and ~270 Ma (e.g. Ross, 1986; Thomas, 2004). The Ouachita Marathon Orogeny was driven by the collision of Laurentia with Gondwana, gave rise to the Ouachita – Marathon Suture (Fig. 11) and is mainly preserved in North America within a foreland basin, and by deformation that is dated between ~325 Ma and ~270 Ma (e.g. Ross, 1986; Thomas, 2004). Deformation was coeval with the deposition of continental sandstones of the ~3350m thick Sabaneta Fm. (Fig. 3; Odremán and Useche, 1997) of the Mérida Andes. The Sabaneta Fm. grades upwards into marine siltstones and carbonates of the ~650m thick Palmarito Fm. (Fig. 3), revealing a transgressive sequence (Laya and Tucker, 2012) until the cessation of sedimentation at ~260 Ma. Laya and Tucker (2012) interpreted the Sabaneta and Palmarito Fms. to have formed in a flexural foreland basin resulting from the Ouachita-

Marathon Orogeny, and assumed that crustal loading derived from a deformation front located between the Venezuelan basement and Yucatán (Maya block; Fig. 11).

7.3. ~257 to ~200 Ma

Sedimentation over the basement rocks of the Mérida Andes was immediately followed between ~257 Ma and ~200 Ma by rapid cooling of ~250°C-300°C (Fig. 9) at minimum rates of >4.4°C/Ma, in areas which are unlikely to have been affected by thermal resetting resulting from the intrusion of magmatic bodies. If we assume that cooling was driven by exhumation through a geothermal gradient of ~25°C/km, exhumation of 10-12 km may have occurred at minimum rates of 0.18 km/Ma. Low grade metamorphism of chlorite schist 08VDL16 (Mucuchachi Fm.) between ~450 Ma and ~254 Ma may have been a result of burial beneath Palaeozoic sedimentary units with a cumulative thickness of >9000m (Fig. 9). The basement rocks of the Mérida Andes may have been separated from the collision zone by the Maya Block and other blocks (Fig. 11), and perhaps they were too far from the Ouachita-Marathon suture (Fig. 11) to be deformed during the amalgamation of Pangaea. Short-lived, high grade metamorphism in the Maya Block resulted in anatexis at ~253 Ma (Weber et al., 2007), which postdates the Ouachita-Marathon Orogeny. This time period corresponds with shortening and crustal thickening in the Maya Block, and is interpreted to have occurred during slab-advance beneath western Pangaea (Weber et al., 2007). The same mechanism may also account for exhumation in the Mérida Andes at ~257 Ma because i) the crustal fragments which host the Mérida Andes may have been juxtaposed against the Maya Block, ii) Permian and Jurassic sequences are well preserved (Fig. 1b), whereas Triassic sedimentary sequences are not documented in the Mérida Andes and the Santander Massif, suggesting that that these regions were topographically high, perhaps due to compression that occurred at the end of the Permian, iii) it is difficult to tie a relationship between the Ouachita – Marathon Orogeny, which had only local low grade effects near the suture zone (e.g. Ross, 1986; Thomas, 2004) and later rapid exhumation in

732 the Mérida Andes, and iv) the arc axis was west of the Mérida Andes at ~255 Ma (Fig. 11). The
733 interpretation of the Sabanata and Palmarito Fms. as foreland basin sequences (Laya and Tucker,
734 2012) implies the proximity of an orogenic belt at ~290 Ma- 260 Ma, which has not been identified
735 in the Maya Block, Colombia or Venezuela.

736

737 Extension along the north-western margin of Gondwana between 240 Ma and 215 Ma is
738 documented by bimodal magmatism, which culminated in increasingly isotopically juvenile mafic
739 dykes and sills, and ultimately an ophiolite sequence, which are exposed in the Eastern Cordillera of
740 Ecuador and the Central Cordillera of Colombia (Fig. 1a; Cochrane et al., 2014b; Spikings et al.,
741 2015). These rocks formed during the separation of Mexican and Central American terranes from
742 the western margin of Gondwana (Fig. 11; Litherland et al., 1994). No magmatism is documented at
743 this time in the Mérida Andes and the Santander Massif, which is consistent with distal slab retreat
744 during extension. However, arc magmatism is documented in the Mérida Andes and the Santander
745 Massif between ~213 Ma and ~196 Ma (van der Lelij et al., 2015), and we hypothesize that the
746 separation of continental blocks was transtentional, thus allowing arc magmatism to resume in these
747 regions shortly after continental separation (Fig. 11). Active extension may not be documented in
748 the Mérida Andes and the Santander Massif after ~257 Ma, until the deposition of the graben-filling
749 La Quinta and Jordan Fms. (Fig. 3) at ~200 Ma (Maze, 1984; Kammer and Sánchez, 2006).
750 Extension at ~200 Ma may have resulted from the combined effects of early continental separation
751 of North and South America, and back-arc extension (e.g. Maze, 1984; Bartok, 1993).

752

753 Rapid alternations between extension and compression over a ~10 Ma period are documented in
754 arc-related magmatism in the Tasmanides of Australia (e.g. Kemp et al., 2009 and references
755 therein), where shorter contractional episodes do not appear to have a major influence on isotopic
756 trends. Therefore, we consider it possible to reconcile high grade deformation in the Maya Block at
757 ~253 Ma (Weber et al., 2007) and exhumation in the Mérida Andes during compression, with

758 extension documented at ~240 Ma elsewhere along the (presumably) same margin (e.g. Litherland
759 et al., 1994; Cochrane et al., 2014b).

