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This is the final peer-reviewed author's accepted manuscript (postprint) of the following publication:

Published Version:

Gennaro, I., Weber, J., Vitale Brovarone, A., Arkle, J., Chu, X.u. (2023). Geothermometric Constraints on the Thermal Architecture, Metamorphism, and Exhumation of the Northern Range, Trinidad. JOURNAL OF METAMORPHIC GEOLOGY, 41(2), 327-349 [10.1111/jmg.12697].

Availability:

This version is available at: <https://hdl.handle.net/11585/905011> since: 2023-03-01

Published:

DOI: <http://doi.org/10.1111/jmg.12697>

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Gennaro, Ivano; Weber, John; Vitale Brovarone, Alberto; Arkle, Jeanette; Chu, Xu:
*Geothermometric Constraints on the Thermal Architecture, Metamorphism, and
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JOURNAL OF METAMORPHIC GEOLOGY vol. 41 ISSN 0263-4929

DOI: 10.1111/jmg.12697

The final published version is available online at:

<https://dx.doi.org/10.1111/jmg.12697>

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**Geothermometric Constraints on the Thermal Architecture, Metamorphism, and Exhumation of
the Northern Range, Trinidad**

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27 **Running Title**

28 Thermal architecture of the Northern Range

29

30 **Keywords**

31 RSCM geothermometry, Trinidad, Geothermometry, Caribbean tectonics

32

33 **Data Availability**

34 The data presented here as supplemental material (Tables S1 and S2) are openly available on the
35 PANGAEA Database and may be accessed under the DOI:

36 <https://doi.pangaea.de/10.1594/PANGAEA.932772>

37

38 **Abstract**

39 The Northern Range of Trinidad is composed of Mesozoic passive margin sedimentary rocks
40 that underwent ductile deformation and sub-greenschist to greenschist facies metamorphism in the early
41 Miocene. Previous studies identified a westward increase in the metamorphic grade but were unable to
42 resolve whether this trend was discrete or continuous. In this study, we use Raman spectroscopy on
43 carbonaceous materials (RSCM) to constrain peak temperatures across the Northern Range with a
44 greater resolution than was available in previous studies. These data are then combined with published
45 thermochronological data to constrain the tectonic history of a range-cutting fault that had been
46 previously inferred in previous geologic mapping. The RSCM temperatures show an abrupt increase
47 from 337 ± 10 °C in the east to 442 ± 16 °C west of the Chupara Fault at Chupara Point. Our RSCM-
48 derived peak metamorphic temperatures are 50–100 °C higher than those from previous estimates,
49 requiring revision of tectonic models to account for deeper burial and greater exhumation. The peak
50 metamorphic conditions determined here, and our inferred timing of faulting, are consistent with the
51 two-stage tectonic model proposed in previous studies. A two-sided wedge formed during oblique plate
52 collision and mountain building (stage I). Cooling rates differed in the Northern Range between eastern
53 and western blocks between ~10–4.5 Ma; this difference is attributed here to the activity along the cross-
54 range Chupara Fault. This pattern of differential cooling, bedrock exhumation, and inferred bedrock and
55 surface uplift may be associated with plate-scale strike-slip tectonics and the passage of a crustal-scale
56 bulge induced by deep tearing of South American lithosphere (stage II).

57

58 1. INTRODUCTION

59 Trinidad, the largest island in the nation of Trinidad and Tobago, is the southeastern-most island
60 in the Caribbean and is located in the Caribbean-South American plate boundary zone (Fig. 1a), where
61 the Caribbean plate currently moves east relative to the South American Plate at a rate of ~20 mm/yr
62 (Weber, Dixon, et al., 2001). Current dextral plate motion is accommodated primarily along the Central
63 Range Fault, which strikes NEE-SWW through the middle of the island (Weber, 2009; Weber, Dixon, et
64 al., 2001). The Northern Range, composed of Mesozoic passive margin sedimentary rocks that were
65 metamorphosed in the early Neogene, exposes pre-transform tectonic structures. Plate convergence in
66 the early Miocene gave rise to shortening in the foreland fold-and-thrust belt and metamorphism at sub-
67 greenschist to greenschist facies conditions in the Northern Range segment of the hinterland belt (Algar
68 & Pindell, 1993; Frey et al., 1988; Weber, Ferrill, et al., 2001). Following the culmination of contraction
69 and burial, exhumation of the hinterland metamorphic rocks took place from the Miocene to Pliocene,
70 which is thought to have resulted from an oblique Caribbean-South American Plate collision (Algar &
71 Pindell, 1993; Weber, Ferrill et al., 2001) and/or propagation of a deep-seated lithospheric tear fault
72 (Arkle et al., 2017, 2021; Clark et al., 2008). The post-peak-metamorphic structures of the Northern
73 Range preserve the history of the regional exhumation kinematics (Weber, Ferrill et al., 2001), but the
74 general low grade of the Northern Range rocks limits the utility of conventional geothermometers for
75 accurately assessing metamorphic conditions and changes. Illite crystallinity measurements, carbonate
76 thermometry, muscovite-paragonite geothermometry, and mineral equilibria were used to establish
77 baseline metamorphic conditions (Frey et al., 1988; Weber, Ferrill, et al., 2001). Using temperature-
78 sensitive quartz and calcite microstructures, Weber, Ferrill, et al. (2001) were the first to recognize a
79 systematic increase in deformation temperatures from east to west. However, this method involved high

80 uncertainties in temperature estimates, making it difficult to identify whether this temperature gradient is
81 continuous or segmented by discrete structures.

82 Raman spectroscopy on carbonaceous material (RSCM) is applicable to a wide range of
83 lithologies within metamorphic terranes; it is particularly powerful in resolving low-grade peak
84 metamorphic temperatures, such as those in the Northern Range. This geothermometric technique
85 utilizes the crystallinity of carbonaceous material (CM) within the temperature range of 150–650 °C
86 (Aoya et al., 2010; Beyssac et al., 2002; Kouketsu et al., 2014; Lahfid et al., 2010). With increasing
87 metamorphic temperatures, amorphous CM evolves towards turbostratic carbon and then graphitic
88 carbon through the process of graphitization, which is reflected in the relative areas of graphite and
89 defect bands in Raman spectra (Beyssac & Lazzeri, 2012; Beyssac et al., 2002; Lahfid et al., 2010;
90 Wopenka & Pasteris, 1993). Because graphitization is theorized to be an irreversible process, the
91 crystallinity and Raman spectra of the CM should record peak metamorphic temperature conditions
92 (Beyssac & Lazzeri, 2012; Beyssac et al., 2003). It is important to ensure quality sample collection and
93 preparation for RSCM analysis, as for example, brittle deformation can affect the CM structure and
94 therefore the calculated metamorphic temperatures (Nakamura et al., 2015; Kirlova et al., 2018). When
95 high-resolution RSCM is combined with a high sample density this geothermometer can identify
96 structural features not seen using less accurate methods (e.g., see Fig. 3 of Vitale Brovarone & Agard,
97 2013). The technique has been applied in a variety of metamorphic environments, from high-pressure
98 terranes to contact aureoles (Bayet et al., 2018; Beyssac et al., 2019; Lahfid et al., 2010; Molli et al.,
99 2018; Rahl et al., 2005; Simoes et al., 2007; Vitale Brovarone & Agard, 2013; Vitale Brovarone et al.,
100 2013).

101 The Northern Range rocks and structures are deeply weathered, covered by dense tropical
102 vegetation, and, aside from excellent coastal and stream exposures, are generally poorly exposed and

difficult to access. Thus, previous structural studies have produced contradicting maps and interpretations (see below). In this study, we approach some of the unsolved structural and kinematic problems from a metamorphic perspective. Specifically, we use RSCM geothermometry to determine the peak metamorphic temperatures for a robust suite of Northern Range samples from east to west across the range (Fig. 2). We aimed to resolve the field gradient and to provide new temperature constraints to better establish the thermal architecture of the Northern Range. These new constraints help us determine whether the Northern Range was exhumed as a single rigid block or as discrete, fault-bounded range segments. Our new RSCM data, combined with previously published geological maps and thermochronological data, confirm the presence of a large-scale, cross-range, crustal-scale fault, the Chupara Fault, which was previously mapped but only poorly defined. Based on our new RSCM constraints, we review and modify existing tectonic models to explain the metamorphic, emplacement, and exhumation history of the Northern Range. The results presented here explore the thermal architecture and tectonic evolution of a mountain range formed in plate boundary that transitioned from collisional to transform, providing a real-world laboratory to examine processes that can then be used to better constrain geodynamic models (e.g. Govers & Wortel, 2005).

2. TECTONIC FRAMEWORK AND GEOLOGIC HISTORY

The Caribbean plate is postulated to have developed from the Caribbean large igneous province, an oceanic plateau or hotspot swell that formed from 139–83 Ma far west of its current location (Burke, 1988; Hoernle et al., 2004). The Caribbean plate began migrating eastward and entered the gap between the North and South American plates near the end of the Cretaceous (~66 Ma), and it eventually collided with northwestern South America (Kennan & Pindell, 2009) and southern North America. The rocks of the Northern Range were presumably derived from passive margin sediments deposited along the

126 northern edge of South America (Frey et al., 1988; Robertson & Burke, 1989). These sedimentary rocks
127 underwent low-grade metamorphism, with sub-greenschist mineral assemblages present in the east and
128 greenschist assemblages present in the central and western Northern Range (Frey et al., 1988, Weber,
129 Ferrill, et al., 2001). The presence of slates, marbles, quartzite and schists, as well as their detrital zircon
130 U-Pb age spectra, supports a passive margin provenance for these rocks (Bartlett et al., 2021).

131 Zircon fission-track thermochronology data was used to suggest that the initial phase of
132 metamorphism and deformation in the Northern Range may have started as early as the late Eocene
133 (Algar & Pindell, 1993), but this interpretation was based on partially reset samples and from only a
134 limited portion of the range. Subsequent studies used apatite fission track and apatite and zircon (U-
135 Th)/He data to revealed that the initial exhumation of the Northern Range occurred ~10–8 Ma in west
136 and ~6–4 Ma in the east (Arkle et al., 2021). Hinterland structural and metamorphic events are most
137 clearly delineated in the western Northern Range. White mica $^{40}\text{Ar}/^{39}\text{Ar}$ ages date peak metamorphism
138 in the Northern Range between 23 Ma and 34 Ma (Foland et al., 1992; Speed et al., 1997; Weber &
139 Arkle, 2015). Syn-metamorphic deformation (isoclinal F_1 folding, transposition of original bedding, and
140 S_1 foliation development) was followed by a second F_2 folding event and the development of S_2
141 crenulation cleavage (Weber, Ferrill, et al., 2001). The general picture is that plate-scale transpression
142 drove high magnitude crustal shortening and exhumation in the southern Caribbean metamorphic
143 hinterland, which includes the Northern Range (Cruz et al., 2007; Weber, Ferrill, et al., 2001). The main
144 stages of deformation and exhumation in the hinterland belt are largely coeval with folding and thrusting
145 in the Trinidad foreland, i.e., the Central and Southern Ranges. Deformation and exhumation in the
146 Central Range is constrained to have occurred in the mid-Miocene (11–18 Ma) by using stratigraphy and
147 zircon (U-Th)/He dating (Giorgis et al., 2017).

148 A late stage of deformation (D_3) within the Northern Range is recorded as a series of steeply

149 dipping normal faults that strike SE-NW and NE-SW and crosscut previous fabric elements (Fig. 2;
150 Kugler, 1961; Weber, Ferrill, et al., 2001). These faults have been interpreted as related to E-W
151 transform shear, and as such, these faults may accommodate transtension associated with relative
152 Caribbean-South American strike-slip motion (Algar & Pindell, 1993). Thus, Flinch et al. (1999) posit
153 that these D₃ faults formed during or after the late Miocene or Pliocene.

154 Many different scenarios have been proposed to explain the metamorphism and exhumation of
155 the metamorphic hinterland belt along northern South America. A few works advocate for the oblique
156 collision and subduction of South America beneath the Caribbean plate (Audemard, 2009; Cruz et al.,
157 2007; Pindell & Kennan, 2001, 2007, 2009). Some oblique collision models require about 240 km of
158 displacement since 12 Ma to bring the Northern Range into its current position (Pindell & Kennan,
159 2001). Exhumation of the Northern Range has been explained by isostatic rebound after forming a deep
160 crustal root that developed during an oblique collision (Algar & Pindell, 1993; Algar et al., 1998; Cruz
161 et al., 2007), or through the activation of late-stage, range-bounding brittle structures. Weber, Ferrill, et
162 al. (2001), for instance, initially called on differential dip-slip along the Arima Fault, a late south-side-
163 down normal fault bounding the southern Northern Range (Fig. 1b & 2), to explain differences in
164 eastern and western metamorphic grades.

