

1 **Active fault scarps in southern Malawi and their implications for the**
2 **distribution and evolution of strain in amagmatic continental rifts**

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10 **Key Points:**

- 11 • In the Zomba Graben, southern Malawi, we identify 5 subparallel, 10-50 km long, 3-
12 20 m high, previously unrecognised fault scarps.
- 13 • ~75% strain since rift initiation was focussed on the border faults, but the intra-rift
14 strain has increased by ~25% during this period.
- 15 • The Zomba Graben is a zone of relatively high seismic hazard linking Lake Malawi to
16 the Urema Graben in Mozambique.
- 17

18 **Abstract**

19 The distribution of deformation during early stages of continental rifting is an important
20 constraint on our understanding of continental breakup. Early stage rifting is typically
21 considered to be dominated by slip along rift border faults. The subsequent transition to
22 focussed extension on axial segments is thought to require thinned crust and active
23 magmatism. Here we study high resolution satellite data of the Zomba Graben in southern
24 Malawi, an amagmatic rift with a short (<10-5 Ma) history of extension, but little prior
25 evidence of active faulting. We discover active fault scarps that probably represent late-
26 Quaternary activity in the graben and represent a previously unrecognised seismic hazard.
27 By comparing total fault displacements since rift initiation with the heights of newly
28 identified, recently formed fault scarps, we constrain the spatial evolution of strain through
29 the early stages of continental rifting. Whereas 75 ± 18 % of the total extension was
30 accommodated by slip along border faults, displacement along three intra-rift fault scarps
31 demonstrate that current extension is distributed throughout the graben. The border faults
32 remained active throughout the period of rifting, but displacement on the intra-rift faults
33 has increased from 25 ± 8 % of the total extensional strain, to 50 ± 25 % of the recent
34 extensional strain. Previous models of the East African Rift suggest that the transition from
35 border fault dominated to axial strain is driven by fluids or runaway lithospheric thinning.
36 Our observations demonstrate that significant intra-rift strain can occur in thick continental
37 lithosphere with no evidence for magmatic fluids.

38 Plain Language Summary

39 When continents begin to stretch, individual faults host earthquake that incrementally
40 accumulate slip to eventually form a rift valley. To estimate earthquake hazard, it is

41 important to understand where these faults are located, and how much stretching they
42 accommodate. We analyse faults in the Zomba Graben, a young rift in southern Malawi
43 using high-resolution satellite data. Since the rift formed, most of the deformation has
44 occurred at the edges of the rift valley. In contrast, steep scarps, indicate that more recent
45 earthquakes have occurred at both the edges and the middle of the rift valley. This shift
46 occurs much earlier than previously thought, and is not associated with volcanic fluids
47 weakening the Earth's crust as suggested by existing models. Instead, the distribution of
48 faults, and hence of earthquakes, is likely due to weaknesses in the middle and lower parts
49 of the Earth's crust.

50 **1. Introduction**

51 The development and interaction of faults dominate the early stages of continental rifting
52 and contribute to the eventual breakup of continental lithosphere (Cowie et al. 2005;
53 Ebinger and Scholz, 2012). Current conceptual models suggest that rifting is initially
54 dominated by activity on border faults (Ebinger, 2005, Muirhead et al 2019). During these
55 early stages, intra-basin faulting is limited to small-displacement faults that accommodate
56 flexure, and strain only begins to migrate to the rift interior when magmatic intrusions
57 weaken the lithosphere and cause runaway thinning (Buck, 2004; Ebinger, 2005). However,
58 because these later phases of magmatic rifting are associated with changes in
59 sedimentation, fault activity, and volcanism, patterns of intra-basin faulting generated
60 during earlier stages of rifting are typically obscured.

61 The overall mode of a rift (e.g. narrow vs wide *sensu* Buck 1991) is thought to be
62 predominately controlled by the interplay between Moho temperature, crustal thickness,
63 and strain rate (Buck, 1991; Huisman and Beaumont, 2007). In contrast, the surface

64 expression of a rift - whether faults form symmetric grabens or asymmetric half-grabens,
65 and the length, orientation and segmentation of major faults - is controlled by crustal
66 rheology (Huismans and Beaumont, 2007; Hodge et al., 2018). Within the East African Rift
67 system, there is evidence for both narrow and wide rift modes (*sensu* Buck, 1991; Ebinger
68 and Scholz, 2012), symmetric and asymmetric grabens (Ebinger et al, 1999; Lao Davila et al,
69 2015) and a range of relationships between fault strike, metamorphic foliation, and regional
70 stresses (Hodge et al., 2018; Williams et al., *in revision*). The lack of constraints on strain
71 distribution, geochronology and geophysical properties in East Africa makes it challenging to
72 ascertain the relative roles of shallow crustal rheology, heatflow, crustal thickness and finite
73 strain in shaping the geometry of incipient rift basins.

74 To investigate the geometry and strain distribution within a youthful rift, we analyse the
75 geomorphic signature of an active normal fault array in the Zomba Graben, which is at the
76 southern end of the incipient, amagmatic Malawi Rift. The graben is dominated by a west-
77 dipping border fault, but we also find evidence for both syn- and antithetic normal faults in
78 its hanging wall. As well as the prominent border faults, we use geomorphic analysis of high-
79 resolution topographic data to detect five previously unrecognised active fault scarps. We
80 compare footwall escarpments that are representative of the long-term pattern of strain in
81 the graben, with fault scarps that represent more recent displacements. This analysis is then
82 used to address how and when rift-related deformation is distributed between the border
83 faults and faults within the rift interior. Our findings have implications for our understanding
84 of the evolution of strain in the early stages of continental rifting and the associated seismic
85 hazard.

86 2. Tectonic and Geological Setting

87 2.1. *The Malawi Rift*

88 The ~ 900 km long Malawi Rift System is located at the southern end of the largely
89 amagmatic western branch of the East African Rift System (EARS; Ebinger et al., 1987; Figure
90 1). The rift along Lake Malawi is defined by a series of asymmetric half-grabens with
91 segmented border faults that offset Proterozoic medium to high grade metamorphic rocks
92 with fabrics formed in multiple Precambrian orogenic events (Fritz et al., 2013). The border
93 faults are ~120 km long with up to ~ 6 km of throw related to the current rift regime
94 (Contreras et al., 2000; Lao-Davila et al., 2015). The locations of faults in Lake Malawi are
95 well constrained by a number of seismic surveys carried out in the lake (Scholz and
96 Rosendahl, 1988; Scholz, 1995; Lyons et al., 2011; Shillington et al., 2016) as well as a recent
97 network of lake bottom and onshore seismometers in northern Malawi and Tanzania
98 (Shillington et al., 2016; Accardo et al., 2018). These seismic surveys, along with the 2009
99 Karonga earthquake sequence that occurred within the hanging wall of the rift-bounding
100 Livingstone Fault, indicate that both rift border and intrabasinal faults are currently active at
101 the northern end of the Malawi Rift (Biggs et al 2010, McCarntney and Scholz 2017, Gaherty
102 et al 2019). Active faulting is also occurring at the southern end of the lake, as demonstrated
103 by the M_w 6.1 1989 Salima Earthquake (Jackson and Blenkinsop 1993).

104 Rifting within the western branch of the EARS is thought to have initiated in the Oligocene
105 (~25 Ma; Roberts et al., 2012). Low-temperature cooling ages from the northern basin of
106 Lake Malawi indicates that regional scale cooling, associated with the onset of rifting,
107 commenced at ~23 Ma (Mortimer et al., 2016), and thus earlier than previously proposed
108 age of ~9 Ma based on radiometric dating of volcanic and volcanoclastic deposits from

109 northern Malawi (Ebinger et al., 1993). The age of the rift in central and southern Malawi is,
110 however, poorly-constrained. A 4.6 Ma age has been proposed for the onset of sediment
111 accumulation in Lake Malawi's central basin (~350 km to the north of the Zomba Graben),
112 from extrapolating the average rates of sediment accumulation in ~1.3 Ma drill core (Lyons
113 et al 2015) to the entire sedimentary sequence (McCartney and Scholz 2016). This would
114 suggest a gradual southward propagation of the rift (Ebinger et al., 1987; Contreras et al.,
115 2000), which is consistent with sediment thickness and footwall topography decreasing
116 from north to south within Lake Malawi (Specht and Rosendahl, 1989; Flannery and
117 Rosendahl, 1990; Lao Davila et al., 2015). An alternative hypothesis is that the onset of
118 extension is uniform along the Malawi Rift, but the extension rate is faster in the northern
119 part of the rift because the Euler pole of the plates is located south of the Malawi Rift (Calais
120 et al., 2006; Saria et al., 2014; Stamps et al., 2018). Although there are currently few
121 constraints on the age of the EARS to the south of Lake Malawi (Dulanya et al 2017), we
122 consider it unlikely that extension initiated in this region prior to the onset of sedimentation
123 in Lake Malawi. This places an approximate maximum age of the rifting in the Zomba
124 Graben as the mid-Miocene to early-Pliocene.

125 The onshore rift south of Lake Malawi consists of three linked half-grabens (Figure 1,
126 Williams et al, in revision). The border faults have escarpment heights of <1000 m, and so
127 are less prominent than those within the lake (Lao-Davila et al., 2015). The initiation and
128 accumulation of displacement on these faults is thought to have led to Lake Malawi
129 becoming externally drained via the River Shire, which links the lake to the Zambezi River
130 (Dulanya, 2017; Figure 1). The NW-SE trending Makanjira graben contains two known active
131 faults: the ~55 km long Malombe fault, and the ~110 km long Bilila-Mtakataka fault, a

132 possible source of the 1989 M_w 6.1 Salima Earthquake (Jackson and Blenkinsop, 1993; Hodge
133 et al., 2018; 2019). Immediately south of the Makanjira graben lies the Zomba Graben,
134 which is the focus of this study and described in more detail in section 2.2. To the south of
135 the Zomba Graben, in the middle Shire Valley, the river drops by \sim 380m in elevation with
136 no evidence of active faulting (Dulanya, 2017; Figure 1). The NW-SE trending Lower Shire
137 graben lies \sim 60 km further south and is a reactivated Karoo-age basin bounded to the east
138 by the 85 km long active Thyolo fault (Figure 1b; Hodge et al., 2019). The EARS continues
139 south into the Urema Graben, the site of repeated seismic activity following the 2007 M_w 7.0
140 Machaze earthquake (Lloyd et al, 2019; Copley et al., 2012).

141 *2.2. The Zomba Graben*

142 The Zomba Graben is a NE-SW trending segment of the onshore Malawi Rift where the
143 dominant structure is the NW-dipping Zomba fault. The Zomba Graben has traditionally
144 been considered a half-graben similar to those in Lake Malawi (e.g. Ebinger et al., 1987), but
145 Lao Davila et al. (2015) map it as a \sim 50 km wide full-graben, bounded to the west by the E-
146 dipping Lisungwe fault.

147 The topography of the Zomba Graben is influenced by structures within the basement
148 complex, which comprises Proterozoic metamorphic rocks of the Southern Irumide Belt and
149 subsequent intrusions. On the western side of the rift, the mountainous Kirk range is
150 composed of metasedimentary schists, paragneisses and granulites, and contrasts with the
151 lower elevation eastern side, which is composed of meta-igneous charnockitic granulites
152 (Figure 2; Bloomfield and Garson, 1965). Syn and postkinematic intrusions form local
153 regions of high topography. The most notable of these are the Proterozoic Chingale Ring

154 Complex, and the Upper Jurassic-Lower Cretaceous Chilwa Alkaline Province, which includes
155 the Zomba Massif quartz-syenite and granite intrusion (Bloomfield 1965, Eby et al 1995).

156 Despite the apparent presence of border faults within the Zomba Graben, little is known
157 about the distribution of faulting in the region and the way in which strain is distributed. A
158 set of escarpments in the centre of the graben that offset fluvio-lacustrine sediments have
159 been variously mapped as 'terrace features' (Bloomfield, 1965; Figure 2), active fault scarps
160 (Dixey, 1926) and inactive late-Jurassic or early Cretaceous faults (Dixey, 1938).