760

761 The post-200 Ma thermal histories in the Mérida Andes are poorly constrained by weighted mean
762 $^{40}\text{Ar}/^{39}\text{Ar}$ dates from disturbed age spectra (Fig. 5). Biotite $^{40}\text{Ar}/^{39}\text{Ar}$ plateau dates of >200 Ma
763 suggest that temperatures were below the bulk closure temperature of $320\pm 30^\circ\text{C}$ after ~200 Ma, in
764 spite of burial beneath Mesozoic sedimentary rocks with a cumulative thickness of ~6500m (Fig.
765 10). Assuming a geothermal gradient of $\sim 25^\circ\text{C}/\text{Ma}$ and a homogeneous distribution of Mesozoic
766 cover, this suggests that all samples were at temperatures of $>180^\circ\text{C}$ at ~65 Ma (Fig. 9). This is
767 supported by the scattered zircon fission track dates between ~61 Ma and ~172 Ma reported by
768 Kohn et al. (1984), which they interpret to reflect partial annealing (180°C - 260°C ; Bernet, 2009).
769 Consequently, the few weighted mean $^{40}\text{Ar}/^{39}\text{Ar}$ dates younger than ~200 Ma are affected by
770 mineral alteration, and cannot be interpreted as reliable points in t-T space.

771

772 **7.4. ~40 Ma (Santander Massif)**

773 Zircon fission track dates from the Santander Massif vary with distance from the Santa-Marta
774 Bucaramanga Fault (Fig. 10). A thermal pulse associated with intrusive activity at ~200 Ma resulted
775 in total annealing of zircon fission tracks in Early Palaeozoic samples and the resetting of $^{40}\text{Ar}/^{39}\text{Ar}$
776 dates. Rapid cooling through the zircon partial annealing zone at ~200 Ma may be recorded in
777 samples which were collected more than 20 km from the Santa-Marta Bucaramanga Fault. In
778 contrast, samples close to the fault may have cooled during asymmetric exhumation of the
779 Santander Massif, which was highest near the fault. This suggests that the Santa-Marta
780 Bucaramanga Fault was active at ~40 Ma, which is consistent with the proposed Late Eocene onset
781 of exhumation obtained from the analyses of syn-orogenic basins surrounding the Eastern
782 Cordillera and the Santander Massif (Nie et al., 2012; Horton et al., 2010).

783

784 Rapid exhumation at ~40 Ma is well documented along the western margin of South America over
785 an along-strike distance of >2000 km, from the Central Cordillera of Colombia (e.g. Restrepo-
786 Moreno et al., 2009; Villagómez and Spikings, 2013) to Peru (e.g. Barnes et al., 2006) via Ecuador
787 (Spikings et al., 2000, 2001, 2005), and this widespread compressive phase can be related to an
788 episode of accelerated convergence of ~15 cm/a (e.g. Pardo-Casas and Molnar, 1987) between the
789 South American margin and the Farallón Plate.

790

791 **7.5. ~18 Ma to present**

792 Apatite fission track data from the Santander Massif show rapid exhumation to the east of the
793 SMBF since at least $\sim 17.6 \pm 3.3$ Ma (10VDL05; Figs. 8 and 10). Exhumation may have been
794 controlled by strain partitioning along the numerous secondary faults (Fig. 2), and some fault blocks
795 remained approximately isothermal subsequent to rapid exhumation, while others were exhuming
796 rapidly (Fig. 8). The spatial trends of AFT dates and the time-temperature paths of granodiorites
797 from the west of the Santa-Marta Bucaramanga Fault (Fig. 10) exhumed at lower rates of 0.25-0.5
798 km/Ma than rocks to the east, which exhumed at rates of 0.5-1.0 km/Ma. The onset of exhumation
799 east of the fault is not constrained by our AFT data, but may well have been continuous since ~40
800 Ma. The evidence that many samples were in or below the zircon partial annealing zone at ~40 Ma
801 suggests that average exhumation rates between 40 Ma and 18 Ma did not exceed ~0.15 km/Ma,
802 assuming a geothermal gradient of 25°C/km. Present data do not constrain any inflection points in
803 the thermal histories before ~18 Ma, and the timing of the onset of rapid cooling is unknown. There
804 is also limited evidence for ongoing exhumation between ~27 Ma and ~15 Ma in the Caparo Block
805 of the Mérida Andes (Bermúdez et al., 2010), although the onset of exhumation may be
806 unconstrained in the Caparo Block. Accelerated exhumation in the Santander Massif may have been
807 driven by indentation of the Panama-Chocó arc to the western Colombian margin at ~24 Ma (e.g.
808 Farris et al., 2011) and accelerated convergence between the Nazca and South American Plates
809 between ~20 Ma and ~10 Ma (e.g. Pardo-Casas and Molnar, 1987).

810

811 **8. Conclusions**

812 A comprehensive, multi-method thermochronological study of the basement rocks of the Mérida
813 Andes and the Santander Massif shows that distinct periods of cooling driven by exhumation can be
814 resolved, and solve key issues towards understanding the development of the Northern Andes:

815 1) The collisions between north-western Gondwana and southern Laurussia to form Pangaea
816 resulted in the low intensity Ouachita-Marathon orogeny between ~325 and ~270 Ma, which is
817 recorded near the collision zone by deformed foreland basin deposits. The basement of the Northern
818 Andes appears to not have been affected by this orogeny. The Northern Andes were separated from
819 the collision zone by the Maya Block (Fig. 11), where deformation during Pangaea amalgamation is
820 also not identified.

821 2) Exhumation by > 10 km in the Mérida Andes occurred between ~258 Ma and ~200 Ma, and is
822 recorded by i) apatite U-Pb thermal history solutions which suggest that rapid cooling started
823 between ~258 Ma and ~250 Ma, iii) several biotite and muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ plateau dates between
824 ~257 Ma and ~200 Ma, and iii) a sedimentary hiatus of ~60 Ma spanning this period. This is coeval
825 with shortening and high-grade metamorphism at ~253 Ma in the Maya Block (Weber et al., 2007),
826 which may have been juxtaposed against the Mérida Andes prior to Pangaea breakup. High grade
827 metamorphism in the Maya Block and coeval exhumation in the Mérida Andes are consistent with a
828 compressive regime resulting from flat slab subduction (Weber et al., 2007). High temperature
829 conditions prior to ~257 Ma may have been facilitated by burial beneath Palaeozoic sequences,
830 whose total thickness exceeds 9000m. However, the absence of Permian anatectites in the Mérida
831 Andes or the Santander Massif suggests that temperatures did not exceed ~600°C in the Late
832 Palaeozoic.

