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25 **Abstract**

26 We present new insights on the geometry, initiation and growth of the Thyolo fault,
27 an 85 km long active border fault in the southern Malawi Rift, from high-resolution
28 topography, field and microstructural observations. The Thyolo fault is located
29 towards the edge of the Proterozoic Unango Terrane, and is the border fault of the
30 Lower Shire Graben, which has experienced four phases of extension since the
31 Jurassic. Recent activity is demonstrated by an 18.6 ± 7.7 m high fault scarp, with
32 two substantial reductions in scarp height along strike. However, the segment
33 boundaries suggested by these displacement measurements do not coincide with
34 changes in fault strike. Elsewhere, a ~5 km long fault perpendicular scarp joins two
35 overlapping sections, yet the scarp height in this linking section is similar to the
36 bounding sections, and there is no evidence of significant pre-linkage strain
37 accumulation. Microstructural analyses along the fault show a 15-45 m thick footwall
38 damage zone with a 0.7 m thick core. We suggest that favourably-oriented, pre-
39 existing shallow structures control changes in surface geometry and the narrow fault
40 core, whereas exploitation of weak ductile zones at depth, possibly associated with
41 the terrane boundary, control the displacement profile of the fault.

42

43 1. Introduction

44 Narrow amagmatic rifts (*sensu* Buck, 1991) are typically characterised by a ≤ 100 km
45 wide graben or half graben where the greatest cumulative displacement is
46 accommodated on large offset normal fault systems, known as rift border faults, that
47 bound a region of distributed but relatively small displacement brittle deformation
48 (Ebinger, 1989; Gawthorpe and Leeder, 2000; Muirhead et al., 2019). These basin-
49 bounding faults are thought to be most active prior to any magmatic influence on
50 rifting (Ebinger, 2005; Muirhead et al., 2019), have a distinctive impact on basin
51 geomorphology (Leeder and Gawthorpe, 1987; Gawthorpe and Leeder, 2000) and
52 can accumulate sufficient displacement to induce flexural bending within the hanging
53 wall basin (Turcotte and Schubert, 2002). Border faults can penetrate the entire

54 depth of the crust, and in East Africa are probably the source of some of the deep
55 earthquakes that occur within the 30-40 km thick seismogenic layer (Jackson and
56 Blenkinsop, 1993; Craig et al., 2011; Lavayssière et al., 2019).

57

58 How faults grow from nucleation to crustal scale features is a long-standing topic of
59 research (Cowie and Scholz, 1992b; Cowie, 1998; Walsh et al., 2002; Nicol et al.,
60 2005; Worthington and Walsh, 2017; Rotevatn et al., 2019), and numerous studies
61 have mapped fault trace geometry and measured displacement-length profiles to
62 discuss the mechanism and timing of how long faults develop through segment
63 initiation, growth, and linkage (Cowie and Scholz, 1992a, 1992b, 1992c; Scholz et
64 al., 1993; Dawers et al., 1993; Dawers and Anders, 1995; Schlische et al., 1996;
65 Walsh et al., 2003; Nicol et al., 2005, 2017; Giba et al., 2012; Rotevatn et al., 2018).

66 Structural heterogeneities at segment boundaries, such as fault bends and step-
67 overs, are thought to influence the propagation and termination of earthquake
68 ruptures (Segall and Pollard, 1980; Zhang et al., 1991; Wesnousky, 2006, 2008).

69 However, recent earthquakes (e.g. 2010 M_w 7.2 El Mayor-Cucapah Earthquake,
70 Mexico – Wei et al., 2011 and the 2016 M_w 7.8 Kaikoura Earthquake, New Zealand –
71 Hamling et al., 2017) have propagated across multiple segment boundaries, making
72 it unclear how to best assess fault segmentation for seismic hazard purposes
73 (DuRoss et al., 2016). Border faults are now generally thought to develop through
74 the accumulation of displacement on fault segments that formed and linked during
75 the early stages of rifting (Gawthorpe et al., 2003; Rotevatn et al., 2019; Muirhead et
76 al., 2019); however, the effects of this linkage on the displacement profile and
77 surface trace of a fault are commonly long-lasting (McLeod et al., 2008). Minima in
78 fault displacement profiles are persistently observed at segment boundaries

79 (Machette et al., 1991; Gupta and Scholz, 2000; Mortimer et al., 2007, 2016) as are
80 relay ramps, increased fault complexity, step-overs, and changes in fault orientation
81 (Leeder and Gawthorpe, 1987; Crone and Haller, 1991a; Peacock and Sanderson,
82 1991; Crider and Pollard, 1998; McLeod et al., 2008; Fossen and Rotevatn, 2016;
83 Hodge et al., 2018a). Thus, observations of fault segmentation provide a permanent
84 record of processes that occurred during the formation and linkage of fault
85 segments, and consequently they offer insights into the fundamental processes of
86 fault growth and the controls on the limits of earthquake rupture propagation.

87

88 Rifts rarely initiate and grow in isotropic crust, and therefore it is important to
89 understand the effect of pre-existing heterogeneities and structures on the growth
90 and segmentation of faults. These pre-existing structures are often cited as the
91 predominant control on rift geometry, the distribution of strain within rifts, and the
92 orientation of magmatic bodies, magmatic rift segments and faults within rifts
93 (McConnell, 1967; Daly et al., 1989; Ebinger et al., 1997; Morley, 2010; Henstra et
94 al., 2015; Robertson et al., 2016; Muirhead and Kattenhorn, 2018). At the scale of an
95 individual fault, analogue models have shown that reactivation of pre-existing fabrics
96 affects fault growth by influencing fault orientation, relay zone geometry and the
97 distribution of basins (Bellahsen and Daniel, 2005; Henza et al., 2011). However,
98 comparisons of these models with natural faults is challenging as it can be difficult to
99 differentiate between contemporary and pre-rift heterogeneities that have similar
100 geometries (Smith and Mosley, 1993; Holdsworth et al., 1997), especially using
101 seismic reflection and aeromagnetic surveys, which can only resolve features at
102 scales >10 m.

103

104 Investigating the interactions between pre-existing fabrics and strain localisation on
105 rift border faults also requires understanding the structure and mechanics of these
106 faults. In general, as faults grow, the rocks surrounding the fault accumulate damage
107 (Cowie and Scholz, 1992b; Caine et al., 1996; Shipton and Cowie, 2003). However,
108 the structure of a rift border fault has only been described in a limited number of
109 cases (Ord et al., 1988; Wheeler and Karson, 1989; Kristensen et al., 2016;
110 Hollinsworth et al., 2019), with most models of normal fault structural evolution based
111 on studies of small displacement (<100 m) faults within high porosity sedimentary
112 rock (Shipton and Cowie, 2003; Childs et al., 2009; Torabi and Berg, 2011; Savage
113 and Brodsky, 2011). Consequently, it remains unclear whether these models are
114 applicable to large offset rift border faults where the footwall is composed of foliated
115 crystalline metamorphic rocks.

116

117 In this paper, we analyse the Thyolo fault, the border fault of the Lower Shire Graben
118 in southern Malawi (Figure 1). The fault is an ideal location to study the effects of
119 pre-existing structures on fault geometry and structure as the graben has a well-
120 documented polyphase history of extension (Castaing, 1991; Chisenga et al., 2019).
121 Furthermore, thin syn-rift sediments mean that fault exposures formed during the
122 current rift phase are not hidden by post-rift sediments (e.g. Hodge et al., 2019;
123 Williams et al., 2019). We begin by describing the tectonic history of the region,
124 before analysing the current activity, geometry and structure of the fault. In doing so,
125 we assess how reactivation of pre-existing fabrics and heterogeneities at different
126 scales affects the evolution of a rift border fault.

127

128 2. Tectonic History

129 The Thyolo fault bounds the north-eastern edge of the Lower Shire Graben, which is
130 located at the southern end of the largely amagmatic Western branch of the East
131 African Rift (EAR; Figure 1). Extension within the Western branch of the EAR
132 initiated ~25 Ma (Roberts et al., 2012) and within southern Malawi, the current
133 horizontal geodetic extension rate is ~2 mm yr⁻¹ (relative to the stable Nubian plate;
134 Stamps et al., 2018; Figure 1). It is not known whether this extension rate has been
135 constant during the current phase of rifting (the last ~25 Ma).

136

137 The footwall of the Thyolo fault is composed of charnockitic gneiss and granitic
138 granulites of the Mesoproterozoic Unango Terrane, part of the Mozambique Belt,
139 with the fault located towards the southwestern edge of the terrane (Figure 2;
140 Fullgraf et al. *in press*; Bloomfield, 1965; Johnson et al., 2005). The Unango Terrane
141 likely formed in a continental volcanic arc setting at ~1 Ga, and experienced
142 granulite facies metamorphism associated with ductile deformation shortly after
143 emplacement (957 ± 27 Ma; Bingen et al., 2009). The metamorphic foliation and
144 migmatitic banding in the Thyolo fault's footwall dips moderately SW. The foliation
145 formed at granulite facies metamorphic conditions coincident with partial melting
146 during a series of collisional events at a convergent continental margin in the Pan-
147 African Orogeny (~710-555 Ma) associated with the amalgamation of Gondwana
148 (Kröner et al., 2001; Johnson et al., 2005; Manda et al., 2019). In the region of the
149 Thyolo fault, the edge of the Unango Terrane is in contact with the basement of the
150 South Irumide Belt, which underwent peak metamorphism between 1.06 and 1.05
151 Ga (Johnson et al., 2005; Westerhof et al., 2008; Karmakar and Schenk, 2016). The
152 terrane boundaries have been roughly mapped based on exposures within Malawi
153 (Manda et al., 2019), but because younger cover sequences (~330-180 Ma Karoo-

154 age sedimentary rocks) have obscured the basement, the unit boundaries are largely
155 extrapolated from neighbouring Mozambique, where mapping was supported by
156 geochemical and airborne magnetic data (Bingen et al., 2009; Macey et al., 2010).

157

158 2.1 Previous phases of rifting

159 The Lower Shire Graben contains Phanerozoic sedimentary and volcanic deposits
160 related to three regional phases of extension that occurred prior to the current rifting
161 (i.e. prior to ~25 Ma): two distinct events during the Karoo-age breakup of Gondwana
162 (~330-180 Ma), and a later phase at the end of the Jurassic and into the early
163 Cretaceous (~140-110 Ma; Castaing, 1991; Figure 2).

164

165 NW-SE Karoo-age transtension in the Lower Shire Graben created space to deposit
166 a sequence of Late Ecca (Middle Permian) to Late Beaufort (Early Triassic) coal
167 shales, coarse grained grits, mudstones and sandstones (orange units in Figure 2c).
168 These sedimentary rocks are bound by E-W striking normal faults and NW-SE
169 striking dextral strike-slip faults including the Mwanza and possibly the Thyolo fault
170 (Figure 2c; Habgood, 1963; Habgood et al., 1973; Castaing, 1991).

171

172 NW-SE extension continued into the late Karoo period, when it was associated with
173 basaltic volcanism and contemporaneous emplacement of NE-SW striking dolerite
174 dykes. These dykes and volcanic deposits are collectively known as the Stormberg
175 Volcanics, which are widely observed in the footwalls of the Mwanza, Thyolo and
176 Mtumba faults (Figure 2d; Habgood, 1963; Habgood et al., 1973; Woolley et al.,
177 1979; Castaing, 1991).

178

179 After the Karoo period, during the Late Jurassic – Cretaceous, the extension
180 direction rotated from NW-SE to NE-SW and reactivated NW-SE structures
181 established in the earlier phase of NW-SE transtension as dip-slip normal faults
182 (Figure 2e; Castaing, 1991). In the Lower Shire Graben, remnant deposits from the
183 NE-SW extension are limited to sandstones in the hanging wall of the Panga and
184 Chitumba faults (Figure 2e) and siliceous fault rock along the Namalambo Fault
185 (Habgood, 1963; Habgood et al., 1973). These sedimentary deposits form part of the
186 Lupata series, a mix of coarse grained sandstones, and rhyolitic and alkaline lavas
187 found extensively in Mozambique (Dixey and Campbell Smith, 1929; Habgood,
188 1963), and were emplaced contemporaneously with the Chilwa Alkaline Province,
189 which involves intrusive rocks that crosscut the Stormberg dykes (Macdonald et al.,
190 1983; Woolley, 1987; Castaing, 1991; Eby et al., 1995). Cretaceous activity on the
191 Thyolo and/or Mwanza faults cannot be ruled out as current syn-rift sediments will
192 likely have buried any Cretaceous sedimentary deposits.

