

1 **The tectonic history of a crustal-scale shear zone in the Tanzania**
2 **Craton from the Geita Greenstone Belt, NW-Tanzania Craton**

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9
10 **Abstract**

11 In this contribution, we present for the first-time field based evidence of a crustal scale
12 shear zone from the southern margin of Geita Greenstone Belt. The Geita Shear Zone is a broad
13 (~ 800 m wide) ductile, high-strain, deformation zone that can be traced for at least 50 km
14 along the southern margin of the Geita Greenstone Belt. It is near vertical, trends ~ E-W and
15 separates the mafic volcanics of the Kiziba Formation, to the north, from the TTG gneisses that
16 crop out south of the shear zone. The shear zone is hosted almost entirely by the TTG gneisses
17 and is characterised by a well-developed mylonitic foliation near the greenstone margin that
18 transitions into a gneissic foliation and eventually becomes a weakly developed foliation
19 further south. It contains approximately equal amounts of dextral, sinistral, and asymmetric
20 shear sense indicators suggesting that the shear zone accommodated mainly flattening strain
21 while the mineral stretching lineation defined by quartz and feldspar ribbons and stretched
22 biotite selvages plunges shallowly W. A series of younger, sub-vertical, NW trending brittle-

23 ductile, strike-slip shear zones truncate and displace the Geita Shear Zone with dextral
24 displacement in the order of 2-4 km. Deformed tonalite interpreted to predate the shear zone
25 yielded U-Pb zircon ages of ~2710 Ma while synshearing granodiorite samples have zircon
26 ages between 2680 Ma and 2660 Ma. The ~2630 Ma age of the undeformed Nyankumbu
27 granite is interpreted to mark the minimum age of movement on the shear zone. The presence
28 of 3000 Ma and 3200 Ma zircon xenocrysts in the tonalite and granodiorite opens the possibility
29 that older basement rocks underlie the greenstone belts in the northern half of Tanzania Craton.
30 Whether or not the greenstone belts were erupted on older basement, thrust on top of older
31 basement rocks or incorporated older basement fragments has profound implications for the
32 tectonic framework and evolution of the Tanzania Craton.

33

34 **1. Introduction**

35

36 Large scale shear zones are well documented features in Archean terrains worldwide
37 (e.g. Bédard et al., 2003; Cassidy et al., 2006; Chardon et al., 2008; Jelsma and Dirks, 2002;
38 Dirks et al., 2013) and their structural style has implications for the tectonic history and crustal
39 growth models of Archean cratons. One of the most common Archean structural styles is the
40 dome and keel geometry (e.g. Pilbara, Kaapvaal, Western Dharwar), linked to gravity-driven
41 tectonics and redistribution of rock masses through vertical processes (e.g. Choukroune et al.,
42 1995; Chardon et al., 1996; Collins et al., 1998; Bédard et al., 2003; Van Kranendonk, 2011).
43 The majority of these types of structures are characterised by gneiss domes with radial
44 elongation lineations and normal shear sense indicators suggesting diapiric ascent through the
45 crust. Between the gneiss domes, synformal-shaped greenstone belts are preserved. In general,
46 the geometry of the greenstone belts follows the contours of the gneiss domes resulting in cusp-

47 like shapes. Typical examples include the Barberton Greenstone Belt in South Africa (e.g.
48 Dirks et al., 2013; Van Kranendonk et al., 2014; Brown, 2015) and the greenstone belts in the
49 Pilbara Craton of Western Australia (e.g. Collins et al., 1998; Van Kranendonk et al., 2004).
50 Alternative interpretations for dome and keel geometries include core complexes exposed by
51 extensional unroofing (Kloppenburger et al., 2001) or antiformal culminations exposed by cross-
52 folding (Blewett et al., 2004).

53 Another common Archean structural style is represented by linear belts separated by
54 transcurrent shear zones, which are typically interpreted to indicate lateral terrain accretion
55 (e.g. Polat et al., 1998; Dirks et al., 2002; Blewett et al., 2010; Kabete et al., 2012a). Typical
56 examples include the Yilgarn Craton (e.g. Cassidy et al., 2006; Blewett et al., 2010) the
57 Superior Province (e.g. Card, 1990; Polat et al., 1998) and the Dharwar Craton (e.g. Chadwick
58 et al., 2000; Manikyamba and Kerrich, 2012). The transcurrent shear zones form an
59 anastomosing pattern encompassing elongated greenstone belts, many with similar
60 stratigraphy, bordered by granite-gneiss terranes of younger age. For example, in the eastern
61 Yilgarn Craton, detailed stratigraphic analyses showed that the stratigraphy of the greenstone
62 belts, although disrupted by shear zones, can be correlated for hundreds of kilometres (Hayman
63 et al., 2015).

64 The structural and tectonic evolution of the Archean Tanzania Craton is poorly
65 understood (Kabete et al., 2012a; Kabete et al., 2012b) and in many instances stratigraphic
66 correlations based on sparse geochronological data are contradictory (e.g. Borg and Krogh,
67 1999; Many et al., 2006; Sanislav et al., 2014). A first attempt to define tectonic and structural
68 boundaries within the Tanzania Craton was made by Kabete et al. (2012a) based on existing
69 geological maps and the interpretation of geophysical datasets. A series of NW trending shear
70 zones were proposed and interpreted to delineate superterrane boundaries (Fig. 1). However,
71 the existence of these large-scale shear zones, their continuity, style, age and kinematic sense

72 was never confirmed with field studies, and the terrane-accretion model, therefore, remains
73 entirely speculative.

74 In this contribution, we provide the first field-based description of a regional scale shear
75 zone from the northern half of the Tanzania Craton and present new zircon age data from
76 igneous rocks that intruded along and across the shear zone to constrain its kinematic history.
77 Results will be discussed in terms of the regional significance for the structural and tectonic
78 evolution of this part of the Tanzania Craton.

79

80 **2. Regional geology**

81

82 The geology of the Archean Tanzania Craton is generally described in terms of three
83 main stratigraphic and tectonic units, the Dodoman Supergroup, The Nyanzian Supergroup and
84 the Kavirondian Supergroup (e.g. Many and Maboko, 2008; Kabete et al., 2012a; Cook et al.,
85 2016; Sanislav et al. 2017). The Dodoman Supergroup is the oldest tectonic unit and its
86 occurrence is restricted mainly to the central part of the Tanzania Craton (Kabete et al., 2012a).
87 This unit consists of high-grade mafic and felsic granulite with subordinate lower-grade schist
88 and thin slivers of greenstone. A limited number of zircon ages confirm that the Dodoman
89 Supergroup forms the older part of the Tanzania Craton with ages ranging between 3 Ga and
90 3.6 Ga (Kabete et al., 2012b).