833 3) Extension documented between ~240 Ma and ~215 Ma in the Central Cordillera of Colombia
834 (Cochrane et al., 2014b) may not have affected basement rocks of the Mérida Andes and the
835 Santander Massif until after ~200 Ma, when an extensional regime resulted in the formation of
836 grabens hosting the La Quinta, Jordan and Girón Fms. (Fig. 3).

837 4) The medium to high temperature thermochronological data from the Santander Massif are too
838 sparse to clearly identify pre-Jurassic periods of cooling. The pre-Mesozoic, <550°C thermal
839 histories of Palaeozoic igneous and metamorphic rocks in the central Santander Massif were
840 overprinted by reheating during intrusion of numerous granitoids at ~205-196 Ma. The easternmost
841 and northernmost areas of the Santander Massif were located further from the main locus of
842 magmatism (Fig. 2), and may be more amenable to future high-temperature thermochronology
843 studies.

844 5) The NW-SE trending Santa Marta – Bucaramanga Fault became an active structural feature at
845 ~40 Ma, and resulted in asymmetric exhumation of the Santander Massif. Total exhumation since
846 ~40 Ma was highest near the fault, and may have exceeded ~10 km. The onset of exhumation of the
847 Santander Massif and the Eastern Cordillera (Horton et al., 2010) was coeval, and may have been
848 driven by accelerated convergence between the Farallón / Nazca Plate and the South American
849 margin (Pardo-Casas and Molnar, 1987). Coeval deformation is not documented in the Mérida
850 Andes.

851 6) Exhumation patterns in the Santander Massif are strongly affected by secondary faults, and small
852 structural blocks exhumed diachronously since at least ~18 Ma. It is unclear whether the indentation
853 of the Panama Arc to the Western margin of Colombia at ~25 Ma (e.g. Farris et al., 2011) had
854 significant effects on exhumation in the Santander Massif, because exhumation started at ~40 Ma
855 and accelerated exhumation rates can only be determined since ~18 Ma.

856

857 **Acknowledgements**

858

859 The authors are grateful to Prof. Tabata Hoeger, Daniela Tazzo and Maria Carolina Maninat from
860 ULA Mérida for field assistance in Venezuela, and to Prof. Mauricio Bermúdez from UCV Caracas
861 for logistical support. Field work and logistics in Colombia were greatly facilitated by the generous
862 assistance of Ecopetrol. Andrew Kerr is thanked for editorial handling, and the authors are grateful
863 to two anonymous reviewers for providing constructive criticism which helped us to improve the

864 manuscript. This project was funded by the Swiss National Science Foundation fund
865 200021_129497, which was awarded to RS.

866

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1176 **Figure Captions**

1177

- 1178 Figure 1A: Digital relief map (Amante and Eakins, 2008) showing the tectonic provinces of
1179 northern South America modified from Feo-Codecido et al. (1984) and Ostos et al. (2005). CC:
1180 Central Cordillera, CM: EC: Eastern Cordillera, FM : Floresta Massif; LB : Llanos Basin, MA:

1181 Mérida Andes, MTB : Maracaibo Triangular Block, MV : Magdalena Valley, PP: Paraguana
 1182 Peninsular, SM: Santander Massif, SNSM: Sierra Nevada de Santa Marta, WC: Western Cordillera.
 1183 B: Geological map of the Mérida Andes, Venezuela, simplified and modified from Hackley et al.
 1184 (2005). Locations of all analysed samples are shown. Sample codes in parentheses (XX) correspond
 1185 to full sample codes 08VDLXX. Zircon U-Pb ages in bold text are weighted mean zircon ^{238}U - ^{206}Pb
 1186 dates with uncertainties at the 2σ confidence level (Van der Lelij et al., 2015). $^{40}\text{Ar}/^{39}\text{Ar}$ dates
 1187 obtained during this study are colour coded according to the corresponding mineral phase. Apatite
 1188 U-Pb data was collected from sample 08VDL44 in the north-east of the map and are shown in
 1189 Figure 4.

1190

1191 Figure 2: Geological map of the Santander Massif, Colombia, simplified and modified from Gómez
 1192 et al. (2007). The localities of the stratigraphic columns of the Middle Magdalena Valley Basin and
 1193 the Eastern Piedmont (Fig. 3) are shown. Sample codes in parentheses (XX) correspond to full
 1194 sample codes. Weighted mean ^{238}U - ^{206}Pb dates of concordant zircons (Van der Lelij et al., 2015)
 1195 with uncertainties at the 2σ confidence level (Van der Lelij et al., 2015) are marked in bold text.
 1196 Regular text indicates a sample's zircon fission track date with 2σ uncertainty, while italic text
 1197 indicates a sample's apatite fission track date with 2σ uncertainty, followed by mean C-Axis
 1198 projected track length and standard deviation of the fission track lengths. Colour coded text without
 1199 sample numbers indicate previously reported $^{40}\text{Ar}/^{39}\text{Ar}$ plateau dates and the phase from which they
 1200 were extracted (Restrepo-Pace, 1995). Weighted mean $^{40}\text{Ar}/^{39}\text{Ar}$ dates from published disturbed
 1201 spectra and previous $^{40}\text{Ar}/^{39}\text{Ar}$ data without reported sampling localities (Cordani et al., 2005) are
 1202 not indicated.

1203

1204 Figure 3: Simplified pre-Pliocene stratigraphic columns of the Middle Magdalena Valley basin, the
 1205 Santander Massif, the Eastern Piedmont, and the Mérida Andes, compiled from various sources
 1206 (Ward et al., 1974; Gonzales de Juana et al., 1980; Shagam et al., 1984; Cooper et al., 1995;
 1207 Odremán and Useche, 1997; Gómez et al., 2005; Nie et al., 2012; Van der Lelij et al., 2015).