165 Alternatively, metamorphism, deformation, and exhumation of the Northern Range could be
166 associated with the passage of a lithospheric tear fault under Trinidad (Arkle et al., 2017, 2021; Clark et
167 al., 2008; Levander et al., 2014). Currently, the subduction of the Atlantic oceanic lithosphere (of both
168 the North and South American plates) under the Caribbean plate is accommodated by tearing of the
169 lithosphere along a subduction-transform edge propagator (STEP, Fig. 1c; Govers and Wortel, 2005).
170 STEP faults mark the transition between subduction and transform plate boundaries and are
171 characterized by non-uniform relative plate motion; deformation, rotation, and uplift are often induced

172 along the transform side of the transition as the subducting plate is torn at the STEP edge (Govers &
173 Wortel, 2005). The active southeastern Caribbean STEP fault is marked by the Paria cluster of depth
174 earthquakes beneath the northern coast of eastern Venezuela (Fig. 1a), and the active STEP edge is
175 currently proposed to be positioned offshore eastern Trinidad, about 175 km east of the Paria cluster (c.f.
176 Fig. 2 of Nijholt & Govers, 2015). Some recent interpretations have postulated that exhumation of the
177 Northern Range was driven by STEP fault processes including the detachment of the South American
178 lithosphere, asthenosphere flow, and flexural bulging ahead of the STEP tear (Arkle et al., 2017; 2021;
179 Clark et al., 2008; Levander et al., 2014). Accordingly, eastward propagation of the Caribbean plate and
180 associated STEP fault caused the observed asymmetric exhumation, and the thermochronological ages
181 from the region that increase eastward and post-date the presumed oblique collision (Arkle et al., 2021).
182 STEP-induced deformation also fits the observed Quaternary tectonic inversion and asymmetry of slow
183 eastern and fast western exhumation rates (Arkle et al., 2017; Clark et al., 2008; Levander et al., 2014).
184 Differences between regional tectonic models germane to this study are whether the hinterland
185 metamorphic rocks are allochthonous and disconnected with folding and thrusting in the foreland or
186 parautochthons and connected with foreland deformation in space and time.

187

188 **3. GEOLOGIC SETTING AND SAMPLE DESCRIPTION**

189 **3.1 Structures and Stratigraphy**

190 The Northern Range of Trinidad is bounded by the Arima Fault to the south and the Northern
191 Coast and Sub-Tobago Terrane faults offshore to the north (Fig. 1b) (Robertson and Burke, 1989; Weber
192 et al. 2015, 2020). The Arima Fault was originally mapped by Kugler (1961) and corresponds to
193 observable meso-scale faults and cataclastic zones (Weber, Ferrill et al., 2001). Previously hypothesized

194 to be an eastward extension of the El Pilar transform fault, the Arima Fault is actually a south-side-
195 down, inactive normal fault that accommodated N–S extension (Weber, Ferrill et al., 2001). The
196 subsurface Caroni fault may instead represent the eastern El Pilar extension (Pindell & Kennan, 2007),
197 though active strike-slip motion now occurs along the Central Range Fault to the south (Fig. 1b; Weber,
198 Dixon, et al., 2001). The North Coast and Sub-Tobago Terrane fault zones (Fig. 1b) accommodate active
199 strike-slip and normal dip-slip between the Northern Range and Tobago terrane (Robertson & Burke,
200 1989; Weber et al. 2015, 2020). This offshore fault zone also contains *en echelon* normal faults that
201 strike northwest and have moderate to steep northeastern dips (Robertson & Burke, 1989). These *en*
202 *echelon* structures mirror the late normal faults mapped and observed in the Northern Range at the
203 meso- and macro-scales (Fig. 2) and are attributed by Robertson & Burke (1989) to be related to broad,
204 plate-scale, transform shear. Such structures are also mapped in eastern Paria (Cruz et al., 2007).

205 At Chupara Point, both Kugler (1961) and De Verteuil et al. (2005) infer the presence of the
206 Chupara Fault, a northwest-striking, steeply dipping, west-side-up normal fault. The inferred Chupara
207 Fault trends NNW-SSE from Chupara Point on the north coast, across the range, potentially extending
208 across the entire range southward to where it is truncated by the Arima Fault near the town of Maturita
209 (Fig. 2; Kugler, 1961; De Verteuil et al., 2005). On the other hand, Potter (1968) inferred a series of
210 short, disconnected, NW-striking fault segments at Chupara Point, Verdant Vale, and along the southern
211 foot of the range. Kugler (1961) mapped the Chupara Fault as one in a series of southeast- and
212 southwest-striking normal faults. Some of the other geologic maps of the range do not show the Chupara
213 Fault (e.g., Saunders et al., 1997). Mapping the Northern Range has been a significant challenge.
214 Kugler's (1961) biostratigraphic approach that proved so successful in Trinidad's sedimentary rocks
215 further south, together with related attempts to establish a protolith stratigraphy in these metamorphic
216 rocks, has largely failed (see below).

217 The metasedimentary rocks of the Northern Range include predominantly slates, quartz and mica
218 schists, quartzites, and metacarbonates. Mesozoic fossils have been reported from only about a dozen
219 localities scattered in the range (e.g., Dighton-Thomas, 1935; Imlay, 1954; Kugler, 1961; Saunders,
220 1972; Spath, 1939; Trechmann, 1935); these ages range from Tithonian (Jurassic) (?) to Maastrichtian
221 (Upper Cretaceous) (Fig. 2). Significantly, relict fossils have only been found in the low-grade rocks
222 east of the inferred Chupara Fault and in the fault-bounded, low-grade Laventille metalimestone and
223 Lopinot phyllite along the range front. Apparently, no relict fossils have survived in the higher-grade
224 metamorphism west of the Chupara Fault, which have been isoclinally folded, structurally transposed
225 (i.e., original bedding has been rotated into, cut by, and highly stretched in the S_1 foliation), and highly
226 recrystallized. In addition to the relict fossils, two radiometric ages that are highly disparate have been
227 reported from the San Souci metabasalt, which is a small and significant, but poorly studied, poorly
228 mapped, and poorly understood meta-igneous unit in the range (Fig. 2). Wadge & Macdonald (1985)
229 report a whole rock K-Ar protolith age of 87 ± 4.4 Ma, whereas Neill et al. (2014) dismiss the earlier
230 age as erroneous due to alteration and report a 135 ± 7.3 Ma zircon U-Pb crystallization age.

231 Traditionally it has been assumed, though not demonstrated, that a simple protolith stratigraphy
232 still exists in these metamorphic formations, despite the limitations discussed above, and the fact that
233 they are indeed metamorphic rocks, not sedimentary rocks (Algar & Pindell, 1993; Barr, 1965; Kugler,
234 1961; Potter, 1973; Saunders, 1997). Even in the western Northern Range, where relict fossils are
235 completely absent, this imagined protolith stratigraphy leads to the erroneous, widespread, and still
236 recently circulating interpretation that the megastructure of the Northern Range is a northward-vergent,
237 overturned, macroscopic anticline (Kugler, 1961; Potter, 1973; Algar and Pindell, 1993; Babb & Mann,
238 1999). Instead, the sheet dip of the bedding-parallel S_1 foliation is consistently toward the south there
239 and forms a homocline in which the resistant metamorphic units (quartz schists and quartzites) form a

240 series of south-dipping dip-slopes and hogbacks. The De Verteuil et al. (2005) map is the first attempt to
241 abandon the hypothesized protolith stratigraphy and simply map metamorphic rock types as directly
242 observed in the field. We follow a similar approach below as we describe and treat the units in the
243 Northern Range as low-grade metamorphic terranes.

244 **3.2 Petrography and Sample Descriptions**

245 Northern Range protoliths were likely derived from passive margin sediments deposited along
246 the northern edge of South America during and after the breakup of Pangea in the Jurassic and
247 throughout the Cretaceous (Speed, 1985; Frey et al., 1988; Robertson & Burke, 1989; Bartlett et al.,
248 2021). These protoliths were likely carbonates, mudstones, and sandstones, which have been
249 metamorphosed at sub-greenschist to greenschist facies conditions (Frey et al., 1988). The protoliths
250 represent a mixture of continental shelf sedimentary rocks that were deformed and exhumed during and
251 after Miocene collision (Pindell & Kennan, 2001, 2007). The twenty-six samples we collected and
252 analyzed are representative of the six major lithologies from across the Northern Range (Fig. 2) (1)
253 high-grade, foliated marble, (2) low-grade metacarbonate, (3) mica schist, (4) mica-rich phyllite, (5)
254 metasandstone, and (6) carbon-rich slate (Fig. 2 & 3).

255 The high-grade marbles (Fig. 3a) are all strongly foliated, exhibiting dark bands rich in CM and
256 lighter, carbonate-rich bands. Many of these samples also contain minor sulfides, mainly pyrite and
257 arsenopyrite; some of the pyrite grains have been weathered and broken down into iron oxides. The low-
258 grade metacarbonates (Fig. 3b; metacarbonate in Fig. 2) generally preserve primary sedimentary
259 structures and textures (Fig. 3c). Fossil shells and ooids are common in many of these rocks, which
260 exhibit low-strain, ductile deformation at the micro-scale (Fig. 3c). Some low-grade metacarbonates are
261 crosscut by calcite-filled veins indicative of brittle extensional deformation and precipitation from
262 metamorphic fluids. Further, later stage brittle deformation is recorded by the cross-cutting of calcite

263 veins by fractures.

264 The quartz-mica schists (Fig. 3d) mainly crop out in the western Northern Range. These schists
265 are well-foliated and consist of a greenschist-facies mineral assemblage (muscovite + chlorite + quartz +
266 plagioclase + CM \pm biotite \pm oxides or sulfides). These rocks are variable in both composition and
267 structure, showing a wide range of mica content (5–25%). Coarser-grained rocks tend to have less mica
268 and a less evident foliation. These are the only rocks seen to host metamorphic biotite in substantial
269 quantities, indicating they are likely of the highest grade in the study area. Quartz and calcite grains in
270 these rocks are thermally recrystallized, as indicated by the presence of undulose extinction in quartz
271 and the presence of triple-point grain boundaries (Fig. 3d). Recrystallization likely occurred during
272 deformation, as evidenced by the oriented stretching elongation of quartz parallel to the S_1 foliation
273 plane (Fig. 3d). This elongation is consistent with the observations of Weber, Ferrill, et al. (2001), who
274 attributed the foliation to the first stage, D_1 , of Northern Range deformation. At the microscopic scale,
275 CM is observed to be concentrated in mica-rich bands related to the tight F_1 folds that transpose bedding
276 (S_0) into S_1 and later kink bands.

277 The phyllites (Fig. 3e) are extremely rich in sericite, and are fine-grained, and well foliated. The
278 phyllitic foliation is defined by sericite \pm chlorite, with some degree of post- S_1 crenulation common in
279 these rocks. The multiple foliation planes recorded in these phyllites reflect the first two phases of
280 Northern Range deformation, which predominantly occurred in the ductile regime (Weber, Ferrill, et al.,
281 2001). The CM content of the phyllites is variable; only a few samples had sufficient CM for RSCM
282 analysis. Carbonaceous material with radial growth textures 0.1–0.3 mm in length is also observed in
283 some of these rocks and it appears to be post-kinematic as it crosscuts foliation (Fig. 3f); multiple
284 explanations are possible for their formation (cf. Luque et al., 1998; Rumble, 2014). Their composition
285 was confirmed to be carbonaceous through electron probe microanalysis (EPMA).

286 Metasandstones are composed mainly of subangular quartz grains (Fig. 3g), but also contain
287 minor mica and calcite along with CM. The weakly-metamorphosed rocks of this type come from the
288 eastern and central Northern Range and are typically non-foliated to weakly foliated. Atypically strong
289 foliation and folding were observed in sample NR-1 (Fig. 3h). Sample NR-1 (equivalent to NR-3)
290 displays isoclinal microfolds, likely representing parasitic folds created during early semi-ductile
291 deformation. Metasandstones from the central Northern Range display a weak foliation, but some quartz
292 grains experienced modest shape changes, likely through a combination of pressure solution and low-
293 temperature crystal plasticity (Weber, Ferrill, et al., 2001). Minor chlorite is seen in central Northern
294 Range samples. However, it is generally absent in the eastern metasandstones, reflecting either a
295 variation in composition or metamorphic grade; limited evidence suggests the former (Frey et al., 1988).

296 The dark, carbon-rich slates (Fig. 3i) are compositionally and texturally similar to the phyllites,
297 but tend to have a lower mica content and higher content of carbonate minerals (Fig. 4a) and CM. These
298 rocks are very fine-grained and well foliated, containing minor amounts of quartz, plagioclase, and
299 sulfides that mainly occur as framboidal pyrite. These rocks have a well-developed, early S_1 slaty
300 cleavage that is weakly crenulated by subsequent deformation.

301 **4. METHODS**

302 **4.1 Electron Probe Microanalysis**

303 Quantitative wavelength-dispersive spectrometry and backscattered electron (BSE) imaging were
304 carried out using a JEOL JXA-8230 electron probe microanalyzer (EPMA) at the University of Toronto.
305 Quantitative analyses employed a 10 nA beam current, 15 kV accelerating voltage, and off-peak
306 background corrections. The electron beam was defocused (10 μm) to minimize beam damage on
307 carbonate minerals.