91 The Nyanzian and the Kavirondian Supergroups are concentrated in the northern half
92 of the Tanzania Craton and were placed stratigraphically above the Dodoman Supergroup (e.g.
93 Quenell et al 1956; Gabert, 1990). The Nyanzian Supergroup is overlain unconformably
94 (Gabert, 1990) by the Kavirondian Supergroup. The lowermost unit of the Nyanzian is

95 dominated by mafic volcanics and volcanoclastics with minor felsic volcanics erupted at ~
96 2820 Ma (e.g. Borg and Krogh, 1999; Many and Maboko, 2003; Cook et al., 2016). These are
97 overlain by felsic volcanic and pyroclastic units inter-bedded with banded ironstone,
98 volcanoclastic sequences and immature turbiditic sediments (e.g. Borg, 1992; Borg and
99 Shackleton, 1997; Sanislav et al., 2015). All greenstone sequences from the northern half of
100 Tanzania were assigned to the Nyanzian Supergroup. Six major greenstone belts were
101 described (Fig. 1): the Sukumaland Greenstone Belt (SGB), The Nzega Greenstone Belt, the
102 Iramba-Sekenke Greenstone Belt, the Shynianga-Malita Greenstone Belt, the Kilimafedha
103 Greenstone Belt and the Musoma-Mara Greenstone Belt (Borg and Shackleton, 1997). The
104 greenstone belts are fragmented by shear zones and intruded by ~2800 to 2600 Ma granitoids
105 (Kabete et al., 2012b; Sanislav et al., 2014). Most of the greenstone belts are oriented along ~
106 E-W lineaments (Fig. 1), one of these lineaments forming the southern margin of the Geita
107 Greenstone Belt. The Kavirondian Supergroup consists mainly of coarse-grained
108 conglomerate, grit and quartzite (Gabert, 1990) and its areal extent is poorly constrained.

109 The Geita Greenstone Belt (GGB) is a large greenstone fragment forming most of the
110 northern part of what has previously been defined as the SGB (Fig. 1). It contains elements of
111 the Nyanzian and the Kavirondian stratigraphy (Fig. 2). The base of the greenstone belt is
112 dominated by lower-amphibolite facies metamorphosed mafic volcanics forming the Kiziba
113 Formation (Cook et al., 2016). Whole rock Sm-Nd dating of the mafic volcanics yield model
114 ages of ~2820 Ma (Many and Maboko, 2008; Cook et al., 2016). The mafic volcanics of the
115 Kiziba Formation are positioned below a complexly deformed sequence of intercalated
116 volcanoclastics, ironstones and immature turbidites, which experienced greenschist facies
117 metamorphism (Cook et al., 2016; Sanislav et al., 2015, 2017). The greenstone sequence was
118 intruded syntectonically by diorite (D₂ to D₄) and by late-syntectonic felsic porphyries (D₆ to
119 D₇). The northern, eastern and western margins of the greenstone are intruded by 2660 to 2620

120 Ma high-K granites (Sanislav et al., 2014). The southern margin of the Geita Greenstone Belt
121 is bounded by gneiss and deformed granite across a steeply dipping shear zone, which was
122 intruded by the undeformed high-K Nyankumbu granite (Fig. 2). A few outcrops of
123 Kavirondian dominated by quartzitic conglomerate with minor sandstone occur in the north-
124 eastern part of the greenstone belt.

125 Sanislav et al. (2015; 2017) proposed that the GGB was affected by at least eight,
126 Archean deformation events. The first event (D_1) produced mainly layer parallel shears in what
127 may have been an extensional setting, while the second (D_2) to fourth (D_4) deformation events
128 were responsible for the formation of large-scale folds and associated shear zones, which in
129 turn were refolded by open to gentle, sub-horizontal (D_5) folds. These ductile events were
130 followed by two brittle-ductile shearing events that produced localised reverse faults (D_6) and
131 large scale NW trending dextral strike slip faults (D_7). The last event (D_8) is an extensional
132 event that produced steeply dipping and commonly ~ E-W trending normal faults that
133 reactivated earlier D_6 - D_7 structures.

134

135 **3. Geita Shear Zone**

136

137 The Geita Shear Zone (GSZ) is a mylonitic shear zone that defines the southern margin
138 of the GGB and can be mapped along strike for at least 50 km (Fig. 2). The general orientation
139 of the GSZ is ~ E-W (Fig. 2) and the overall straight map pattern and gravity data for the
140 greenstone belt suggest that the shear zone is steeply dipping. Regional aeromagnetic data
141 suggest that the GSZ extends further to the east and west (Fig. 1) outside of the area mapped
142 in Figure 2. In the Geita area the shear zone marks the boundary between mafic volcanics of
143 the Kiziba Formation to the north and TTG granitoid-dominated gneiss and granites to the

144 south (Fig. 2). The shear zone is generally poorly exposed, and is characterised on
145 aeromagnetic data sets by a 700-900 m wide zone with an intensely developed planar fabric
146 (Fig. 2), and in the field by scattered outcrops of partly recrystallised mylonite and striped
147 gneiss (Fig. 3a and b) Aeromagnetic datasets further indicate the presence of well-foliated
148 gneiss across a width of at least 0.8 km to the south of the main shear zone, whilst to the north,
149 the shear zone appears to transect well-foliated meta basalt with the northern boundary of the
150 GSZ defined by a low-angle truncation zone within deformed meta-basalt. The contact between
151 the Kiziba Formation and the TTG gneisses is not exposed anywhere in the Geita Greenstone
152 Belt, but the two units crop out within 50 meters of each other SW of Geita Town (GR 411520-
153 9679010 - all grid references in WGS84-36S), which represents one of the few areas with
154 excellent outcrop in the shear zone.

155 Near the contact with the Kiziba Formation the TTG gneiss units preserve a mylonitic
156 fabric characterised by an intense, mm-scale foliation and a generally near-horizontal mineral
157 stretching lineation defined by quartz-feldspar rodding (Fig. 3a). The mylonitic fabric has been
158 recrystallised, resulting in recovery of the deformation textures on thin-section scale, and
159 partial destruction of the mineral lineation especially in compositionally homogenous (i.e. bt-
160 amp-poor) portions of the gneiss. Highly boudinaged quartz veins and flattened feldspar
161 porphyroclasts (Fig. 3c) with symmetrical and asymmetrical tails (Figs. 3d and e) are locally
162 preserved within the mylonitic fabric, and shear bands are common some containing granite
163 veins. Going south, away from the contact with the greenstone belt, the mylonitic fabric
164 remains strong for about 400m beyond which the deformation fabric becomes less intense and
165 takes the form of a well-developed gneissic foliation with alternating dark biotite selvages and
166 cm-scale quartzo-feldspathic lithons that retain the stretching lineation (Fig. 3f). At about 1 km
167 south of the contact only a weak gneissic foliation is preserved (Fig. 3g) and feldspar

168 phenocrysts appear largely undeformed (Fig. 3h and i). The shear fabric is truncated by the
169 undeformed Nyankumbu granite (Fig. 3j).

170 The TTG gneiss units and granite intrusions south of the GSZ display a single dominant
171 gneissic foliation that is steep with a dominantly southerly dip and an E-W trend (average dip
172 / dip direction: $86^{\circ}/170^{\circ}$; Fig. 4a). The stretching lineation in the gneisses and granites
173 generally plunges shallowly W (Fig. 4b). Plagioclase porphyroclasts are mostly symmetrical,
174 but some porphyroclasts with asymmetric tails indicate variable shear directions with dextral
175 (Fig. 3d) appearing to be more common on a regional scale. Dextral and sinistral shear bands
176 occur side-by side. This suggests that the shear zone mainly accommodated flattening strain
177 with N-S shortening, and stretch along a horizontal, E-W direction, possibly with a minor
178 dextral component.