1208 Periods of magmatism (Van der Lelij et al., 2015) are marked in pink. The Pre-Jurassic and pre-
1209 Carboniferous stratigraphies of the Eastern Piedmont and the Middle Magdalena Valley basin,
1210 respectively, are poorly known and not indicated, but presumably correlative with the basement of
1211 the Mérida Andes and the Santander Massif. The locations of the Middle Magdalena Valley Basin
1212 and the Eastern Piedmont are indicated in figure 2.

1213

1214 Figure 4: Apatite U-Pb data from granodiorite 08VDL44. Analyses denoted in red are considered to
1215 be compositional outliers and may have different diffusion kinetics, and were consequently not
1216 included in thermal history models. All uncertainties are quoted at the 2σ level. A: Concordia plots
1217 of U-Pb data from all measured apatites; b: $^{238}\text{U}/^{206}\text{Pb}$ dates plotted against grain size; c: $^{238}\text{U}/^{206}\text{Pb}$
1218 dates plotted against Th/U ratios, which we use as a proxy for grain composition. See text for
1219 details.

1220

1221 Figure 5: Summary of new $^{40}\text{Ar}/^{39}\text{Ar}$ dates obtained from the Mérida Andes and the Santander
1222 Massif. Dates are regrouped by sample, and crystallisation ages of corresponding samples are
1223 indicated with red boxes. The thickness of each box corresponds to its 2σ uncertainty. A P denotes
1224 statistically robust plateau dates, and numbers indicate the Mean Squared Weighted Deviates
1225 (MSWD) if the corresponding $^{40}\text{Ar}/^{39}\text{Ar}$ date is a statistically less robust weighted mean date
1226 (MSWD>2.5). A % symbol indicates that consecutive steps in the age spectrum are statistically
1227 equivalent but that they do not represent more than 50% of the total measured ^{39}Ar . $^{40}\text{Ar}/^{39}\text{Ar}$ dates
1228 with mean square weighted deviates > 50 are not shown. The location of the Caparo Block is shown
1229 in Figure 1B. Indicated closure temperatures were calculated using a range of cooling rates of
1230 $100^\circ\text{C}/\text{Ma} - 1^\circ\text{C}/\text{Ma}$ and a diffusion length of 100 μm . See text for details

1231

1232 Figure 6: Representative $^{40}\text{Ar}/^{39}\text{Ar}$ plateau spectra from the Mérida Andes (samples 08VDLXX)
1233 and the Santander Massif (samples 10VDLXX). Corresponding sample localities are indicated in

Fig. 1b. Isochron data not shown when isochrons did not converge due to insufficient spread in ^{36}Ar contents between degassing steps. See appendix for $^{40}\text{Ar}/^{39}\text{Ar}$ spectra from all samples.

Fig. 7: Results of the apatite U-Pb thermal history modelling using the seven grains with similar Th/U ratios from granodiorite 08VDL44. 7a: comparison between modelled and measured apatite U-Pb dates for the best fit solution. 7b: Thermal history model. The good fit envelope is indicated in dark blue and the acceptable fit envelope is indicated in light blue, while the thick black line indicates the best fit solution. A red line centered on the muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ plateau date of 234.8 ± 1.8 Ma obtained from the same granodiorite is plotted, and its temperature range corresponds to a range of closure temperatures of 440 ± 50 Ma determined for cooling rates of $100^\circ\text{C}/\text{Ma}$ to $1^\circ\text{C}/\text{Ma}$ and a diffusion length of $100\ \mu\text{m}$. See text for details.

Fig. 8: Summary of best fit thermal history solutions for apatite fission track data from the Santander Massif, sorted by locality. APAZ: apatite partial annealing zone, inside of which the apatite fission track data provide reliable constraints (60°C - 110°C ; Laslett et al., 1987). Numbers “XX” adjacent to identically coloured best fit lines indicate abbreviated sample codes, where the full sample codes are 10VDLXX. An example of full thermal history envelope with typical uncertainty is indicated in the top graph for granodiorite 10VDL55.

Fig. 9: Schematic thermal history envelope for samples from the Mérida Andes. Dotted lines are speculative thermal history envelope limits. The minimum temperatures of the thermal history envelope are based on the assumption that sedimentary sequences were deposited homogeneously over the basement sequences, resulting in homogeneous burial by sedimentary thicknesses from Odreman and Useche (1997), using a geothermal gradient of $25^\circ\text{C}/\text{km}$. Periods of cooling are defined by i) the apatite U-Pb thermal history good fit envelope; ii) $^{40}\text{Ar}/^{39}\text{Ar}$ plateau dates from muscovite and biotite, iii) previously published AFT data (Bermúdez et al., 2010; Bermúdez et al., 2011; Erikson et al., 2012). Unconstrained are i) thermal histories of samples yielding only

1261 discordant $^{40}\text{Ar}/^{39}\text{Ar}$ data; ii) thermal histories of the Caparo Block after ~410 Ma; and iii) thermal
1262 histories of samples which may have cooled at ~350 Ma (Fig. 5) because of a paucity of statistically
1263 robust analyses.

1264

1265 Fig. 10: Fission track dates relative to the linear distance from the Santa Marta- Bucaramanga Fault
1266 (SMBF) in the Santander Massif. Plotted uncertainties are 2σ level, and indicated numbers XX are
1267 partial sample codes corresponding to full codes 10VDLXX. a) Zircon fission track dates compared
1268 to distance from the SMBF. Red dots denote the zircon U-Pb ages obtained from these samples if
1269 they plot on the same scale. Uncertainties on U-Pb ages are too small to be plotted. b) Apatite
1270 fission track dates compared to linear distance from the SMBF.