308 **4.2 RSCM Geothermometry**

309 Thin sections were polished to a thickness of ~35 μm and made in accordance with the
310 specifications of Beyssac et al. (2003) for RSCM analysis. Rocks were cut perpendicular to bedding or
311 foliation planes and parallel to lineation in order to reduce within-sample heterogeneity caused by the
312 anisotropy of CM (Beyssac et al., 2003). The carbonaceous inclusions in transparent minerals were
313 analyzed using two Raman spectrometers: 15 samples were analyzed at the Department of Earth
314 Sciences at the University of Turin (UNITO) in Turin, Italy, and another 18 samples were analyzed at
315 the Royal Ontario Museum (ROM) in Toronto, Canada. Seven samples were analyzed on both
316 spectrometers to ensure inter-instrument consistency. Only one sample (DT-93-1) was omitted due to a
317 lack of CM.

318 Raman spectra were collected using a LabRAM HRVIS from Horiba Jobin Yvon Instruments at
319 UNITO and a Horiba LabRAM ARAMIS micro-Raman spectrometer at the ROM. Both spectrometers
320 were calibrated with a silicon standard and used a 532 nm solid-state neodymium laser. The lasers were
321 dispersed using 1200 gr/mm gratings and focused on the samples using 100x objectives with a
322 numerical aperture of 0.9. In order to avoid laser-induced heating, a low laser power (<5 mW) was used
323 during analysis (Beyssac et al., 2003). At UNITO, no filter was required; at the ROM, the D1 or D0.6
324 filters (10% and 25% transparency respectively) were used to ensure a laser power below 5 mW.

325 A combination of transmitted and reflected light was used to target CM located below the
326 surface of the host minerals to avoid artificial defects induced by sample preparation (Beyssac &
327 Lazzeri, 2012; Beyssac et al., 2003). At least 15 spectra were collected for each sample. The spectral
328 baselines were corrected, and the spectra were fitted using the Peakfit software (AISN Software Inc.,
329 1995). Wherever possible, a linear baseline was used for the correction, but micaceous samples

330 sometimes required a best-fit baseline correction as the fluorescence of mica is detectable within the
331 range of 700–2000 cm^{-1} (Beyssac & Lazzeri, 2012).

332 For samples showing CM Raman spectra characteristic of high-temperature conditions ($>330\text{ }^{\circ}\text{C}$;
333 e.g., SC-2b, Fig. 5a), we apply the 532 nm laser-based calibration of Aoya et al. (2010) (Eq. 1) along
334 with the 514 nm laser-based calibration of Beyssac et al. (2002) (Eq. 2) as a reference. These spectra
335 were fitted with three bands and Voigt peaks, which converged to a unique solution (Beyssac & Lazzeri,
336 2012; Lahfid et al., 2010). For samples showing CM Raman spectra characteristic of low-temperature
337 conditions ($<350\text{ }^{\circ}\text{C}$; e.g., T94-3, Fig. 5b), we apply the 514 nm laser-based calibration of Lahfid et al.
338 (2010) (Eq. 3) and the two 532 nm laser-based calibrations of Kouketsu et al. (2014) (Eq. 4 & 5). For the
339 Lahfid et al. (2010) calibration, these spectra are decomposed into five bands; the use of Lorentzian
340 profiles was necessary to reduce the degree of freedom and allow a unique solution to be obtained
341 (Beyssac & Lazzeri, 2012; Lahfid et al., 2010). The samples were then refit following Kouketsu et al.
342 (2014) to use their calibrations (see below). The temperatures were then calculated using the area ratios
343 of the peaks based on the respective thermometer calibrations (see Aoya et al., 2010; Beyssac et al.,
344 2002; Lahfid et al., 2010).

345 Most samples displayed solely high- or low-temperature spectra, whereas nine samples of
346 intermediate temperature (300–360 $^{\circ}\text{C}$) feature a mixture of 3- and 5-band spectra. In the case of the
347 nine samples with both types of spectra, the Aoya et al. (2010) calibration was applied to 3-band spectra
348 and the Lahfid et al. (2010) calibration was applied to 5-band spectra. We note that only four reported
349 temperatures are the averages of both calibrations that statistically overlap. The second calibration
350 applied to the other five samples is rejected as outliers (Table 1). For example, among the fifteen
351 datapoints of 93-T-CB, three 5-band spectra yield much lower temperatures ($284 \pm 10\text{ }^{\circ}\text{C}$) than the
352 twelve 3-band spectra which make up the majority ($366 \pm 5\text{ }^{\circ}\text{C}$; Table S1). On the other hand, one

353 datapoint of MAT-1(3) shows a 3-band spectrum that yields a temperature of 472 °C, >100 °C higher
354 than the average of the 5-band spectra (340 ± 4 °C); this CM inclusion, consistent with the temperature
355 calculated using the Beyssac et al. (2002) calibration, is interpreted as detrital CM (Table S1). In both
356 cases, the outliers were omitted when reporting the final temperature.

357 A second fitting of the data was done on samples with peak temperatures below 400 °C in
358 accordance with the specifications of Kouketsu et al. (2014). These calibrations utilize 3–5 peaks which
359 allow for the calculation of peak temperature based on the full width at half maximum (FWHM) of the
360 D1 and D2 bands (Kouketsu et al., 2014). A second fitting ensured all spectra treated were fit with a
361 linear baseline in the range of 1000–1750 cm^{-1} with pseudo-Voigt (Gaussian-Lorentzian sum) functions
362 in accordance with Kouketsu et al. (2014). Some peaks were fixed based on the characteristics of the
363 Raman spectra, such as the intensity ratio of the main D and G bands, to allow for convergence to a
364 unique solution (Kouketsu et al., 2014).

365 Following these calibrations, R2 and RA1 represent the area ratios between the main D band and
366 the entire spectrum in high- and low-temperature samples, respectively, with FWHM_{Dx} representing the
367 full width at half maximum of the corresponding defect band (D1 or D2) (Beyssac et al., 2002;
368 Kouketsu et al., 2014; Lahfid et al., 2010).

369
$$T(^{\circ}\text{C}) = 221.0(R2)^2 - 637.1(R2) + 672.3 \quad (1)$$

370
$$T(^{\circ}\text{C}) = -445(R2) + 641 \quad (2)$$

371
$$T(^{\circ}\text{C}) = (RA1 - 0.3758)/0.0008 \quad (3)$$

372
$$T(^{\circ}\text{C}) = -2.15(\text{FWHM}_{\text{D1}}) + 478 \quad (4)$$

373
$$T(^{\circ}\text{C}) = -6.78(\text{FWHM}_{\text{D2}}) + 535 \quad (5)$$

374

375 **5. RESULTS**

376 **5.1 Carbonate Thermometry**

377 Weber, Ferrill et al. (2001) presented comprehensive documentation and analyses of the calcite
378 microstructures in the Northern Range metasedimentary rocks (see their Figure 8). Depending on the
379 metamorphic grade, the calcite microstructures range from thin calcite twins in rocks with intact and
380 preserved sedimentary textures to plastically deformed, fully recrystallized, mechanical twins. Given
381 that these dynamically recrystallized calcite grains are commonly present in crosscutting veins, the
382 highest temperatures derived from calcite microstructures in that study can thus be considered as a lower
383 bound for peak metamorphic temperatures. Here, we attempt Ca-Mg-Fe thermometry between calcite
384 and dolomite in a slate sample with disseminated calcite and dolomite (NR-5, equivalent to NR-2), and
385 show that this thermometer is not applicable due to a lack of chemical and textural equilibrium.

386 A BSE image representative of the mineral assemblage and texture of NR-5 is presented in
387 Figure 4a. Ankerite grains (dol in Fig. 4) are larger and more euhedral than calcite. The schistosity bends
388 and forms pressure shadows around ankerite micro-porphyroblasts. The ankerite has a $Mg\# (=Mg / [Mg + Fe]) \sim 0.4$ and Ca content ~ 0.54 per 3 O. The calcite contains minor (~ 0.04 per 3 O) Fe, Mg and Mn
389 (Fig. 4b). The bending of schistosity around both calcite and ankerite grains suggests that both are likely
390 pre-kinematic, potentially pointing to the presence of both minerals during peak metamorphism.
391 However, the two minerals show contrasting crystal habits, with euhedral ankerite exhibiting stability
392 while anhedral calcite grains likely reequilibrated with the surrounding matrix during retrograde
393 metamorphism.
394

395 Along with the disequilibrium textures, the analyzed carbonates display a high degree of
396 scattering on the carbonate phase diagrams, yielding a large temperature range (Fig. 4b). While the
397 thermodynamic models (Anovitz & Essene, 1987) include uncertainties and equilibrium assumptions,
398 the variability of measured carbonate concentrations adds to the concern of applying this particular
399 mineral-pair thermometer. Calcite microstructural geothermometry from the adjacent area yields 200–
400 250 °C (Weber, Ferrill, et al., 2001), but inferred phase relations suggest disequilibrium between calcite
401 and dolomite at 250 °C (Fig. 4b). In contrast to the >300 °C temperature recorded by RSCM (see
402 below), the calcite composition and microstructures must have been modified during retrograde
403 metamorphism. In sum, as in Frey et al. (1988), conventional geothermometers could not resolve the
404 field gradient across the low-grade eastern Northern Range due to the violation of several of the required
405 fundamental assumptions.

406 **5.2 RSCM Data and Thermometry**

407 Representative low-temperature and high-temperature Raman spectra are presented in Figure 5
408 to illustrate the decomposition of Raman spectra for RSCM calculation. A complete set of fitted spectral
409 data are provided in the supplementary material (Tables S1 & S2; Gennaro et al., 2021), and a complete
410 set of results based on spectra fitting are presented in Table 1. In the first-order region (700–2000 cm⁻¹),
411 the low-temperature and high-temperature samples show characteristic RSCM features such as the
412 presence of five (G, D1, D2, D3, D4) and three (G, D1, D2) major Raman bands, respectively (Fig. 5;
413 Beyssac & Lazzeri, 2012; Beyssac et al., 2002; Lahfid et al., 2010). Samples of intermediate
414 temperature (~340–380 °C) were also fitted with four bands in accordance with Kouketsu et al. (2014).
415 Along with the major D and G peaks, spectra from carbonate-rich samples show calcite peaks at ~1100
416 cm⁻¹, which were minimized as much as possible in measurements. Values of the R² fitting correlation

417 coefficient on samples with calcite peaks may be low as this peak is removed after fitting to allow for
418 calculations using only peaks associated with CM.

419 The centers of fitted G bands show moderate variation (Table S1; Fig. S1) from fully ordered
420 graphite (1582 cm^{-1} ; Wang et al., 1990), especially the low-temperature samples analyzed at UNITO
421 (Fig. S1). The downshifted Raman spectra might reflect amorphization by laser heating (Kagi et al.,
422 1994; Iwasaki et al., 2013; Nakamura et al., 2019), as Raman spectra of low-grade CM are ideally
423 obtained with laser powers of $<1\text{ mW}$ to avoid laser-induced artifacts (Nakamura et al., 2019).
424 Nonetheless, we note that the apparent temperatures calculated using 3- and 5-band calibrations (Aoya
425 et al., 2010 and Lahfid et al., 2010, respectively) do not show any systematic correlation with the shifts
426 of G bands (Fig. S1). Given that the intrinsic uncertainty of RSCM on low-grade samples is unquantified
427 and potentially considerable, we reasonably regard the temperatures as robust results.

428 The temperatures from the two Kouketsu et al. (2014) calibrations display up to $100\text{ }^{\circ}\text{C}$
429 variations between the temperatures calculated using the two calibrations on the same sample (Fig. 6
430 inset). The FWHM of G and D bands has been known to vary as a result of laser-induced heating at laser
431 powers above 1 mW , thus laser-induced heating is the likely cause of this discrepancy (Iwasaki et al.,
432 2013, Nakamura et al., 2019). This is especially evident if the downshift of the G band is used as a proxy
433 for laser-induced heating. Samples with higher scatter in the position of the G band tend to show greater
434 disagreement between the D1- and D2-based calibrations of Kouketsu et al. (2014) (Fig. S1). Therefore,
435 while laser-induced heating does not significantly affect the temperatures of the Aoya et al. (2010) or
436 Lahfid et al. (2010) calibrations in our analysis, the temperatures derived from the Kouketsu et al.
437 (2014) thermometer are considered to be significantly altered and are not used for interpretation.

438 Our new RSCM temperatures range from $310 \pm 13\text{ }^{\circ}\text{C}$ to $465 \pm 30\text{ }^{\circ}\text{C}$ (Lahfid et al. 2010 and
439 Aoya et al. 2010 calibrations; 1σ), with an outlier at $497 \pm 27\text{ }^{\circ}\text{C}$ (sample M-4). Figure 5c shows the

440 evolution of the acquired spectra with increasing metamorphic temperature. The spectral evolution
441 shows an increase in the intensity ratio of the main D and G bands (I_D/I_G) until about 350 °C, where the
442 main G peak then begins increasing in intensity and area as the D peak decreases. In low-temperature
443 spectra, the D3 and D4 bands decrease in area as temperature increases, and they disappear at around
444 350 °C. Above this threshold, the D1 and D2 bands decrease in area relative to the G composite band,
445 suggesting progressive graphitization (Beyssac et al., 2003).