161 **3. Active faults within the Zomba Graben**

162 Within Malawi, faults have been identified as active based on the presence of a continuous,
163 steep scarp at the base of the footwall at the surface (Jackson and Blenkinsop, 1997; Hodge
164 et al., 2018; 2019). Observations of soft hanging wall sediments, preserved uplifted river
165 terraces in the footwall, and knickpoints in streams close (within <100 m) to the fault scarps
166 were also used as evidence to confirm the faults are active (Jackson and Blenkinsop, 1997).
167 These criteria have been used prior to this study to identify three active faults south of Lake
168 Malawi but outside the Zomba Graben: The Bilila Mtakataka, Malombe and Thyolo faults
169 (Figure 1c; Hodge et al., 2019). As no active faults had previously been identified in the
170 Zomba Graben, geological maps were used to first identify faults where Bloomfield (1965)
171 identified tertiary-recent hanging wall sediments (Figure 3a). The detailed topographic
172 analysis described below was then used to determine whether a fault has formed an active
173 fault scarp, in which case the fault would be considered currently active. We supplement
174 these observations with fieldwork, from a 6-week campaign in 2018.

175 *3.1. Topographic analysis to assess fault activity*

176 We used a TanDEM-X digital elevation model (DEM) of the Zomba Graben to identify and
177 analyse fault scarps and the river channels that cross them (Figure 3). TanDEM-X DEMs have
178 a horizontal resolution of 12.5 m and an absolute vertical mean error of ± 0.2 m (RMSE < 1.4
179 m; Wessel et al., 2018), which is sufficient for measuring the meter-scale fault scarps in the
180 Zomba Graben. We consider the presence of a linear scarp coinciding with changes in
181 channel incision and width, and hanging wall sediment deposition, as evidence for active
182 faulting during the current rifting episode (Figure 3).

183 We produced slope maps from the DEM by calculating the scalar magnitude of the slope
184 derivative using the `gdgradient` tool in the Generic Mapping Tools routines (Wessel and
185 Smith, 1998; Figure 3b) and use these maps to identify the location of the active fault
186 scarps. We noted any gaps in the scarps, or peaks and troughs in the displacement profiles,
187 that may be indicative of fault segmentation (following Hodge et al., 2018). We extracted
188 500 m long fault-perpendicular topographic profiles every 12 m, and stacked them at 100 m
189 intervals. The stacking has the effect of removing short-wavelength random topographic
190 features not related to the fault such as local sedimentation and erosion or human
191 settlements and vegetation close to the fault. The surface offset was calculated by fitting
192 regression lines to the footwall and hanging wall topography of the stacked profiles, and
193 measuring the vertical difference between the extrapolated regression lines at the point of
194 maximum steepness on the scarp (Avouac, 1993; Figure 4a). To estimate the uncertainty,
195 we applied a Monte Carlo approach by selecting 10,000 random subsets of points from the
196 footwall and hanging wall, and allowing the exact location of the fault to vary. The variation
197 between profiles is due to a combination of differential geomorphic degradation of the fault

198 scarp and variation in fault offset that forms during an earthquake. Example profiles for
199 each fault are shown in Figure 4a. We filter the resulting measurements along strike using a
200 3 km wide moving median (Hodge et al., 2018; 2019).

201 Rivers that cross normal faults record information about the timing and magnitude of active
202 faulting in their long profile (i.e. their change in elevation with distance along the channel)
203 and can be used as an indicator of fault activity in regions where the location of active faults
204 is poorly constrained (Boulton and Whittaker 2009). We extracted long profiles of rivers in
205 the Zomba Graben with a drainage area greater than 6,000 m² using TopoToolbox
206 (Schwanghart and Scherler, 2014). We identified changes in channel steepness that are
207 evidence of perturbations to the power-law relationship, $S = k_{sn} A^{-\theta_{ref}}$, between local channel
208 gradient, S , and upstream drainage area, A , in detachment limited rivers (Whipple and
209 Tucker 1999). We use a reference concavity index, $\theta_{ref} = 0.45$ to calculate a normalised
210 channel steepness index value, k_{sn} , using the 12.5 m resolution DEM, thus facilitating a
211 comparison between streams with a large range of drainage areas (Wobus et al., 2006).
212 Consistent along-strike increases in k_{sn} suggest a perturbation of the stream power law due
213 to changes in uplift rates caused by active normal faulting (Wobus et al., 2006; Figure 3d).

214 *3.2. Description of fault scarps*

215 Based on the topographic analysis we found evidence for five active fault scarps within the
216 Zomba Graben that we subsequently confirmed during a 6-week field campaign in 2018. In
217 this section, we describe the tectonic geomorphology and field observations of the five
218 faults which includes two border faults, Zomba and Lisungwe, and three intrarift faults,
219 Chingale Step, Mlungusi and Mtsimukwe. For each fault, we report the dimensions of the
220 fault scarps discovered at the base of footwall escarpments (all measurements are listed in

221 Table S1-S11 in the supplementary material), and describe the other features that led us to
222 conclude they are active. The measurements are shown in Figure 5 and summarised in Table
223 1.

224 The NW-dipping Zomba fault borders the eastern side of the Zomba Graben and has a ~50
225 km long escarpment with an active fault scarp at the base (Figure 5a and Figure S1). The
226 mean height of the fault scarp is 15.6 ± 5.2 m (Figures 4a & 5a) and normalised channel
227 steepness values (K_{sn}) increase where rivers cross the fault (Figure S1d). Higher K_{sn} values 3-
228 4 km into the hanging wall at the northern end of the fault are associated with the distal
229 edges of alluvial fans sourced from the Zomba Plateau (Figure S1d). The scarp is noticeably
230 steeper adjacent to the Zomba Plateau and a ~2 km step to the north-west occurs at the
231 northern end of the fault in front of the Zomba Plateau (Figure S1b-c). At the northern end
232 of the fault, triangular facets were observed in the field (Figure 6a). Large alluvial fans in the
233 hanging wall are composed of material derived from the Zomba Plateau and appear to have
234 been offset by the fault (Figure 6b). In the middle of the fault, a ~17 m high fault scarp was
235 observed during fieldwork (Figure 6c). At the base of the scarp, a zone of highly fractured
236 gneiss was observed, consistent with a fault zone (Figure 6d).

237 The NW-dipping Chingale Step fault (Figures 4a & 5b and Figure S2) is located approximately
238 10 km into the hanging wall of the Zomba fault and has formed a ~40 km long composite
239 scarp with evidence for multiple offsets (Figure 4a and Figure S2) and two segments (Figure
240 5b and Figure 7). Although the average scarp height along the whole fault is 19.6 ± 12.1 m
241 (black line Figure 5b), the scarp in the northern segment (25.2 ± 11.7 m) is higher on
242 average than the scarp in the southern segment (15.1 ± 10.6 m). In contrast, the height of

243 the lowermost offset is approximately constant along the entire length of the fault (5.7 ± 2.5
244 m; red line in Figure 5b).

245 Where river channels cross the Chingale Step fault, K_{sn} increases (Figure S2) and the
246 upstream long profiles are oversteepened (Figure 7c-d and Figure S3). Plotting the elevation
247 of the top of the oversteepened reaches shows two fault segments separated by a zone of
248 linkage (Figure 7). The river channel within this linkage zone shows two knickpoints, likely
249 related to the separate initiation of faulting of the southern and northern segments (Figure
250 7c). The consistent offset measured across the lowermost fault scarp (red line in Figure 5b)
251 suggests that these segments have linked in the time since the active fault scarp began to
252 be preserved, and are now operating as a single fault. A scarp was visible along all sections
253 visited, with a zone of fault gouge and fractured basement rocks visible at the base of the
254 fault scarps in exposed stream beds (Figure 8). At the Kalira river site, the fault surface itself
255 was visible with a polished surface and slickensides (Figure 8). The composite scarp was not
256 visible in the field, although most accessible locations were on the southern end of fault
257 where the difference between the total scarp and the lowermost scarp was smaller than
258 along the northern segment of the fault.

259 The SE-dipping Mlungusi fault has formed a ~20 km long scarp in the centre of the Zomba
260 Graben (Figures 4a & 5c) and has a mean scarp height of 6.9 ± 3.1 m. The slope map shows a
261 prominent scarp along the entire length of the fault (Figure S4), whereas K_{sn} values are only
262 elevated in the centre of the fault (Figure S4d). The Shire River crosses the southern end of
263 the Mlungusi fault, where it changes from a wide, meandering channel with a floodplain in
264 the hanging wall, to a narrow, incised channel with an associated set of rapids as it crosses
265 into the footwall (Figure 9). Along strike from this location, a steep fault scarp was observed

266 in the field (Figure 9b) which has offset alluvial-lacustrine deposits found on the floor of the
267 graben. The central segment of the fault scarp is covered by lacustrine beach deposits
268 characterised by rounded to subrounded pebbles within a clast-supported sandy matrix
269 (Figure 9b). Long-term footwall uplift has led to drainage reorganisation as rivers draining
270 into the axial Mtsimukwe River in the footwall of the Mlungusi fault are restricted to the
271 western side of the channel, distal to the fault (Figure S4d).

272 The E-dipping Mtsimukwe fault has formed a ~13 km long scarp trending ~N-S (Figures 4a
273 5d). The mean scarp height is 3.6 ± 0.7 m. The slope map shows that the fault scarp is best
274 preserved in the central part of the fault except a ~2 km long section where the fault
275 intersects a road (Figure S5b). The streams that cross the fault show a small increase in K_{sn}
276 value, with the largest increase in centre of the fault, but there is no evidence of
277 segmentation (Figure S5).

278 The E-dipping Lisungwe fault is the western border fault of the Zomba Graben (Figure 2 and
279 Figure 5e). High slope values have a linear trend aligned with an increase in K_{sn} in the
280 streams that cross the escarpment (Figure S6). The mean offset across the late-Quaternary
281 fault scarp is 10.0 ± 6.7 m (Figures 4a & 5e). No fieldwork was conducted on the Mtsimukwe
282 and Lisungwe faults fault due to access reasons, but the remote sensing observations
283 described above and seen in Figure S5 and Figure S6 are similar to the observations on faults
284 (e.g. Zomba, Chingale Step and Mlungusi faults) where fieldwork was able to confirm the
285 evidence gained remotely.

286 *3.3. Fault Kinematics and Extension Direction*

287 The extension direction within the Zomba Graben is approximately NW-SE, as inferred from
288 measurements of slickensides along the Chingale Step Fault (Figure 8b-d; plunging 52°

289 towards 301°) and elsewhere in the graben (Bloomfield, 1965; Chorowicz and Sorlien, 1992).
290 The trend of these slickensides is orthogonal to the surface traces ($015 \pm 11^\circ$) of the 5 faults,
291 indicating dip-slip displacement. This contradicts previous rift-wide estimates of NE-SW
292 extension (Delvaux and Barth, 2010; Figure 1). However, these estimates likely reflect that
293 normal faulting events in the Malawi Rift are approximately purely dip-slip despite regional
294 changes in fault strike (Williams et al., *in review*), as also indicated by the 2009 Karonga
295 Earthquakes (Biggs et al., 2010), the 1989 Salima earthquake (Jackson and Blenkinsop,
296 1993), and a $M_w=5.6$ earthquake in March 2018 (red focal mechanism in Figure 1b).
297 Previous rift-wide estimates of NE-SW extension do therefore not necessarily apply to the
298 Zomba Graben.

299 *3.4. Distribution of contemporary strain in the Zomba Graben*

300 We found evidence for five active faults within the Zomba Graben. The largest fault scarp
301 was found along the Chingale Step fault, which is located in the hanging wall of the border
302 fault. The distribution of fault scarps within the graben is a notable feature. The across strike
303 spacing between the active faults is approximately 10-15 km and the height of each fault
304 scarp ranges between ~4 and ~20 m. Thus, the strain within the Zomba Graben over the
305 time period that the active fault scarps have formed is not localised on a single major border
306 fault but instead distributed across the width of the rift. To explore whether this pattern of
307 distributed faulting is a long-lived feature of the rift, we make an assessment of the longer
308 term deformation in the region and compare the patterns of strain over both timescales.

309 **4. Fault activity since the onset of rifting**

310 All five faults have formed active fault scarps at the base of a larger footwall escarpment,
311 which represents the longer-term topographic record of faulting. In this section, we

312 measure this longer-term record of faulting which allows us to compare the distribution of
313 strain associated with the active fault scarps (Section 3) with the strain in the Zomba Graben
314 since the onset of rifting. We consider the height of the footwall escarpments to represent
315 the minimum throw on each fault since the onset of rifting, as erosion of the footwall and
316 deposition of sediment in the hanging wall have likely reduced the total faulted relief. As
317 before, we extracted fault-perpendicular topographic profiles, but in this case, we stacked in
318 1 km wide bins and the profiles were 6 km long. As progressive fault slip results in rotation
319 of the footwall block (Wernicke and Axen, 1988), we used the maximum elevation of the
320 footwall within 3km of the fault (relative to the base of the scarp), rather than fitting
321 regression lines to the footwall and hanging wall slopes (see Figure 4b for examples from
322 the centre of each fault).