179 In the metabasalts to the north of the contact the GSZ is not exposed but geophysical
180 images indicate that near the contact with the TTG the fabric from the Kiziba Formation is
181 truncated and deflected into parallelism with the GSZ suggesting that the shear zone has only
182 a limited extent to the north. Rare metabasalt outcrops preserve evidence of intrafolial folding
183 (Fig. 6a; Cook et al., 2016) and a foliation defined by the preferential alignment of amphibole
184 and locally by a well-defined mm-scale banding of alternating layers of amphibole and
185 plagioclase. The foliation trends ~E-W and dips mainly north (average dip / dip direction
186 $73^{\circ}/352^{\circ}$; Fig. 4c). These mafic schists locally contain a moderately E to NE plunging (Fig.
187 4d) mineral lineation defined by hornblende and a plagioclase stretching lineation (Fig. 2), and
188 numerous asymmetric, boudinaged quartz veins (Fig. 6b) indicative of a N-up, sinistral
189 movement sense. Aeromagnetic images suggest that locally, large-scale folds occur within the
190 Kiziba formation north of the GSZ with E-W trending fold axes and axial planar surfaces (Fig.
191 3g). In the western part of the GGB, within metabasalts that outcrop to the north of the GSZ,

192 flattened pillow basalts occur with well-preserved vesicles filled with quartz and carbonate
193 indicating younging to the north (Cook et al., 2016).

194 Along the entire length of the GGB, the GSZ has been truncated and displaced by NW-
195 trending, near-vertical, D₇ faults (Fig. 2; Sanislav et al., 2015; 2017). These faults contain near-
196 horizontal lineations defined by quartz fibres and striations and brittle-ductile deformation
197 textures including well preserved S-C fabrics indicative of a dextral sense of shear. The D₇
198 shear zones are characterised by numerous, planar milky quartz veins that intruded parallel to
199 the strike of the D₇ faults. Networks of conjugate pairs of NW- and NE-trending quartz veins
200 are also common in the mylonitic gneisses of the GSZ within the vicinity of D₇ faults. Dextral
201 displacements of the GSZ along the principle D₇ faults are in the order of 2-4 km.

202 Within the felsic gneiss along the Geita Shear zone, tonalitic to granodioritic gneiss is
203 the dominant rock type, but less deformed, porphyritic to coarse-grained granite and
204 granodiorite bodies (Figs. 3h and i) are common as dykes and small intrusions. In the zone of
205 most intense deformation a number of over-printing dyke-like (Fig. 3b) to lensoidal bodies of
206 granite and granodiorite can be recognised, most of which are positioned in near-parallelism to
207 the main gneissic fabric and all of which contain an S-L fabric with constant lineation
208 directions. This suggests that shearing coincided with the emplacement of several generations
209 of felsic intrusions, which can provide a direct age estimate for the principle period of
210 movement along the shear zone.

211 Away from the core of the shear zone, some of the truncating granite dykes show only
212 weak fabrics or no deformation features at all. Some granite dykes did not develop a foliation,
213 but preserve a stretching lineation parallel to the dominant lineation direction in the shear zone,
214 defined by quartz aggregates, aligned biotite grains and weakly deformed feldspar phenocrysts.
215 The TTG gneisses are medium- to coarse-grained and their mineralogy is dominated by quartz

216 and plagioclase with minor amounts of k-feldspar, biotite and hornblende. The cross cutting
217 granite dykes and bodies intruding the gneisses of the GSZ during the shearing events are
218 quartz-rich and have approximately equal amounts of plagioclase and k-feldspar with only
219 minor amounts of biotite and hornblende. Based on their mineralogy the granites have been
220 classified as monzogranites. It is common for these granites to contain large k-feldspar
221 phenocrysts (Figs. 3h and i).

222 A younger, undeformed granite, the Nyankumbu granite (Fig. 3j) intruded the GSZ
223 (Fig. 2) in the central southern part of the greenstone belt, and crosscuts the mylonitic fabrics
224 in the TTG gneisses and the Kiziba Formation. This granite provides a lower limit for the time
225 of shearing along the GSZ. The mineralogy of the Nyankumbu granite is dominated by quartz
226 and k-feldspar with lesser plagioclase, biotite and minor hornblende, and it contains numerous
227 xenoliths. The xenoliths vary in size from a few centimetres to tens of centimetres in diameter
228 and their composition is similar to the composition of the TTG gneiss units and granites that
229 are intruded by the Nyankumbu granite.

230

231 **4. Geochronology**

232

233 Nine samples were collected for zircon dating along and across the GSZ (Fig. 2) to
234 constrain the time of shearing. Seven samples come from the granitoids south of the GSZ and
235 two samples (K81 and IH3) come from small tonalite bodies that intruded into the Kiziba
236 Formation. These tonalite bodies are part of the greenstone sequence affected by the GSZ and
237 their time of emplacement is interpreted to provide an upper age limit for the GSZ. From south
238 of the GSZ two samples were collected from the TTG gneisses (SM02 and SM25), which may
239 provide a maximum time for shearing. Three samples were collected from the deformed

240 granites that intruded into the TTG gneisses and that retain the shear fabric, i.e. these granites
241 may have been intruded during shearing (SM07, SM23 and SM27). Two samples were
242 collected from the undeformed Nyankumbu granite (SM56 and SM57) that crosscuts the GSZ
243 and constrain the cessation of shearing along the GSZ.

244 All samples were dated at the Advanced Analytical Centre hosted by James Cook
245 University. All zircon grains were imaged with a Jeol JSM5410LV with attached
246 cathodoluminescence detector. Analytical spots for each zircon grain were selected based on
247 internal zonation. U-Pb isotope analyses were obtained using a GeoLas 200 Excimer Laser
248 Ablation System in a He ablation atmosphere, coupled to a Varian ICP-MS 820 series
249 instrument. The detailed analytical procedures used for U-Pb dating in the James Cook
250 University laboratory are outlined in Sanislav et al. (2014). Data were processed using the
251 software package GLITTERTM (Jackson et al., 2004), and age calculations were done using
252 Isoplot (Ludwig, 2003). The zircon standard GJ1 (ID-TIMS $^{207}\text{Pb}/^{206}\text{Pb}$ age = 608.5 ± 0.4 Ma;
253 Jackson et al., 2004) was used as the primary standard while the zircon standards FC1 (ID-
254 TIMS $^{207}\text{Pb}/^{206}\text{Pb}$ age = 1099.0 ± 0.6 Ma; Paces and Miller, 1993) and 91500 ($^{207}\text{Pb}/^{206}\text{Pb}$ age
255 = 1065 ± 0.3 Ma; Wiedenbeck et al., 1995) were used as secondary standards.