446 We also assess the inter-instrument correlation between the spectra collected at UNITO and the
447 ROM, plotting the calculated temperatures from both spectrometers against a 1:1 line (Fig. 6). The
448 reproducibility of the high-temperature spectra is very good. However, the lower temperature spectra
449 display a larger inter-instrument error regardless of the calibration used. A degree of inter-instrument
450 error is to be expected, has been observed in the past, and is caused by differences in analytical settings
451 such as laser power, the instrument's CCD, and slit spacing on the diffraction grating (Lünsdorf et al.,
452 2013). The increased inter-instrument error in the low-temperature region may be due to differences in
453 laser power, despite being kept below 5 mW, as laser-induced heating may still induce defects in the
454 Raman spectrum at this energy, especially in highly amorphous CM (Beyssac et al., 2003, Kagi et al.,
455 1994; Iwasaki et al., 2013; Nakamura et al., 2019). With regards to the scatter of low-temperature data,
456 the calibrations for low-temperature spectra are more complex as graphitization below 330 °C is affected
457 by variables other than just temperature (Lahfid et al., 2010). Furthermore, the RSCM calibrations,
458 especially at low temperatures, are highly sensitive to the fitting and baseline correction method used
459 (Beyssac & Lazzeri, 2012; Lünsdorf et al., 2013). For consistency, and to reduce the variability caused
460 by inter-instrument error and laser-induced heating, samples from the ROM were used in all cases where
461 samples were analyzed using both spectrometers.

462 The spatial variation of temperatures exhibits an increase from east to west across the Northern
463 Range (Fig. 7a), in agreement with previous studies (Frey et al., 1988; Weber, Ferrill, et al., 2001). The
464 Lahfid et al. (2010) RSCM results yield temperatures around 320 °C in the easternmost parts of the
465 Northern Range, which increase westward to about 360 °C in the central Northern Range (Fig. 7b).
466 Temperatures of ~360 °C in the central Northern Range are predicted by both low-temperature
467 calibrations. An abrupt temperature difference is seen in the central Northern Range across the
468 previously mapped Chupara Fault (De Verteuil et al. 2005; Kugler, 1961). To the west of this fault,
469 RSCM records peak temperatures of around 450 °C, except for the two samples at ~420 °C in the
470 vicinity of the fault. The temperature difference between the eastern (337 ± 10 °C; Lahfid et al. (2010)
471 calibration) and western segments (442 ± 16 °C; Aoya et al. (2010) calibration), within ~5 km, is $105 \pm$
472 19 °C, and a two-tailed Student's t-test indicates that the means are statistically different at the 99%
473 confidence level. Qualitative observation of the Raman spectra corroborates the significant temperature
474 difference between the eastern and western samples. The presence of significantly different Aoya
475 temperatures in the east and west further corroborates this difference as these were derived using the
476 same calibration and represent a degree of internal consistency. The temperatures are internally
477 consistent within the western block, with the eastern block showing higher variation, particularly in the
478 easternmost regions of the Northern Range. Samples in the east gradually increase in temperature from
479 east to west, though they remain close to 340 °C. In addition, samples in the southwestern Northern
480 Range are bounded by the Arima Fault and perhaps additional bounding faults, and typically show lower
481 temperatures (350–410 °C) than their northwestern counterparts (~450 °C; Fig. 7a).

482 **6. Discussion**

483 **6.1 Metamorphic Field Gradient and Faulting**

484 Our RSCM data not only corroborate the existence of a geothermal gradient across the Northern
485 Range, but now resolve that this gradient is discrete. The thermal discontinuity in the central Northern
486 Range (Fig. 7a, b) juxtaposes high- (442 ± 16 °C) and low-temperature (337 ± 10 °C) blocks at Chupara
487 Point. Given the abrupt temperature change we observe, a structural discontinuity must exist to
488 accommodate it. The most probable candidate for such an intra-range boundary is the Chupara Fault as
489 mapped by Kugler (1961) and De Verteuil et al. (2005). The Chupara Fault is one in a series of SE-
490 striking normal faults that were hypothesized by Algar & Pindell (1993) to represent the final,
491 transtensional stage of Northern Range deformation and presumed to be related to a change in the
492 direction of Caribbean-South American plate motion from collisional to right-lateral strike-slip
493 (transform).

494 Much of our RSCM data comes from the north coast, leaving the southern mountain front and
495 internal segment of the metamorphic discontinuity more poorly constrained. Here, we refine slightly the
496 Chupara Fault trace as mapped by De Verteuil et al. (2005), which was based on differences in
497 metamorphic rock types and inferred differences in grade using microstructures. Topographic radar
498 images show possible minor differences across the northern trace of the Chupara Fault; these diminish to
499 the south (Fig. 7a), suggesting that the Chupara Fault is likely inactive. Future sampling in the range
500 interior and more RSCM data can help to better define the position of the southern extent of the thermal
501 gradient we observe as well as that of the Chupara Fault. We also note that we lack the data needed to
502 speculate whether or how the onshore Chupara Fault relates to any offshore faults (e.g., Robertson and
503 Burke, 1989).

504 In the southwestern Northern Range, the lower temperature samples of 350–410 °C in the
505 Chancellor schist are separated from the main western block by a fault previously mapped by De
506 Verteuil et al. (2005) (Fig. 7a). This fault may accommodate up to 4 km of vertical slip based on a

507 typical passive margin geotherm of 25 °C/km (Weber, Ferrill, et al., 2001). The temperature difference
508 between the northwesternmost rocks and the Chancellor schist in the southwest mirrors the S–N grade
509 increase observed in Paria, eastern Venezuela by Cruz et al. (2007, and references therein). Two samples
510 ~3–5 km west of the Chupara Fault display slightly lower peak temperatures (~420 °C) than others
511 surrounding them and thus may be separated from the rest of the western block by an additional splay
512 fault(s) (e.g., Fig. 3b in Algar & Pindell, 1993) or some other unrecognized fault(s) (Fig. 2; De Verteuil
513 et al., 2005). Finally, one outlier in the west, sample M-4, displays scattered, high peak temperatures of
514 497 ± 27 . This sample was collected from a shear zone where processes such as frictional heating during
515 faulting may have altered the CM structure, leading to an increased RSCM-derived temperature
516 (Furuichi et al., 2015).

517 The RSCM results from within the eastern block display an additional, less pronounced,
518 systematic scatter (Fig. 7b), which might reflect second-order gradients or discontinuities that are
519 difficult to quantify. This can be seen qualitatively by comparing the CM Raman spectra of samples
520 from the eastern and central Northern Range (following, e.g., Kouketsu et al., 2014). We also note that
521 the increased elongation of quartz grains in central metasandstones indicates a higher degree of quartz
522 ductility and therefore, a potential for deeper burial and higher-grade metamorphism than rocks further
523 east (Weber, Ferrill et al., 2001). Furthermore, the Sans Souci metabasalt (Fig. 2) along the northeast
524 coast has a prehnite-pumpellyite mineral assemblage, while the rest of the eastern and central Northern
525 Range appears to be mineralogically closer to or within the higher-grade greenschist facies (Frey et al.,
526 1988). The RSCM data in the eastern block are sparse, so we are unable to rule out faulting as a cause of
527 the potential eastern thermal gradient, and therefore further work is needed to confirm and more fully
528 quantify the nature of this gradient.

529 **6.2 Timing of Metamorphism and Fault-related Exhumation**

530 We next used metamorphic ages and low-temperature thermochronology data, together with our
531 RSCM results, to infer the timing of fault slip, differential bedrock cooling, and inferred exhumation
532 along the block-bounding Chupara Fault. Our RSCM data provide new constraints and support for the
533 previously proposed Miocene age of the Northern Range. $^{40}\text{Ar}/^{39}\text{Ar}$ spectra from the Northern Range,
534 introduced in abstract form (Foland et al., 1992) and in a field guide (Weber & Arkle, 2015) all indicate
535 ages of ca. 25–30 Ma (Fig. 8b). In addition, Speed et al. (1997) determined the $^{40}\text{Ar}/^{39}\text{Ar}$ age of well-
536 ordered (10\AA), syn-metamorphic white mica of the Dragon Gneiss in Paria, Venezuela, presenting an
537 age of 23.3 ± 0.2 Ma (between 8% and 95% Ar release). A second age in that study yielded a total gas
538 age of 34.3 Ma with a plateau around 21 Ma (Speed et al., 1997). We note that the temperature estimates
539 in other geothermometry studies ($<400\text{ }^{\circ}\text{C}$; Frey et al., 1988; Weber, Ferrill, et al., 2001) are all lower
540 than, or close to, the closure temperature of the muscovite $^{40}\text{Ar}/^{39}\text{Ar}$ system ($380 \pm 30\text{ }^{\circ}\text{C}$; Harrison et
541 al., 2009). Because the $^{40}\text{Ar}/^{39}\text{Ar}$ ages are mainly taken from metamorphic micas, previous studies
542 regard them as representative of the timing of peak metamorphism. However, some of the younger ages
543 (21–23 Ma) might have experienced Ar loss or resetting during cooling.

544 Zircon fission track (ZFT) data from the Northern Range show a general pattern with reset ages
545 of ~ 12 Ma in the west (Fig. 8b), whereas most eastern zircons are un-reset and display ages ranging
546 from ~ 80 – 200 Ma (Algar et al., 1998; Arkle et al., 2021; Weber, Ferrill, et al., 2001). Our new RSCM
547 temperatures from both the east and west (310 – $450\text{ }^{\circ}\text{C}$) all exceed the ZFT closure temperature of $240 \pm$
548 $30\text{ }^{\circ}\text{C}$ (Brandon et al., 1998). This produces some discrepancy because of the presence of detrital ZFT
549 ages at these temperatures; this relationship is also seen in studies of Taiwan's mountain ranges (Beyssac
550 et al., 2007; Simoes et al., 2007). These discrepancies are thought to result from the complexities of
551 zircon annealing, including effects related to variable thermal histories and a range of physical and
552 chemical properties of zircon grains (Brandon et al., 1998).

553 Post-metamorphic thermochronology data from the Northern Range indicate significant
554 differential exhumation of the eastern and western blocks between ~10–4.5 Ma (Arkle et al., 2021),
555 which we suggest corresponds to the period of most active slip of the Chupara Fault (Fig. 8). Arkle et al.
556 (2021) document rapid bedrock cooling and exhumation (~1.5 mm/yr) of the western Northern Range,
557 wherein >7 km of rock was exhumed from ~10–4.5 Ma (Fig. 8b). During the same time period, thermal
558 models indicate that pre-Pliocene exhumation of the eastern Northern Range was only <2 km (Fig. 8a).
559 Although exhumation was clearly focused within the western and eastern crustal blocks of the Northern
560 Range, mechanisms of differential exhumation such as discrete faults or other structures that dissect the
561 mountain range could not previously be identified. We propose that the east-to-west offset of RSCM
562 temperatures we observe need to be consistent with the magnitude and style of offset on the Chupara
563 Fault. The 4 Ma–Recent differential exhumation between the eastern and western Northern Range could
564 be accommodated by faulting, such as reactivation of the Chupara Fault. However, no geomorphic or
565 geodetic evidence supports its Quaternary to Recent slip.

566 In the discussions above, we assume that cooling was mainly induced by exhumation. Evidence
567 at the regional scale suggests that a majority of the differential cooling in the Northern Range can most
568 reasonably be attributed to exhumation driven by normal faulting and erosion. Bedrock exhumation by
569 erosion off the top of the North Range is supported by the initiation of high sedimentation rates in the
570 Gulf of Paria basin starting at ~4 Ma and in the North Coast basin at ~12 Ma (Flinch et al., 1999). The
571 late-stage normal faults pervasive throughout and surrounding the Northern Range (Algar & Pindell,
572 1993; Kugler, 1961; Weber et al., 2015) also indicate favorability for fault-driven exhumation following
573 peak burial. Furthermore, the timing of increased sedimentation, e.g., as described by Flinch et al.
574 (1999), matches the timing of accelerated cooling in the west (late Miocene–Pliocene), suggesting a
575 coupling between sedimentation and bedrock cooling due to erosional exhumation of the Northern

576 Range. On the other hand, Pindell and Kennan (2007) suggest that the underplating of cold proto-
577 Caribbean material beneath the western Northern Range is responsible for its cooling (also see 6.3).
578 However, this sort of regional-scale thermal conduction would presumably produce a more gradual
579 temperature profile than that seen in our new RSCM data. In addition, such underthrusting is postulated
580 to have occurred before the ~12 Ma Caribbean-South American collision in Trinidad (Algar & Pindell,
581 1993; Pindell & Kennan, 2007). If underthrusting did cause significant cooling, its effect should
582 therefore have been strongest before 12 Ma, precluding it from contributing significantly to the younger
583 thermal gradient that we observe. Advection of cold fluid can, in principle, also produce steep field
584 gradients (e.g., Rogiers et al., 2014), though its durations are shorter (< 0.1 Myr) and depths are typically
585 much shallower (< 2 km) than the t - T histories documented herein.