323 We assume that there was little significant relief across the escarpment at the onset of
324 faulting and that there is no spatial variation in erosion and deposition rates across the
325 Zomba Graben. Neither of these assumptions apply to areas that have experienced local
326 intrusions into the Precambrian basement, therefore we disregard measurements made in
327 these areas. To test these assumptions, we calculate the local relief: the range of
328 topography within a 1 km window. Although local relief is affected by basin infilling, low
329 values of local relief both inside and outside the graben would suggest a lack of pre-existing
330 topography, and spatially uniform values of local relief in the hanging wall and footwall of
331 the faults would suggest a lack of significant erosion or sedimentation. Meanwhile, low
332 values of local relief are thought to be consistent with low erosion rates (Montgomery and
333 Brandon, 2002). To assess the depth of hanging wall sediments we used logs from drinking
334 water boreholes described on the geological maps (Figure 2a; Bloomfield 1965).

335 *4.1. Footwall escarpments in the Zomba Graben*

336 The most prominent footwall escarpment in the Zomba Graben is associated with the
337 Zomba fault. The mean relief of the footwall of the Zomba fault is 326 ± 113 m (Figure 5a).
338 The Zomba Plateau lies to the north of the fault, and the footwall here is about 1 km higher
339 than in the basement gneisses to the south. As the Zomba Plateau has a higher elevation
340 than the surrounding landscape, we suspect it violates our assumption of no significant
341 topography prior to rifting hence we do not include it in our measurement of the mean
342 footwall uplift.

343 Along the Chingale Step fault, the mean height of the footwall escarpment is 84 ± 18 m
344 (Figure 5b; excluding measurements associated with pre-existing topography from the
345 Chingale Ring Structure and the Chinduzi Hill nepheline-syenite province; Figure S2).
346 Boreholes drilled on the hanging wall reached depths of 41.5 m and 49.3 m without
347 reaching basement rock (Figure S2; Bloomfield, 1965).

348 The mean height of the footwall of the Mlungusi fault is 19 ± 5 m (Figure 5c). The low
349 topographic expression of this fault may be due to higher sedimentation rates within the
350 centre of the graben as no footwall bedrock is mapped on the geological maps (Figure 2a).
351 However, during fieldwork, bedrock was observed in the footwall of the fault in all locations
352 visited, shown in Figure 19.

353 The mean height of the footwall along the Mtsimukwe fault is 38 ± 12 m (Figure 5d). The
354 footwall of the fault is made up of Proterozoic gneisses, and two boreholes drilled through
355 the hanging wall sediments in 1954 (see Figure S5d for locations) penetrated the basement
356 gneisses at depths less than 37 m (Bloomfield, 1965).

357 The Lisungwe fault, the western border fault of the Zomba Graben, is divided into two
 358 segments separated by an area of marble and metadolomite within the basement gneiss.
 359 However, this segmentation visible in the surface trace of the fault is only weakly visible in
 360 the height of the footwall escarpment. The footwall escarpment is 277 ± 76 m high (Figure
 361 5e) with Tertiary-recent sediments in the hanging wall (Figure S6b-f). As the fault is located
 362 within the topography of the Precambrian Kirk Range, we removed the pre-existing
 363 topographic signal by differencing the long-wavelength topography (calculated using a low-
 364 pass Gaussian spatial filter where $6\sigma = 50$ km) and the present-day topography. This residual
 365 relief picks out the high frequency fault-related topography, which is identified by the sharp
 366 elevation contrast between the footwall and the back-tilted hanging wall block, but does
 367 not significantly affect the estimated height of the footwall escarpment (Figure S6g-h).

368 **5. Strain distribution across the Zomba Graben**

369 We calculate the cumulative strain since: i) the onset of faulting; and ii) since the formation
 370 of the fault scarps (assuming an approximately uniform age of the fault scarps) which allows
 371 us to assess any changes in the relative distribution of strain across the rift through time.

372 We calculate finite strain across the Zomba Graben using the England and Molnar (1997)
 373 adaptation of the Kostrov (1974) expression of strain rate:

$$374 \quad \bar{\epsilon}_{ij} = \frac{1}{2a} \sum_{k=1}^k \frac{L^k s^k}{\sin \vartheta^k} (\hat{u}_i^k \hat{n}_j^k + \hat{u}_j^k \hat{n}_i^k) \quad (1)$$

375 where a is the surface area of the region, L^k is the length of fault k in that region, s^k is the
 376 slip of fault k in that region, ϑ^k is the dip of fault k , \hat{u}^k is a unit vector in the direction of slip
 377 and \hat{n}^k is a unit vector normal to the fault plane. The advantage of this adaption is that it is
 378 independent of seismogenic thickness or shear modulus. We assume pure dip-slip faults

379 that strike parallel to each other (see section 3.3), such that ε_{22} is equal to zero, and ε_{11} is
380 perpendicular to the mean strike of the faults. The only unknown term is fault dip, for which
381 we assume an Andersonian value of 60° , although using a randomly selected fault dip,
382 within the usual range for normal faults (45° - 60° ; Collettini and Sibson 2001), produces
383 similar results (Figure S7). We calculated the strain within 5 km (across strike) by 50 km
384 (along strike) rectangular boxes with long axes orientated 100° .

385 We find differences between the distribution of strain since the onset of faulting and strain
386 over the time period since fault scarps have been preserved (Figure 10). 75 ± 18 % of strain
387 since the onset of faulting is focussed on the border faults (Zomba and Lisungwe faults;
388 Figure 10b). The relief demonstrates that significant motion has occurred on both these
389 faults, but the graben is asymmetric, with 55 ± 16 % of the strain in the region of eastern
390 border fault (the Zomba fault), and 19 ± 3 % of strain across the western border fault
391 (Lisungwe fault; Figure 10b) with the remaining strain accommodated on intrabasin faults.
392 Although the Zomba and Lisungwe faults only partially overlap across strike, which may
393 indicate they initially formed as separate half-grabens, both the profiles of strain (Figure
394 10b-c) and topography (Figure 2c) suggest that the Zomba Graben is a full-graben as
395 inferred by Lao-Davilila et al. (2015). However, the eastern border fault accommodates more
396 strain than the western border fault resulting in the asymmetric characteristics of a half-
397 graben (Figure 10c; see also Ebinger et al., 1987).

398 In contrast, over the time period during which the active fault scarps have been preserved,
399 only 50 ± 17 % of strain has occurred across the two border faults (Figure 10a). In this time
400 period, the strain dominantly occurred across the Chingale Step fault (35 ± 21 %) and the
401 Zomba border fault (42 ± 15 %), both W-dipping faults on the eastern side of the graben.

402 This relative change is consistent with either a decrease in the absolute strain across the
403 Zomba border fault or an increase in the absolute strain across the Chingale Step fault, and
404 without better time constraints, we cannot distinguish between these possibilities.
405 However, we can conclude that a lower proportion of the strain occurred across the border
406 fault in this period relative to the time since the onset of faulting (Figure 10b-c). The high
407 uncertainty of the late Quaternary strain calculation is caused by high scarp height standard
408 deviation on the Chingale Step fault, the fault which accommodates the majority of the
409 intra-rift strain. This high standard deviation is due to segment linkage during the late
410 Quaternary which has caused an along strike displacement profile with dual maxima (Figure
411 5b & 7a).

412 A pattern of increasing strain within the rift interior also occurs on the western side of the
413 graben, where the proportion of strain across the Lisungwe border fault decreased from $19 \pm 3\%$
414 to $6 \pm 2\%$, while the proportion across the Mlungusi fault increases from $2 \pm 0.5\%$ to $8 \pm 3\%$
415 and the proportion of strain across the Mtsimukwe fault remains the same within error
416 (from $11 \pm 6\%$ to $8 \pm 3\%$; Figure 10b-c). Although the lowest topography is currently found in
417 the centre of the rift, the graben still retains its asymmetric character, with $78 \pm 37\%$ of the
418 strain across W-dipping faults in the east and $22 \pm 8\%$ across E-dipping faults in the west.
419 The distributed pattern of strain means that neither full- or half-graben models apply.

420 Due to a lack of geochronological dates in the Zomba Graben, we do not know the exact
421 timing of either the onset of faulting or the age of the fault scarps. Therefore, we are unable
422 to calculate the strain rate within the region or comment on changes in strain rate through
423 time. As a consequence, we have limited our analysis and discussion to changes in the
424 *relative* distribution of strain in the two different time periods identified. We discuss the

425 validity of this comparison before commenting on the implications that can be drawn from
426 our results for the dynamics of strain in early stage rifts.

427 **6. Discussion and the timing of rifting in the Zomba Graben**

428 *6.1. Preservation of fault related topography since the onset of rifting*

429 The Zomba Graben is an ideal location for using topography to study the activity of faults
430 since the onset of rifting, as a lack of pronounced pre-existing topography is coupled with
431 low sedimentation and erosion rates. In this section we verify that these conditions apply
432 within the graben and show how our measurements of footwall topography allow a
433 comparison between the activity of each fault since the onset of rifting.

434 In order for our measurements of footwall relief within 3 km of the surface trace of each
435 fault to be a valid marker of faulting since the onset of rifting, we assumed that there was a
436 lack of significant topography in the graben prior to the current rifting episode. We also
437 assumed that rates of erosion and sedimentation across the graben are approximately
438 uniform. If this assumption is valid, then while the absolute values of footwall topography
439 that we measure may have been affected by erosion and sedimentation, the spatial
440 patterns of fault activity since the onset of rifting should have been preserved. We tested
441 these assumptions by calculating the local relief (Figure 11). In areas away from the faults
442 and intrusions, the local relief (range of topography within a 1 km window of each point) is
443 generally < 50 m and there is no discernible spatial pattern of different local relief values
444 within the Zomba Graben. In the footwall of the Zomba fault, the local relief is more variable
445 but still includes large areas where the local relief is < 50 m with similar values to the floor
446 of the Zomba Graben. Locally high values are associated with intrusions into the
447 Precambrian basement. The low values in the footwall of the Zomba fault (Figure 11a) and
448 the preserved back tilted footwall topography (Figure 2c) is consistent with the preservation

449 of a low-relief pre-rift surface that has experienced little erosion since the onset of faulting.
450 The similarity of the low relief values between the footwall of the Zomba fault and the floor
451 of the Zomba Graben suggests regionally low erosion and sedimentation rates, with the
452 exception of the areas immediately surrounding the faults where values are locally elevated
453 (Figure 11a). The median local relief in the region is 47 m (Figure 11b), which is consistent
454 with low average erosion rates of $< 1 \text{ mm yr}^{-1}$ when compared to global comparisons of
455 local relief and mean erosion rates (Montgomery and Brandon, 2002).

456 Where we were able to compile data from drinking boreholes drilled in the 1950s,
457 basement depth in the centre of the graben (in the hanging wall of the Mtsiumkwe fault),
458 was $< 40 \text{ m}$, providing further evidence for low sedimentation rates since the onset of
459 rifting. If the Mtsimukwe fault has been active since the Pliocene ($\sim 5 \text{ Ma}$; the minimum age
460 of the rifting in the region), then the sedimentation rate would be $8 \times 10^{-3} \text{ mm yr}^{-1}$, which is
461 approximately 1-2 orders of magnitude lower than sedimentation rates recorded in
462 northern Lake Malawi (Johnson et al., 2002), but approximately equal to our minimum
463 estimate of the throw rate of the fault ($5 - 10 \times 10^{-3} \text{ mm yr}^{-1}$, assuming the footwall
464 topography we have measured is an estimate of the minimum fault throw). Although this
465 leaves open the possibility that some faults in the centre of the rift may have been buried by
466 sedimentation, the lack of any more fault-related geomorphology along the River Shire
467 (other than that seen in Figure 9), suggests we have not missed any major faults. Moreover,
468 combining size-frequency and displacement-length fault scaling relations indicates that with
469 a maximum fault length of $\sim 50 \text{ km}$ and a lower fault length of $\sim 10 \text{ km}$ we will have
470 accounted for $\sim 90 \%$ of the tectonic strain in the region (Scholz and Cowie, 1990).