256 The weighted average ages (errors at 95% confidence) for the standards were as
257 follows: GJ-1 - 610 ± 10 Ma for $^{207}\text{Pb}/^{206}\text{Pb}$ and 600 ± 2 Ma for $^{206}\text{Pb}/^{238}\text{U}$; FC-1 - 1101 ± 9
258 Ma for $^{207}\text{Pb}/^{206}\text{Pb}$ and 1103 ± 15 Ma for $^{206}\text{Pb}/^{238}\text{U}$; 91500 - 1071 ± 17 Ma for $^{207}\text{Pb}/^{206}\text{Pb}$ and
259 1060 ± 5 Ma for $^{206}\text{Pb}/^{238}\text{U}$. Analyses with significant discordance ($>10\%$) and those with
260 elevated common Pb, where $^{206}\text{Pb}/^{204}\text{Pb}$ (background corrected) is <1000 , were excluded from
261 age calculations.

262

263 **5. Results**

264

265 *5.1. The TTG samples predating the GSZ*

266

267 Samples SM02 and SM25 were collected from the TTG gneisses that crop out in the lower-
268 strain, southern margin of the GSZ, where the shear fabric is only weakly developed. Both
269 samples are coarse-grained and based on their mineralogy can be classified as tonalite. The
270 zircon grains separated from these samples are, in general, euhedral to subhedral with a
271 prismatic to needle-like shape and low cathodoluminescence luminosity (Fig. 6). The zircon
272 grains separated from sample SM02 have a low luminescence core, without any clear zoning,
273 surrounded by a thin rim overgrowth with a slightly higher luminescence (Fig. 6a). In some
274 zircons the rim overgrowth is delineated from the core by a thin high luminescence band.
275 Because the rim overgrowth was too thin to analyse by laser, all the measurements were
276 performed on the cores. Eight U-Pb analyses with low ^{204}Pb yielded concordant ages and were
277 used for age calculations. The $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average of this sample is 2705 ± 14 Ma (Fig.
278 6b) with an identical concordia age of 2705 ± 18 Ma (Fig. 6c), interpreted to represent the
279 emplacement age.

280 The cathodoluminescence images of the zircon grains from sample SM25 revealed the
281 presence of grains with complex internal growth domains (Fig. 6d). The majority of the zircons
282 grains contain a low luminescence core with a weakly developed concentric zoning surrounded
283 by a very thin, high luminescence rim (Fig. 6d). Two concordant and nine nearly concordant
284 analyses from this domain have a $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average of 2706 ± 13 Ma (Fig. 6e) and
285 an identical concordia age of 2706 ± 13 Ma (Fig. 6f), which are interpreted to represent the
286 emplacement age. One large zircon grain (Fig. 6d) contains an elongated and sub-rounded

287 inherited core with low luminescence and without any internal zoning surrounded by a thick
288 median overgrowth with higher luminescence and concentric zoning. The median overgrowth
289 has a thin and weakly developed high luminosity rim. Two U-Pb analyses from the inherited
290 core yielded a concordant analysis with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2840 ± 22 Ma and one nearly
291 concordant analysis with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2826 ± 21 Ma. One nearly concordant analysis
292 from the median overgrowth gave a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2710 ± 22 Ma, which is similar to the
293 emplacement age. Three zircon grains contained large, low luminescence cores surrounded by
294 well-developed, high luminescence rims (Fig. 6d). Only one core analysis returned a nearly
295 concordant $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2702 ± 20 Ma, similar to the emplacement age, while the rims
296 returned nearly concordant analyses with a $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age of 2661 ± 24 Ma
297 (Fig. 6e) and a concordia age of 2665 ± 41 Ma (Fig. 6f). The rim age is interpreted to represent
298 zircon overgrowth during a later thermal event related to the emplacement of the granites that
299 intruded the TTG's during shearing along the GSZ.

300

301 *5.2. Tonalite intrusions internal to the greenstone belt predating the GSZ*

302

303 Samples IH3 and K81 were collected from two small tonalite bodies that intruded into the
304 mafic rocks of the Kiziba Formation (Fig. 2). Both tonalite outcrops contain a well-developed
305 foliation having a similar orientation to the foliation found in the surrounding mafic volcanics,
306 which occur to the north of the GSZ. Sample IH3 was collected near the contact with the TTG
307 gneisses where the foliation is sup-parallel to the trend of GSZ. Sample K81 was collected
308 further north of the GSZ boundary where the foliation forms a small angle to the trend of the
309 GSZ. The tonalite bodies are interpreted to have been emplaced before development of the
310 GSZ. The zircon grains separated from these samples are generally medium- to small-sized,

311 prismatic, and mostly euhedral in shape. Cathodoluminescence images of zircon grains from
312 sample IH3 (Fig. 7a) show that most of the zircon grains have low luminescence and poorly
313 developed, concentric zoning. It is common for these zircon grains to display a thin, high-
314 luminescence band, which only partly surrounds the grain, and which, in turn, is overgrown by
315 a narrow rim (Fig. 7a). Ten concordant and near concordant analyses from these zircons gave
316 a $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age of 2708 ± 15 Ma (Fig. 7b) and a similar concordia age of
317 2708 ± 25 Ma (Fig. 7c). These ages are interpreted to indicate the emplacement age of the
318 tonalite bodies. The rim overgrowth was too narrow to be analysed. Sample IH3 also contained
319 a few higher luminescence zircon grains (Fig. 7a) with some of them having a narrow rim
320 overgrowth. Four analyses of these higher luminescence zircon grains returned concordant or
321 near concordant ages. Their $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age is 2831 ± 21 Ma (Fig. 7b) while
322 the concordia age is 2831 ± 22 Ma (Fig. 7c). These ages are similar to the ages of the inherited
323 zircon cores found in sample SM25 suggesting that these two intrusions may have sampled a
324 similar reservoir.

325 Sample K81 yielded a large number of zircon grains. Cathodoluminescence images
326 revealed that the majority of zircon grains are similar to the zircon grains separated from sample
327 IH3; in general, they display low luminescence, with poorly developed concentric zoning, with
328 or without a thin outer high-luminescence rim and a further, narrow rim overgrowth (Fig. 7d).
329 Twenty-three concordant and nearly concordant analyses from these zircons gave a $^{207}\text{Pb}/^{206}\text{Pb}$
330 weighted average age of 2711 ± 8 Ma (Fig. 7e) and an identical concordia age of 2711 ± 8 Ma
331 (Fig. 7f), which are interpreted to indicate the emplacement age. However, many of these
332 zircon grains contain irregularly shaped, inherited cores with very low luminescence and no
333 internal zoning (Fig. 7d). These cores always contain a thin and highly luminescent outer band
334 that may or may not completely surround the core and have a narrow rim overgrowth. A few
335 of these cores were large enough to be analysed. One nearly concordant analysis from a low

336 luminescence core yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3231 ± 19 Ma. Six other analyses performed on
337 inherited cores yielded almost identical ages so that they were treated together. The $^{207}\text{Pb}/^{206}\text{Pb}$
338 weighted average age of these six analyses is 3004 ± 21 Ma (Fig. 7e) while their concordia age
339 is 3005 ± 26 Ma (Fig. 7f).