193

194 2.2 Current rifting

195 Previous studies have interpreted the Thyolo fault as an active, reactivated dextral
196 strike-slip fault linking the Urema Graben (the southern active continuation of the
197 EARS in Mozambique) and the Zomba Graben (Castaing, 1991; Chorowicz and
198 Sorlien, 1992; Chorowicz, 2005). In other studies, the Thyolo fault is considered
199 inactive (Laõ-Dávila et al., 2015; Prater et al., 2016). Hodge et al. (2019) combined
200 remote sensing observations with field observations to show that the fault is active,
201 by documenting a pseudo-continuous fault scarp and triangular facets at the
202 southern end of the fault. However, the fault was divided into two separate faults, the
203 Thyolo and Muona faults, based on the separation distance between their scarps

204 (Hodge et al., 2019). A M_w 5.6 earthquake in March 2018 had a normal faulting focal
205 mechanism with nodal planes aligned with the surface traces of faults in the Lower
206 Shire Graben (Figure 1; Ekström et al., 2012). Williams et al. (2019) suggest that the
207 Thyolo fault is currently active as a dip-slip normal fault oriented obliquely to the
208 ENE-WSW regional extension direction. Below, we describe in detail the dimensions
209 and geometry of the fault scarp, and the fault damage zone along the Thyolo fault,
210 and analyse the factors that control fault segmentation, orientation and structure.

211

212 3. Methodology

213 3.1 Fault segmentation analysis

214 We used a high resolution 12 m TanDEM-X digital elevation model (DEM) to identify
215 different indicators of fault segmentation based on two distinct sets of criteria: map-
216 view geometry and scarp height. We mapped the fault trace in high resolution and
217 noted prominent changes in fault strike, map-view fault steps, fault branches,
218 increased fault complexity or gaps, which are considered geometrical indicators of
219 fault segmentation (Zhang et al., 1991; Wesnousky, 2008; DuRoss et al., 2016).

220

221 Fault segmentation can also be defined from the along-fault displacement profile
222 (e.g. Dawers and Anders, 1995; Willemse et al., 1996; Willemse, 1997; Walsh et al.,
223 2003). In a plot of fault displacement vs. fault-parallel distance, the segment
224 boundaries are usually located at local displacement minima (Crone and Haller,
225 1991a; Dawers and Anders, 1995; Walsh et al., 2003). We used scarp height as a
226 proxy for displacement (e.g. Morewood and Roberts, 2000; Hodge et al., 2018b,
227 2019; Wedmore et al., 2020) and identified segments based on local minima in the
228 along-strike scarp height profile. We used adapted versions of the SPARTA scarp

229 measuring tools (Hodge et al., 2019) to measure the height of the fault scarp along
230 the Thyolo fault on the 12 m DEM. We differ from Hodge et al. (2019) by extracting
231 500 m long fault-perpendicular topographic profiles from the DEM every 12 m along
232 the fault, which were averaged by stacking at 100 m intervals before measuring the
233 vertical difference between regression lines on the footwall and hanging wall
234 surfaces. We estimated the uncertainty of each measurement by applying a Monte
235 Carlo approach to sample 10,000 random subsets of points from the hanging wall
236 and footwall of the fault as well as allowing the location of the fault to vary along the
237 section of the topographic profiles identified as the fault scarp. The resulting
238 measurements of vertical offset were then filtered using a 5 km wide moving median
239 filter along the strike of the fault.

240

241 We also examined the Thyolo fault for any evidence of features associated with fault
242 linkage. Where two un-linked fault segments interact, fractures and faults form such
243 that the faults maintain laterally constant extensional strain (Walsh and Watterson,
244 1991). These features create relay zones or transfer structures, which evolve as the
245 faults overlap and link. Typically relay structures have 10-15° dipping ramps and
246 breach structures that link the segments (Fossen and Rotevatn, 2016). Before a
247 linking fault has breached a relay ramp, propagating and overlapping fault tips
248 induces rotation (about a vertical axis) within the transfer structure (Fossen and
249 Rotevatn, 2016). To identify linking structures, we analysed variations in the strike of
250 the fault by dividing the fault trace into 50 m long sections and measuring the strike
251 of each section from the trend of the surface trace, assuming negligible topography.
252 The orientation of pre-existing basement structures were analysed by digitising the

253 3D foliation measurements and strike of dolerite (Stormberg) dykes in Habgood et al.
254 (1973).

255

256 The fault scarp and geometry of some of the structures that we observed in the
257 remote sensing data and geological maps were ground-truthed during field
258 campaigns in 2017 and 2018, during which additional observations were made
259 regarding the structure and slip sense of the fault.

260

261 3.2 Fault zone meso- and microstructural analysis

262 We made lithological and structural observations of the footwall damage zone of the
263 Thyolo fault along four transects, Kalulu, Kanjedza, Mbewe, and Muona (Figure 3),
264 which extended from the fault scarp to distances up to 280 m into the footwall. In
265 addition, we collected samples at Kalulu ($n = 5$) and Kanjedza ($n = 11$) for
266 microstructural and compositional analyses. Individual samples were accurately
267 located using digital elevation models and orthophotos constructed from drone-
268 based photogrammetry, as well as handheld GPS measurements. Sample thin
269 sections were photographed in plane-polarised and cross-polarised light, and
270 fracture density was calculated from manually traced fractures in three 10-15 mm²
271 areas in each sample using FracPaQ 2.2 (Healy et al., 2017). We only traced
272 fractures in quartz or feldspar grains, to allow comparison between lithologically
273 diverse samples, and used an area-weighted average as the fracture density value
274 for each thin section (see supplementary text for full details of sample collection,
275 preparation and analysis).

276

277 4. Results

278 4.1 The Thyolo Fault

279 The Thyolo fault has a recently active fault scarp at the base of a ~1 km high footwall
280 escarpment (Figure 3 & 4). Note that we consider the previously described Thyolo
281 and Muona faults as a singular 'Thyolo fault', rather than two separate structures (in
282 contrast to Hodge et al., 2019) because the Chisumbi link section connects the
283 Thyolo and Muona sections of the fault without a break in the scarp (Figures 3-5 and
284 Section 4.4). Triangular facets are visible within the high-resolution topography along
285 the southeastern end of the Thyolo section and the northwestern end of the Muona
286 section (Figure 3), and were also observed in the field (Figure 4). We observed no
287 systematic deflection of river channels or any other geomorphological features that
288 might indicate strike-slip faulting (Figure S1), and we therefore consider the Thyolo
289 fault to be currently accommodating pure normal dip-slip displacement (see also
290 Williams et al., 2019).

291

292 4.2 Map View Geometry

293 The Thyolo fault is ~85 km long and has a mean strike of $139 \pm 15^\circ$ (1 standard
294 deviation) dipping to the southwest (Figure 3 & 4). A fault scarp is visible in the high-
295 resolution topography along the length of the fault, with gaps observed where ≤ 100
296 m wide rivers cross the fault and have eroded the scarp (Figure 3b). However, the
297 effect of rivers on the observations in Figure 5 is negligible as we did not measure
298 scarps in these locations, and the final scarp height is averaged over 5 km along
299 strike to smooth short-wavelength features such as these rivers. We find seven
300 sections along the fault trace that trend approximately perpendicular to the average
301 fault strike (Figure 5c). These NE-SW oriented sections have a mean strike of $044 \pm$
302 28° (black lines in Figure 5) with five sections dipping to the northwest and two

303 sections dipping to the southeast. The dip angle of these NE-SW oriented sections is
304 unknown but is likely $>30^\circ$ based on the slope of the facet in the escarpment above
305 (Figure 6; Tucker et al., 2011). The most prominent of these NE-SW sections forms
306 a 4.8 km orthogonal link between the Thyolo and Muona sections (Figure 6). The
307 ~ 69 km long Thyolo and the ~ 28 km long Muona sections overlap by ~ 10 km and are
308 connected by this 4.8 km long section, which we refer to as the Chisumbi section.
309 The six other scarp sections that trend perpendicular to the average fault strike are
310 each <500 m long.

311

312 The mean strike of the pre-existing metamorphic foliation, within the footwall
313 basement of the Thyolo fault, is $140 \pm 37^\circ$ with a dip of $56 \pm 12^\circ$ SW (Figure 5 & 6).
314 This is subparallel to the mean strike of the fault scarp ($139 \pm 15^\circ$; Figure 5). A fault
315 dip could only be constrained in the field at Kalulu, nevertheless, this measurement
316 (60°) and the range of active normal fault dips elsewhere in Malawi (45° - 60° ; Hodge
317 et al., 2018b; Kolawole et al., 2018; Williams et al., 2019) and globally (Colletini and
318 Sibson, 2001) suggest that the Thyolo fault dips subparallel to the foliation.
319 Conversely, the mean strike of the dolerite dykes in the fault's footwall (calculated by
320 digitising the orientation of the contact between the dyke and basement rock) is 037
321 $\pm 9^\circ$, which is within error of the strike-perpendicular sections of the fault ($044 \pm 28^\circ$;
322 including the Chisumbi section). Our field measurements at four localities along the
323 Thyolo Fault indicate that the dykes are subvertical (Figure 3a).

324

325 4.3 Scarp Height

326 The median height of the fault scarp along the Thyolo fault is 18.6 ± 7.7 m (Figure 5).
327 The along-strike profile of the scarp height measurements shows two scarp height

328 minima (besides the fault tips; Figure 5b). The distance from the north-western fault
329 tip to the first minimum is 28 km with a median scarp height of 24.9 ± 9.0 m in this
330 portion of fault (Figure 5b). The middle portion of the fault is 15 km long and has a
331 median scarp height of 20.8 ± 6.3 m. The south-eastern portion is 48 km long with
332 median scarp height of 17.8 ± 6.5 m (Figure 5b). None of the scarp height minima
333 identified from the scarp height profile coincide with changes in fault strike, i.e. the
334 short segments that trend perpendicular to the average fault strike (Figure 5). The
335 scarp height minimum that occurs at ~ 53 km on Figure 5 is adjacent to a region
336 where the foliation strike varies within a range of $\sim 70^\circ$ around the mean. The region
337 contains both typical and somewhat anomalous orientations, and although it may be
338 an area of structural complexity, it does not stand out as markedly different from
339 elsewhere along the fault.

340

341 4.4 The Chisumbi Section

342 The 4.8 km Chisumbi section links the Muona and Thyolo sections (Figure 4b) and
343 has a strike of $046 \pm 31^\circ$, which is approximately orthogonal ($93 \pm 34^\circ$) to the
344 average strike of the main fault ($139 \pm 15^\circ$) but subparallel to the average strike of
345 the dolerite dykes ($037 \pm 9^\circ$; Figure 6). Along this linking section, we observed a 19.0
346 ± 4.2 m high scarp (median value; see example profile D in Figure 3b; Figure 5). This
347 height (yellow triangle in Figure 5b) is within the error bounds of the scarps found
348 along the adjacent Muona and Thyolo sections. Thus, the fault scarp along Chisumbi
349 section trends approximately perpendicular to the bounding sections, but has a
350 similar scarp height.