471 *6.2. Age of the fault scarps*

472 The faults in the Zomba Graben have offset the sedimentary cover, which includes coarse-
473 grained sandstones of the Matope beds, as well as sandy, pebbly, fluvio-lacustrine deposits
474 in the centre of the rift (Figure 2). However, the ages of these sediments have not been
475 verified by geochronological dating. Dulanya (2017) correlated the Matope beds with the
476 Plio-Pleistocene development of the graben (the 'lower Upper Shire section' in Dulanya,
477 2017) based on stratigraphic mapping and age estimates from Bloomfield (1965). Optically-
478 Stimulated Luminescence (OSL) ages for surficial palaeo-fluvio-lacustrine deposits near Lake
479 Chilwa, 20-30 km to the east of the Zomba Graben, are <50 ka (Thomas et al., 2009; Figure
480 1).

481 The floor of the Zomba Graben, which is covered by both fine-grained fluvio-lacustrine
482 sediments, and the coarse-grained Matope beds, lies ~10 m above the current level of lake
483 Malawi. Since the Mid-Pleistocene Transition (MPT; ~800 ka), the level of Lake Malawi has
484 fluctuated with high-stands up to ~150 m higher than the present lake level and low-stands
485 up to ~500 m lower (Lyons, 2015; McCartney and Scholz, 2016; Owen et al, 1990). Although
486 the exact magnitudes of these high- and low-stands are difficult to constrain, the last high-
487 stand is thought to have begun at ~75 ka and continued to the present day (Ivory et al.,
488 2016). Consequently, during high-stands, lacustrine conditions presumably flooded the
489 Zomba graben, whereas low-stands were accompanied by the deposition of coarse-grained
490 sediments (Lyons et al., 2011; 2015). Therefore, the fine- and coarse-grained sediments on
491 the floor of the Zomba graben were likely deposited during the late-Quaternary, which
492 implies that the fault scarps we observe that have offset and been draped by these
493 sediments have formed during the late Quaternary time period: <800 ka and maybe as
494 recently as <75-50 ka. A number of observations suggest that this interpretation is

495 reasonable: i) The fault scarps are generally steep (Figure 3b & Figure 4a); and ii) fieldwork
496 along the Mlungusi fault found the fault draped by a layer of sub-rounded to rounded
497 cobbles (3-6 cm) in a clast-supported sandy matrix that is consistent with a lacustrine
498 palaeobeach deposit, likely deposited during a high-stand within Lake Malawi during the
499 late Quaternary (since the MPT ~800 ka). The heights of the measured scarps are generally
500 greater than the displacements seen in single events for normal faults (e.g. Leonard, 2010),
501 therefore the scarps should be viewed as composite scarps representing the cumulative
502 offset from multiple earthquakes.

503 **7. Strain Migration and Fault Evolution during rifting**

504 Current conceptual models suggest that the transition from rift border faults to intrarift
505 deformation occurs over timescales of 10-15 Ma in the East African Rift as lithospheric
506 thinning leads to asthenospheric upwelling and production of magmatic fluids (Ebinger,
507 2005; Ebinger and Scholz, 2012; Buck, 2004). However, both our study and seismic
508 reflection profiles of Lake Malawi (McCartney and Scholz, 2016) show that both border
509 faults and intrabasin faults are active even in this amagmatic rift segment. The reflection
510 studies have been unable to constrain the timing of the transition to axial deformation, but
511 here we show that the proportion of strain accommodated in the interior of the Zomba
512 Graben has increased from $25 \pm 8\%$ since the onset of rifting to $50 \pm 25\%$ over the since the
513 fault scarps have been preserved in the late Quaternary (Figure 10). We discuss four
514 mechanisms that may explain this changing strain pattern across a distributed network of
515 faults in an apparently non-volcanic rift: 1) cessation of border fault activity; 2) lithospheric
516 flexure; 3) a hidden fluid phase; and 4) transient changes associated with fault linkage. We
517 then discuss whether the pattern of strain can be explained by the lithospheric structure,
518 including crustal heterogeneities.

519 *7.1. Cessation of Border Fault Activity and Lithospheric Flexure*

520 The maximum amount of displacement a normal fault can accumulate, before it becomes
521 more favourable to form a new fault, is thought to be controlled by a combination of
522 effective elastic thickness, seismogenic thickness, and surface processes, such as footwall
523 erosion and hanging wall deposition (Scholz and Contreras, 1998; Olive et al., 2014). The
524 exact mechanism of fault abandonment is debated but it generally requires large total
525 displacements (>5 km) that lead to an increase in the flexural restoring force and/or rotation
526 of the fault dip to unfavourable angles (Scholz and Contreras, 1998; Goldsworthy and
527 Jackson, 2001).

528 When rift border faults are abandoned, migration of fault activity into the hanging wall of
529 the previously active fault is widely observed (see Goldsworthy and Jackson, 2001, for
530 further discussion). Accardo et al. (2018) propose that the ~6 km of throw on the border
531 faults in the northern and central basins of Lake Malawi indicates that these faults are
532 approaching their maximum size and that the migration of strain into the rift interior is
533 imminent or now occurring (e.g. the Karonga earthquake sequence; Biggs et al., 2010;
534 Kolawole et al., 2018). However, this mechanism is unlikely to cause the late-Quaternary
535 intrarift faulting in the Zomba Graben, because both the border and intrarift faults have
536 been active during the late-Quaternary and the topographic relief is relatively small (<1 km).
537 Similarly, bending forces associated with flexure of the border fault hanging wall can induce
538 strain in the intrarift region within the upper crust, and although this has been proposed for
539 Lake Malawi (Kolawole et al., 2018), the low throws and thick elastic crust (~30 km) in the
540 Zomba Graben would generate negligible flexural strain (Jackson and Blenkinsop 1997,
541 Muirhead et al 2016).

542 *7.2. Influence of fluids*

543 Magma-assisted rifting, where magmatic fluids and volatiles enable extension at lower
544 stresses than the available tectonic forces, is thought to play an important role in facilitating
545 rifting in thick continental lithosphere (Buck, 2004; Ebinger et al., 2017). There is no
546 evidence of active magmatic activity or more than a few kilometres of crustal thinning in the
547 Zomba Graben (Reed et al., 2016, Wang et al 2019). Furthermore, the geochemistry of hot
548 springs in southern Malawi does not suggest a magmatic influence (Dulanya et al., 2010)
549 and the nearest active volcano, Rungwe, is located ~700 km to the north (Figure 1).
550 However, decompression melting or lower crustal magmatic intrusion might not lead to
551 perceptible surface effects, and there is evidence for non-zero crustal thinning beneath
552 southern Malawi (Wang et al., 2019). In an example further south and west in the EARS, low
553 seismic velocities suggest decompression melting in the upper asthenosphere of the
554 Okavango Delta, despite a very low level of extension and no surface volcanism or evidence
555 for mantle upwelling (Yu et al., 2017).

556 Earthquakes occur in Malawi throughout the 38-42 km thick crust (Tedla et al., 2011;
557 Jackson and Blenkinsop, 1993; Craig et al., 2011), consistent with estimates of effective
558 elastic thickness in excess of 30 km (Ebinger et al., 1999). This can occur in magmatic rift
559 zones if border faults penetrate into the lower crust, or if melt and volatile migration into
560 the lower crust causes localised weakening, which can also trigger seismicity (Ebinger et al.,
561 2017). However, current imaging of the lower crust in Malawi does not allow us to
562 discriminate between these mechanisms and alternative explanations for the deep
563 seismicity, that include 1) a dry, strong, granulite facies lower crust (Jackson et al., 2004), 2)
564 above average lithospheric thickness (Chen and Molnar, 1983), or 3) localised zones of weak
565 rheology within a strong, elastic lower crust (Fagereng, 2013).

566 *7.3. Transient Changes Associated with Fault Evolution*

567 The process of fault segment linkage can lead to an increase in fault slip rates over
568 timescales of ~ 100 ka (Taylor et al., 2004). This can subsequently enable a newly linked fault
569 to accommodate a greater proportion of the regional extension rate (Taylor et al., 2004;
570 Cowie et al., 2005). We infer segment linkage on the Chingale Step fault that may have
571 occurred as recently as the late-Quaternary (Figure 7), suggesting that faults in this region
572 grow through linkage that occurs before they have accumulated significant (i.e. >1 km)
573 offsets and on short timescales of < 800 ka (and possibly $< \sim 50$ ka). This rapid linkage, before
574 continued slip accumulation, is consistent with the recent hybrid fault growth model where
575 slip accumulates at a constant fault length after an initial growth phase (Rotevatn et al., *in*
576 *press*). The segment linkage we observe may have driven the apparent increase in strain
577 within the rift interior. However, when faults are closely spaced, as they are in the Zomba
578 Graben, across-strike co-seismic elastic stress changes can also drive transient variations in
579 fault slip rates (Cowie et al., 2012). Elastic interactions between faults can therefore change
580 slip rates on individual faults over the time scale of a few earthquake cycles (Cowie et al.,
581 2012). With only one example, we cannot tell whether the observed increase in intra-rift
582 strain is a long-lived feature of the rift, or a transient feature only representative of the time
583 window we sampled (i.e. since the fault scarps have been preserved in the late-Quaternary).
584 Nonetheless, although elastic fault interactions are likely contributing to the temporal
585 evolution of strain across the network of faults in the Zomba Graben, they cannot explain
586 the initial formation of a distributed fault network.

587 *7.4. Lithospheric Structure*

588 The roles of crustal thickness, lower crustal rheology, and lithospheric thermal structure in
589 developing end-member narrow or wide rifts have been well studied (e.g. Buck, 1991;

590 Huismans and Beaumont, 2007). Typically, strain in narrow rifts concentrates in the weakest
591 part of the lithosphere while in wide rifts, lower crustal viscous flow drives distributed
592 deformation in the upper crust (Buck, 1991). Southern Africa has unusually thick continental
593 lithosphere (140-180 km; Craig et al., 2011), resulting from multiple episodes of orogenic
594 thickening (Fritz et al., 2013), and 38-42 km thick crust (Tedla et al., 2011, Wang et al 2019).
595 The available constraints on the lithospheric properties of the Zomba Graben suggest that
596 conditions are similar to the very young (120-40 Ka) Okavango Rift (Craig et al., 2011).
597 However, we have demonstrated that strain within the narrow (<50 km) Zomba Graben is
598 distributed whereas the Okavango Rift is >150 km wide and deformation is localised to <50
599 km wide zones at the edges of the rift (Ebinger and Scholz, 2012). Thus, the currently-
600 available constraints on lithospheric properties cannot explain the differences between
601 these two rifts.

602 Numerical simulations demonstrate that a strong, ductile lower crust promotes distributed
603 faulting due to enhanced upper crustal fault interactions during rifting (Heimpel and Olson,
604 1996). The crust in southern Malawi is made up of high grade metamorphic fabrics that lack
605 hydrous minerals with low friction coefficients (Hellebrekers, et al. *in review*), and exposed
606 fault rocks are not demonstrably different in composition from these high grade basement
607 rocks (Williams et al., *submitted*). Thus, distributed faulting in the Zomba Graben cannot be
608 attributed to reactivation of pre-existing frictionally weak planes in the brittle regime.

609 However, lateral heterogeneity in the lower crust, such as an anastomosing shear zone
610 system, would enable strain localisation in lower viscosity zones in an otherwise cold, strong
611 layer (Fagereng, 2013). Such localisation at depth may guide deformation patterns in the

612 overlying brittle crust, such as the oblique strike of faults in southern Malawi, relative to the
613 regional extension direction (Hodge et al., 2018; Williams et al., *in review*).

614 The hypothesis of a lower crust with rheological heterogeneity: i) satisfies the requirements
615 for a dominantly strong lower crust; ii) can explain the deep seismicity observed in the
616 Malawi rift; iii) facilitates the formation of a distributed fault network at the surface; and iv)
617 does not require pre-existing frictional weaknesses in the upper crust. This suggests that
618 patterns of strain distribution in amagmatic rifts may not be solely controlled by rift
619 maturity; instead, rheological heterogeneity in the lower crust and elastic fault interactions
620 in the upper crust may also have important effects.