586 **6.3 Mechanisms of Hinterland Metamorphism**

587 Most models attribute metamorphism and subsequent exhumation of the Northern Range, as well
588 as of Venezuela's Cordillera de la Costa, to the oblique collision between the Caribbean and South
589 American Plates (Algar & Pindell, 1993; Garciacaro et al., 2011; Pindell & Kennan, 2001; Robertson &
590 Burke, 1989). It is also generally agreed upon that exhumation occurred in Trinidad in the mid-Miocene,
591 although its mechanism is disputed (Algar & Pindell, 1993; Arkle et al., 2017; 2021; Clark et al., 2008;
592 Cruz et al., 2007; Speed, 1985; Weber, Ferrill, et al., 2001). Our new RSCM temperature estimates
593 exceed temperature estimates from all previous metamorphic studies in the Northern Range (300–350
594 °C, Frey et al., 1988; 250–400 °C, Weber, Ferrill, et al., 2001). If attributed to depth, the higher
595 metamorphic grades call for revision of tectonic models to accommodate deep burial; here, we suggest
596 that the required deep burial likely occurred through 1) the incorporation of additional shelf sediments
597 into the fold-thrust or accretionary wedge (e.g., Pindell & Kennan, 2001), and 2) transpression acting to

598 squeeze this material between the two plates, allowing for both increased crustal thickness and orogenic
599 uplift (e.g., Cruz et al., 2007).

600 The Northern Range rocks likely began as a mixture of passive margin (metamorphosed
601 mudstone and turbidites) and continental shelf (becoming marbles and metasandstones) sedimentary
602 protoliths (Fig. 9a). The sequence was incorporated into a fold-thrust wedge along the leading edge of
603 the Caribbean plate with varying degrees of transport having been inferred (Fig. 9b; Pindell and Kennan,
604 2001, 2007; Speed, 1985). The two end-members contrasting oblique collision models depict the entire
605 south Caribbean metamorphic hinterland, including the Northern Range, either as an allochthon
606 (Audemard, 2009; Avé Lallemant, 1997; Pindell & Kennan, 2001) or a parautochthon (Cruz et al., 2007;
607 Russo & Speed, 1992, 1994).

608 Tomographic imaging beneath northern South America reveals a sinking mass, which has been
609 associated with either a southward-subducting proto-Caribbean plate (Pindell and Kennan, 2001, 2007)
610 or the removal of South American lithosphere along a near vertical crustal STEP (Levandar et al., 2014).
611 The allochthonous model attributes deformation to events occurring since the early Paleocene due to the
612 subduction of a proto-Caribbean plate beneath South America. Pindell and Kennan (2007) suggest that
613 the Northern Range represents the accretion of both proto-Caribbean and Caribbean trench sediments.
614 However, no magmatism associated with the southward subduction of a proto-Caribbean lithosphere
615 beneath South America has been found on the South American continent. The lack of upper-plate
616 volcanism and differing seismic tomographic interpretations leave the existence of a subducted proto-
617 Caribbean plate open to debate (Levander et al., 2014; Pindell & Kennan, 2001, 2007). In addition, the
618 accretionary model (Pindell & Kennan, 2001, 2007) requires about 200 km of displacement since 10 Ma
619 to bring the Northern Range to its current position. This value contrasts with <100 km of displacement
620 along the El Pilar and associated transform plate boundary faults (Audemard, 2009, and references

621 therein). We also note that the 450 °C temperature from our new RSCM results suggests deeper burial
622 than is typical for both sedimentary and structural burial within accretionary wedges. In addition, the
623 lack of high-pressure minerals like sodic-calcic amphiboles and garnet rules out potential high-P/T
624 metamorphism of an oceanic slab such would be expected from the subduction of a large, mature
625 “proto-Caribbean” plate. Thus, we prefer a model with significant crustal shortening associated with
626 tectonic burial and crustal downflow to bring the Northern Range protoliths down to depths of ~17 km
627 (25 °C/ km geotherm; c.f. Cruz et al., 2007), although we cannot rule out the possibility of some pre-
628 collisional influence from a proto-Caribbean plate.

629 Cruz et al. (2007) suggest that the Venezuelan metamorphic belt was exhumed as a two-sided
630 wedge, such as those described in the numerical models of Willet et al. (1993); as the Northern Range
631 represents an extension of this belt, the same mechanism may apply. The geometry of a two-sided
632 wedge (e.g., Cruz et al. 2007) could have been induced in Trinidad by northward subduction of the pre-
633 collisional oceanic portion of the South American plate beneath the Caribbean plate. The model is
634 supported by the oblique-normal shear indicators and SW-plunging lineations observed in the Cordillera
635 de la Costa, Venezuela (Cruz et al., 2007). In a two-sided wedge, the material is accreted in the foreland
636 and moves towards the retro-wedge of the hinterland, and thus, the metamorphic grade increases
637 towards the retro-wedge (Willett et al., 1993). Topographically, a two-sided wedge presents a low-angle
638 taper on the “pro-wedge” portion on the side of the subducting slab, while the “retro-wedge” is on the
639 side of the overriding plate and displays a high-angle taper (Willet et al., 1993). The RSCM data from
640 the southwestern Northern Range are consistent with this model, which predicts a metamorphic
641 temperature increase towards the retro-wedge. The Chancellor schist displays lower RSCM temperatures
642 than those in the rest of the western block (340–410 °C, purple unit in Fig. 7a). The range-front position
643 of the Chancellor schist, and the fault-bounded, low-grade Laventille metalimestone and Lopinot

644 phyllite (Fig. 2), which display a range-front fold geometry of upright NE-SW trending folds, could
645 reflect the modified southern side of a bivergent wedge (e.g., De Verteuil et al., 2005; Teyssier et al.,
646 2002; Weber, Ferrill, et al. 2001). Farther south in the Central Range, which represents a major foreland
647 fold-thrust structural culmination, <4 km of exhumation occurred from 18–11 Ma (Giorgis et al., 2017).
648 Crustal shortening, development of a deep crustal root, and the deformation of the South American
649 margin (Cruz et al., 2007), likely produced the observed metamorphic grades and significant burial
650 depths of the Northern Range rocks (Fig. 9b). In sum, the lithological, structural, and thermal data of the
651 Northern Range are all consistent with a two-sided wedge model and collision mechanism.

652 The bedrock cooling models of Arkle et al. (2021) indicate that a significant amount of post-
653 collisional cooling and exhumation occurred in the western Northern Range from ~10–4.5 Ma, while
654 little cooling and exhumation occurred in the east during that same time period. Arkle et al. (2021)
655 further speculate that the STEP edge passed eastward under Trinidad following oblique collision (Fig.
656 9c), creating a series of late-stage, en echelon normal faults that accommodated the overall strike-slip
657 (transform) motion that is also pervasive throughout the Northern Rin Trinidad (Algar & Pindell, 1993;
658 De Verteuil et al., 2005) and eastern Venezuela (Cruz et al., 2007). We propose that the Chupara Fault
659 may be such a post-orogenic, extensional feature (Figs. 2, 7a, 9c). Thus, the activity of the Chupara
660 Fault may be linked to the exhumation of the deeply buried rock in the western Northern Range that we
661 document using RSCM data, perhaps as the STEP front migrated eastward (Fig. 9c). Thermal models,
662 erosion data, and other regional geomorphic data also indicate that differential cooling and exhumation
663 inverted around 4.5 Ma (Arkle et al., 2017: 2021). The STEP edge at this time (~4.5 Ma) would have
664 moved sufficiently far to the east, became insignificant in the Northern Range, and brought the activity
665 along the Chupara Fault to an end (Fig. 9d).

666 7. CONCLUSIONS

667 RSCM geothermometry provides important constraints on the peak metamorphic temperatures of
668 key lithologic units across the Northern Range and helps to fill vital gaps in our understanding of its
669 thermal history. The RSCM data collected across a broad swath of the range reveals a discrete thermal
670 gradient with peak metamorphic temperatures of 337 ± 10 °C in the east and 442 ± 16 °C in the west,
671 confirming the field gradient proposed by Frey et al. (1988) and demonstrated by Weber, Ferrill, et al.
672 (2001). The abrupt temperature discontinuity of 105 ± 19 °C located at Chupara Point corresponds to the
673 location of the previously mapped Chupara Fault. Further field work is required to better constrain the
674 precise location, timing, and kinematics of the Chupara Fault. Our new RSCM temperature estimates
675 surpass all previous estimates and thus require greater burial depths of Northern Range rock. We
676 speculate that Miocene oblique plate motion created a two-sided wedge in which parautochthonous
677 Northern Range rocks were incorporated into a fold-thrust hinterland wedge. Our new peak
678 temperatures, together with thermochronological data, lead us to interpret that the Chupara Fault is
679 likely the main structure that accommodated differential exhumation between the eastern and western
680 Northern Range. The distinctly different cooling trajectories suggest that the Chupara Fault was likely
681 active from ~10–4.5 Ma, though activity before or after this time period cannot be completely ruled out.

682 **Acknowledgments**

683 This study was inspired by a departmental field trip to the Caribbean, organized and led by
684 Professor Emeritus Edward Spooner and Laurent de Verteuil, along with undergraduate students Colin
685 Roth and Jason Hinde. The Raman analyses and training at the University of Turin were supported by an
686 Undergraduate Student Research Award of the Natural Sciences and Engineering Research Council
687 (NSERC USRA, reference number 501984) to I.G. and a MIUR grant Levi Montalcini to A.V.B. The
688 study is also supported by NSERC (Discovery Grant RGPIN-2018-03925 to X.C.). This work is part of
689 a project that has received funding from the European Research Council (ERC) under the European

690 Union's Horizon 2020 research and innovation programme (Grant agreement No. 864045). We thank
691 Veronica DiCecco and Yanan Liu for help with Raman spectroscopy and the electron microprobe
692 analysis, at the Royal Ontario Museum and the University of Toronto respectively. Sample collection by
693 Adam Brudner is gratefully acknowledged.

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Table 1. Rock Type, Location, and RSCM Temperatures of Northern Range Samples

Sample	Spectro- meter	Lithology	UTM X coordinate ^a	UTM Y coordinate ^a	5-Peak Spectra Temperature (SE) ^f		3-Peak Spectra Temperature (SE) ^g		RSCM Temperature (SE) ^e		FWHM-D1 Temperature (SE) ^h		FWHM-D2 Temperature (SE) ^h	
DT-93-1 ^b	-	Phyllite	672200	1189900	-	-	-	-	-	-	-	-	-	-
145	UNITO	Schist	672500	1178400			348	2.9	348	2.9	362	3.6	277	18.3
HT-161	ROM	Phyllite	672500	1178400	341	4.0	355	2.1	347	4.7	360	3.4	383	3.6
CF-94-1	ROM	Marble	672698	1178373			353	1.8	353	1.8	367	2.6	332	6.8
T94-3	ROM	Slate	723638	1198298	325	4.4			325	4.4	287	5.0	320	1.1
T97- MAT3	ROM	Slate	706235	1196714	339	2.9			339	2.9	298	2.1	322	0.9
M-3	UNITO	Schist	671697	1189674	<i>1 outlier</i>		461	4.7	461	4.7	-	-	-	-
	ROM	Schist	671697	1189674			447	3.6	447	3.6	-	-	-	-
M-4	UNITO	Schist	671697	1189674	<i>1 outlier</i>		497	7.9	497	7.9	-	-	-	-
	ROM	Schist	671697	1189674			492	10.6	492	10.6	-	-	-	-
Ar-17	ROM	Metasandstone	683682	1193264	364	3.9			364	3.9	334	3.5	310	3.6
LC-91a	UNITO	Metasandstone	675697	1192174			413	3.6	413	3.6	-	-	-	-
T90-9b	ROM	Phyllite	672820	1178826			340	1.4	340	1.4	354	3.1	298	7.1
LN-94-1	UNITO	Metacarbonate	721934	1197921	327	3.1			327	3.1	243	5.3	256	11.7
	ROM	Metacarbonate	721934	1197921	272	3.5			272	3.5	270	5.8	292	9.1
SC-2	UNITO	Marble	666223	1185430			465	8.6	465	8.6	-	-	-	-
T94-1	UNITO	Metacarbonate	723497	1198074	322	2.7			322	2.7	227	7.9	323	5.5
	ROM	Metacarbonate	723497	1198074	302	2.5			302	2.5	245	3.5	360	3.5
SC-2b	UNITO	Marble	669482	1189870	<i>1 outlier</i>		449	4.1	449	4.1	-	-	-	-
MV-94-3	ROM	Metacarbonate	717297	1197675	310	3.5			310	3.5	263	1.6	319	7.2
PG-94-5 ^c	ROM	Metacarbonate	650597	1180273			315 ^c		315^c		314 ^c		287 ^c	
93-T-CB	UNITO	Marble	679302	1194346	<i>3 outliers</i>		366	4.6	366	4.6	354	2.4	248	8.9
MAT-1	ROM	Slate	704866	1196982	341	4.1	<i>1 outlier</i>		341	4.1	292	3.2	320	3.7
CB-2a	UNITO	Marble	678698	1195474	348 ^d		344	1.2	344	1.2	349	2.1	233	10.9
	ROM	Marble	678698	1195474			341	1.4	341	1.4	366	2.3	302	5.8
DT-93	UNITO	Schist	672763	1189887			454	2.9	454	2.9	-	-	-	-
	ROM	Schist	672763	1189887			453	3.3	453	3.3	-	-	-	-
MB-93-1	ROM	Schist	650900	1187570			444	5.2	444	5.2	-	-	-	-
LC-91b	UNITO	Schist	676033	1192116			422	3.9	422	3.9	-	-	-	-
LC-3	ROM	Phyllite	661869	1180984			410	2.2	410	2.2	-	-	-	-
NR-1	UNITO	Metasandstone	720092	1185918			338	0.3	337	2.2	343	2.0	244	7.2
	ROM	Metasandstone	720092	1185918	336	4	341	1.1	342	1.1	355	1.5	287	3.8
NR-2 (-5)	UNITO	Slate	724652	1192884	351	1.5	340	***	350	4.5	311	1.9	311	4.1
NR-3 (-1)	UNITO	Metasandstone	720092	1185918	327	1.3	335	0.3	329	3.7	318	5.5	174	10.8