351

352 Within the footwall of the Chisumbi section, where the Muona and Thyolo sections
353 overlap, the dolerite dykes have a strike of $031 \pm 9^\circ$ whereas the strike of the dolerite
354 dykes outside the overlap zone is $038 \pm 9^\circ$ (Figure 6b). Thus, as these values are
355 within the error bounds of each other, the average trace of dykes within the
356 overlapping zone have either no rotation or a slight anticlockwise rotation around a
357 vertical axis (Figure 6b). The dip of the topography in the footwall of the Chisumbi
358 section (excluding the facet slope above the fault scarp) is 2° (Figure 6c),
359

360 4.5 Damage zone and fault core structure of the Thyolo fault

361 4.5.1 Lithology

362 The contact between hanging-wall sediments, and footwall gneisses and granulites
363 of the Unango Terrane, is exposed at Kalulu (Figure 7), where these two units are
364 separated by an approximately 60° dipping, 0.7 m thick, incohesive white to minty
365 green fault gouge. The gouge contains a brown clay-rich matrix (70-90% by area)
366 with subangular to subrounded quartzofeldspathic clasts (10-30% by area) up to 3
367 mm in size (Figures 7b and S2). A 5-15 m thick zone of incohesive
368 quartzofeldspathic granulite and hornblende gneiss borders the fault gouge. At the
369 other three localities where field exposures of the fault were studied, a 15-45 m wide
370 unit of incohesive biotite \pm hornblende \pm pyroxene gneiss is present adjacent to the
371 scarp, beyond which, the gneiss appears more intact (Figures 8 and S3). The
372 gneissic foliation is defined by alternating quartzofeldspathic and biotite \pm hornblende
373 \pm garnet bands, in which elongate biotite grains are aligned to and also define a
374 foliation subparallel to the compositional banding. Subvertical, NE-SW striking
375 Stormberg dolerite dykes crosscut the gneisses at all distances from the fault.

376

377 4.5.2 Fault zone structure

378 Quartzofeldspathic clasts within the fault gouge at Kalulu are intensely fractured and
379 sheared (Figure 7), and therefore we interpret this 0.7 m zone as the fault core
380 (sensu Caine et al., 1996). In the incohesive gneisses and granulites in the footwall
381 beyond the fault core, the metamorphic foliation and pegmatite veins are generally
382 preserved (Figure 9 and S4), although locally offset by < 0.6 m along minor faults
383 (i.e. discrete secondary faults within the footwall of the main fault; Figure 8d). At
384 Kanjedza, a 2 m wide ductile reverse shear zone has been exploited by a NW-SE
385 striking dyke whose emplacement predates movement on the Thyolo fault, as it has
386 been offset by a minor normal fault (Figure 8c). At Mbewe, a 50 cm thick, scarp-
387 parallel, steeply dipping foliated fault gouge is present 10 m into the footwall, and
388 juxtaposes charnockite and locally folded hornblende gneisses (Figure S3).

389

390 In the incohesive gneisses < 50 m from the Thyolo fault, quartz and feldspar grains
391 exhibit fracture densities of $2.3-4.8 \text{ mm}^{-1}$, with most fractures oblique to the foliation
392 (Figure 9 and Figure S5). These fractures are generally intragranular and closed, but
393 there are rare cases of biotite and calcite mineralisation, which are most prevalent
394 close to minor faults and/or dykes (Figure 9b). Microscale fracture density within the
395 intact gneisses 50-280 m from the Thyolo fault is $0.9-2.2 \text{ mm}^{-1}$, and fractures are
396 almost exclusively parallel to the foliation (Figure 9e). We interpret the 15-45 m wide
397 unit of incohesive gneiss with a relatively high density of foliation-oblique
398 microfractures, but little evidence of shear displacement, as the footwall damage
399 zone (sensu Caine et al., 1996). At distances > 45 m from the fault, the gneiss is
400 intact, and there is no major changes in composition. Therefore, we are confident
401 that the incohesive nature of the damage zone is related to brittle deformation and

402 fracturing around the fault, and any physical weathering or chemical alteration is
403 enhanced by and linked to fault-related fracture damage.

404

405 We do not observe a systematic decay in fracture density with distance from the fault
406 within the 15-45 m wide damage zone of the Thyolo fault (Figure 9). This may reflect
407 damage around minor faults and dykes within the damage zone (e.g. Figure 9b), the
408 lack of consistently oriented samples with respect to the fault, a sample bias that
409 misses the most damaged rocks because they are impossible to sample intact,
410 and/or variations in grain size and composition. The microfracture density increase
411 inside the damage zone relative to the background level is relatively minor (Figure
412 9g; compare with Wilson et al., 2003; Mitchell and Faulkner, 2009). It is difficult to
413 assess if this small increase in fracture density represents a relatively low fracture
414 density in the damage zone, or selective sampling of more cohesive portions of the
415 damage zone for analysis.

416

417 5. Discussion

418 The 18.6 ± 7.7 m fault scarp and triangular facets indicate that the Thyolo fault has
419 been reactivated during the current stage of East African Rifting (since 25 Ma),
420 although the steepness of the scarp indicates that it probably formed in the Late-
421 Quaternary or it would have degraded further (Hodge et al., 2020). Whereas the
422 Thyolo fault is dominantly subparallel to the metamorphic foliation, there are notable
423 sections where the strike turns by $\sim 90^\circ$ and the fault scarp trends subparallel to the
424 strike of Stormberg-age subvertical dolerite dykes (Figure 6).

425

426 5.1 Fault segmentation

427 Scarp height minima and changes in surface fault trend are generally considered
428 indicators of fault segment boundaries (Crone and Haller, 1991b; Machette et al.,
429 1991; Peacock and Sanderson, 1991; Crider and Pollard, 1998; Mortimer et al.,
430 2007, 2016; Fossen and Rotevatn, 2016). However, along the Thyolo fault, the
431 locations of scarp height minima do not coincide with changes in the trend of the
432 fault scarp (Figure 5).

433

434 The sections that trend perpendicular to the overall strike of the Thyolo fault range in
435 length from 170 m to 4.8 km. Because of our method of measuring strike as a 50 m
436 average, we may have missed < 50 m long strike-perpendicular sections that
437 coincide with the scarp height minima. However, we do not see short fault-
438 perpendicular sections near the scarp height minima on the scarp maps made from
439 the 12 m DEM. Only one of the fault-perpendicular sections (the Chisumbi section)
440 would normally be considered long enough to be a segment boundary based on
441 geometrical criteria (i.e. $\geq \sim 3\text{-}5$ km; Wesnousky, 2008). This geometry has been
442 used to argue that the Thyolo and Muona sections are different faults (Hodge et al.,
443 2019). However, a previously unrecognised fault scarp along the Chisumbi section
444 links the Thyolo and Muona sections, and the height of this scarp is in the same
445 range as scarps along the bounding Thyolo and Muona sections (Figure 3b, profile
446 D; Figure 5b). The presence of this continuous scarp implies that during the recent
447 earthquakes that formed the scarp, slip likely propagated along and through the 4.8
448 km long, $\sim 90^\circ$ kink in the fault. Given the ~ 600 m high escarpment and triangular
449 facets along the Chisumbi section it is also likely that slip has persistently
450 propagated along this section over longer geological time, through what would
451 typically be considered geometrical segment boundaries (Figure 10c). Consequently,

452 on faults that have reactivated pre-existing fabrics, purely geometrical criteria for
453 identifying fault segmentation may not adequately identify segment boundaries.

454

455 The Chisumbi section lacks evidence for distributed strain in the area between the
456 tips of the Thyolo and Muona sections it links (Figure 10). There is no or minor
457 anticlockwise rotation of dykes in the footwall of the Chisumbi section, and the
458 topographic slope dips at a much shallower angle ($\sim 2^\circ$; Figure 10) compared to a
459 global study of the dip of breached relay ramps (10-15°; Fossen and Rotevatn,
460 2016). Although the comparatively gentle ramp dip may be because most examples
461 in the global catalogue are from sedimentary rocks, it remains clear that little strain
462 accumulated within the Chisumbi section prior to the bounding Thyolo and Muona
463 sections becoming linked. The morphology of the Chisumbi linkage section is most
464 similar to a mid-ramp relay zone breach (Figure 10), which appears to be favoured in
465 locations with brittle basement rock and/or where pre-existing structures are
466 reactivated (Fossen and Rotevatn, 2016). When the distance between bounding
467 sections is less than 20% of their total fault length, Hodge et al. (2018a) suggest that
468 breaching occurs as slip is promoted by positive Coulomb stress changes in the
469 linking region. As the ~ 5 km Chisumbi section is $<20\%$ of the ~ 80 km Thyolo fault,
470 we propose that the Thyolo and Muona sections are linked by weak structures that
471 were activated by Coulomb stress changes in the shallow upper crust. This linkage
472 occurred prior to the accumulation of significant strain within the relay zone, and the
473 link section no longer operates as a permanent barrier to earthquake rupture and
474 propagation (Figure 10d).

475

476 5.2 Thyolo fault zone structure

477 The Thyolo fault has a ~1 km high footwall escarpment, which suggests that the total
478 displacement across the fault is at least 1.2 km (assuming a 60° fault dip) and the
479 damage zone documented here reflects fault-related deformation within a kilometre
480 of the surface. The 15-45 m thick footwall damage zone is within the range of other
481 faults with km-scale displacement in global comparisons, whereas the 0.7 m fault
482 core is relatively narrow (Torabi and Berg, 2011; Savage and Brodsky, 2011; Torabi
483 et al., 2019). However, there is considerable scatter in these global comparisons
484 owing to variations in fault kinematics, lithology and depth of faulting. A more
485 instructive comparison may be to the Dombjerg fault, Greenland, which is another
486 well exposed rift border fault of similar scale to the Thyolo fault (3 km throw, ~100 km
487 long), with a footwall hosted in crystalline metamorphic basement rocks that contain
488 a fault-parallel gneissic foliation. The Dombjerg fault footwall damage zone is 600 m
489 wide, ~10 times wider than the Thyolo fault, and the core comprises several <0.5 m
490 thick strands of gouge and breccia in a ~200 m wide zone, (Kristensen et al., 2016),
491 whereas the single Thyolo fault core is 0.7 m thick. The thicker and more distributed
492 nature of the Dombjerg fault core compared to the Thyolo fault core may reflect that
493 the Dombjerg fault was studied near a step between two fault segments (Kristensen
494 et al., 2016). On the other hand, our observations from the Thyolo fault were taken
495 from sites where there is no obvious geometrical complexity. The smaller fault zone
496 thickness and localised, single fault core in the Thyolo fault support theories where
497 the distribution of fault complexities during initiation and growth control fault zone
498 width (Kim and Sanderson, 2005; Childs et al., 2009). Localised slip and a narrow
499 damage zones are observed for other faults that follow pre-existing discontinuities
500 (Heermance et al., 2003; Zangerl et al., 2006), and so the fault-parallel foliation may
501 have also contributed to the Thyolo fault's relatively narrow structure. Therefore,

502 where a fault such as the Thyolo fault initiates, grows and lengthens rapidly in
503 mechanically anisotropic crust, a narrow damage zone and core initially forms, prior
504 to any increase in damage that may occur with the accumulation of larger
505 displacements (e.g. Kolyokhin and Torabi, 2012).

506

507 5.3 Mechanism of fault reactivation

508 Within amagmatic portions of the East African Rift System, immature faults (Biggs et
509 al., 2010), strong, cold intact crust (Fagereng, 2013) and low b-values recorded
510 during seismic sequences (Gaherty et al., 2019; Lavayssière et al., 2019) are all
511 suggestive of high differential stress in the region. Furthermore, wall-rock and gouge
512 samples from the Thyolo fault at Kalulu do not contain significant amounts of
513 frictionally weak, authigenic phyllosilicate minerals (Williams et al., 2019),
514 deformation experiments on representative lithologies from the Malawi Rift indicate
515 that they are frictionally strong (coefficient of friction, $\mu_s > 0.55$; Hellebrekers et al.,
516 2019), and the thick (30-40 km) seismogenic crust suggests that the frictional-
517 viscous transition does not occur at shallow depths (Craig et al., 2011; Fagereng,
518 2013). The Thyolo fault is, however, generally oriented parallel or subparallel to
519 basement foliation and possibly also NW-SE striking dykes (Figure 5-6, 8c), implying
520 that reactivation of these pre-existing structures is preferred over initiation and
521 growth of new fault surfaces. There is no evidence that the faults host authigenic
522 minerals and clays as described in other, long-term, frictionally weak faults described
523 in the shallow crust elsewhere (e.g. Holdsworth, 2004; Jefferies et al., 2006).