621 **8. Implications for Seismic Hazard.**

622 Our analysis of high-resolution satellite topography has identified five active faults in the
623 Zomba Graben, but current assessments of seismic hazard in Malawi do not extend south of
624 Lake Malawi (Midzi et al., 1999; Hodge et al., 2015). Based on standard scaling laws
625 (Leonard, 2010), these 15-50 km long faults could host earthquakes of Mw 6.3-7.1,
626 assuming that faults do not fail in smaller individual segments, and that multiple separate
627 faults do not slip in the same event. The administrative regions within 30 km of the Zomba
628 Graben contains ~22% of the Malawian population (~ 4 million people), including the major
629 population and administrative centres of Blantyre (population: ~800,000) and Zomba
630 (population: ~100,000; Malawi National Statistics Office, 2018). The proximity of these
631 newly-identified faults to this number of people presents a significant challenge for both
632 seismic risk and regional development. Furthermore, the path of the Shire River is strongly
633 affected by rift topography, and a large earthquake on the Mlungusi fault in particular could
634 affect both irrigation and flooding on a regional scale as well as dam stability on the Shire

635 River. This is particularly important as ~80% of Malawi's electricity is generated by
636 hydroelectric dams within the middle Shire valley (Taulo et al., 2015).

637 In slowly deforming regions and/or regions with poor seismic detection infrastructure, even
638 decadal long instrumental catalogues are unlikely to record a representative enough sample
639 of seismicity to sufficiently assess seismic hazard. Including fault maps in probabilistic
640 seismic hazard analysis can overcome this problem (Pace et al., 2018), but this requires
641 estimates of both earthquake magnitude and recurrence interval. In Lake Malawi, where
642 fault slip rates have not been measured, Hodge et al. (2015) assigned the plate motion to
643 individual faults in their hazard assessment. However, the discovery of multiple, subparallel
644 fault scarps in the Zomba Graben suggests longer earthquake recurrence times and lower
645 probabilities of peak ground acceleration for a given return period. Furthermore, increases
646 in the number of faults within a region also leads to more variable fault slip rates, which in
647 turn can result in earthquake clusters (Cowie et al., 2012). The 2009 Karonga earthquake
648 sequence in northern Malawi illustrates that such clusters occur in the East African Rift
649 (Biggs et al., 2010), but instrumental records are not sufficient to determine how
650 widespread this behaviour is. Earthquake clusters present an additional challenge for
651 seismic hazard assessment as they are not captured by time-independent probabilistic
652 seismic hazard assessments.

653 **9. Conclusion**

654 We analyse high-resolution TanDEM-X data to identify late-Quaternary faults scarps on five
655 faults in the Zomba Graben, southern Malawi. We compared the activity of the faults in the
656 graben since the onset of rifting with the displacement accumulated since fault scarps have
657 been preserved at the base of footwall escarpments. We show how 75 ± 18 % of strain was

658 distributed on the border faults since the onset of rifting whereas in the more recent past,
659 since the active fault scarps have been formed, ~50% of the strain has accumulated within
660 the rift interior. This presents new insights into the behaviour of rifts during the incipient
661 stages of continental extension in a region of thick lithosphere and no active volcanism. In
662 contrast to the prevailing paradigm, it suggests that early, amagmatic stages of continental
663 extension can be distributed across both rift border and intra-rift faults. We find evidence
664 for linkage of fault segments within the rift interior that occurs prior to the accumulation of
665 significant fault offset, and possibly on rapid, late-Quaternary timescales. While the overall
666 mode of rifting is likely to be controlled by the rheology of the lithosphere, we suggest that
667 upper-crustal fault interactions and strength variations within the lower crust can lead to
668 spatially distributed and temporally transient faulting within early stage rifts. Finally, we find
669 that the onshore rift in Southern Malawi represents a significant, but previously
670 unappreciated source of seismic hazard within the East African Rift.

671 **10. Acknowledgements**

672 This work was funded by the EPSRC project 'Prepare' (EP/P028233/1), funded under the
673 Global Challenges Research Fund. We thank Kondwani Dombola for his assistance with
674 fieldwork planning and logistics. TanDEM-X data were obtained via DLR proposal
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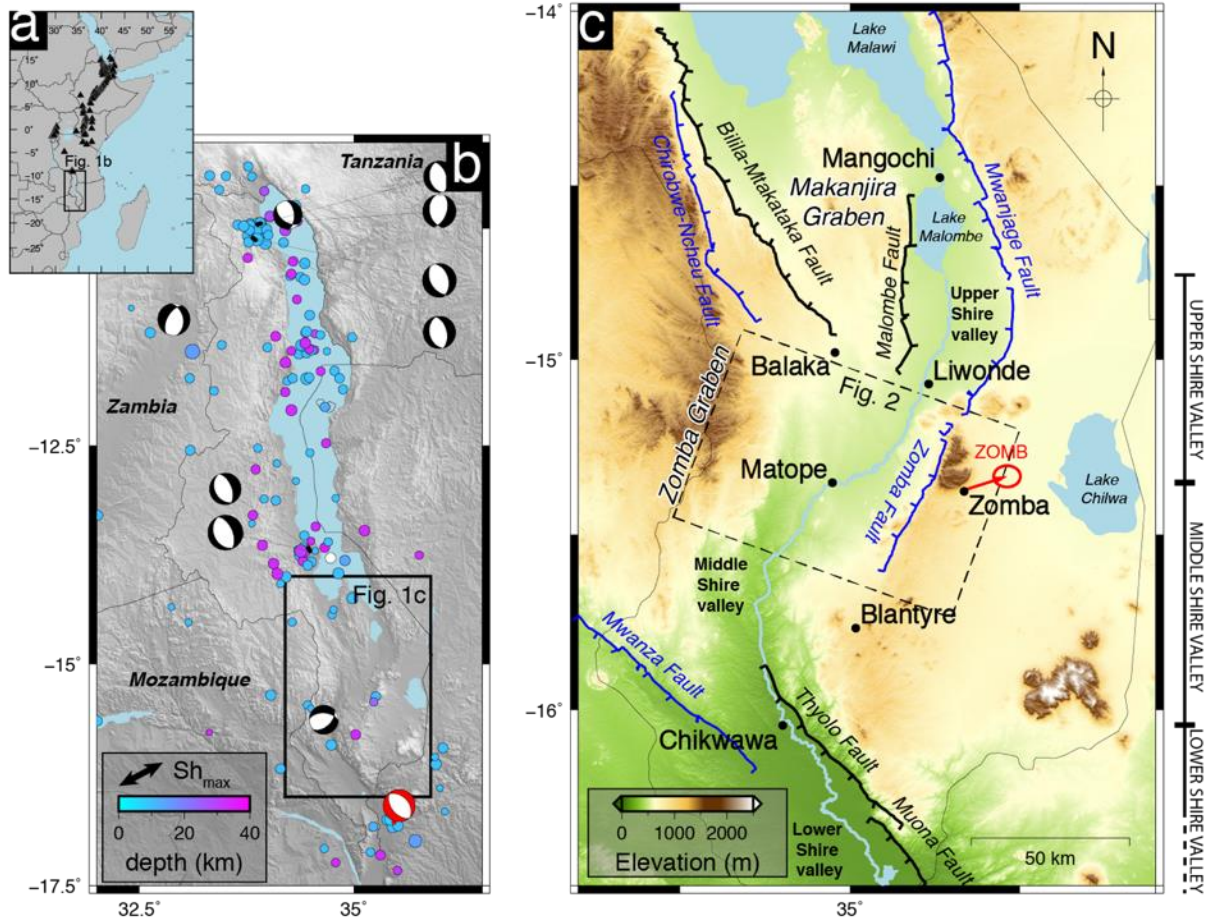
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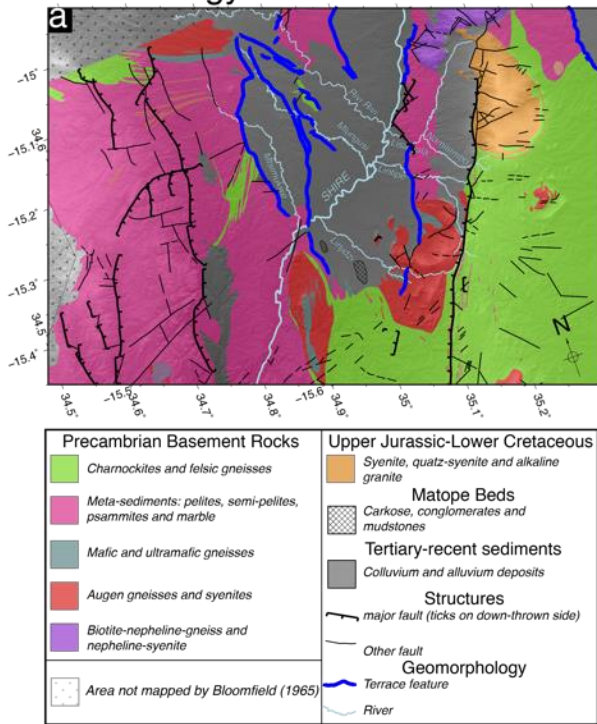
952 Figure 1: The location of the Zomba Graben within the East African Rift and Malawi. (a) The
 953 location of Malawi within the East African Rift. Black triangles show the active volcanoes
 954 within the rift. (b) Seismicity in the Malawi rift and the location of the Zomba Graben. Dots
 955 show National Earthquake Information Centre (NEIC) earthquakes from 1971-2018 coloured
 956 by depth. Focal mechanisms for all events greater than $M_w 5.0$ are from Craig et al. (2010)
 957 with the exception of the red focal mechanism which shows the location and CMT
 958 mechanism of the 17th March 2018 Nsanje earthquake. The Sh_{max} direction is from Delvaux
 959 and Barth (2010). (c) Overview map of southern Malawi showing the location of the Zomba
 960 Graben relative to other known faults in the region. Faults where scarps have been detected
 961 and measured are indicated in black. Faults which are suspected as active but with no

962 measurements of throw or where no fault scarp has previously been detected are shown in

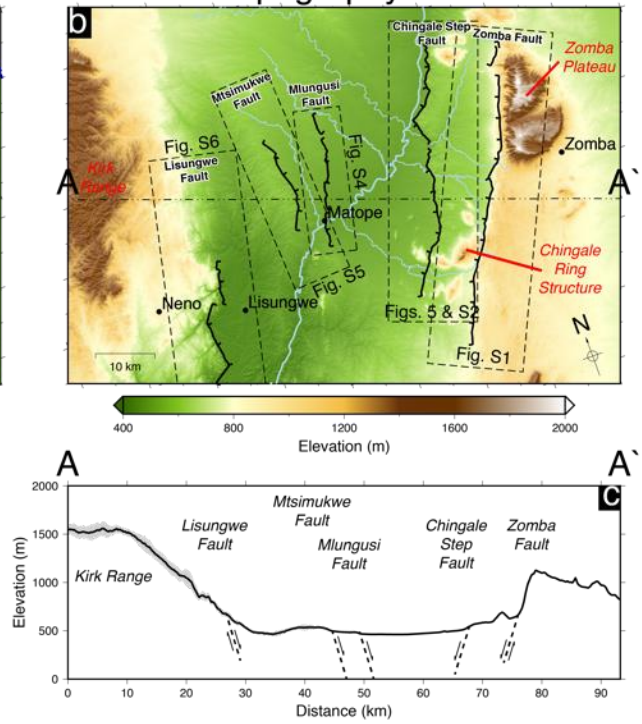
963 blue. GPS vector from Stamps et al. (2018) is shown in red.