1271

1272 Fig. 11: Tectonic model illustrating the positions of relative tectonic blocks discussed in the text,
1273 modified and simplified from Weber et al. (2007) to focus on the Mérida Andes and the Santander
1274 Massif. MA: Mérida Andes; SM: Santander Massif; ME-CA: Mexican and Central American
1275 Terranes except for the Maya Block. i) 255 Ma: Configuration immediately prior to orogenesis
1276 recorded in the Maya Block at ~253 Ma (Weber et al., 2007) and coeval exhumation recorded in the
1277 Mérida Andes. The location of the Ouachita – Marathon Suture (~290 Ma; Ross, 1986) and
1278 recorded Permian arc magmatism is shown. ii) 210-195 Ma: Configuration after the rifting recorded
1279 in basement rocks of the Central Cordillera of Colombia at ~216 Ma (Cochrane et al., 2014b). The
1280 localities of arc rocks with crystallisation ages of ~200 Ma are indicated. After removal of the
1281 conjugate margin at 216 Ma (Cochrane et al., 2014b), the magmatic arc must have migrated inland
1282 to account for voluminous arc rocks in the Mérida Andes and the Santander Massif between ~213
1283 Ma and ~196 Ma (Van der Lelij et al., 2015).

1284

1285 Table 2: Apatite U-Pb data from granodiorite 08VDL44. See text for details.

1286

1287 Table 1: Summary of $^{40}\text{Ar}/^{39}\text{Ar}$ data obtained from the Mérida Andes during this study. Zircon U-
1288 Pb dates are from Van der Lelij et al., 2015. MSWD= Mean Squared Weighted Deviates. n=
1289 number of steps in the plateau or weighted mean date calculation.
1290
1291 Table 3: Zircon fission track data from basement rocks of the Santander Massif. See text for details.
1292
1293 Table 4: Apatite fission track data and HeFTy goodness of fit parameters from basement rocks of
1294 the Santander Massif. MTL: Mean Track Length; GOF: Goodness of fit. Minimum and maximum
1295 cooling rates were estimated by determining minimum and maximum slopes through good fit
1296 HeFTy envelopes, and could not be determined for granodiorite 10VDL22 because it does not yield
1297 good fits. Maximum cooling rates of over $>100^\circ\text{C}/\text{Ma}$ are not reliable, but cannot be determined
1298 more accurately with our data. The oldest track records the entry of a sample into the Partial
1299 Annealing Zone during cooling, according to the best fit model.

Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 7

Figure 8

Figure 9

Figure 10

Figure 11

Table 1. Apatite U-Pb data from granodiorite 08VDL44, Mérida Andes. Zircon U-Pb age: 466.8±2.5 Ma

Sample data			Isotopic Ratios							Dates (Ma)							Apatite Composition					
Fraction	Radius	Length	²⁰⁶ Pb/ ²⁰⁴ Pb ^a	²⁰⁶ Pb/ ²³⁸ U ^b	±2σ	²⁰⁷ Pb/ ²³⁸ U ^b	±2σ	²⁰⁷ Pb/ ²⁰⁶ Pb ^b	±2σ	²⁰⁶ Pb/ ²³⁸ U	±2σ	²⁰⁷ Pb/ ²³⁵ U	±2σ	²⁰⁷ Pb/ ²⁰⁶ Pb	±2σ	Corr. coef.	Th/U	U (ppm)	Th (ppm) ^c	Pb* (pg) ^d	Pbc (pg) ^e	Pb*/Pbc ^f
08VDL44	(μm)	(μm)			(%)		(%)		(%)		(abs)		(abs)		(abs)							
1	250	722	31.88565	0.04211	2.73	0.3074	21.97	0.052954	19.86	265.9	7.1	272.2	52.5	326.8	450.7	0.80	0.07	6.49	93.11	182.28	914.99	0.20
2	206	265	62.21391	0.05472	0.84	0.4139	6.53	0.054862	5.88	343.4	2.8	351.7	19.4	406.6	131.5	0.80	4.08	7.28	1.79	186.75	132.12	1.41
3	100	284	35.42516	0.04324	2.14	0.3047	17.88	0.051099	16.22	272.9	5.7	270.0	42.4	245.3	373.6	0.80	2.31	4.96	2.14	34.45	82.36	0.42
4	41	125	21.60124	0.04104	9.18	0.2934	81.09	0.051856	74.37	259.3	23.3	261.2	186.8	279.1	1702.5	0.76	0.24	2.98	12.50	0.24	4.75	0.05
5	42	241	25.32428	0.04075	4.76	0.3077	37.80	0.054758	34.17	257.5	12.0	272.4	90.3	402.4	765.4	0.79	0.20	18.72	91.51	0.98	9.20	0.11
6	34	234	25.65412	0.04020	4.44	0.3058	35.61	0.055162	32.25	254.1	11.1	270.9	84.7	418.8	720.2	0.78	0.23	8.43	36.96	0.85	7.58	0.11
7	65	313	29.61318	0.03985	3.11	0.2779	26.37	0.050578	23.96	251.9	7.7	249.0	58.2	221.6	554.3	0.80	0.04	10.66	275.24	3.05	18.61	0.16
8	82	355	29.95557	0.04001	3.06	0.2906	24.89	0.052670	22.53	252.9	7.6	259.0	56.9	314.6	512.4	0.80	0.24	6.11	25.56	4.43	24.69	0.18
9	94	365	35.76316	0.04009	2.06	0.2867	17.01	0.051873	15.42	253.4	5.1	256.0	38.5	279.8	352.9	0.80	0.35	15.44	44.64	9.15	33.02	0.28

^a Measured ratio corrected for fractionation and spike contribution only.

^b Measured ratios corrected for fractionation, tracer, blank and initial common Pb.

^c Th contents calculated from radiogenic 208Pb and the 207Pb/206Pb date of the sample, assuming concordance between U-Th and Pb systems.

^d Total mass of radiogenic Pb.

^e Total mass of common Pb.

^f Ratio of radiogenic Pb (including 208Pb) to common Pb.