340

341 *5.3. The granite samples emplaced during movement on the GSZ*

342

343 Three samples (SM07, SM23 and SM27) were collected from variably deformed,
344 coarse-grained to porphyritic granite that intruded into the TTG gneisses south of the GSZ (Fig.
345 2). They appear as dykes and lensoidal intrusions having a similar, but less intense shear fabric
346 with the host TTG. Some thin dykes (~10 cm) of similar composition have been rotated into
347 near parallelism with the shear fabric suggesting that these granites were emplaced during
348 movement on the GSZ. The zircon grains are small- to medium-sized, in general prismatic and
349 euhedral in shape (Figs. 8 and 9). They display low luminescence with a weakly defined growth
350 zoning. All three samples contain zircon grains with inherited cores that are surrounded by a
351 narrow high luminescence rim. However, only sample SM07 contained a zircon grain with an
352 inherited core large enough to be analysed. This zircon grain (Fig. 8a) contains three growth
353 domains: a large core surrounded by a very thin luminescent rim followed by a narrow median
354 overgrowth, also surrounded by a thin high luminescent rim, and finally a rim overgrowth. One
355 nearly concordant analysis from the core returned a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 3264 ± 20 Ma, which is
356 similar to the age of one of the inherited cores found in sample K81. This sample also contained
357 a zircon grain (Fig. 8a) with different cathodoluminescence properties compared to the
358 remaining grains, including a well-defined growth zoning and medium luminescence. Two
359 analyses were performed on this grain: one concordant analysis returned a $^{207}\text{Pb}/^{206}\text{Pb}$ age of

360 2841±22 Ma and one nearly concordant analysis returned a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2866±23 Ma.
361 These ages are similar to the ages of inherited cores found in samples SM25 and IH3. Analyses
362 performed on the remaining low luminescence grains yielded similar ages with a $^{207}\text{Pb}/^{206}\text{Pb}$
363 weighted average age of 2667±19 Ma (Fig. 8b) and a concordia age of 2667±27 Ma (Fig. 8c),
364 which are interpreted to indicate emplacement ages. The zircon grains from samples SM23 and
365 SM27 had small inherited cores, which could not be analysed. All zircon grains from sample
366 SM23 displayed low luminescence (Fig. 8d) and yielded a $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age of
367 2671±18 Ma (Fig. 8e) and a concordia age of 2673±22 Ma (Fig. 8f). The zircon grains from
368 sample SM27 have well defined growth zoning (Fig. 9a), and ten analyses yielded a $^{207}\text{Pb}/^{206}\text{Pb}$
369 weighted average age of 2680±13 Ma (Fig. 9b) and a concordia age of 2678±27 Ma (Fig. 9c).

370

371 *5.4. The Nyankumbu granite emplaced into the GSZ*

372

373 Two samples (SM56 and SM57) were collected from the Nyankumbu granite to constrain
374 the minimum age of the GSZ. Sample SM57 yielded a large number of zircon grains while
375 sample SM56 yielded only few zircon grains. Both samples contain a mixed zircon population
376 with inherited cores and xenocrysts. Cathodoluminescence images show that in sample SM56
377 the majority of the grains are in general high luminescence with well-defined growth zoning
378 (Fig. 10a). The inherited cores have low luminescence and in many instances are surrounded
379 by rim overgrowths with a higher luminescence. A few zircon grains with low luminescence
380 and a weakly defined growth zoning were interpreted as xenocrysts. Five high luminescence
381 zircon grains yielded nearly concordant results with a $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age of
382 2638±16 Ma (Fig. 10b) and an identical concordia age (Fig. 10c). Two low luminescence
383 zircon xenocrysts yielded a combined $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2655±25 Ma (Fig. 10b). Three zircon

384 grains contained inherited cores that were large enough to be analysed (Fig. 10a). Their
385 $^{207}\text{Pb}/^{206}\text{Pb}$ ages are similar so that they were treated together and their $^{207}\text{Pb}/^{206}\text{Pb}$ weighted
386 average age is 2709 ± 22 Ma (Fig. 10b). The 2638 ± 16 Ma age is interpreted to indicate the
387 emplacement age.

388 The zircon grains from sample SM57 have similar luminescence (Fig. 10d) and shape,
389 which made it difficult to identify the zircon xenocrysts. Only a few grains contained inherited
390 cores, but they were too small to be analysed. A number of 22 zircon grains from this sample
391 returned concordant (15 grains) or nearly concordant (7 grains) ages. The youngest concordant
392 grain returned a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2615 ± 21 Ma while the oldest concordant grain returned a
393 $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2684 ± 20 Ma. The 2615 ± 21 Ma grain is similar within error to the weighted
394 average of high luminescence grains from sample SM56, while the 2684 ± 20 Ma grain is similar
395 within error to the ages of the deformed granites (Fig. 10e). Taking into account that the
396 Nyankumbu granite intruded into the TTG and the deformed granites, we can assume that all
397 grains with ages that are within error similar to the youngest age of the deformed granites or of
398 the TTG's are xenocrysts, while all grains with ages that are within error similar to sample
399 SM56 and younger than the syn-shearing granites belong to the Nyankumbu granite (Fig. 10e).
400 Based on this assumption, 7 grains can be interpreted as Nyankumbu granite zircons and the
401 remaining 15 grains can be interpreted as xenocrysts inherited from the surrounding older
402 granites. The $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age of the 7 zircon grains interpreted to be related
403 to the emplacement of the Nyankumbu granite is 2632 ± 16 Ma (Fig. 10e) with a similar
404 concordia age of 2637 ± 22 Ma (Fig. 10c). These ages are identical within error to the age of the
405 high luminescence zircon grains from sample SM56, and indicate that the emplacement of the
406 Nyankumbu granite took place between ~ 2632 and 2638 Ma.

407

408 **6. Discussion**

409

410 *6.1. Structural history of the Geita Shear Zone*

411

412 The history of the GSZ can be reconstructed based on the field relationships and the
413 new age data. Although, the contact between the TTG gneisses and the Kiziba Formation is not
414 exposed, the map pattern (Fig. 2), suggests that the deformation fabric in the Kiziba Formation
415 predates the deformation fabric in the TTG gneisses. For example, in the western part of the
416 GGB (Fig. 2) the foliation in the Kiziba Formation is truncated by the GSZ while further east
417 the foliation is deflected from NW trending into sub-parallelism with the foliation in the TTG,
418 suggesting that at least near the shear zone margin the GSZ affected the Kiziba Formation.