939 ^a UTM coordinates are in Zone 20.
940 ^b Sample DT-93-1 was unable to be analyzed due to a lack of carbonaceous material.
941 ^c Sample PG-94-5 has a temperature based on one spectrum as all other spectra collected contained large calcite peaks and could
942 not be fit accurately.
943 ^d One spectrum only.
944 ^e These temperatures are calculated using the Aoya et al. (2010) or Lahfid et al. (2010) calibrations. In samples displaying both 3-
945 and 5-band spectra both calibrations were used, with the reported value representing the average temperature of all spectra
946 associated with that sample (exculding outliers). Consult the supplementary material for a full breakdown of the calibrations used
947 for each spectrum.
948 ^f Aoya et al. (2010).
949 ^g Lahfid et al. (2010).
950 ^h Kouketsu et al. (2014).

Figure Captions

Figure 1: a) Schematic tectonic map of the eastern Caribbean showing major plate boundaries. Trinidad (grey) is currently located on the right-lateral strike-slip transform boundary along the southern edge of the Caribbean Plate. PC – Paria cluster of seismicity. b) Map of Trinidad emphasizing the Northern Range study area (grey) and major faults in the region. The active faults (Central Range and Los Bajos faults, bold lines) accommodate most of the present plate motion between the Caribbean and South America (Weber, Dixon, et al., 2001). The Northern Range is bound by the inactive North Coast and Sub-Tobago Terrane fault zones to the north and Arima Fault to the south. The El Pilar Fault zone is active in eastern Venezuela but is inactive or not present in Trinidad. c) Cross section of the subduction-transform edge propagator (STEP) plate boundary and associated lithospheric tear located below the Caribbean Plate. The subduction-transform boundary is located at the letter “S” while the site of active and deep tearing is located at the Paia Cluster, which is a major zone of earthquakes in the area as a result of the slab tear (Russo & Speed, 1992). Modified after Arkle et al. (2021).

Figure 2: Simplified geologic map from De Verteuil et al. (2005) of the Northern Range showing metamorphic units, sample localities, and relict fossil ages. Apart from the Sans Souci Metabasalt, all lithologic units are metamorphosed from sedimentary protoliths (Frey et al., 1988), with relict fossil ages only available from eastern block, lower-grade units (Dighton-Thomas, 1935; Imlay, 1954; Kugler, 1961; Saunders, 1972; Spath, 1939; Trechmann, 1935). A total of 26 RSCM samples were collected and grouped into six representative rock types. Sample distribution was based on obtaining a broad range of compositions and textures. PS – Port of Spain; CP – Chupara Point; VV – Verdant Vale; M – Maturita; T – Toco.

Figure 3: Photomicrographs of petrographic thin sections displaying representative features of each rock type. a) calcite (Cal) marble with carbonaceous material (Gr) concentrated into dark bands; b) calcite (Cal) vein in a metacarbonate rock indicating late extensional brittle deformation and infilling via precipitation from a CO₂-rich fluid; c) low-grade metacarbonate marble displaying primary sedimentary structures (ooids) which have been moderately strained by ductile deformation but have not been recrystallized; d) quartz-mica (Qz-Ms) schist with foliation and minor calcite, likely representing a sandstone protolith; e) mica-rich (Ms) phyllite displaying crenulation after multiple deformation events; f) radial growth of carbonaceous material (CM) superimposed on schistosity; g) metasandstone with minor chlorite (Chl), muscovite, and quartz (Qz); h) folded quartz (Qz)- and CM (Gr)-rich layers in a metasandstone; i) slate rich in carbonaceous material, quartz (Qz), and mica (Ms). PPL: plane-polarized light, XPL: cross-polarized light, RL: reflected light.

Figure 4: a) Backscattered electron (BSE) image of a foliated slate. The dolomite (dol) in the sample has a euhedral shape whereas the calcite (cc) appears more subhedral-anhedral, indicating a lack of equilibrium between the two phases. Pyrite (py) framboids are also present. b) CaCO₃-MgCO₃-FeCO₃

phase diagrams calculated for different temperatures (Anovitz & Essene, 1987), with carbonate compositions plotted in red.

Figure 5: Peak-fitted Raman spectra of (a) high- and (b) low- temperature samples. The high-temperature spectra are fit with Voigt bands while Lorentzian bands are used for the low-temperature spectrum; the area ratios between these bands are used to calculate the peak metamorphic temperature of a sample (Aoya et al., 2010; Lahfid et al., 2010). c) Raman spectra showing decreasing band complexity, correlating to increasing crystallinity and peak metamorphic temperature, from bottom to top. The temperatures shown are calculated using the ¹Lahfid et al. (2010) and ²Aoya et al. (2010) calibrations.

Figure 6: Inter-instrument error between Raman spectrometers at UNITO and the ROM. The low-temperature calibrations of Lahfid et al. (2010) and Kouketsu et al. (2014) show a lower reproducibility than the high-temperature calibration of Aoya et al. (2010). Inset: variability between the two calibrations of Kouketsu et al. (2014) which are up to 100 °C.

Figure 7: a) RSCM results given in °C plotted on a geologic map of Trinidad overlying topographic (radar) data. The abrupt temperature change in the central Northern Range is marked by a major lithologic boundary along the previously mapped Chupara Fault (Kugler, 1961; De Verteuil et al., 2005). The temperatures shown are calculated using the Aoya et al. (2010) and Lahfid et al. (2010) calibrations. b) Temperature (± 1 s.e.) for samples along the north coast of the Northern Range calculated using a variety of RSCM calibrations. Lower temperature samples from the southwestern Chancellor Schist are excluded as they are separated from the higher temperature samples by the E-W trending Arima Fault. Average temperatures (1σ) of the eastern and western blocks are displayed as the blue-shaded regions.

Figure 8: Cooling history of the a) eastern and b) western Northern Range modified from the QTQt bedrock cooling models of Arkle et al. (2021). Timing of peak metamorphism is estimated by ⁴⁰Ar/³⁹Ar ages through the Northern Range (Speed et al., 1997, Weber & Arkle, 2015). In the east, the cooling path following peak metamorphic temperature, and possibly its timing, remains largely unconstrained due to the disparity between un-reset ZFT ages and high RSCM temperatures, potentially indicating a short-lived thermal event. Dark and light grey lines are the expected cooling histories of the high and low elevation samples, respectively, with the shaded regions representing a 95% confidence interval (Arkle et al., 2021).

Figure 9: Cartoon showing tectonic evolution of the Northern Range and Araya-Paria, Venezuela, metamorphic hinterland, modified from Arkle et al. (2021). a) Following Jurassic-Cretaceous rifting, the area was dominated by passive margin deposition of sediment along continental South America. b) Oblique plate collision led to peak metamorphism and thrust faulting in the Northern Range in the late Oligocene. This time marks the inception of foreland and hinterland structures and of hinterland

1028 metamorphism. c) A transition to strike-slip plate motion between the Caribbean and South American
1029 plates produced structures such as the Gulf of Paria pull-apart basin. Eastward propagation of the STEP
1030 edge induced greater exhumation of the western Northern Range upon its arrival in Trinidad, which was
1031 accommodated by the Chupara Fault. d) Continued STEP propagation along the plate boundary led to an
1032 inversion in Northern Range exhumation and late Pliocene to recent surface uplift, with subsidence
1033 currently dominant in the west as the east experiences surface uplift (e.g., Arkle et al. 2017, 2021). NR –
1034 Northern Range; SdI – Serranía del Interior; MB – Maturin Basin; CR – Central Range; SR – Southern
1035 Range; CF – Chupara Fault; NCFZ – Northern Coastal fault zone; EPFZ – El Pilar Fault zone; CB –
1036 Columbus Basin.

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1038 **Figure S1:** Plots of nominal temperatures calculated using 3- (Aoya et al., 2010; square symbols) and 5-
1039 band (Lahfid et al., 2010; plus signs) calibrations against the centers of fitted G bands. The G band of
1040 fully ordered graphite (1582 cm^{-1} ; Wang et al., 1990) is labeled by dashed lines for reference. The
1041 analyses at UNITO and ROM are denoted by black and blue colors, respectively.