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Geology of the Zomba Graben



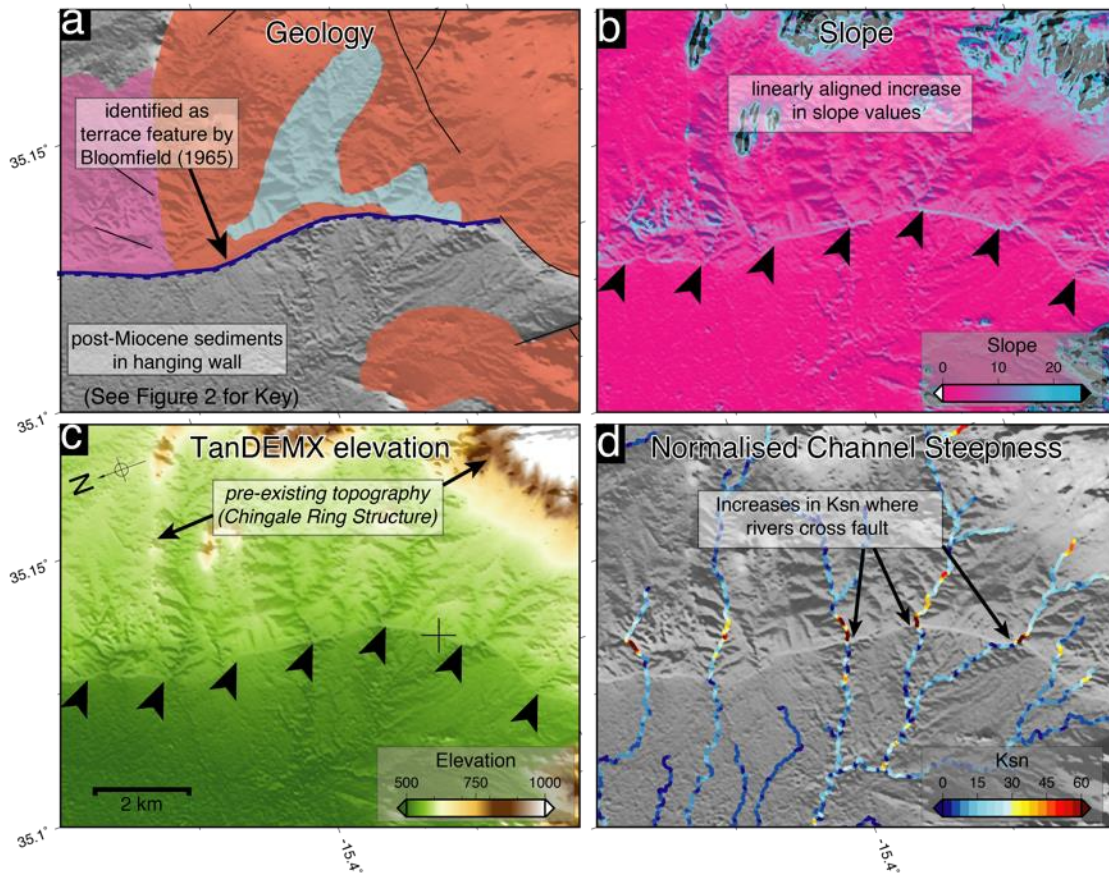
TanDEMx Topography & Active Faults



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966 Figure 2: Geology and topography of the Zomba Graben. (a) Geological map of the Zomba
 967 Graben adapted from Bloomfield (1965). (b) TanDEMx digital elevation model of the Zomba
 968 Graben. (c) Swath topographic profile across the Zomba Graben. The gray shading indicates
 969 the maximum and minimum topography in a 5km either side of the dashed line in part b.
 970 The solid line indicates the mean elevation.

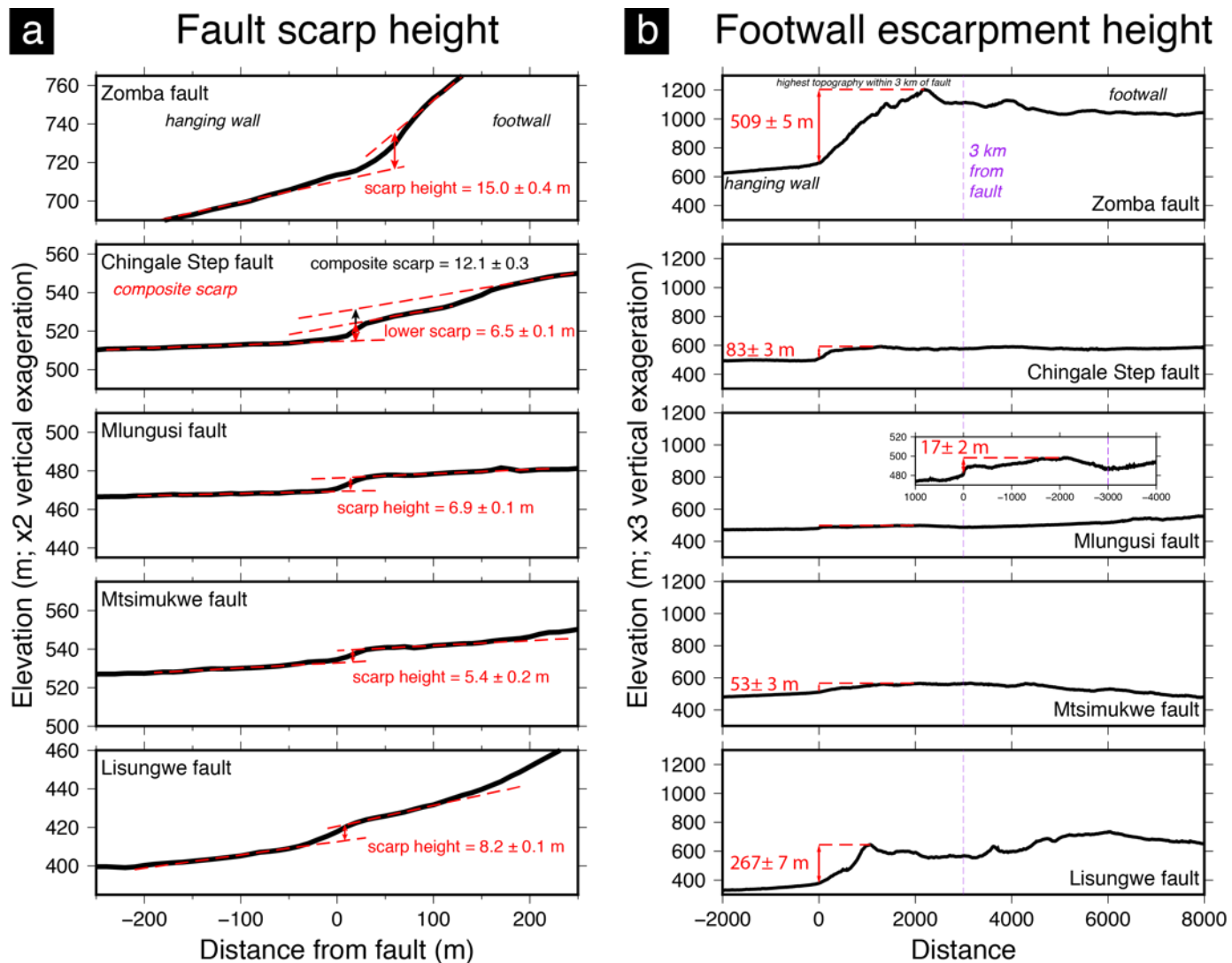
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972

973 Figure 3: The method used to identify and measure the activity of the faults in the Zomba
 974 Graben. A small section of the Chingale fault is used as an example. For full details of each
 975 fault see the supplementary material. (a) Geological maps (Bloomfield, 1965) were used to
 976 identify locations where scarps or terrace features had Tertiary-recent sediments in their
 977 hangingwall. (b) TanDEMx DEM, with the location of the fault identified by the change in
 978 elevation and indicated by the black arrows. (c) Slope map that shows a linearly aligned
 979 increase in slope values that correspond to the terrace feature in part a and the increase in
 980 topography in part b. (d) The normalised channel steepness index increases in the footwall
 981 of the fault as the river channels cross the fault.

982



983

984 Figure 4: Examples of topographic profiles used to assess fault activity in the Zomba Graben.

985 (a) Example topographic profiles used to calculate the height of the fault scarps in the

986 Zomba Graben. The fault scarps are characterised by locally steep slopes at the base of the

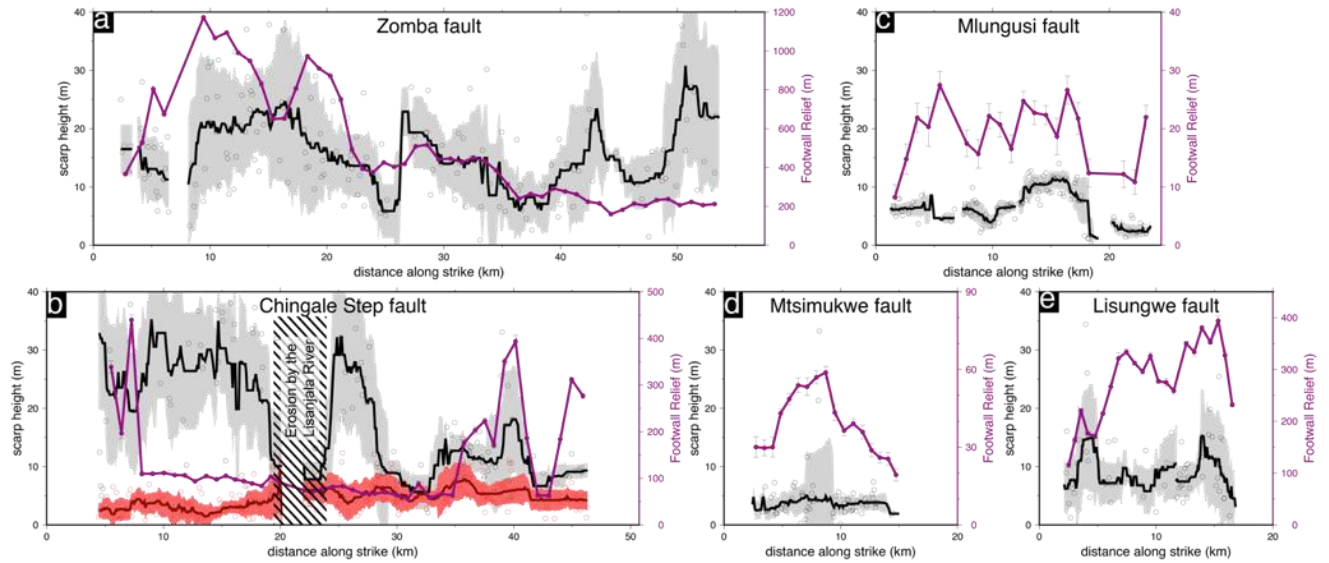
987 larger footwall escarpments. Along the Chingale Step fault, a composite scarp is observed

988 with the lower and upper slopes offset in at least two difference events. For all other, only a

989 single offset was observed. (b) Example of the topographic profile used to calculate the

990 height of the footwall escarpment, which we use as a proxy for minimum throw. The height

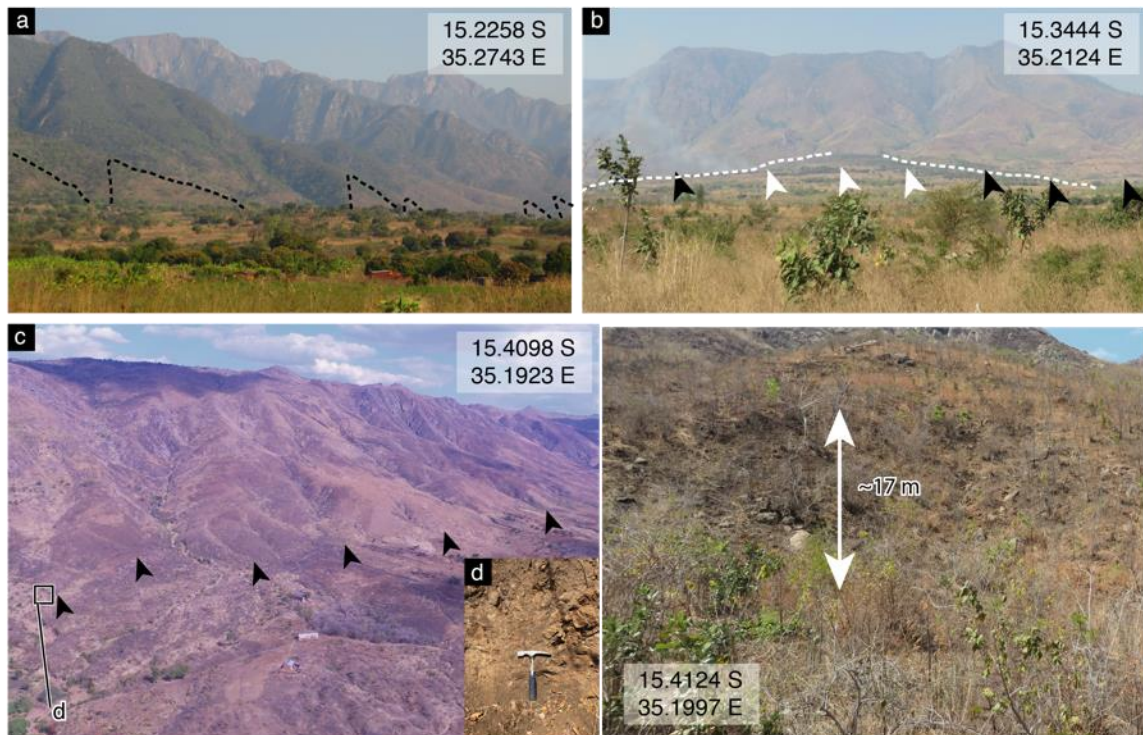
991 is measured as the maximum relief within 3 km of the footwall.