419 The TTG gneisses and the deformed granites, south of the shear zone, and the mafic
420 volcanics of the Kiziba Formation north of the shear zone, have similarly trending foliations
421 (Figs. 4a and c), but differently oriented mineral lineations (Figs. 4b and d) and associated
422 kinematic history, further supporting a different origin for the two deformation fabrics. The
423 Kiziba Formation is characterised by moderately NE plunging lineations with top to the south
424 shear-sense indicators, consistent with SW directed sinistral reverse movement while the
425 shallowly W plunging lineations and the lack of a dominant shear sense direction along the
426 mylonitic margin of the GSZ are consistent with N-S directed flattening strain. The kinematics
427 of the Kiziba formation is similar to the main period of deformation of the greenstone belt,
428 which produced the dominant D₂-D₃ folding with SW-directed vergence and the sinistral-
429 reverse D₆ shears present in the Geita Hill and Nyankanga deposits (Sanislav et al., 2015;
430 2017). Both the GSZ and the Kiziba Formation are affected and displaced by a series of dextral-
431 NW trending brittle-ductile shear zones (Fig. 2). These NW trending shear zones are of D₇

432 origin (Sanislav et al., 2015; 2017) meaning that the GSZ must predate D₇ deformation. In the
433 Geita Hill deposit, 2699 Ma (Borg and Krogh, 1999) diorite dykes are folded by the D₃ folds
434 suggesting that the main period of folding occurred after 2700 Ma. The high-K granites that
435 intruded the eastern, northern and western part of the greenstone belt between 2660 Ma and
436 2620 Ma (Sanislav et al., 2014) truncate all the deformation fabric in the greenstone including
437 the D₇ structures indicating a minimum age of 2660 Ma for the D₇ deformation. The ~ 2700
438 Ma (Fig. 6) tonalite bodies that intruded the Kiziba Formation preserve the same deformation
439 fabric with the host metabasalts indicating that the deformation of the Kiziba Formation
440 occurred most probably after 2700 Ma. Taking into account the sinistral-reverse kinematics
441 with top to the SW shearing of the Kiziba Formation it is likely that the deformation of the
442 Kiziba Formation also occurred between 2700 Ma and 2660 Ma. The two tonalite samples
443 collected from the TTG gneisses have similar ages (Fig. 6) with the tonalite samples that
444 intruded the Kiziba Formation, thus, constraining the maximum age for the GSZ to ~2700 Ma.
445 The three granite samples that intruded the TTG's and contain variably developed deformation
446 fabrics, and were emplacement during movement on the GSZ, thus, their emplacement ages of
447 2667 and 2680 Ma (Figs. 8 and 9) mark the main activity on the shear zone. The undeformed
448 Nyankumbu granite constrains the minimum age movement on the GSZ at ~2640 Ma (Fig. 10).
449 It is worth noting that the NW-trending D₇ shears do not appear to offset the contact of the
450 Nyankumbu granite, indicating that shearing on the GSZ ceased before the emplacement of the
451 Nyankumbu granite. Since, the D₇ shears may predate the emplacement of the 2660-2620 Ma
452 high-K granites the activity of the GSZ can be constrained to between 2690 Ma and 2660 Ma.

453

454 *6.2. Significance of the large-scale structure for the Archean Tanzania Craton*

455

456 The GSZ has a strike length of at least 50 km (Fig. 1) and geophysical data suggest that the
457 shear zone extends further W and E. An eastward extension of the GSZ will end up against the
458 boundary between the Mwanza-Lake Eyasi Superterrane and the Lake Nyanza Superterrane
459 (terminology based on Kabete et al., 2012a) while a westward extension will end up against
460 the Proterozoic Nyakahura-Burigi Terrane. This makes the GSZ an important shear zone with
461 large-scale consequences for the structure and tectonics of the Tanzania Craton. Assuming that
462 the large-scale subdivision of the Tanzania Craton into a series of sub-parallel terranes
463 separated by NW trending crustal-scale shear zones (Kabete et al., 2012a) is valid, then the
464 GSZ will be a second order regional shear zone. However, the existence and the nature of these
465 large-scale shear zones are yet to be confirmed by field studies and it is likely that the GSZ
466 may represent a first order structure. It is worth noting that at least parts of the stratigraphy
467 across the Sukumaland Greenstone Belt can be correlated. For example, the age and
468 geochemistry of the mafic volcanics in the SGB are similar so that the Kiziba Formation north
469 of the GSZ can be correlated with the mafic volcanics cropping out further south of the GSZ
470 (Manya and Maboko, 2003; Manya and Maboko, 2008; Cook et al., 2016). The detrital zircon
471 ages across the SGB and the intrusion of diorite sills suggests deposition of the Nyanzian
472 sediments by ~2700 Ma (e.g. Borg and Krogh, 1999; Sanislav et al., 2014). The main phase of
473 ductile deformation in the GGB is constrained by the ~ 2700 Ma maximum depositional age
474 of the meta-volcanics, and the intrusion of the high-K granites starting at ~ 2660 Ma (Sanislav
475 et al., 2014). The activity of the GSZ ceased at ~2660 Ma, which coincides with a switch in
476 magmatism from a period dominated by deeply sourced magmas such as diorite and TTG to a
477 period dominated by shallow level sourced magmas such as the high-K granites.

478

479 *6.3. The significance of the inherited cores*

480

481 Tonalite samples collected from both sides of the GSZ not only yield similar emplacement
482 ages, but also contain similarly aged zircon xenocrysts and inherited cores. This suggests that
483 the TTG gneisses that crop out south of the GSZ most probably extend northward beneath the
484 GGB. The TTG gneisses south of the GSZ were initially interpreted to represent Dodoman
485 age basement rocks on which the greenstone sequences were deposited (Bell and Dodson,
486 1981; Gabert, 1990; Kabete et al., 2012a). However, zircon dating of a migmatitic gneiss from
487 the southern end of the SGB returned an age on 2680 ± 3 Ma (Borg and Krogh, 1999) indicating
488 that at least some of the gneisses in the SGB are not of Dodoman age and do not constitute
489 basement rocks to the greenstone sequences. The zircon ages presented in this study show that
490 the age of emplacement of the TTG gneisses along the GSZ is very similar, at ~ 2710 Ma,
491 confirming previous interpretations that they do not represent Dodoman age rocks and cannot
492 be the basement to the greenstone sequence. For example, whole rock Sm-Nd model ages from
493 the mafic volcanics that form the base of the greenstone sequence indicate eruption at ~ 2820
494 Ma (Manya and Maboko, 2003; Manya and Maboko, 2008; Cook et al., 2016) making the
495 deposition of the mafic volcanics at least 100 Ma older than the emplacement of the TTG's.
496 The eruption age of the mafic volcanics is corroborated by zircon dating of interlayered rhyolite
497 and felsic tuffs dated at 2808 ± 3 Ma and 2821 ± 30 Ma respectively (Borg and Krogh, 1999;
498 Sanislav et al., 2014). It is worth noting that zircon xenocrysts and inherited cores from samples
499 collected from both sides of the GSZ yielded ages between 2826 and 2840 Ma which are similar
500 to the eruption age of the mafic and the interlayered felsic volcanics indicating the possibility
501 that partial melting of the base of the greenstone sequence may have contributed to the
502 petrogenesis of the TTG's. However, zircon cores as old as 3000 Ma or 3200 Ma found in
503 samples SM07 and K81 cannot be correlated, at the moment, with any of the rock units from
504 the northern half of the Tanzania Craton and raises the possibility of Dodoman age basement
505 rocks being present at depth (Fig. 11). Kabete et al., (2012a) proposed that at least part of the