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Figure 1

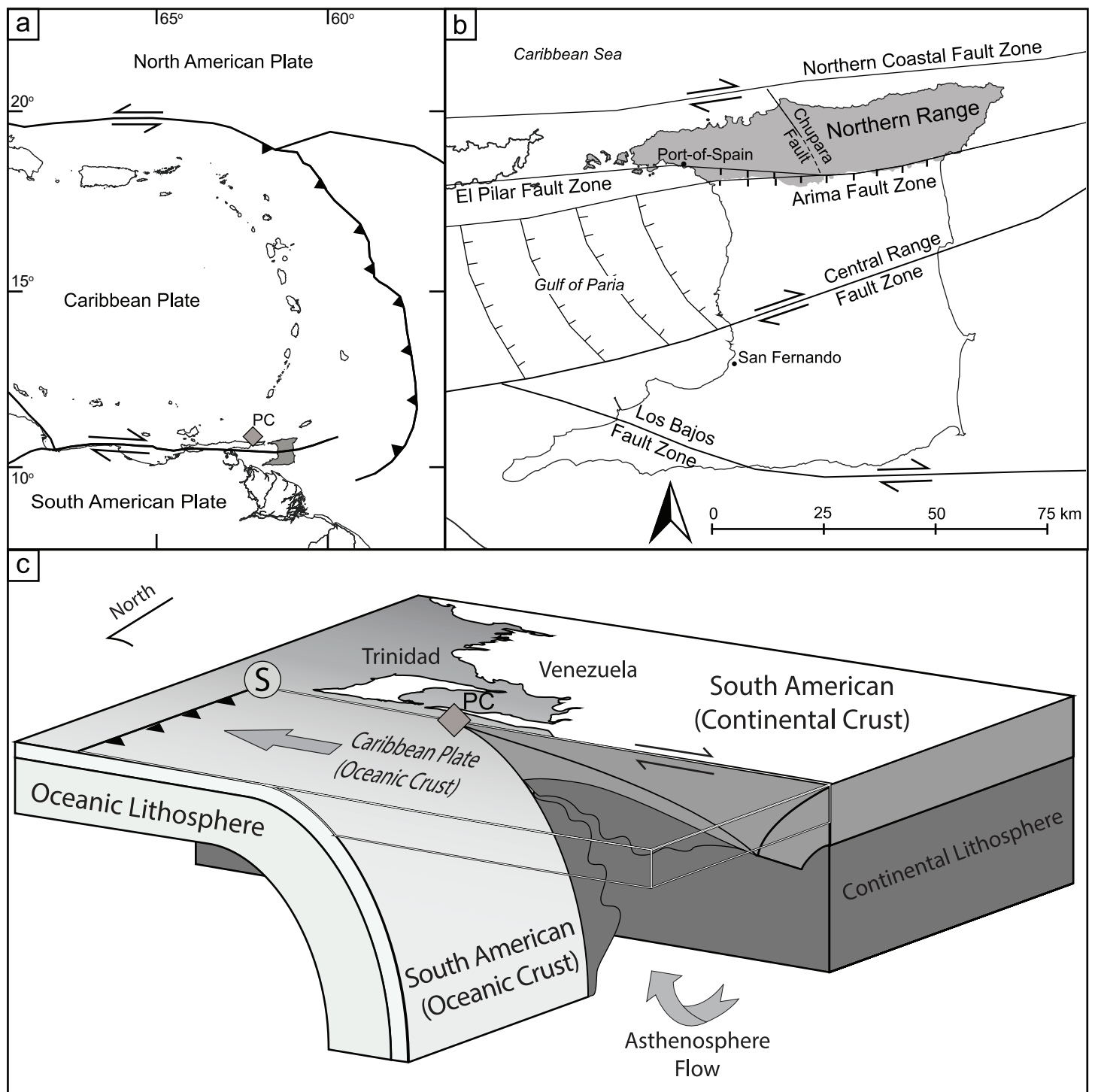


Figure 2

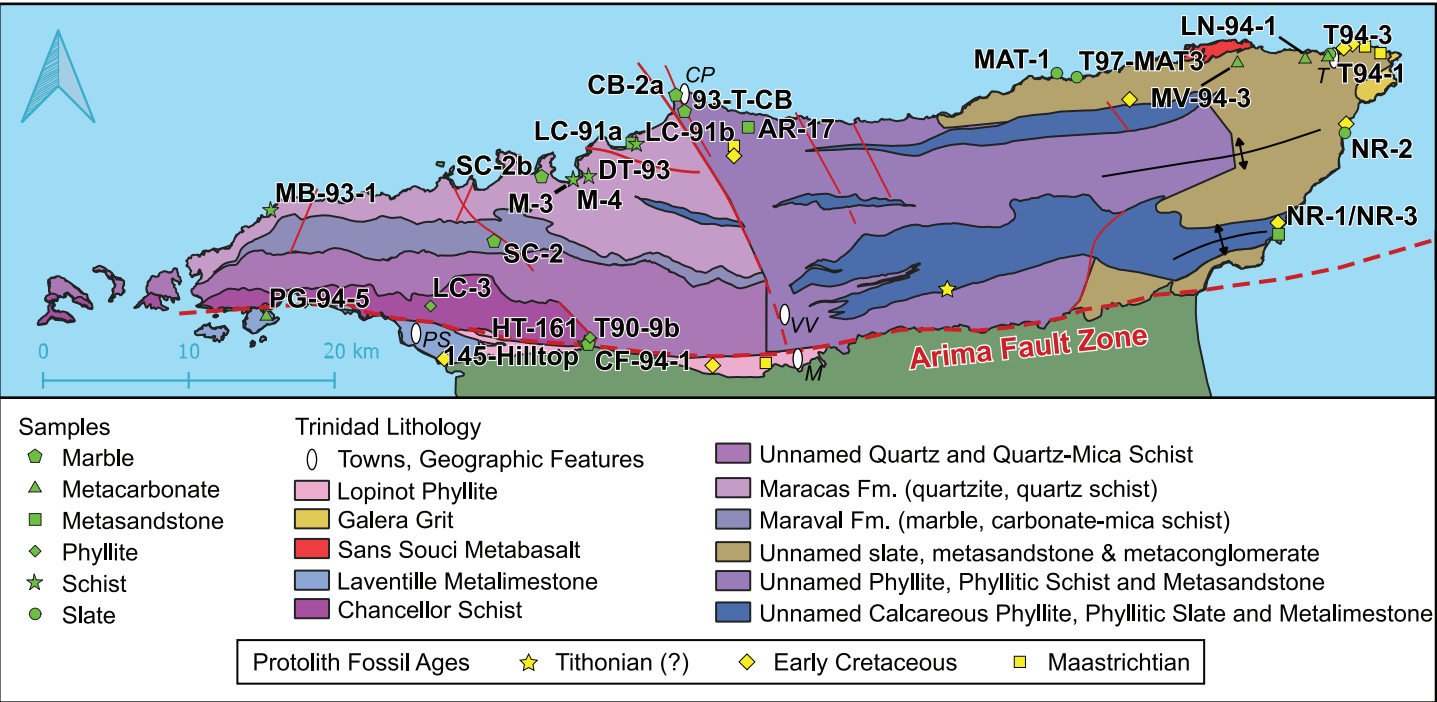


Figure 3

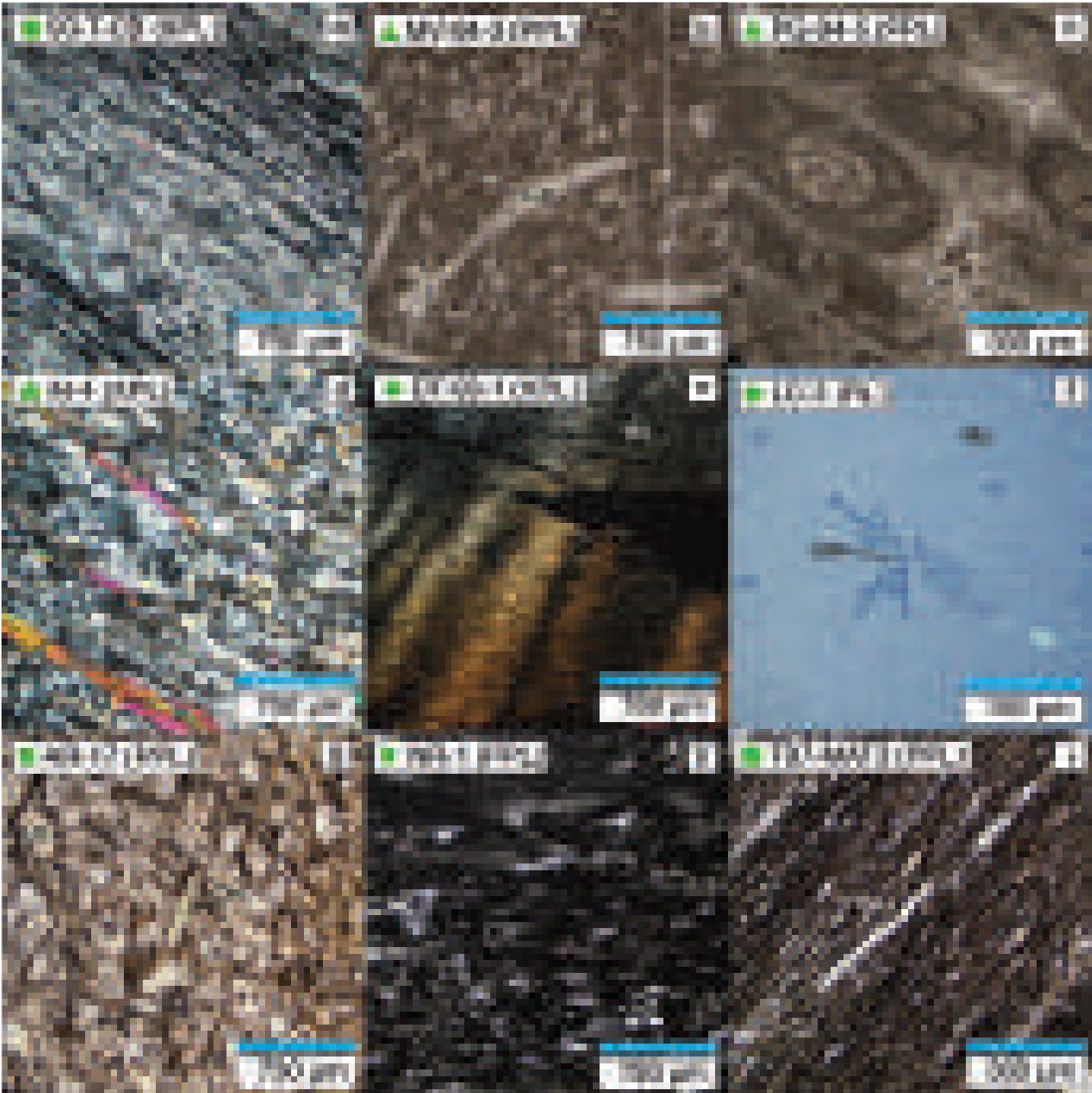


Figure 4

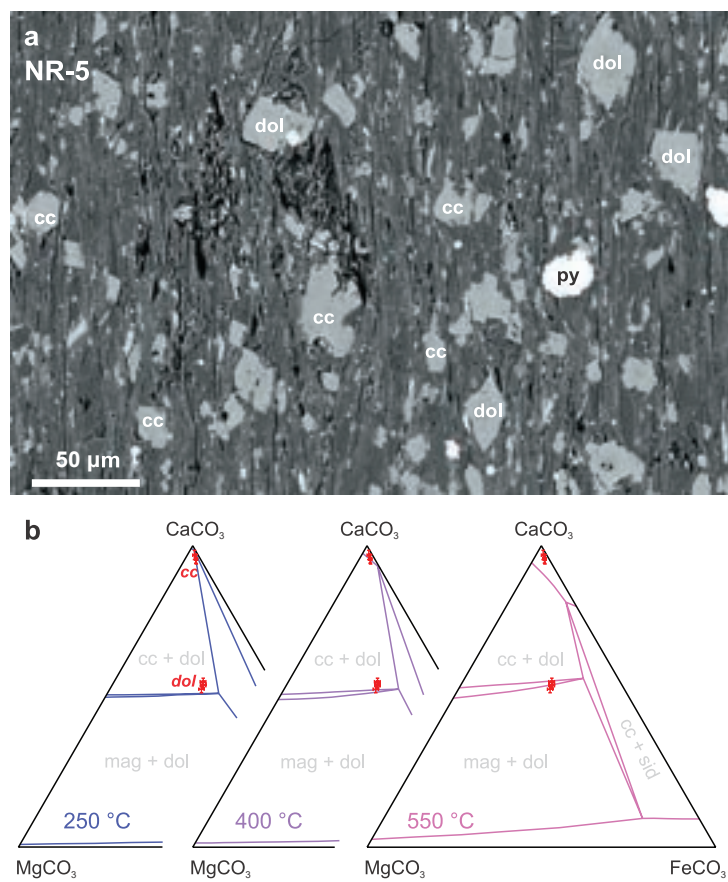


Figure 5