524

525 We suggest that the foliation and dykes play an important role in controlling the
526 geometry of the Thyolo fault in the shallow crust by providing relatively well-oriented

527 planes for reactivation, on which the sum of cohesion and frictional resistance in the
528 current stress field is less on the pre-existing plane than the intact rock. Although
529 well-oriented interconnected biotite foliation may be slightly weaker than the wall-
530 rock (Behnsen and Faulkner, 2012), such planes are not obviously controlling the
531 fault (Figure 7,8 & 9), and a combination of greater phyllosilicate content and low
532 cohesion, compared to the wall-rock, is likely why the foliations and dyke margins
533 are reactivated near-surface. This is consistent with fault reactivation analysis from
534 southern Malawi, which suggests that moderately dipping NW-SE striking incohesive
535 surfaces will reactivate even if they are slightly oblique to the ENE-WSW to E-W
536 trending minimum principal compressive stress (Williams et al., 2019). The Thyolo
537 fault zone contains an incohesive damage zone and fault core (Figures 7 and 8), and
538 we do not see fault zone fluid flow indicators in our microstructural and field
539 observations (e.g. vein networks or fault zone alteration; Wästeby et al., 2014;
540 Williams et al., 2017) that could result in fault cohesion recovery or the growth of
541 frictionally weak minerals (Tenthorey and Cox, 2006; Holdsworth et al., 2011)

542

543 We also suggest that interlinked mechanisms of reactivation of pre-existing surfaces
544 and dynamic stress reorientation along the Thyolo fault may explain why some fault
545 sections are orientated nearly perpendicular to the strike of the main fault. The
546 overlapping geometry between the Thyolo and Muona sections may have been
547 established early in the growth history of the Thyolo fault by the exploitation of pre-
548 existing NW-SE striking surfaces, on which reactivation is preferred over creating
549 new failure surfaces in the current stress field (Williams et al., 2019). Coulomb stress
550 changes within relay zones are known to favour the creation of zigzag fault patterns
551 (Crider and Pollard, 1998), and coseismic Coulomb stress changes around bounding

552 faults with large overlaps favour high angle link structures rather than oblique
553 breached relay ramps or the creation of a fault bend (Hodge et al., 2018a). These
554 links may have originated as transform faults, as they are better oriented for
555 transform motion, and later reactivated as normal faults, although no evidence for
556 transform motion is preserved. Slip on orthogonal structures may also have been
557 favoured by the presence of Stormberg dolerite dykes striking perpendicular to the
558 Thyolo fault (Figure 6a). Although linking segments coinciding with a pre-existing
559 dyke have not been directly observed, the margins of dykes in swarms cutting
560 basement terrains are commonly reactivated during later deformation events
561 (Holdsworth et al., 2020), and dolerite dykes in South Africa are known to have
562 increased brittle damage along the dyke-basement contact zone that reduces
563 cohesion (Senger et al., 2015). In summary, we propose that co-seismic stress
564 changes on overlapping faults favoured shallow activation of high-angle low-
565 cohesion zones at the contact between the pre-existing dykes and the basement, as
566 opposed to an oblique relay zone; in this context, linkage of the Thyolo faults proto-
567 sections could have occurred relatively quickly.

568

569 Although the Thyolo fault surface trace follows near-surface weaknesses, this
570 mechanism is less applicable at depths where cohesion is maintained, the relative
571 strength differences between foliations, dykes, and intact rock are smaller, and/or
572 confining stresses too high for frictional failure. Because variably striking normal
573 faults at shallow depth in anisotropic crust are thought to link to and initiate from
574 more continuous structures at depth (e.g. Graymer et al., 2007; Walker et al., 2017;
575 Hodge et al., 2018b), these near-surface low cohesion or weaker (relative to the
576 wall-rock) fabrics alone are not sufficient to account for the geometry of the Thyolo

577 fault at depth. In this case, it is notable that the Thyolo fault is located at or towards
578 the edge of the Unango Terrane. Although the exact nature and location of this
579 boundary is uncertain, it represents a high metamorphic grade boundary at the edge
580 of the Pan-African orogenic belts. Another regional example of these boundaries
581 includes the Lurio Belt, which borders the Nampula Complex in NE Mozambique and
582 is observed in outcrops of mylonitic sheared leucogneiss (Bingen et al., 2009; Macey
583 et al., 2010). If the Unango Terrane boundary is similar to this other high
584 metamorphic grade boundary, it could represent an existing shear zone that is
585 viscously weak because of small grain size (Watterson, 1975; Fliervoet et al., 1997;
586 Stenvall et al., 2019), foliation of interconnected low viscosity minerals (Handy, 1990;
587 Montési, 2013), crystal-preferred orientations conducive to plastic flow (Poirier,
588 1980), or provide a competency contrast across the boundary that leads to increased
589 stress and therefore a localisation of strain (Goodwin and Tikoff, 2002). At least
590 some of these features apply to the quartz-feldspar-biotite gneissic foliation in the
591 Thyolo fault footwall, and would make the foliation conducive to viscous reactivation.
592 Consequently, we consider it likely that the Thyolo fault follows a more continuous
593 deep-seated structure (Figure 11), possibly associated with ductile reactivation of the
594 boundary of the Unango Terrane at mid-crustal level.

595

596 Although many faults, including the Bilila-Mtakataka fault in the Makanjira Graben,
597 Malawi (Figure 1; Hodge et al., 2018b), show both displacement minima and
598 geometrical changes (or structural complexity) at the same locations (Peacock and
599 Sanderson, 1991; Dawers and Anders, 1995; Walsh et al., 2003), this is clearly not
600 the case for the Thyolo fault (Figure 5). Dyke margins and foliation planes that shape
601 the surface trace of the Thyolo fault have not influenced the along-strike scarp height

602 profile (Figure 5). Instead, we suggest that if the Thyolo fault is also following a more
603 continuous structure at depth, possibly a ductile weakness associated with the edge
604 of the Unango Terrane, then it is the heterogeneities in this weakness that dictate
605 where along-strike variations in fault displacement are located. We refer to this as
606 depth-dependent reactivation.

607

608 Where depth-dependent reactivation occurs, a combination of structural controls
609 affect the fault geometry and displacement in different ways (Phillips and McCaffrey,
610 2019). The Thyolo fault is an example of where the fault structure at the surface is
611 guided by shallow pre-existing fabrics and structures, which have influenced the fault
612 orientation, the fault core and damage zone and the relay zone evolution. In contrast,
613 deeper reactivated structures, which we suggest may be associated with ductile
614 reactivation of shear zones at high-grade metamorphic boundaries, can explain the
615 slightly oblique orientation of the Thyolo fault to the regional extension direction yet
616 apparent dip-slip kinematics (see modelling by Philippon et al., 2015; Hodge et al.,
617 2018b; Williams et al., 2019, and observations from Phillips et al., 2016; Phillips and
618 McCaffrey, 2019), the location of scarp height minima, and its continual reactivation
619 under a diverse range of previous extensional directions within the Lower Shire
620 Graben (Castaing, 1991). We acknowledge the model where the primary control on
621 rift growth is likely to be lithospheric strength (Ebinger et al., 1991); however, while
622 the total fault length and displacement profile may indeed reflect a thick elastic crust,
623 the detailed surface fault geometry appears affected by a combination of shallower
624 brittle and deeper viscous structural elements.

625

626 5.4 Comparison with other continental rifts and grabens

627 Many of the aspects of reactivation that we observe along the Thyolo fault resemble
628 features observed in other parts of the active East African Rift System. Localised
629 deformation, and fast growth and linkage of the Thyolo border fault is comparable to
630 the Okavango rift, which is also inferred to be localised along a long-lived pre-
631 existing crustal-scale weak zone (Kinabo et al., 2007, 2008). If local fabrics only
632 control the shallow orientation of the fault, this also explains why individual faults in
633 Malawi can both crosscut and follow the metamorphic foliation (Hodge et al., 2018b).
634 Furthermore, our model explains the difference between the Lower Shire Graben,
635 where the largest topographic relief indicates that the majority of displacement
636 occurs on the border fault (the Thyolo fault; Figure 1), and the Zomba Graben to the
637 north, where displacement is distributed more evenly between border and intrabasin
638 faults (Wedmore et al., 2020). Lateral heterogeneity within the lower crust beneath
639 the Zomba Graben has been inferred to cause this more heterogeneous strain
640 distribution, possibly by multiple localised shear zones at depth guiding distributed
641 deformation in the upper crust and at the surface (Wedmore et al., 2020). This is a
642 preferred explanation for strain distribution in the Zomba Graben, as it is located
643 *within* the Unango Terrane. In contrast, the Lower Shire Graben is located towards
644 the edge of the terrane and hence the deformation may localise towards the terrane
645 boundary.

646

647 The northern North Sea basin is another example of a multiphase rift where faults
648 are hosted in crystalline basement rocks. Here, lithospheric thinning and heating, as
649 well as stress feedbacks between growing faults, control the rift-scale localisation of
650 strain, with pre-existing shallow brittle faults thought to have little control on
651 reactivation (Cowie et al., 2005; Claringbould et al., 2017). Along the Thyolo fault, we

652 have shown that shallow features affect the geometry but not the displacement
653 profile of the fault. This is consistent with the results from the North Sea basin,
654 suggesting that reactivation of shallow pre-existing structures and fabrics may have
655 only a surficial role in controlling the geometry but not the accumulation of
656 displacement of faults in rifts within crystalline, dry, continental crust. This differs
657 from studies where a major role in rift evolution has been suggested for upper crustal
658 faults (e.g. Bellahsen & Daniel, 2005; Duffy et al., 2015; Heilman et al., 2019;
659 Katumwehe et al., 2015; Lañ-Dávila et al., 2015; Whipp et al., 2014). This confirms
660 the need to consider the scale- and depth-dependence of pre-existing structures
661 when assessing fault reactivation, where the pre-existing weaknesses may control
662 macro- but not meso-scale structural development (Kirkpatrick et al., 2013; Samsu et
663 al., 2020).

664

665 5.5 Implications for seismic hazard assessment

666 Geometrical criteria to define fault segments are commonly used to assess rupture
667 scenarios for seismic hazard assessments (Crone and Haller, 1991a; Lettis et al.,
668 2002; Wesnousky, 2008). We find that earthquake ruptures have persistently
669 propagated through significant changes in fault geometry and suggest that shallow
670 brittle structures only have a superficial, geometric effect on fault segmentation.
671 Instead, the displacement profile may provide a better indication of fault segment
672 boundaries controlled by a deeper, more continuous structure (Figure 11). This result
673 differs from findings on the Wasatch fault, USA, where DuRoss et al. (2016) suggest
674 that displacement profiles have limited value for identifying segment boundaries that
675 restrict earthquake ruptures. Thus, our findings may only apply in regions where pre-
676 existing fabrics play an important role in guiding the surface geometry of a fault. This

677 presents a challenge when segmentation criteria based on shallow structures is used
678 for assessing earthquake magnitudes for seismic hazard analyses (e.g. Field et al.,
679 2009; Petersen et al., 2015; Valentini et al., 2019): where depth-dependent
680 segmentation is not correctly identified, multi-segment and multi-fault ruptures such
681 as those observed in the 2016 earthquakes in central Italy (M_w 6.2, 6.1 & 6.6) and
682 Kaikoura, New Zealand (M_w 7.8) or the 2010 M_w 7.2 El Mayor-Cucapah, Mexico
683 earthquake (Wei et al., 2011; Hamling et al., 2017; Walters et al., 2018) may become
684 more likely than is apparent from surficial indicators of fault segmentation.

685

686 6. Conclusion

687 The Thyolo fault is the major border fault of the Lower Shire Graben, which has
688 experienced at least three previous phases of Phanerozoic rifting. Long sections of
689 the fault have a NW-SE strike, but these are separated by short sections that strike
690 NE-SW. The largest NE-SW section, the Chisumbi section, is 4.8 km long, which is
691 normally considered long enough to define a separate fault segment that
692 accumulates displacement differently from adjacent segments. However, the location
693 of both the Chisumbi section and other shorter sections with a prominent change in
694 strike do not align with two segment boundaries identified by along-strike variations
695 in the height of the active fault scarp. We find that the fault and the pre-existing
696 foliation are broadly parallel, whereas the strike of the short sections orientated NE-
697 SW matches the strike of dykes emplaced during a previous period of Karoo-age
698 rifting. Using field and microstructural observations of the Thyolo fault's footwall, we
699 estimate that the footwall fault zone is between 15-45 m wide, considerably narrower
700 than another example of a rift bounding fault in crystalline metamorphic basement
701 (the Dombjerg fault, Greenland; Kristensen et al., 2016). All these observations

702 suggest that the shallow, near-surface portions of the fault are reactivating well-
703 oriented foliation planes and near-perpendicularly oriented dyke contacts that act as
704 weak surfaces in the shallow crust compared to the crystalline basement. However,
705 these shallow pre-existing structures have not affected the distribution of the most
706 recent, near-surface displacement recorded by the scarp along the fault. Instead, we
707 suggest that the fundamental control on the growth and displacement accumulation
708 of this rift border fault is controlled by a broadly continuous structure at depth, which
709 is likely to be controlled by viscous reactivation of mid-crustal ductile heterogeneities,
710 possibly associated with the edge of the Unango Terrane.