506 greenstone sequence in the SGB was deposited on rifted Dodoman basement. It is possible that
507 >3Ga old basement rocks underlying the greenstone belt and the TTG gneisses (Fig. 11a)
508 constitute the source of the old zircons. However, >3Ga zircon xenocrysts are found only in
509 the proximity of the southern margin of the greenstone belt suggesting that the basement rocks
510 do not extend further north. A second possibility is that basement rocks underlie only the TTG's
511 (Fig. 11b). If this is the case, the GSZ must be dipping steeply N at depth to explain the
512 presence of >3Ga zircons within tonalite that intruded the Kiziba Formation. It also implies
513 that the GSZ has a much older history and may represent a major tectonic boundary separating
514 two distinct terrains. The trace element and isotope geochemistry of the mafic volcanics that
515 crop out north and south of the GSZ and form the base of the greenstone sequences in the SGB
516 show no evidence of crustal contamination and were most probably erupted on oceanic crust
517 (Manya and Maboko, 2003; Manya and Maboko, 2008; Cook et al., 2016). This data excludes
518 the possibility that the greenstone sequence was deposited on old basement. However, given
519 the fact that the geometry of the basin in which the greenstone sequence was deposited is
520 unknown, the possibility of small fragments of rifted Dodoman age rocks incorporated in the
521 greenstone sequence and underlying parts of the SGB (Fig. 11c) cannot be excluded.
522 Alternatively, these 3000 to 3200 Ma zircon cores could have a detrital origin of an unknown
523 source and were recycled through partial melting of amphibolite and sediment.

524 *6.4. Crustal growth and the assembly of Tanzania Craton*

525 The zircon age data (Fig. 12) suggests that the northern part of Tanzania Craton, comprising
526 the six granite-greenstone belts around Lake Victoria (Fig. 1), has a different age structure
527 compared to the central part of the craton. The central part of the craton is dominated by > 3000
528 Ma zircons (Fig. 12a) while the northern part is dominated by < 2850 Ma zircons (Fig. 12b).
529 Indeed, all six greenstone belts (Fig. 1) from the northern part of the craton appear to have
530 evolved within the same time period between 2600 Ma and 2850 Ma (Sanislav et al., 2014).

531 Moreover, new data suggests that some of these greenstone belts share not only a similar age
532 history but also a similar igneous history with TTG magmatism peaking around 2700 Ma and
533 transitioning into high-K magmatism post 2660 Ma. For example, Manya (2016a) showed that
534 in Nzega Greenstone Belt, TTG magmatism occurred at ~ 2710 Ma and was followed by high-
535 K magmatism at ~2670 Ma. In the SGB, the situation is almost identical, TTG magmatism
536 occurred mainly around 2710 Ma (this study) and transitioned into high-K magmatism post
537 2660 Ma (Sanislav et al., 2014). Musoma-Mara Greenstone Belt (Fig. 1), located further north,
538 has a similar igneous history, although, there appears to be evidence of felsic magmatism as
539 early as 2840 Ma (Manya et al., 2006; Sanislav et al., 2014). The central part of Tanzania
540 Craton contains ~2815-2691 Ma greenstone fragments which, are more or less coeval with the
541 greenstones from the northern part of the craton but they are embedded in > 3000 Ma granite
542 and gneisses (Kabete et al., 2012b). Thus the terrane boundary between the Lake Nyanza Super
543 Terrane and the Moyowosi-Manyoni Super Terrane (Fig. 1) could be a fundamental boundary
544 separating two distinct cratonic nuclei. In this context the source of the zircon xenocrysts older
545 than 3000 Ma from the TTGs along the Geita Shear Zone becomes important because may hold
546 the key to the timing of the assembly of the Tanzania Craton. If parts of central Tanzania
547 underlie the Sukumaland Greenstone Belt (Figs. 11a and b) and constitute the source of the >
548 3000 Ma zircon xenocrysts than the assembly of the Tanzania Craton must have occurred prior
549 to the emplacement of the ~2710 Ma TTGs. If that is the case dating of TTG gneisses from the
550 northern half of the craton, particularly near the boundary with the central part of the craton
551 should return an increasing amount of >3000 Ma zircon xenocrysts which is not the case. For
552 example, Manya (2016a) showed that the TTG gneisses from Buzwagi mine in the Nzega
553 Greenstone Belt (Fig. 1), located near the boundary with the central part of the craton, were
554 emplaced at 2713 ± 8 Ma, which is similar to the age of the TTG gneisses found along the Geita
555 Shear Zone. No zircon xenocrysts were reported. Moreover, Borg and Krogh (1999) proposed

556 that prior to 2700 Ma the Nyanzian sedimentation was still active in the Sukumaland
557 Greenstone Belt, a scenario which is at odds with the amalgamation of the craton in the same
558 time. In fact, all shortening occurred after 2700 Ma (Sanislav et al., 2017) and is more likely
559 that this period (<2700 Ma) coincides with the assembly of the Tanzania Craton and the Geita
560 Shear Zone accommodated part of the strain resulted from the amalgamation of the craton.

561 *6.5. Deformation style and Archean tectonics*

562 The Geita Shear Zone structural style, planar fabric with sub-horizontal stretching
563 lineations and a linear map pattern, is similar to other Neoproterozoic shear zone systems (e.g.
564 Dirks et al., 2002; Czarnota et al., 2010; Kabete et al., 2012a) and indicates horizontal
565 shortening as opposed to vertical, gravity-driven, of the dome sliding deformation style typical
566 for earlier Archean terranes (e.g. Collins et al., 1998; Van Kranendonk et al., 2007; Van
567 Kranendonk, 2011). This type of structural style, was interpreted to result from horizontal
568 tectonics responsible for terrane accretion in a geodynamic environment similar to modern day
569 plate tectonics in oblique convergent settings (e.g. Percival et al., 2006; Czarnota et al., 2010;
570 Manykiamba and Kerrich, 2012). It is worth mentioning that the structural styles typical of
571 modern plate tectonics include accretionary mélanges, overthrust ophiolites, paired
572 metamorphic belts and thrust and fold belts (e.g. Hamilton, 1998; Chardon et al., 1999; Stern,
573 2005; Brown, 2007). Based on the scarcity/ lack of the forementioned examples alternative
574 models involving subductionless continental drift were proposed to explain Archean horizontal
575 tectonism and terrane assembly (Bédard et al., 2013). The steep nature of the Geita Shear Zone
576 is not consistent with a thrust and fold belt scenario, even if it had a dip-slip component. This
577 is important because as noticed by Many and Maboko (2008), the mafic volcanics of the
578 Kiziba Formation, which form the base of the stratigraphy, are distributed throughout the
579 Sukumaland Greenstone Belt suggesting that no large scale duplication of stratigraphy

580 occurred. That means the shortening was accommodated mainly by the localization of
581 deformation along steeply dipping deformation zones similar to the Geita Shear Zone.