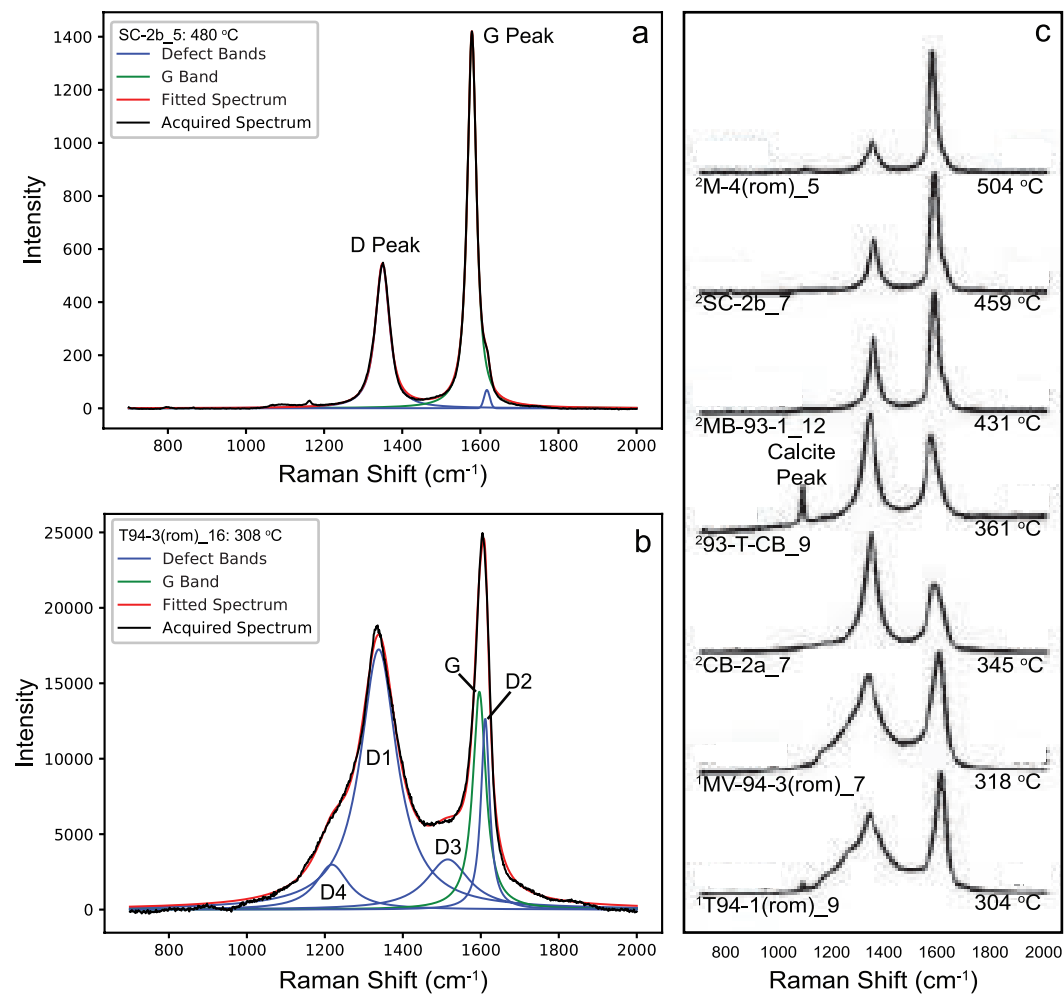


Figure 6

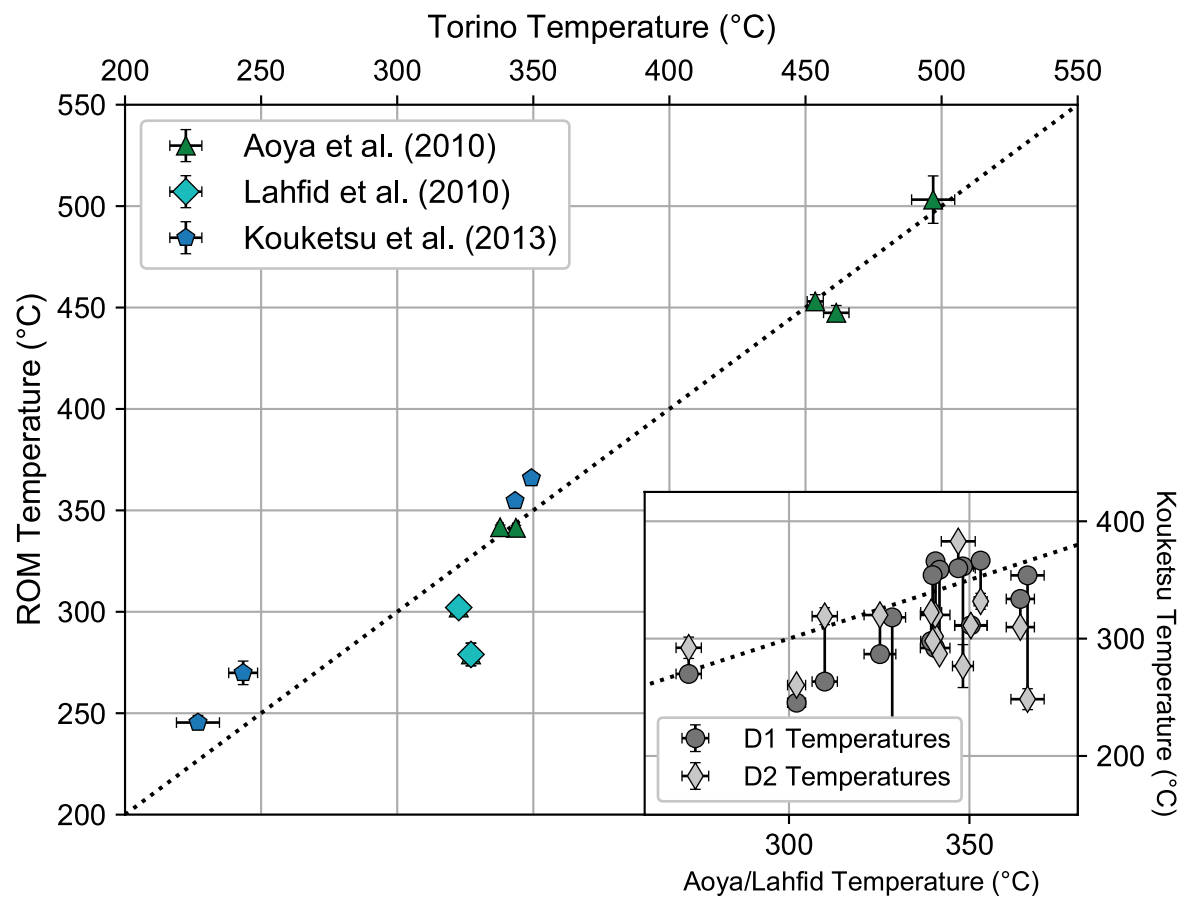


Figure 7

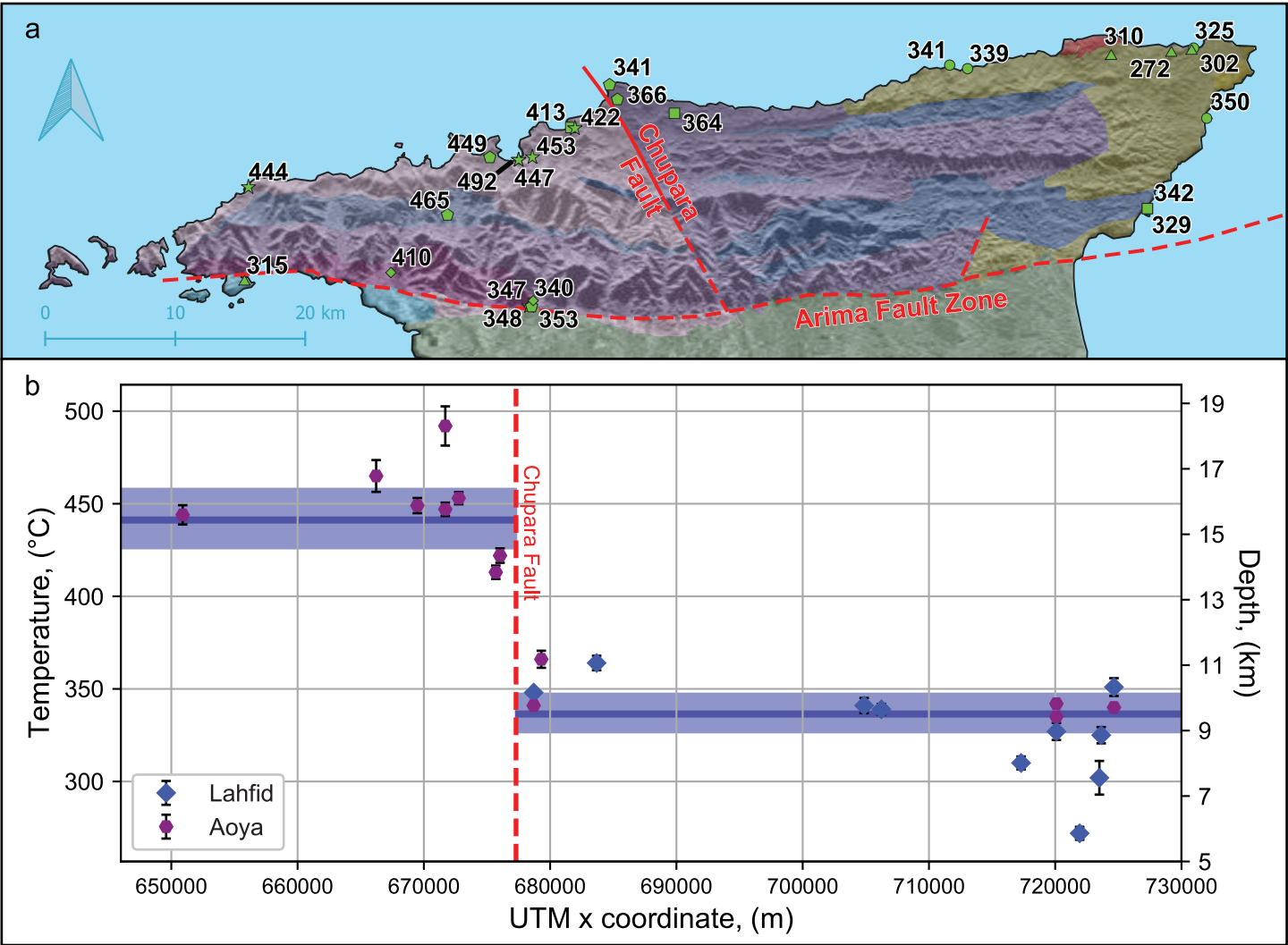


Figure 8

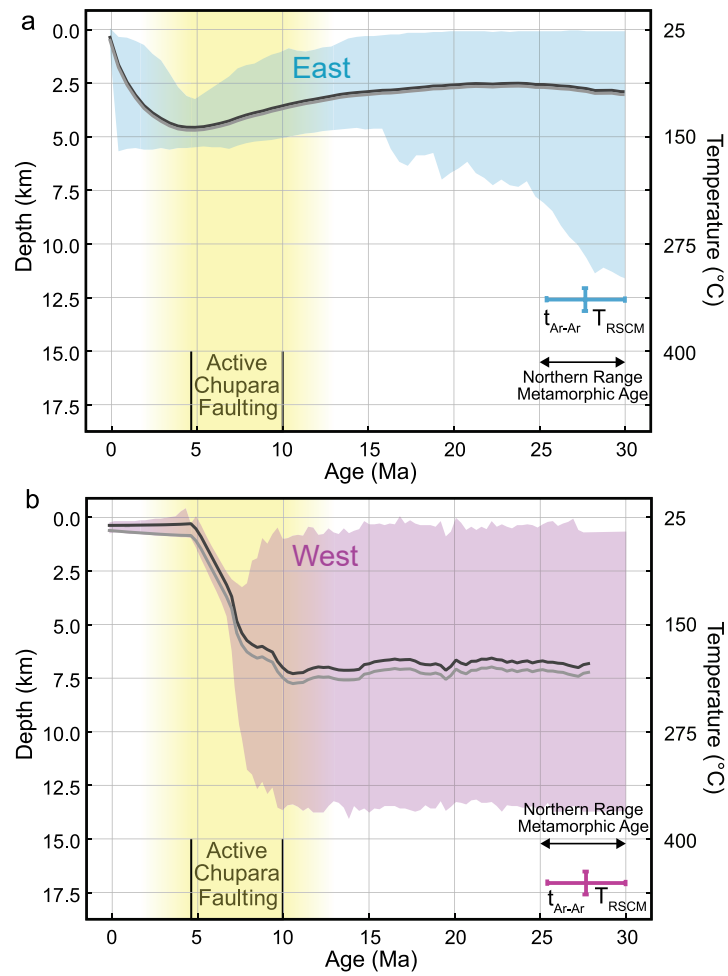


Figure 9

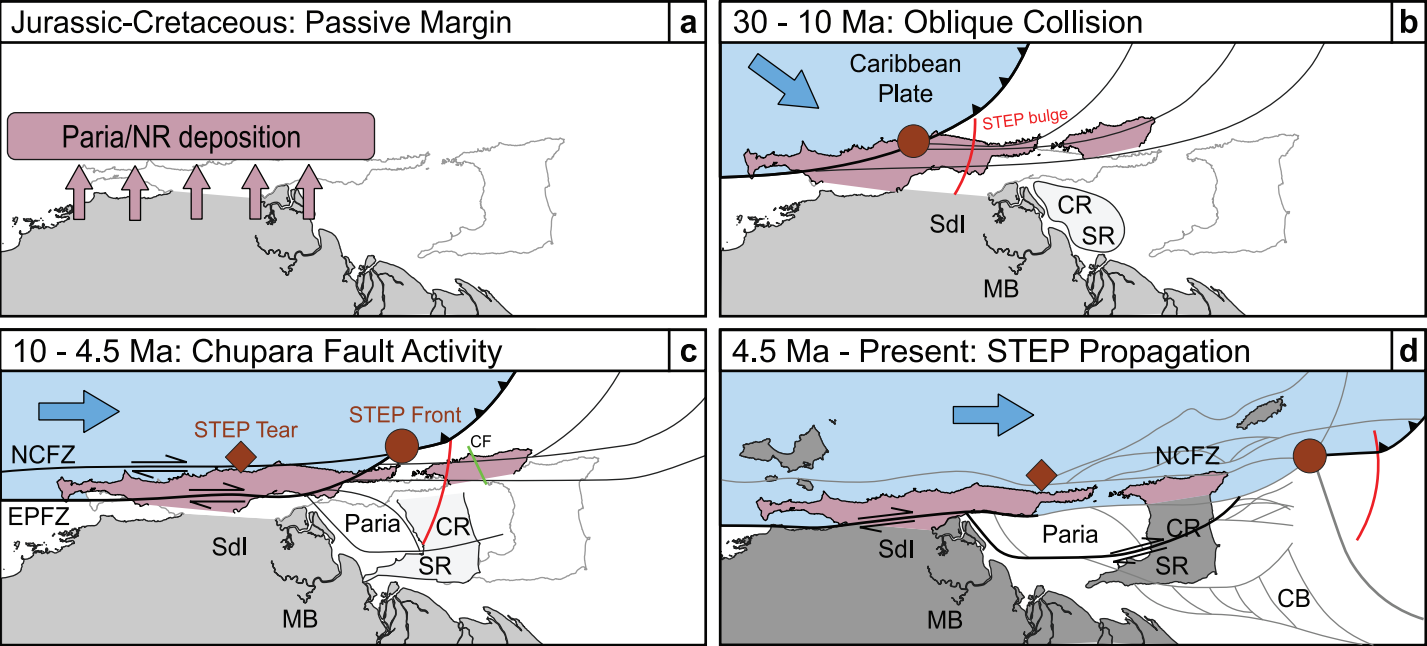


Figure S1