711

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721 measurements, have been archived in the Zenodo data repository at:

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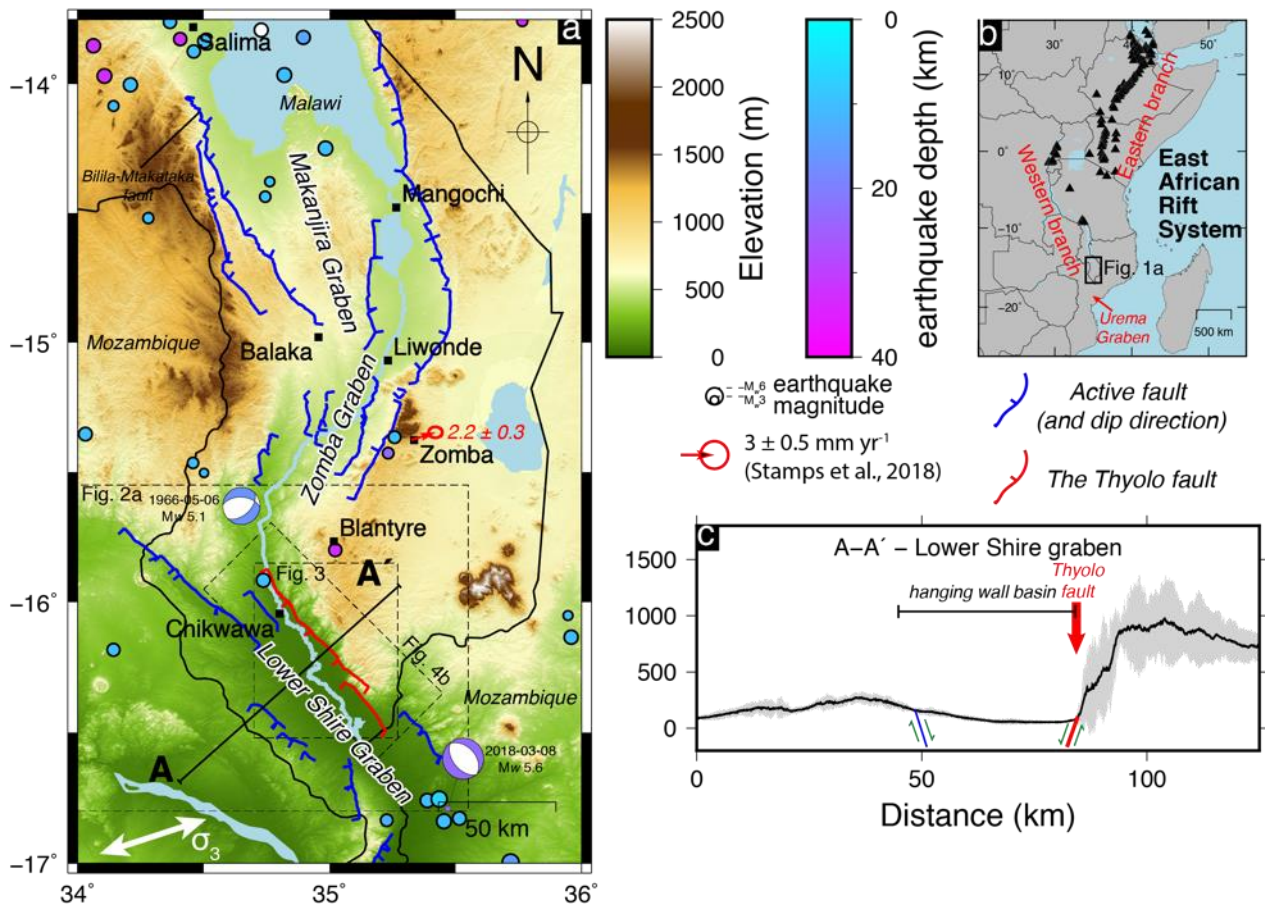
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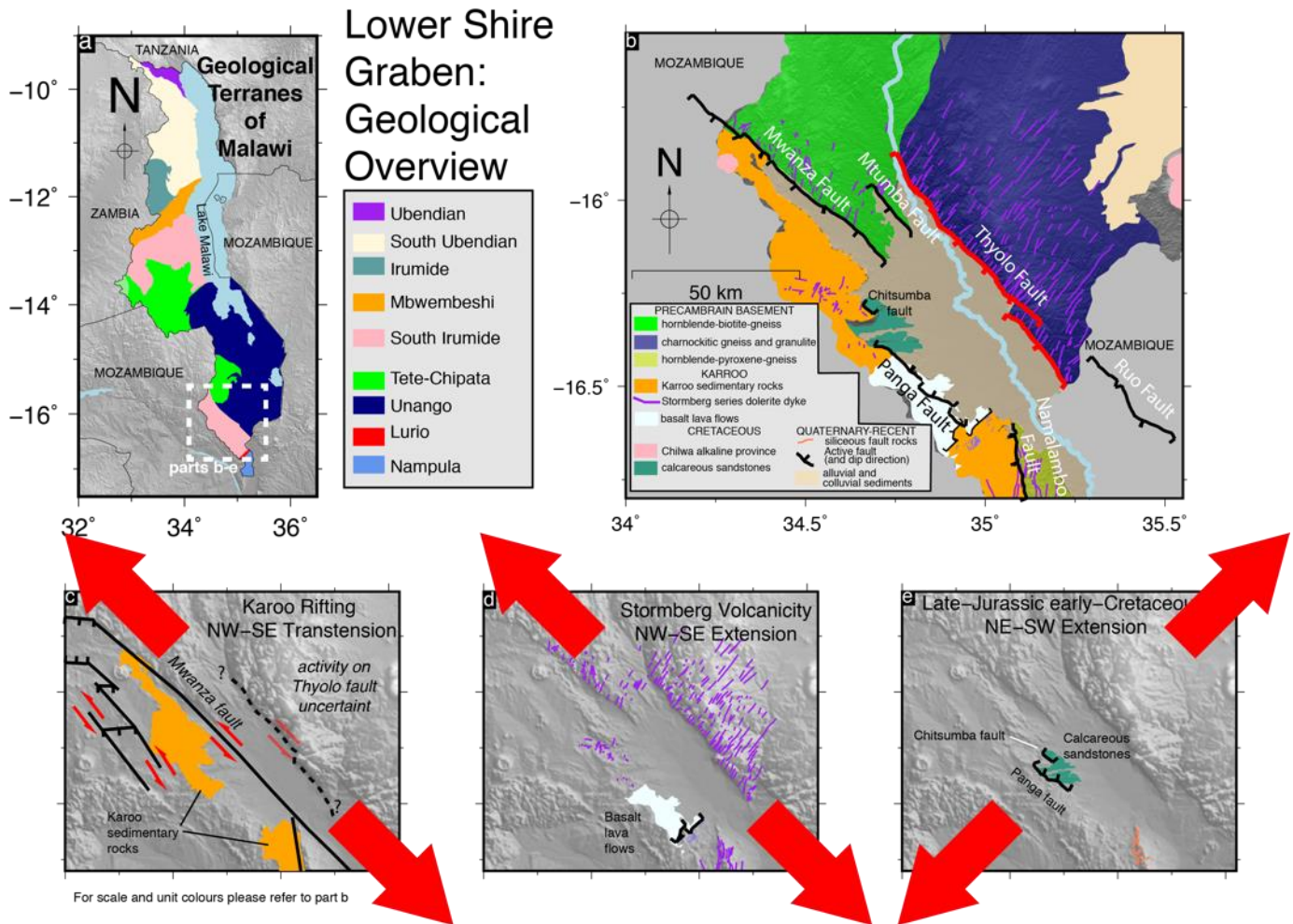
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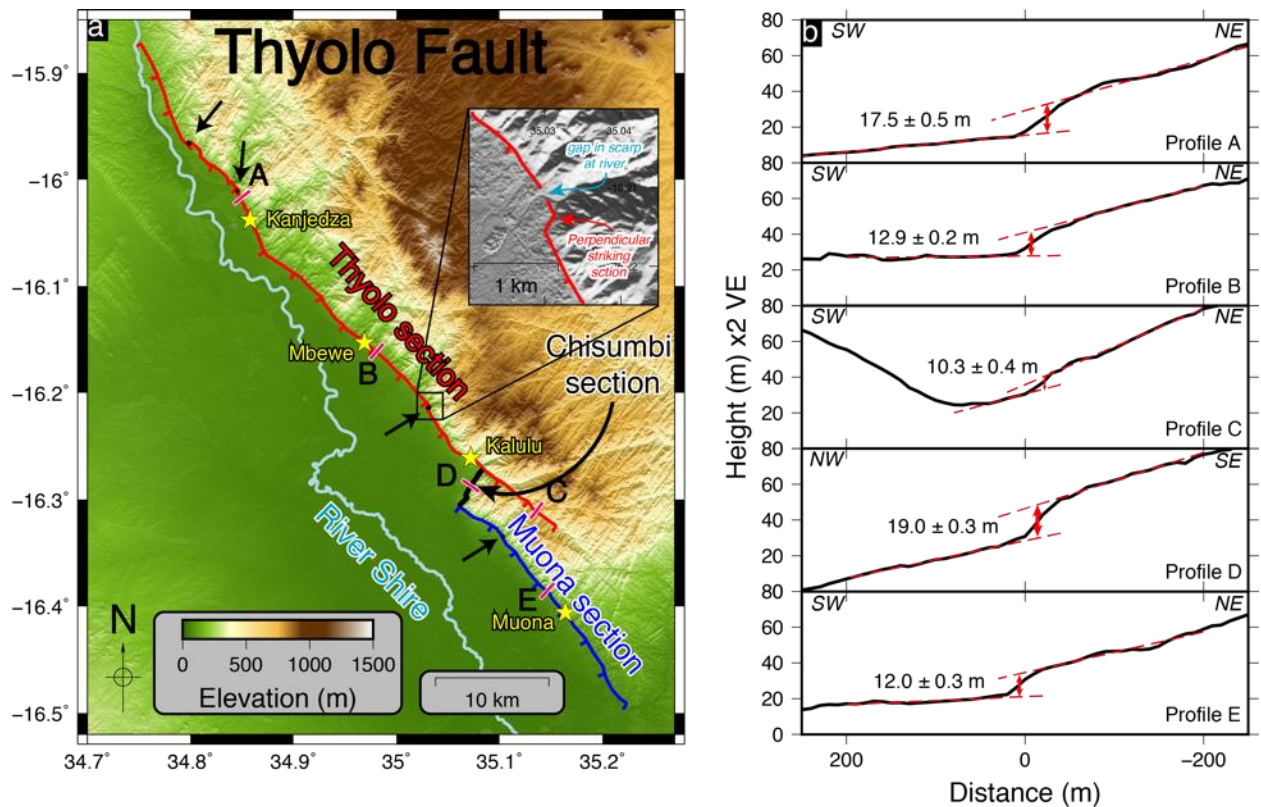


1250 Figure 1. The location and tectonic context of the Lower Shire Graben. (a) The
 1251 southern Malawi rift system with known active fault scarps in blue and the Thyolo
 1252 fault highlighted in red. Also shown is the GPS vector from a station in Zomba,
 1253 National Earthquake Information Centre earthquake locations from 1971-2018
 1254 (circles coloured by depth), and focal mechanisms for the two largest events in the
 1255 region, a M_w 5.1 earthquake in 1966 (from Craig et al., 2011) and the CMT solution
 1256 for the 2018 Nsanje earthquake (M_w 5.6). Extension direction (σ_3 ; 072°) is from a
 1257 moment tensor inversion in Williams et al. (2019) (b) The location of the southern
 1258 Malawi rift system within the East African Rift. Triangles indicate Holocene active
 1259 volcanoes. (c) Swath topographic cross section across the Lower Shire Graben
 1260 extracted from TanDEM-X data. Black line is the median elevation with the grey
 1261 shading the maximum and minimum elevation 10 km either side of profile A-A'
 1262 indicated in part a.