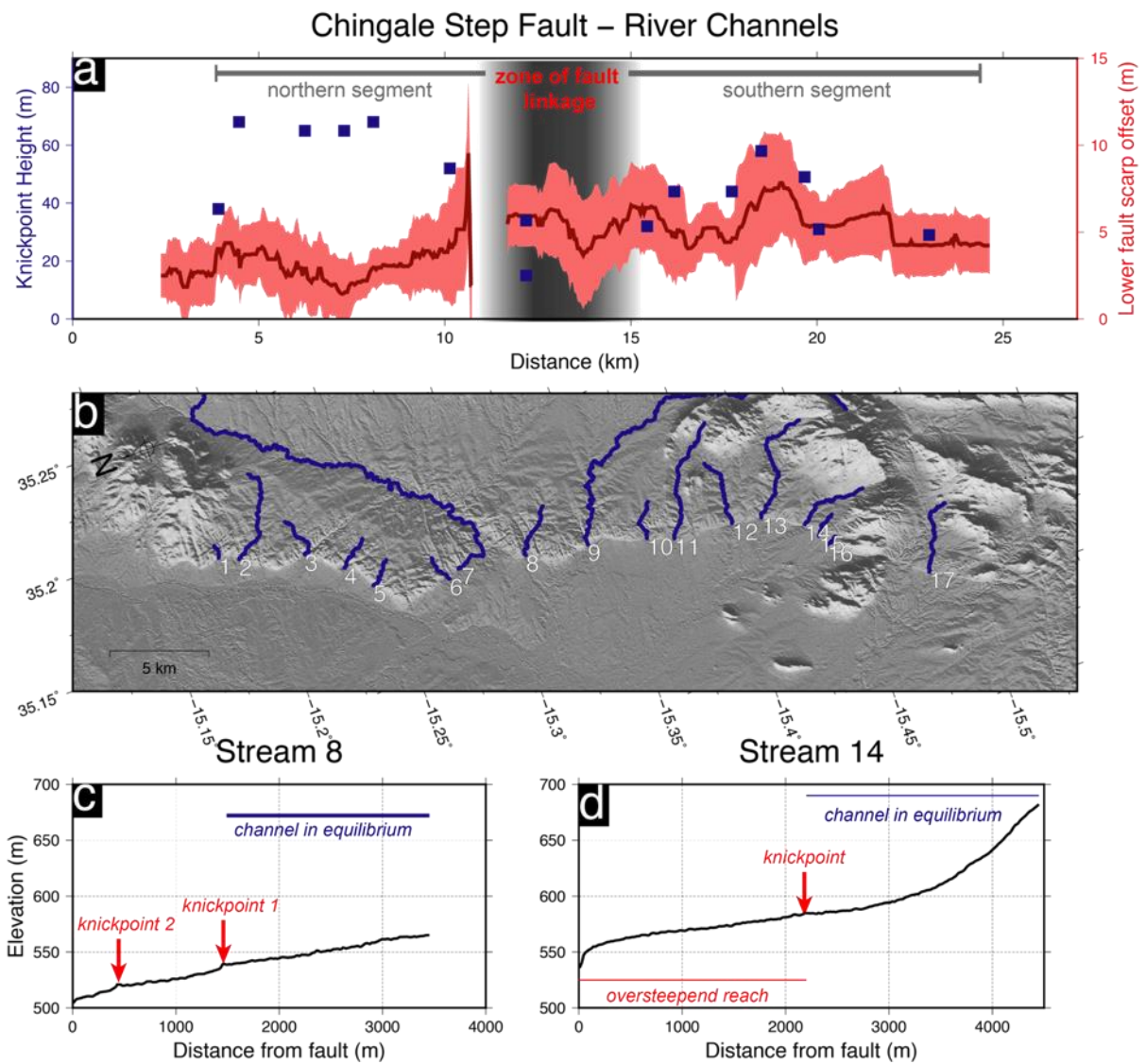
992

993 Figure 5: Fault scarp height (black) and footwall escarpment height (purple) for the five
 994 faults in the Zomba Graben. Note horizontal and scarp height scales are all equal but
 995 footwall relief scales differ in each plot. For the scarp height, the circles show the individual
 996 measurements, the solid black lines is the 3 km wide moving median with the 1σ error
 997 shaded. (a) The Zomba fault. (b) The Chingale Step fault. The red lines and points indicates
 998 the height of the lowest scarp on the composite scarp (see Figure 3 and S2). The black line
 999 and points are the height of the composite scarp. The shaded area is where offsets have
 1000 been eroded by the Lisanjala River. (c) The Mlungusi fault. (d) The Mtsimukwe fault. (e) The
 1001 Lisungwe fault.

1002



1003 Figure 6: Field observations of the Zomba fault. (a) Triangular facets (black dashed lines)
 1004 observed at the northern end of the fault. The hangingwall-footwall contact of the facets is
 1005 not visible in this photo. (b) Alluvial fan observed in the section of the fault where the
 1006 Zomba plateau is in the footwall of the fault. The black arrows indicate the location of the
 1007 fault, the white arrows show where the fault crosses and offsets the alluvial fan. The age of
 1008 the alluvial fan is not known. (c) The fault scarp at the southern end of the Zomba fault
 1009 (indicated by black arrows). (d) Fractured basement rock consistent with a fault zone
 1010 observed where exposed in stream beds at the base of the fault scarp. (e) Steep scarp at the
 1011 base of the footwall escarpment. A scarp height of ~17 m was measured in the field and is
 1012 consistent with the measurements made using remote sensing data (Figure 5).



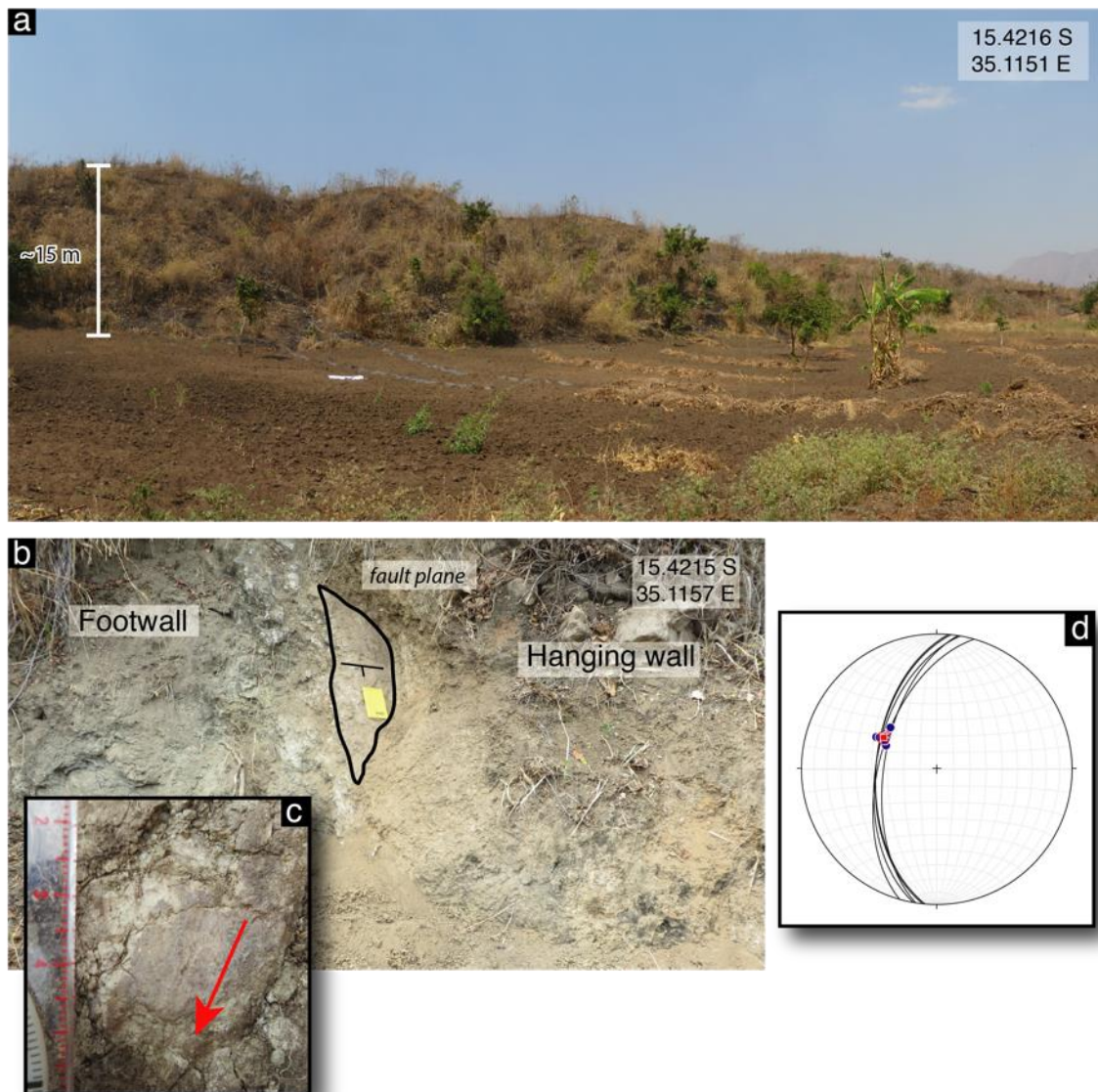
1013

1014 Figure 7: Rivers crossing the Chingale Step fault. (a) The elevation of the knickpoints, at the
 1015 top of oversteepened portions of the river long profiles (see Figure S3), above the fault
 1016 scarp (blue squares). Red shows the offset across the lower fault scarp on the Chingale Step
 1017 Fault (light red is 1 σ error). The gap is due to erosion by the Lisanjala River (stream 7). (b)
 1018 Map of the footwall river channels that cross the Chingale Step fault. Channels 7 and 9 were
 1019 not analysed as they both exhibit behaviour which would suggest that they were not
 1020 detachment limited and both cross multiple lithologies in the footwall of the Chingale Step
 1021 fault. (c) Stream 8 which is found in the linkage zone between the northern and southern
 1022 segment of the fault and displays two prominent knickpoints. (d) Stream 14 shows a clear