Table 2

Sample	Unit name	Latitude N	Longitude W	Altitude (m)	Zircon U-Pb age $\pm 2\sigma$ (Ma) ^a	Phase	⁴⁰ Ar/ ³⁹ Ar date $\pm 2\sigma$ (Ma)	MSWD ^b	n ^c	% ³⁹ Ar	Date type ^d
Mérida Andes											
08VDL03	Verdalito granodiorite	8°28'49"	71°34'13"	363	449.3 $\pm$ 2.5	Bt	251.2 $\pm$ 1.1	0.7	11	91.2	Plateau
08VDL04	Pueblo Hondo granite	8°17'53"	71°53'58"	2010	434.2 $\pm$ 4.4	Bt	257.1 $\pm$ 1.0	2.1	10	86.1	Plateau
08VDL04	Pueblo Hondo granite	8°17'53"	71°53'58"	2010	434.2 $\pm$ 4.4	Fsp	223.0 $\pm$ 0.9	0.7	2	23.8	Disturbed
08VDL05	Sierra Nevada paragneiss	8°29'59"	71°12'04"	2247	>505 Ma	Ms	170.1 $\pm$ 1.0	1.0	4	42.8	Disturbed
08VDL05	Sierra Nevada paragneiss	8°29'59"	71°12'04"	2247	>505 Ma	Bt	168.7 $\pm$ 1.8	44.0	7	74.1	Disturbed
08VDL14	El Carmen granodiorite	8°45'07"	70°57'33"	3195	211.2 $\pm$ 1.2	Ms	200.6 $\pm$ 0.8	1.5	8	95.2	Plateau
08VDL16	Mucuchachi schist	8°11'25"	71°49'06"	2471	No data	Bt	254.7 $\pm$ 1.0	1.2	12	98.5	Plateau
08VDL18	La Quinta rhyodacite	8°08'56"	72°00'55"	1138	202.0 $\pm$ 1.6	Bt	200.4 $\pm$ 0.9	1.7	6	75.7	Plateau
08VDL21	La Grita diorite	8°08'28"	71°56'15"	2240	471.8 $\pm$ 4.0	Bt	182.3 $\pm$ 1.1	7.3	5	54.8	Disturbed
08VDL23	Mucubaji granite	8°48'36"	70°48'47"	3465	237.1 $\pm$ 1.4	Ms	135.6 $\pm$ 1.5	48.6	7	83.2	Disturbed
08VDL23	Mucubaji granite	8°48'36"	70°48'47"	3465	237.1 $\pm$ 1.4	Bt	116.4 $\pm$ 1.1	24.0	5	72.7	Disturbed
08VDL27	La Raya granitic gneiss	8°51'34"	70°34'51"	1427	488.5 $\pm$ 6.5	Ms	353.4 $\pm$ 1.2	2.3	8	85.0	Plateau
08VDL32	La Soledad granite	8°49'48"	70°31'47"	961	414.5 $\pm$ 3.9	Ms	369.6 $\pm$ 1.7	5.8	6	61.7	Disturbed
08VDL32	La Soledad granite	8°49'48"	70°31'47"	961	414.5 $\pm$ 3.9	Bt	323.8 $\pm$ 1.0	0.7	12	88.9	Plateau
08VDL41	Chachopo granite	8°55'55"	70°47'31"	3119	454.5 $\pm$ 3.1	Ms	379.6 $\pm$ 1.5	0.3	3	30.9	Disturbed
08VDL41	Chachopo granite	8°55'55"	70°47'31"	3119	454.5 $\pm$ 3.1	Bt	212.8 $\pm$ 0.9	1.7	3	52.8	Plateau
08VDL44	Miraflores granodiorite	9°10'42"	70°38'08"	867	466.8 $\pm$ 2.5	Ms	234.8 $\pm$ 1.3	2.2	7	98.3	Plateau
08VDL45	Valera granodiorite	9°11'11"	70°37'51"	838	447.9 $\pm$ 1.8	Ms	250.8 $\pm$ 0.8	0.6	6	68.3	Plateau
08VDL46	La Quebrada granite	9°10'25"	70°32'24"	2358	243.5 $\pm$ 3.4	Ms	231.2 $\pm$ 0.8	1.4	8	97.7	Plateau
08VDL47	Carmania monzodiorite	9°15'59"	70°38'18"	789	450.4 $\pm$ 2.5	Bt	245.1 $\pm$ 0.9	1.2	6	79.3	Plateau
08VDL48	Carmania monzodiorite	9°13'27"	70°39'45"	1101	No data	Ms	245.3 $\pm$ 0.8	1.2	9	98.6	Plateau
08VDL48	Carmania monzodiorite	9°13'27"	70°39'45"	1101	No data	Bt	237.3 $\pm$ 0.8	0.2	5	57.6	Plateau
08VDL49	La Puerta granodiorite	9°07'11"	70°42'39"	1920	453.1 $\pm$ 1.8	Ms	225.3 $\pm$ 0.9	3.8	9	90.8	Disturbed
08VDL49	La Puerta granodiorite	9°07'11"	70°42'39"	1920	453.1 $\pm$ 1.8	Bt	205.1 $\pm$ 0.8	2.3	7	83.7	Plateau
08VDL50	Timotes granite	9°00'52"	70°44'25"	1931	445.2 $\pm$ 3.9	Ms	213.7 $\pm$ 1.1	5.9	5	52.7	Disturbed
08VDL50	Timotes granite	9°00'52"	70°44'25"	1931	445.2 $\pm$ 3.9	Bt	203.5 $\pm$ 0.9	3.2	6	65.0	Disturbed
Caparo Block											
08VDL35	Cerro Azul granite	8°53'14"	70°23'34"	848	499.4 $\pm$ 2.7	Hbl	451.4 $\pm$ 5.0	35.3	6	88.1	Disturbed
08VDL55	Islitas tonalite	7°49'01"	71°21'01"	340	473.1 $\pm$ 3.5	Hbl	464.3 $\pm$ 2.4	1.9	6	50.6	Plateau
08VDL57	El Tapo granite	7°51'00"	71°13'41"	259	492.3 $\pm$ 5.3	Bt	404.7 $\pm$ 2.1	3.8	8	90.3	Disturbed
08VDL57	El Tapo granite	7°51'00"	71°13'41"	259	492.3 $\pm$ 5.3	Ms	425.3 $\pm$ 1.9	1.0	7	79.3	Plateau
Santander Massif											
10VDL20	Silgara Formation anatectite	6°48'53"	72°00'28"	601	No data	Ms	210.4 $\pm$ 1.0	1.9	10	97.8	Plateau
10VDL23	Bucaramanga gneiss	6°56'35"	72°58'00"	2055	461.0 $\pm$ 2.1	Hbl	204.1 $\pm$ 3.5	14.6	5	66.9	Disturbed
10VDL32	Mogotes granodiorite	6°25'22"	72°49'29"	1782	198.0 $\pm$ 0.8	Bt	196.6 $\pm$ 1.1	4.1	3	70.4	Disturbed
10VDL36	Silgara garnet metatuff	7°13'19"	73°03'25"	1029	No data	Ms	199.3 $\pm$ 0.7	0.4	9	99.3	Plateau
10VDL37	Berlin orthogneiss	7°11'24"	72°58'41"	1712	451.5 $\pm$ 1.3	Hbl	205.7 $\pm$ 0.9	1.7	4	64.6	Plateau
10VDL45	Bucaramanga gneiss migmatite	7°17'51"	72°35'40"	2269	No data	Ms	450.2 $\pm$ 1.6	2.4	6	77.1	Plateau
10VDL48	Durania anatectites	7°29'20"	72°42'16"	2364	<479.8 Ma	Ms	407.4 $\pm$ 1.4	1.1	4	44.0	Disturbed
10VDL51	Berlin orthogneiss	7°15'17"	72°53'48"	3707	472.5 $\pm$ 3.4	Bt	198.8 $\pm$ 0.7	2.1	7	77.4	Plateau
10VDL55	Tarra granodiorite	8°03'10"	73°04'31"	1506	443.4 $\pm$ 3.2	Bt	428.7 $\pm$ 1.6	0.4	2	26.9	Disturbed