1263 Figure 2. Geological overview of the Lower Shire Graben. (a) Geological terranes
 1264 within Malawi (Fullgraf et al., *in press*). (b) Simplified geological map of the Lower
 1265 Shire Graben adapted from Hapgood 1963. (c) Structures related to NW-SE Karoo-
 1266 age amagmatic transtension. Note the activity of the Thyolo fault during this period is
 1267 uncertain. (d) Dykes and normal faults associated with NW-SE magmatic rifting in
 1268 the late Karoo period. (e) Normal faults and sedimentary deposits related to NE-SW
 1269 rifting during the Late-Jurassic to early-Cretaceous.

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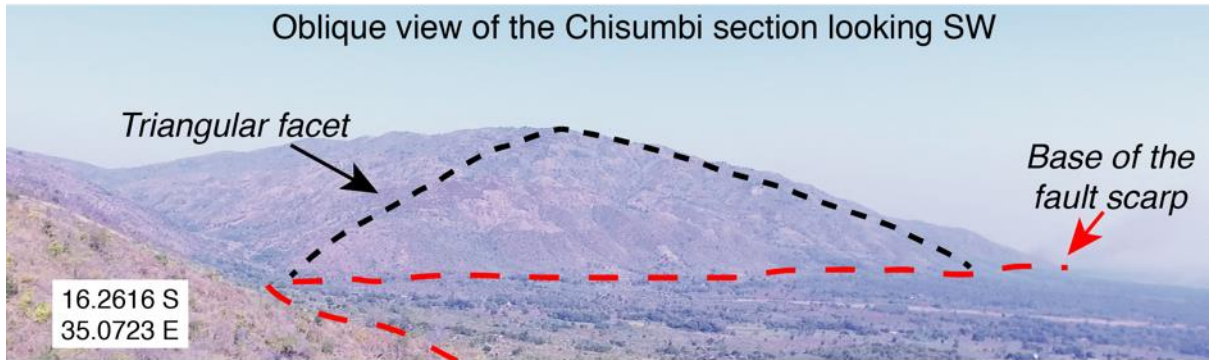
1272 Figure 3 (a) TanDEM-X digital elevation model of the Thyolo fault showing both the
 1273 Thyolo (red) and Muona (blue) sections. The fault sections oriented at $\sim 90^\circ$ to the
 1274 main strike are indicated in black with sections visible at this scale identified by black
 1275 arrows. Yellow stars indicate the locations of field studies reported in this paper. Pink
 1276 rectangles indicate are the locations and orientation of illustrative topographic
 1277 profiles extracted perpendicular to the fault scarp and shown in part b. (b) Example
 1278 topographic profiles extracted perpendicular the fault scarp. All profiles are plotted
 1279 with the footwall on the right-hand side (profile orientation is indicated in the top of
 1280 each panel). Note profile D is located along the Chisumbi section where the strike is
 1281 oriented $\sim 90^\circ$ to the strike of the main fault sections.

1282

Fault scarp along the Thyolo fault looking NE



Oblique view of the Chisumbi section looking SW



1283

1284 Figure 4. Field photos of the Thyolo fault. (a) An oblique view of the fault scarp along

1285 the Thyolo fault taken using a drone, showing an approximately 15 m high scarp.

1286 See the video available in the supplementary material for additional views of the

1287 scarp at this location. (b) An oblique view of the Chisumbi section taken using a

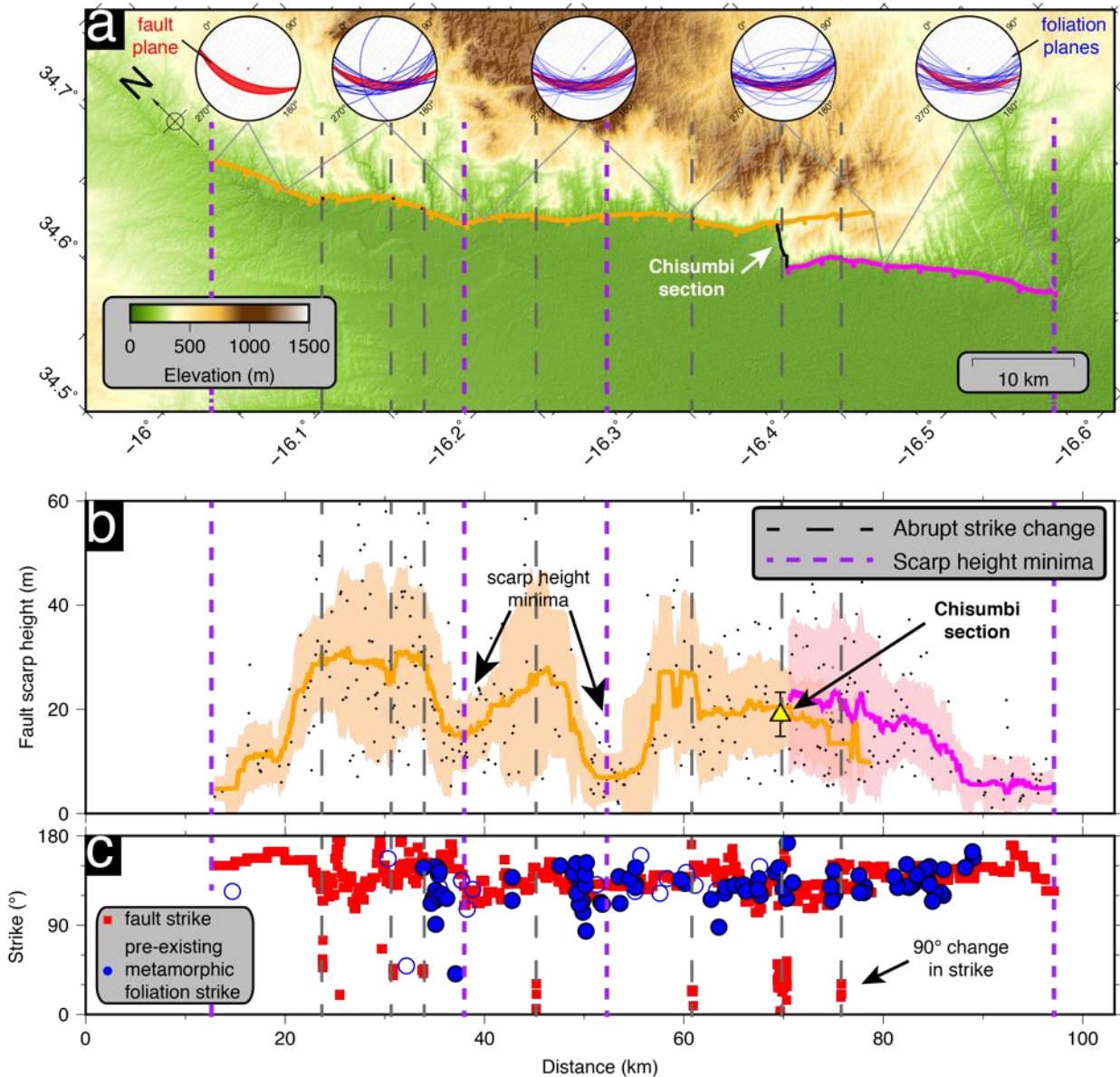
1288 drone, showing one of the triangular facets that are common along south eastern

1289 end of the fault. This section of the fault strikes perpendicular to the main section of

1290 the fault, which is just visible in the bottom left of the photo.

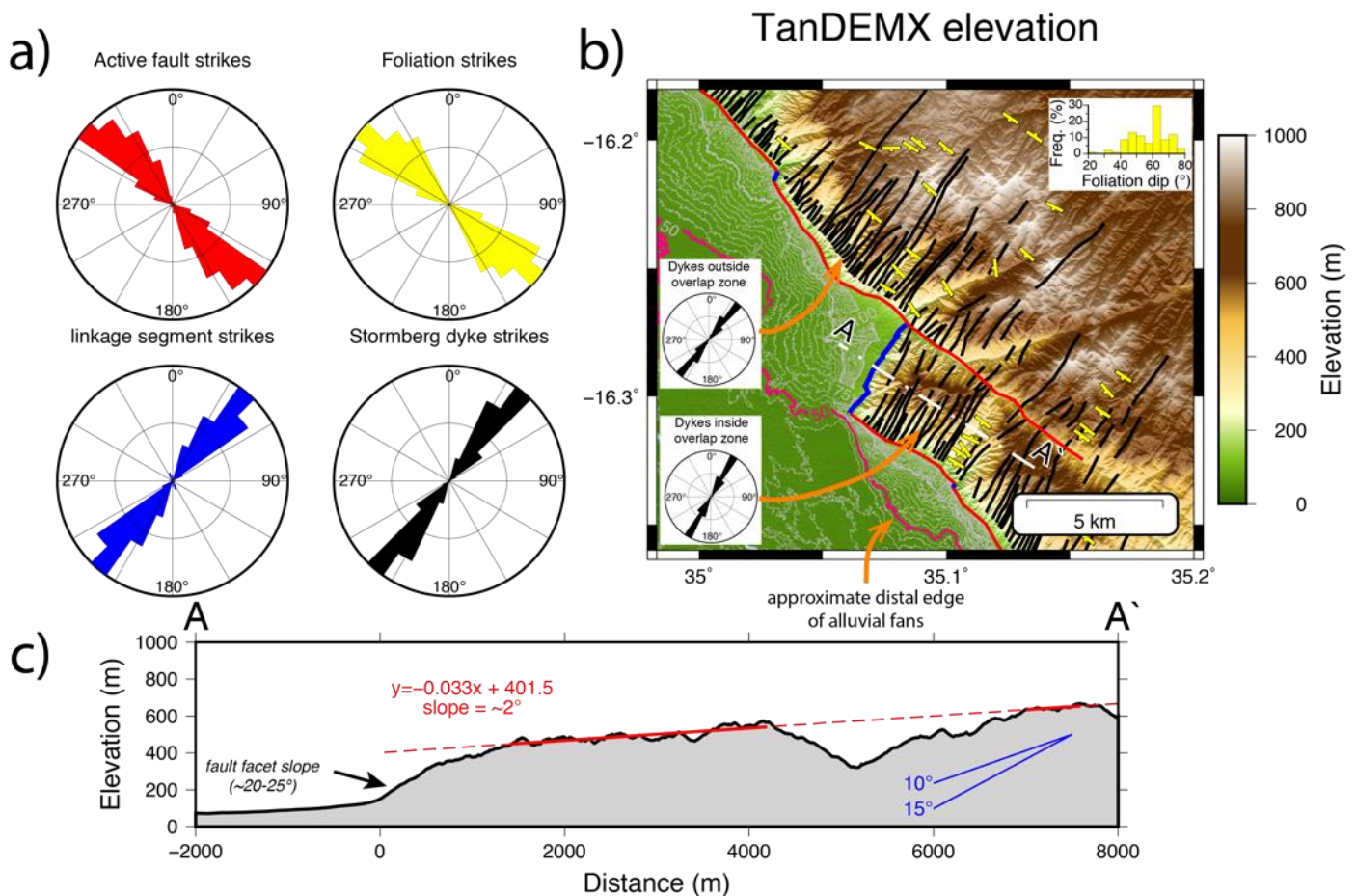
1291

The Thyolo fault scarp and segmentation



1293 Figure 5. Thyolo fault scarp height and segmentation. (a) A rotated view of the
 1294 Thyolo fault showing different indicators of fault segmentation. Inset equal angle,
 1295 lower hemisphere stereonet plots are rotated into the same view as the underlying map.
 1296 Red ellipses show the mean fault orientation measured every 20 km, with a dip
 1297 value plotted between 45°-60°, and the blue lines show foliation orientations. (b) The
 1298 height of the Thyolo fault scarp as a function of distance from the NW to the SE
 1299 along the fault. (c) The strike of the Thyolo fault (measured every 50 m) and foliation

1300 strike measurements (Habgood et al., 1973) as a function of distance from NW to SE
1301 along the fault. Open blue circles represent foliation measurements that did not give
1302 details of the dip of the foliation on the original geological map (Habgood et al.,
1303 1973). Scarp height in b was measured using topographic profiles, perpendicular to
1304 the scarp, extracted every 100 m along strike. Black dots are the individual
1305 measurements with the solid coloured lines the 5 km moving median of these
1306 measurements. The shaded areas represents the 1σ error bars. Red line is the
1307 Thyolo section, blue line is the Muona section. The yellow triangle (with 1σ error
1308 bars) is the scarp height along the ~4 km linking segment.