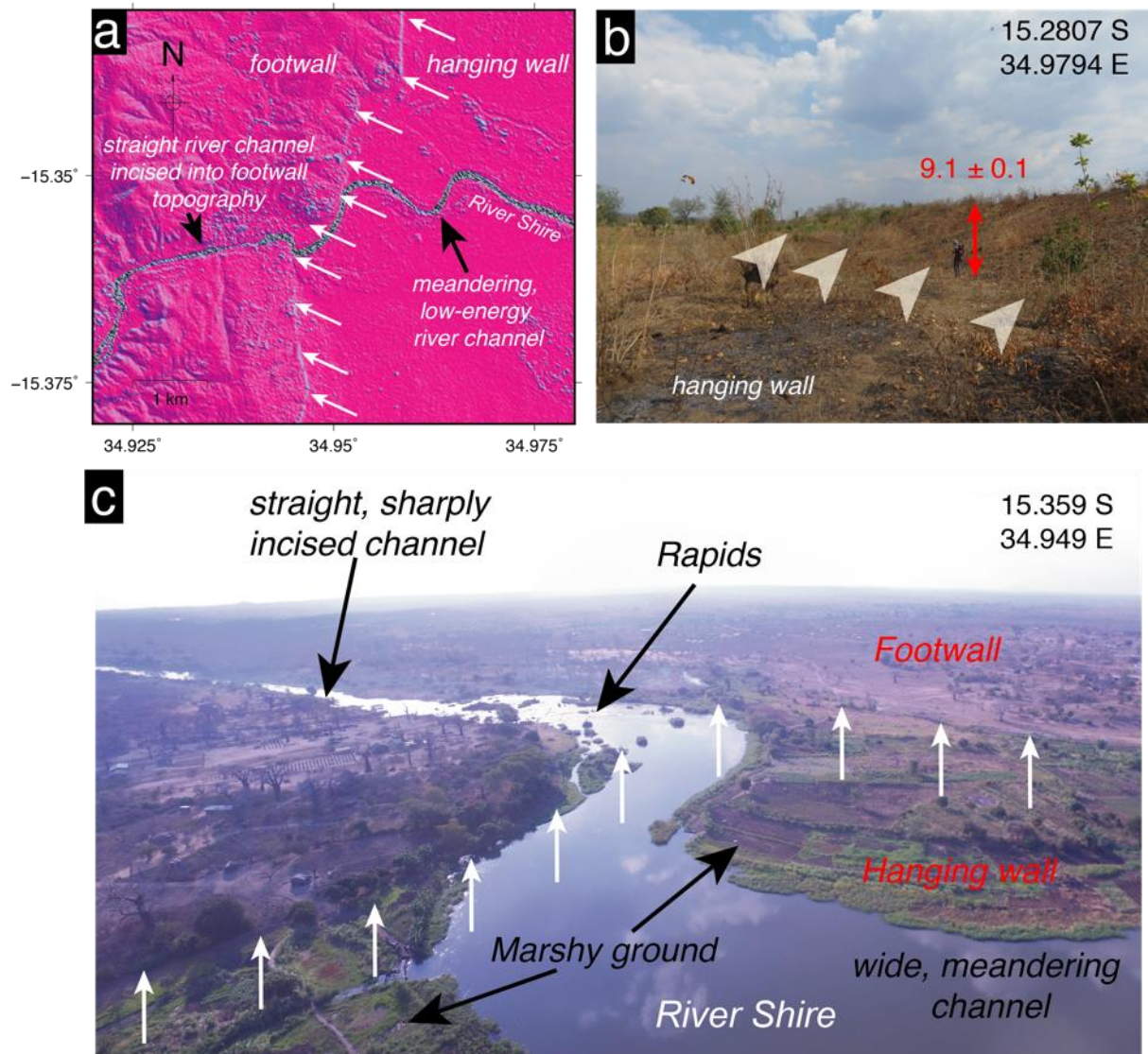
1023 oversteepened reach in the footwall of the fault. Further upstream the channel is in

1024 equilibrium.

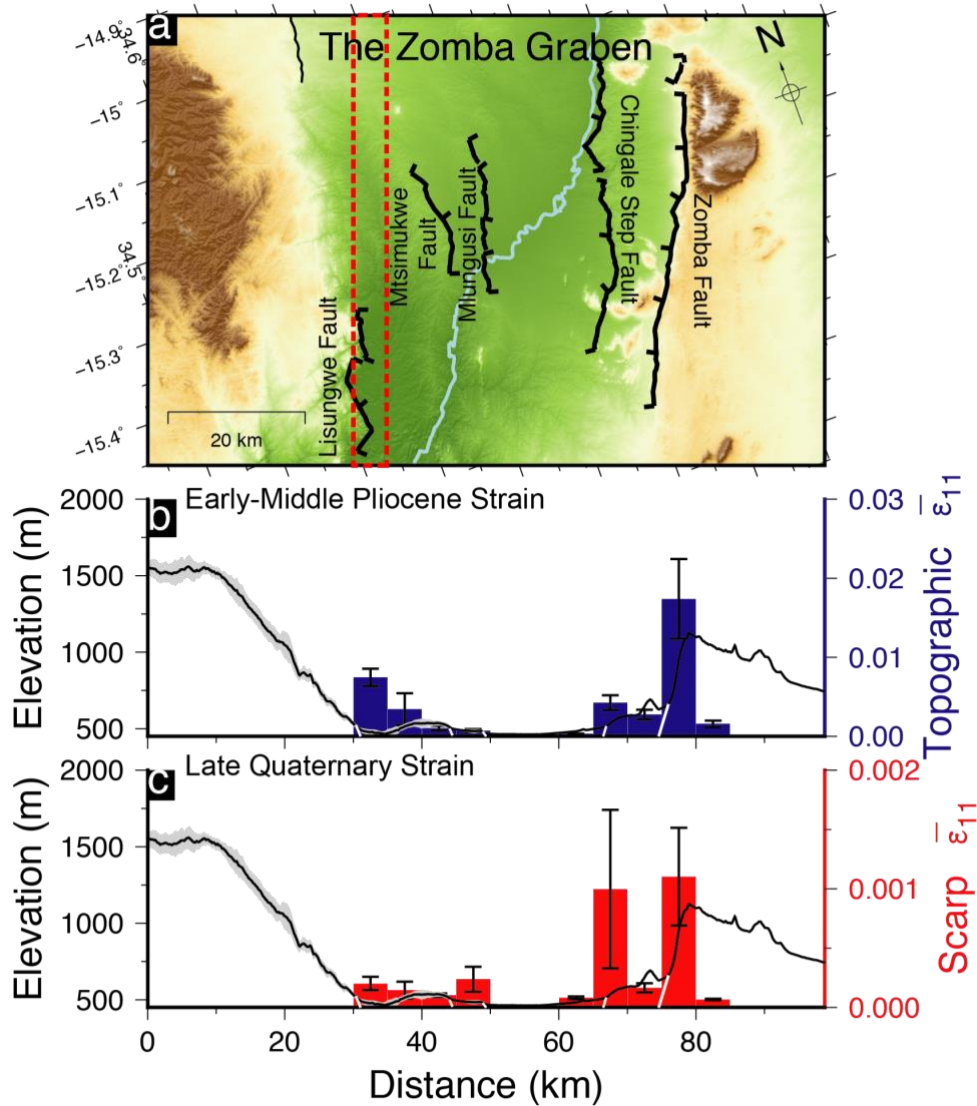
1025



1026 Figure 8. the Chingale Step fault. (a) Fault scarp along the southern section of the Chingale
 1027 Step fault. The hangingwall is comprised of a mix of fluvial, alluvial and lacustrine deposits
 1028 whereas bedrock is exposed in the footwall with a thin soil cover. The slickensides shown in
 1029 part c were observed in a stream bed ~50 m north of this site. (b) The exposure of the fault
 1030 plane at the Kalira River site. (c) Exposed slickensides measured on a polished fault plane
 1031 exposed at the side of the river bank. (d) Stereonet showing the fault plane orientation
 1032 (black lines), the average strike and dip is $189^{\circ}/54^{\circ}$. The trend and plunge of the slickensides
 1033 (blue dots) is shown in part b, the average trend and plunge (red dot) is $52^{\circ} \rightarrow 301^{\circ}$ (N=5).



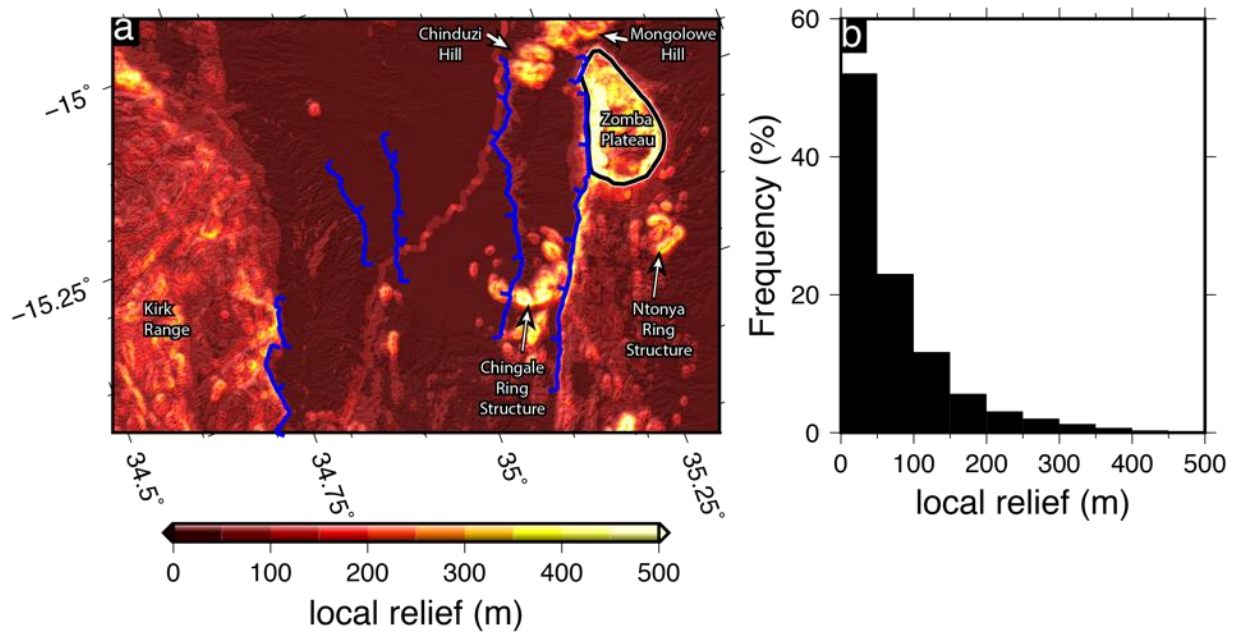
1035 Figure 9: The interaction between the Mlungusi fault and the River Shire. (a) Slope map
 1036 where the River Shire crosses the Mlungusi fault with the fault scarp indicated by white
 1037 arrows. (b) The fault scarp formed by the Mlungusi Fault (~3 km to north of parts a and c).
 1038 The white arrows indicate the base of the fault scarp, the offset is indicated in red. (c) Photo
 1039 taken from a drone of the location where the River Shire crosses the Mlungusi fault. The
 1040 location of the scarp is indicated with white arrows.



1041 Figure 10: Distribution of strain in the Zomba Graben. (a) Active faults and topography of
 1042 the Zomba Graben. The active faults analysed in this paper are indicated with thick black
 1043 lines. ϵ_{11} direction for parts b-c is horizontal in this projection (projection is rotated 10° to
 1044 the West). An example of the 5 km width bins used to calculate the strain is shown with the
 1045 dotted red line. (b) Rift-wide distribution of strain calculated using the footwall relief
 1046 measurements across each fault in 5 km width areas bins. The length of the bins matches
 1047 the area shown in part a. The measured offsets are then averaged in each 5km width bin
 1048 before the strain is calculated in the ϵ_{11} direction. (c) Rift-wide distribution of strain
 1049 calculated using the fault scarp offsets measured across each fault in 5km width bins. For

1050 the Chingale Step fault, the total scarp height measurements are used, rather than the

1051 height of the lower offset measured at the base of the fault scarp.



1052

1053 Figure 11: Local relief in the Zomba Graben. (a) Local relief calculated in 1 km sized windows

1054 within the Zomba Graben. High values associated with the River Shire are due to errors in

1055 the TanDEMx DEM caused by poor coherence of the water in the river which leads to

1056 random height errors (Wessel et al., 2018). Prominent zones of high local relief are related

1057 to pre-existing topographic features such as the Zomba Plateau, the Ntonya Ring Structure

1058 (an intrusion of similar age to the Chingale Ring Structure), the Kirk Range and the Chingale

1059 Ring Structure. The Chinduzi and Mongolowe hills are syenite-granite intrusion of the Chilwa

1060 Alkaline province (the same age as the Zomba Plateau). (b) Histogram of local relief values in

1061 the Zomba Graben (calculated using the area and data shown in part a).

1062

1063

1064 Table 1: The Geometry, activity rates and strain across faults in the Zomba Graben.

Fault	Length (km)	Strike(°)	Dip Direction	Mean Scarp height (m)	Mean Footwall Relief (m)	% of Total Strain	% of late- Quaternary strain
Zomba	51	025	West	15.6 ± 5.2	326 ± 113	56 ± 16	43 ± 15
Chingale Step	39	022	West	19.6 ± 12.1	84 ± 18	12 ± 2	35 ± 22
Mlungusi	22	013	East	6.9 ± 3.1	19 ± 5	2 ± 0	8 ± 3
Mtsimukwe	13	177	East	3.6 ± 0.7	38 ± 12	11 ± 6	8 ± 3
Lisungwe	23	019	East	10.0 ± 6.7	277 ± 76	19 ± 3	6 ± 2

1065