^a Zircon U-Pb ages from the same samples determined in van der Lelij et al. (2015).

^b Mineral phase which yielded the corresponding date. Bt: biotite; Fsp: alkali feldspar; Hbl: Hornblende; Ms: Muscovite

^c Mean Squared Weighted Deviates calculated for n steps

^d Plateau dates are weighted mean dates calculated from >3 consecutive steps with MSWD < 2.5 and ³⁹Ar > 50%. Dates calculated from disturbed spectra are weighted mean dates of n consecutive steps yielding MSWD > 2.5 or < 50% ³⁹Ar

Table 3

Sample data						Zircon fission track data						
Sample	Name	Latitude N	Longitude W	Elevation (m)	Zircon U-Pb Age $\pm 2\sigma$ (Ma) ^d	No. of Grains	Ns ^a	Ω_T (cm ²) ^a	$\Sigma(P_i \cdot \Omega_i) \pm 2\sigma$ (10 ⁻⁷ cm ²) ^b	U (ppm)	P(X ²) (%) ^c	Pooled date $\pm 2\sigma$ (Ma)
10VDL05	Corcova granodiorite	7°06'03"	73°00'36"	2793	198.3 $\pm$ 1.8	27	1670	1.80E-04	15.11 $\pm$ 0.41	270.35	100.0	44.1 $\pm$ 4.6
10VDL22	Pescadero granodiorite	6°49'48"	72°59'27"	636	199.1 $\pm$ 1.3	29	2835	1.77E-04	6.52 $\pm$ 0.18	74.90	40.6	172.0 $\pm$ 16.4
10VDL23	Bucaramanga gneiss	6°56'35"	72°58'00"	2055	461 $\pm$ 2.1	30	2564	1.43E-04	8.15 $\pm$ 0.33	86.38	87.0	124.8 $\pm$ 12.7
10VDL28	Onzaga granodiorite	6°22'31"	72°49'06"	1876	200.45 $\pm$ 0.7	28	2210	1.76E-04	12.07 $\pm$ 0.40	121.51	53.4	72.9 $\pm$ 7.4
10VDL32	Mogotes granodiorite	6°25'22"	72°49'29"	1782	197.97 $\pm$ 0.8	29	2670	1.38E-04	8.01 $\pm$ 0.26	122.55	59.4	132.2 $\pm$ 12.9
10VDL37	Berlin orthogneiss	7°11'24"	72°58'41"	1712	451.5 $\pm$ 1.5	30	4157	2.32E-04	9.84 $\pm$ 0.21	82.65	43.4	167.1 $\pm$ 15.0
10VDL49	Silgara Fm. Gneiss	7°29'20"	72°42'16"	2360	479.8 $\pm$ 3.1	21	2634	1.53E-04	5.26 $\pm$ 0.37	73.12	70.5	197.7 $\pm$ 24.1
10VDL52	Paramo Rico tonalite	7°13'54"	72°53'54"	3731	199.8 $\pm$ 1.2	29	3096	1.81E-04	5.36 $\pm$ 0.19	56.54	79.0	227.2 $\pm$ 22.1
10VDL55	Tarra granodiorite	8°03'10"	73°04'31"	1506	443.4 $\pm$ 3.2	14	1496	8.06E-05	3.05 $\pm$ 0.16	68.47	17.9	193.9 $\pm$ 25.6
10VDL59	Aguablanca granite	7°17'13"	73°08'46"	710	196 $\pm$ 1.1	29	1386	2.15E-04	14.83 $\pm$ 0.53	138.54	78.3	37.4 $\pm$ 4.0
10VDL61	Rio Surata granodiorite	7°09'59"	73°05'17"	827	200 $\pm$ 1.5	25	2040	2.81E-04	21.30 $\pm$ 1.08	134.33	100.0	38.3 $\pm$ 4.3