1309 Figure 6. The Chisumbi linkage section between the Thyolo and Muona sections. (a)

1310 Rose diagrams of the orientation of surface traces of the different structures along

1311 the Thyolo fault. Active faults include the Thyolo and Muona fault sections as

1312 indicated on the map. The fault sections and dykes were divided into 50 m long

1313 sections before calculating the strike of each section. Linkage segments only include

1314 the sections of fault that strike approximately perpendicular to the Thyolo and Muona

1315 sections. Foliation orientations and Stormberg dykes were digitised from Habgood et

1316 al. (1973). (b) TanDEM-X DEM of the Chisumbi linkage section between the Thyolo

1317 and Muona sections. Dykes are indicated with black lines, foliation orientation and

1318 dip direction with yellow lines and ticks, faults with red lines and sections of the fault

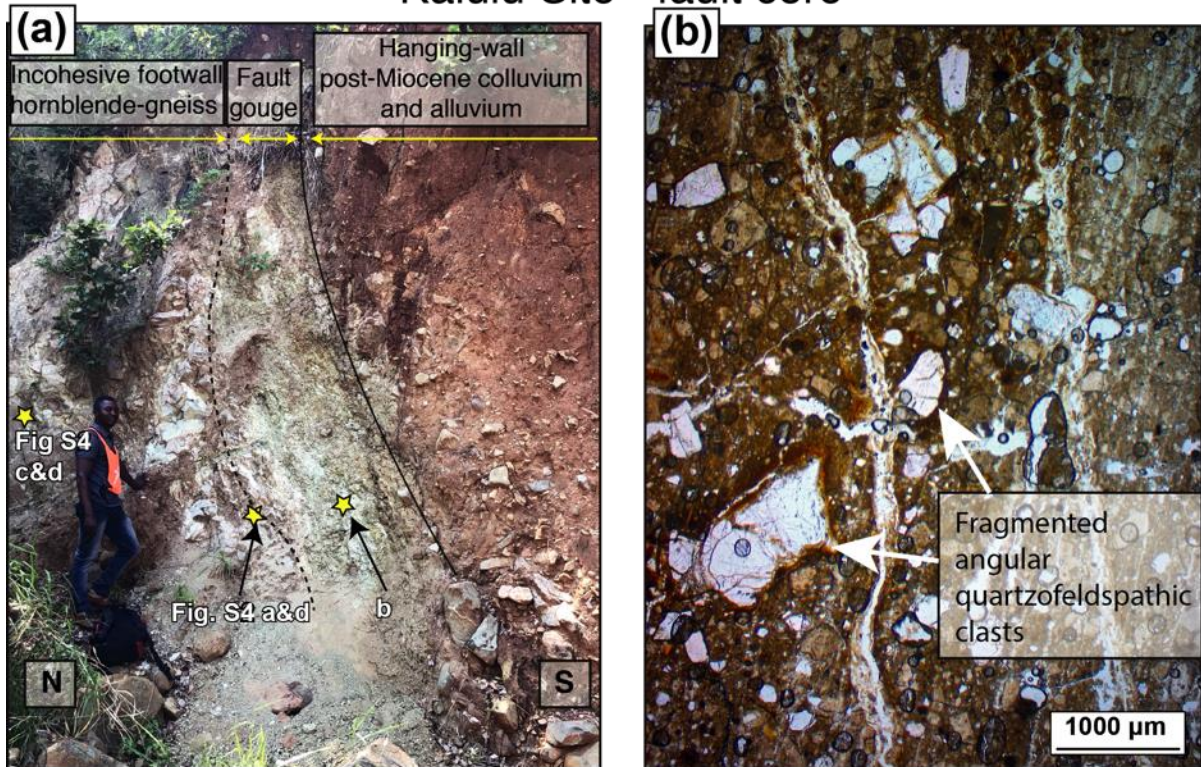
1319 that strike perpendicular to the main fault with blue lines. Grey contour lines are 2.5

1320 m apart, with the 50 m contour, which marks the approximate distal edge of alluvial

1321 fan complexes originating from footwall catchments, marked in pink. The inset

1322 histogram shows the dip of foliation measurements (Habgood et al., 1973). The inset
1323 rose diagrams show the orientation of dykes located inside and outside of the zone
1324 where the Thyolo and Muona sections overlap. (c) Swath topographic extracted
1325 along the transect A-A` shown in part b. The mean topography 1 km either side of
1326 the transect is plotted. The red line is a linear best fit to the slope of the topography
1327 within the portions of the solid red line. The dashed portions are not used as they
1328 have been affected by erosion due to river incision or include the fault scarp and fault
1329 facet slope. Angles which are the normal range of breached relay ramp dips
1330 (according to Fossen and Rotevatn, 2016) are plotted for comparison.
1331

Kalulu Site - fault core



1332

1333 Figure 7 (a) Fault zone exposure at Kalulu showing juxtaposition of hanging wall

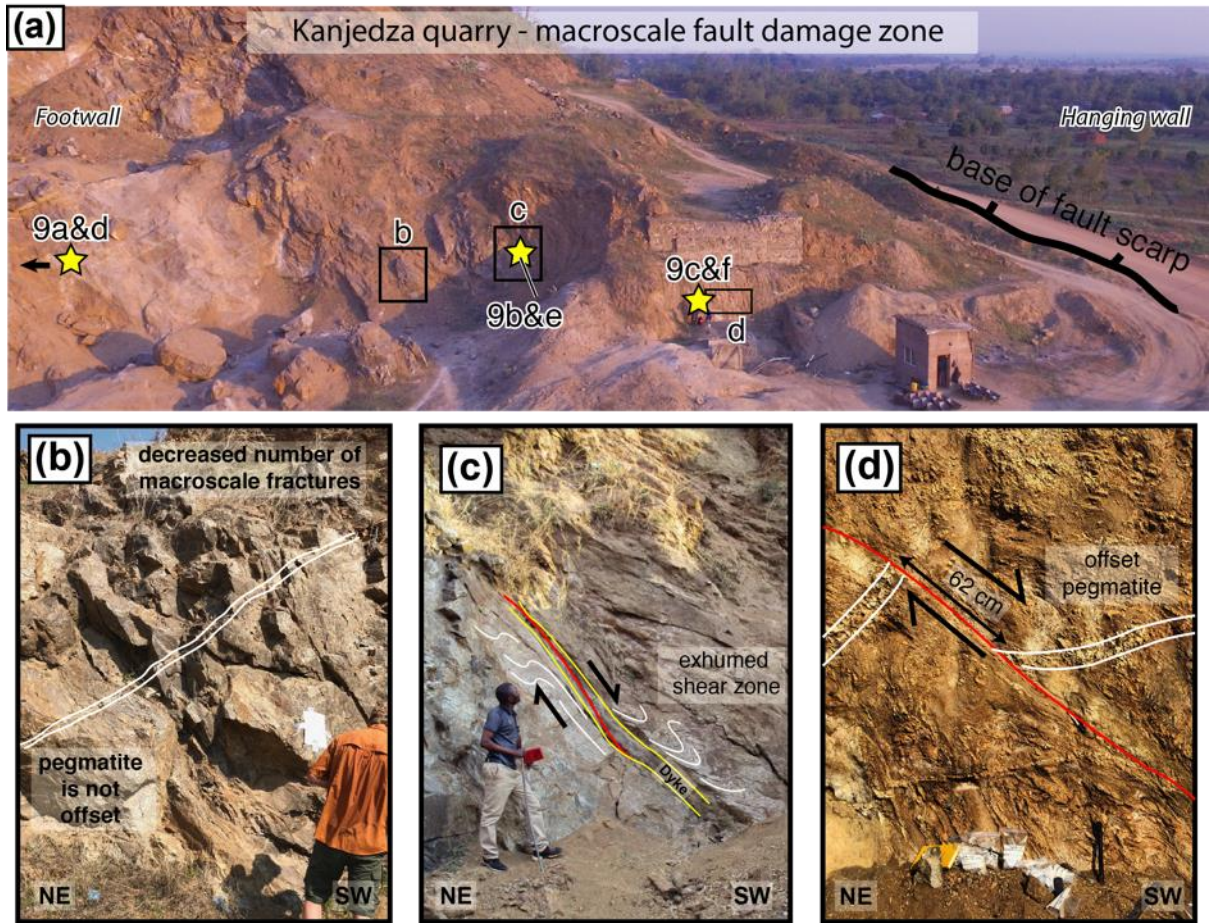
1334 sediments and footwall basement across a 0.7 m unit of fault gouge. Locations of

1335 samples used for photomicrographs in (b-f) shown by yellow stars. (b)

1336 photomicrograph of gouge with fractured quartz clasts and clay-rich brown matrix in

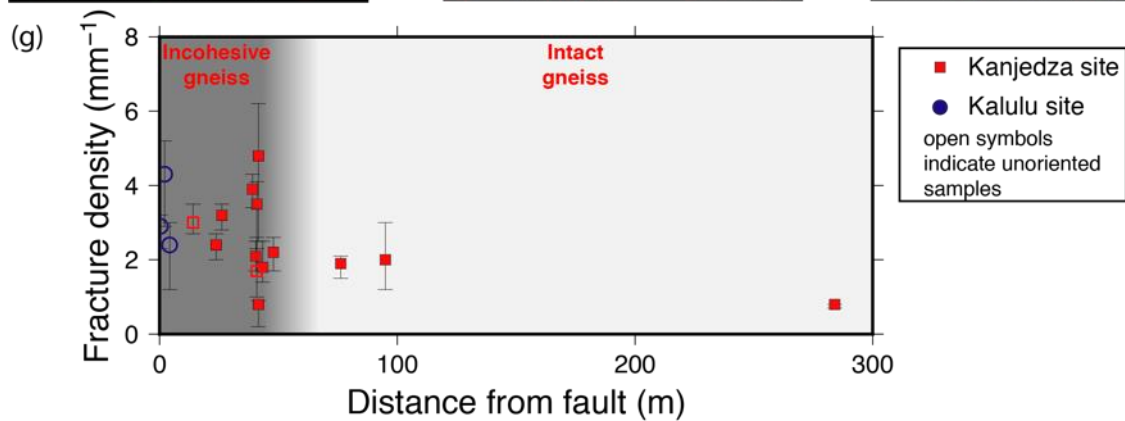
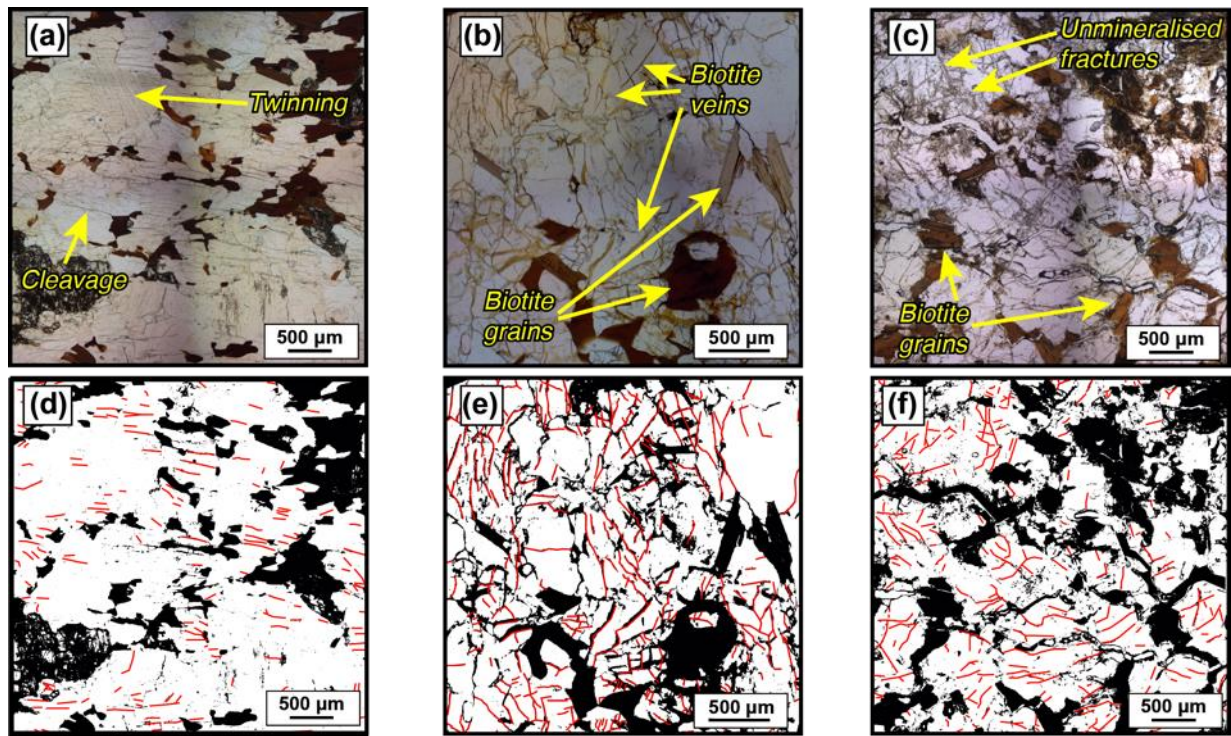
1337 plane polarised light (PPL) in sample from fault contact.