^a Number of spontaneous fission tracks Ns counted over the total area Ω_T

^b Sum of $P_i \cdot \Omega_i$ for all analysed grains, where P_i is the $^{238}\text{U}/^{91}\text{Zr}$ for the zircon grain i and Ω_i is the area over which Ns _{i} and P_i are evaluated

^c Chi-squared test: values greater than 5 % are considered to represent a single population

^d Weighted mean zircon U-Pb ages determined from the same samples, reported in Chapter 2

Table 4

Sample Data					Apatite Fission Track Data													HeFTy best fit model				Cooling rate	
Sample Code	Name	Latitude (N)	Longitude (W)	Elevation (m)	No. of Grains	Ns ^a	Ω _T (cm ²) ^a	Σ(P _i *Ω _i)±2σ (10 ⁻⁵ cm ²) ^b	U (ppm)	P(X ²) ^c	Pooled date (Ma) ±2σ	Dpar (μm) ^d	r _{mro} ^e	MTL (μm) ±σ ^f	S.D. ^g	N ^h	Date (Ma)	Date GOF ⁱ	MTL (μm) ±S.D.	MTL GOF ⁱ	(°C/Ma) ^j		
																					Min.	Max.	
10VDL05	Corcova granodiorite	7°06'03"	73°00'36"	2793	40	120	1.40E-03	5.64 ± 0.18	5.50	43.35	17.6 ± 3.3	1.80	0.847	15.01 ± 0.08 ± 0.95	129	17.7	0.95	15.04 ± 0.97	0.48	12	50		
10VDL22	Pescadero granodiorite	6°49'48"	72°59'27"	636	40	156	1.53E-03	6.45 ± 0.25	5.56	41.77	20.0 ± 3.3	1.98	0.829	15.11 ± 0.07 ± 0.92	195	18.2	0.28	15.13 ± 1.03	0.18	6	ND		
10VDL23	Bucaramanga gneiss	6°56'35"	72°58'00"	2055	39	453	1.57E-03	33.74 ± 1.08	25.62	99.15	11.1 ± 1.1	2.12	0.852	14.94 ± 0.07 ± 0.94	203	11.1	0.90	14.96 ± 1.05	0.38	12	26		
10VDL28	Onzaga granodiorite	6°22'31"	72°49'06"	1876	37	110	1.56E-03	7.52 ± 0.16	4.89	25.61	12.1 ± 2.4	1.97	0.843	14.52 ± 0.10 ± 0.99	103	12.0	0.96	14.55 ± 1.08	0.70	6	12		
10VDL32	Mogotes granodiorite	6°25'22"	72°49'29"	1782	37	123	1.52E-03	5.87 ± 0.20	4.33	24.87	17.3 ± 3.2	1.97	0.831	14.56 ± 0.08 ± 1.00	154	17.2	0.95	14.68 ± 0.96	0.98	4	12		
10VDL35	Surata diorite	7°10'22"	73°05'08"	840	40	137	1.56E-03	13.40 ± 0.35	9.90	63.78	8.5 ± 1.5	1.93	0.845	14.47 ± 0.09 ± 1.06	154	8.3	0.83	14.53 ± 1.05	0.89	9	17		
10VDL37	Berlin orthogneiss	7°11'24"	72°58'41"	1712	29	138	1.38E-03	13.37 ± 0.85	8.46	49.17	8.5 ± 1.5	2.06	0.852	14.82 ± 0.08 ± 0.98	155	8.5	0.95	14.83 ± 1.12	0.51	13	28		
10VDL49	Silgara Fm. Gneiss	7°29'20"	72°42'16"	2360	40	1046	1.88E-03	188.64 ± 12.04	106.44	56.87	4.6 ± 0.4	1.90	0.854	15.02 ± 0.06 ± 0.89	200	4.6	0.96	15.08 ± 0.97	0.90	33	>100		
10VDL52	Paramo Rico Tonalite	7°13'54"	72°53'54"	3731	40	205	1.67E-03	14.59 ± 0.65	9.05	55.52	11.6 ± 1.7	2.08	0.840	15.37 ± 0.06 ± 0.83	172	11.7	0.94	15.38 ± 1.07	0.60	25	>100		
10VDL55	Tarra granodiorite	8°03'10"	73°04'31"	1506	39	100	1.60E-03	6.85 ± 0.36	5.17	55.22	12.1 ± 2.5	1.94	0.838	15.11 ± 0.08 ± 0.82	110	12.2	0.93	15.22 ± 0.97	0.97	15	>100		
10VDL59	Aguablanca granite	7°17'13"	73°08'46"	710	40	106	1.65E-03	8.12 ± 0.37	5.39	62.25	10.8 ± 2.1	1.94	0.846	14.94 ± 0.07 ± 0.85	149	10.8	0.98	15.01 ± 0.99	0.99	10	63		
10VDL61	Surata granodiorite	7°09'59"	73°05'17"	827	38	373	1.56E-03	40.41 ± 1.46	23.63	99.03	7.6 ± 0.9	1.89	0.837	15.01 ± 0.07 ± 1.01	200	7.6	0.98	14.98 ± 1.18	0.60	19	64		

^a Number of spontaneous fission tracks Ns counted over the total area Ω_T

^b Sum of P_i*Ω_i for all analysed grains, where P_i is the ²³⁸U/⁴³Ca for the apatite grain i and Ω_i is the area over which Ns_i and P_i are evaluated

^c Chi-squared test: values greater than 5 % are considered to represent a single population

^d Etch pit length measured parallel to grain C-Axis

^e kinetic parameter for annealing models accounting for apatite chemistry determined during analysis (Ketcham et al., 2007)

^f Mean track length with standard error

^g Standard deviation of mean track lengths

^h Number of measured track lengths

ⁱ Goodness of fit value. See text for details

^j Cooling rates were estimated from the minimum and maximum slopes through good fit thermal history envelopes, and maximum rates could not be determined from 10VDL22 because of an absence of good fits. Cooling rates of >100°C/Ma cannot be reliably determined using our data