1338



1340
1341 Figure 8. Macroscale fault damage zone at the Kanjedza site along the Thyolo fault.

1342 (a) A perspective view of the exposed fault zone indicating the location of sample
1343 macroscale photos shown in parts b-d. Locations of microscale observations shown
1344 in Figure 9 are indicated with yellow stars. (b) Outcrop from outside the macroscale
1345 fault damage zone, note the lack of fracturing within the basement rock when
1346 compared with c and d. (c) Outcrop within the fault damage zone showing an
1347 exhumed shear zone and dyke. We infer that the shear zone had a reverse sense of
1348 shear as we can observe the same ductile structures on both the hanging wall and
1349 footwall sides of the fault. The dyke edge has been reactivated in a normal sense
1350 and acts as a minor slip surface with the same NW-SE strike as the main fault. (d)
1351 Offset pegmatite within the footwall damage zone.



1353

1354 Figure 9. Microscale fault damage zone at Kanjedza Quarry. Photomicrograph of

1355 samples (a) from outside the fault zone; (b) adjacent to a minor footwall slip surface

1356 (Figure 8c); and (c) within the fault damage zone. (d-f) annotated photomicrographs

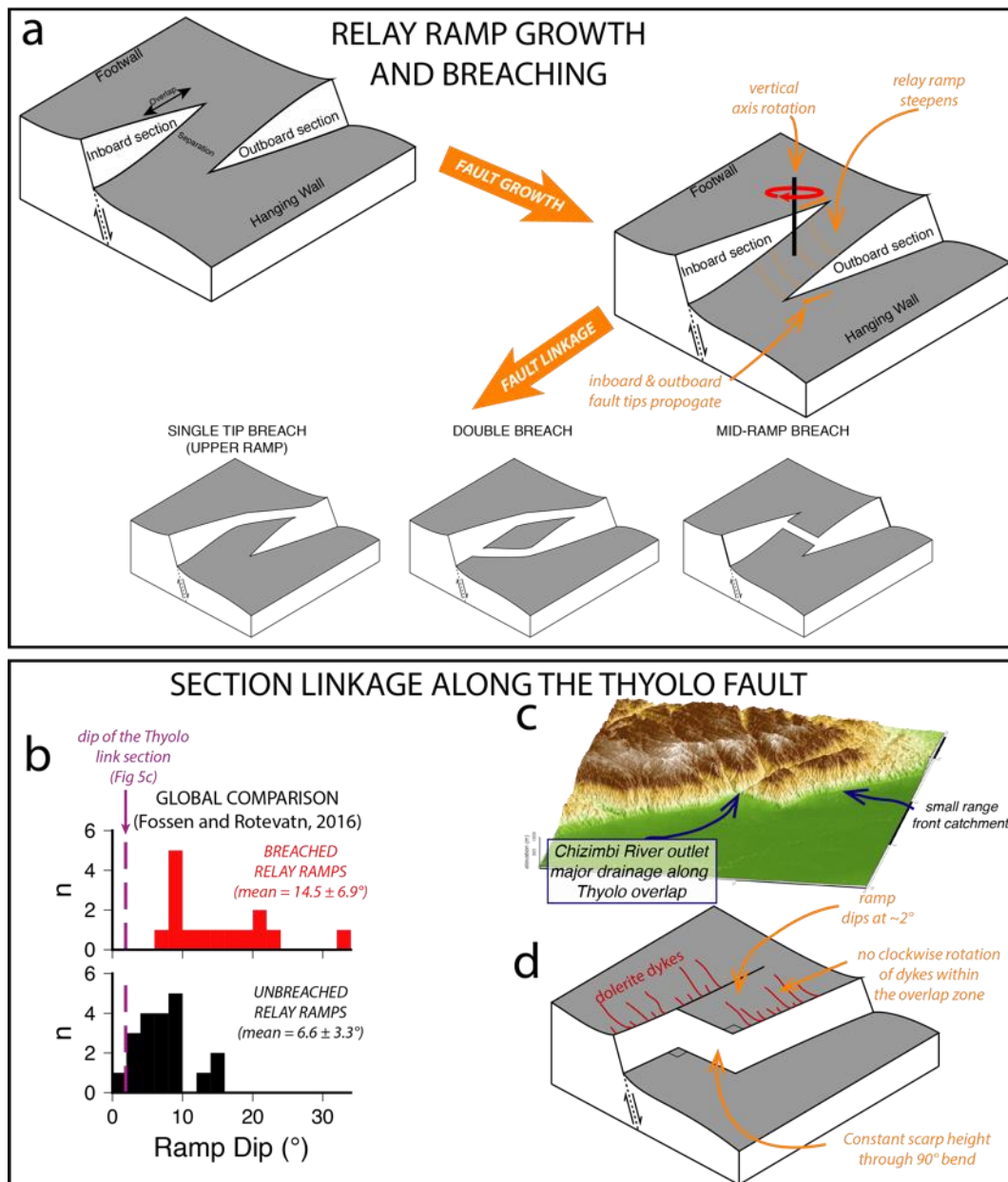
1357 of parts b-c showing the fractures (red lines) identified in each sample. (g)

1358 Compilation of fracture density plotted against distance from the fault for the

1359 Kanjedza and Kalulu sites along the Thyolo fault. The division between intact and

1360 incohesive gneiss is based on field observations (Figure 8).

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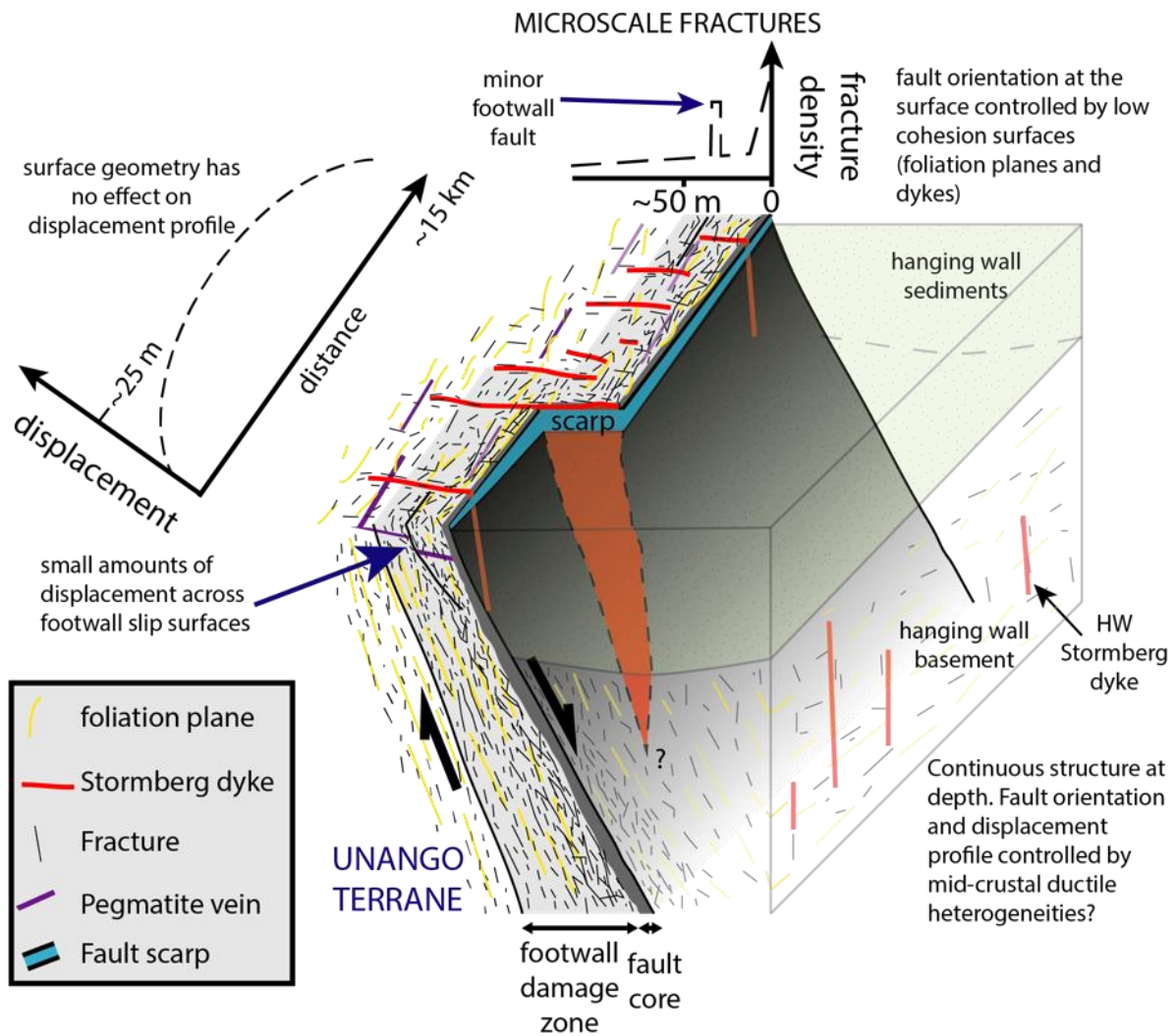
Figure 10. A comparison of relay ramp morphology and the linkage section between the Thyolo and Muona sections. (a) A summary of relay ramp growth and breaching (adapted from Fossen and Rotevatn, 2016). (b) The dip of relay ramp dips from a global compilation of breached and unbreached relay ramps (Fossen and Rotevatn, 2016). The dip of the topography in the section between the Thyolo and Muona sections is indicated with the purple dashed line. (c) A 3d view of the link section between the Thyolo and Muona sections showing the prominent drainage channels including the range front catchments that are predominate in the region and the

1370

1371 triangular facets along the Chisumbi section. (d) A conceptual view of the way the

1372 Thyolo fault has linked between the Thyolo and Muona sections.

1373



1375

1376 Figure 11. A conceptual model of the reactivation of the Thyolo fault towards the
 1377 edge of the Unango Terrane boundary. Shallow structures, including the 90° bend
 1378 and the fault damage zone are controlled by the pre-existing metamorphic foliation
 1379 and dykes. In contrast, the displacement profile is not affected by these shallow
 1380 structures, instead, we suggest that a deeper more continuous structure is
 1381 responsible for the segmentation of the fault observed in the displacement profile.
 1382 This deeper structure may be associated with the edge of the Unango Terrane.