582 The fact that the shear zone is dominated by coaxial deformation and is characterized by a
583 near horizontal mineral stretching lineation implies lateral extrusion of material (e.g. Fossen
584 and Tikoff, 1998). The question on where the laterally extruded material ended up is a complex
585 one and requires knowledge of the geometry of the shear zone at the time it was active. If Geita
586 Shear Zone was linear than the angle of relative motion must change along its strike so that the
587 coaxial component of deformation is transferred into non-coaxial deformation. Alternatively,
588 the Geita Shear Zone had a convex shape with deformation being coaxial along the apex zone
589 and increasingly non-coaxial along the sides. Field data suggests that to the west (Fig. 1) the
590 shear zone preserve its linear character but it is overprinted by the Proterozoic mobile belts. To
591 the east the entire greenstone belt changes orientation and is likely that the Geita Shear Zone
592 curves south with possible lateral extrusion of material towards east.

593 **7. Conclusions**

594

595 In this contribution we provided the first field based description of a large scale shear zone
596 from the Tanzania Craton. The Geita Shear Zone is a major shear zone that occurs in the NW
597 part of the Tanzania Craton. The shear zone trends approximately E-W and has a strike length
598 of at least 50 km and up to a few kilometers in width. The shear zone was active between ~2690
599 Ma and ~2640 Ma, accommodated mostly flattening strain and most likely does not represent
600 a terrane boundary. However, the zircon age distribution from the central and northern part of
601 the Tanzania Craton is different, and suggests the presence of a major terrane boundary a few
602 hundred kilometers south of the Geita Shear Zone. The deformation style of the Geita Shear
603 Zone is similar to other Neoproterozoic shear zone systems and indicates horizontal shortening as
604 opposed to vertical, gravity-driven deformation styles typical for earlier Archean terranes.

605

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761 **Figure captions**

762 **Figure 1.** Simplified geological map of northern half of Tanzania Craton showing the main
763 geological units and the location of the greenstone belts (modified from Sanislav et al., 2015).
764 Super-terrane boundaries are as proposed by Kabete et al. (2012a). SU – Sukumaland
765 Greenstone Belt; NZ – Nzega Greenstone Belt; SM – Shynianga-Malita Greenstone Belt; IS –
766 Iramba-Sekenke Greenstone Belt; KF – Kilimafedha Greenstone Belt; MM – Musoma-Mara
767 Greenstone Belt.: ELVST – East Lake Victoria, MLEST- Mwanza Lake Eyasi, LNST- Lake
768 Nyanza, MMST – Moyowosi-Manyoni, DBST – Dodoma Basement, MAST – Mbulu-Masai,
769 NBT – Nyakahura-Burigi. Inset map of Africa showing the location of Archean blocks.

770 **Figure 2.** Geological map of Geita Greenstone Belt (modified from Sanislav et al., 2015)
771 showing the location of dated samples. The width of the zone showing the foliation trend lines
772 within the TTG gneisses along the southern margin of GGB marks the approximate width of
773 the Geita Shear Zone.

774 **Figure 3.** Photographs showing outcrops of granitoids found within the Geita Shear Zone. a)
775 mylonitic TTG near the greenstone margin; b) mylonitic TTG near the greenstone margin
776 intruded by syn-deformation granite. Note that the granite cuts across and interfingers with the
777 mylonitic foliation. The granite contains a similar but less intense foliation and mineral
778 lineation with the mylonitic tonalite; c) microphotograph of altered and flattened feldspar
779 porphyroclasts from near the contact with the Kiziba Formation. This sample was also affected
780 by D₇ brittle-ductile shearing and overprinted by epidote and chlorite veins; d) symmetric,
781 highly boudinaged quartz vein within the mylonitic TTG; e) asymmetric plagioclase

782 porphyroclasts indicating dextral movement found within the mylonitic TTG; f) TTG gneiss
783 from ~ 400 m south of the greenstone margin having a less intense deformation fabric; g)
784 photomicrograph showing weakly developed deformation fabric in TTG gneiss about 600 m
785 from the greenstone margin; h and i) porphyritic granite dykes and intrusions from ~ 600 m
786 south of the greenstone margin having only a weakly developed foliation. The feldspar
787 phenocrysts appear undeformed; j) outcrop photograph of undeformed Nyankumbu granite that
788 intruded and truncates the mylonitic fabric of Geita Shear Zone.

789 **Figure 4.** Stereoplots showing the poles to foliation planes and the orientation of the mineral
790 lineations for the TTG and granites (a and b) affected by the Geita Shear Zone and for the mafic
791 volcanics of the Kiziba Formation (c and d) that crop out near the shear zone.

792 **Figure 5.** Outcrop photographs of foliated Kiziba formation metabasalts showing an example
793 of internal folding (a) and of asymmetrically boudinaged quartz vein (b). It is uncommon to
794 find examples of internal folding within the metabasalts but geophysical images indicate that
795 along the southern margin of the greenstone belt large isoclinal folds with ~E-W trending axes
796 are present in the un-exposed areas. Asymmetrically boudinaged quartz veins indicating top to
797 the SW movement are common.

798 **Figure 6.** Example of cathodoluminescence images of zircon grains from sample SM02 (a).
799 Diagrams showing the $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age (b) and the concordia plot (c) for
800 sample SM02. Example of cathodoluminescence images of zircon grains from sample SM25
801 (d). Note the presence of inherited cores and of high luminescence rim overgrowths. Diagrams
802 showing the $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age (e) and the concordia plot (c) for sample SM25.
803 Inset diagrams show the results of analyses performed on the high luminescence rims.

804 **Figure 7.** Example of cathodoluminescence images of zircon grains from sample IH3 (a).
805 Diagrams showing the $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age (b) and the concordia plot (c) for

806 sample IH3. Example of cathodoluminescence images of zircon grains from sample K81 (d).
807 Diagrams showing the $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age (e) and the concordia plot (c) for
808 sample K81. Note that both samples contain inherited zircon grains and rim overgrowths. Inset
809 diagrams show the results of analyses performed on the inherited zircon cores.

810 **Figure 8.** Example of cathodoluminescence images of zircon grains from sample SM07 (a).
811 Note the presence of inherited zircon grains and of zircon grains with complex internal pattern.
812 Diagrams showing the $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age (b) and the concordia plot (c) for
813 sample SM07. Example of cathodoluminescence images of zircon grains from sample SM23
814 (d). Note the presence of inherited cores and of high luminescence rim overgrowths. Diagrams
815 showing the $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age (e) and the concordia plot (c) for sample SM23.

816 **Figure 9.** Example of cathodoluminescence images of zircon grains from sample SM27 (a).
817 Diagrams showing the $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age (b) and the concordia plot (c) for
818 sample SM027.

819 **Figure 10.** Example of cathodoluminescence images of zircon grains from sample SM56 (a).
820 Diagrams showing the $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average ages (b) and the concordia plot (c) for
821 different zircon populations from sample SM56. Example of cathodoluminescence images of
822 zircon grains from sample SM57 (d). Diagrams showing the $^{207}\text{Pb}/^{206}\text{Pb}$ weighted average age
823 (e) and the concordia plot (c) for the youngest zircon population for sample SM57. Both
824 samples contain inherited zircon grains having similar ages with the surrounding TTG and
825 granites.

826 **Figure 11.** Series of cartoons illustrating three possible scenarios involving the presence at
827 depth of > 3 Ga basement rocks to explain the old zircon ages found in the TTG and granite
828 samples from both sides of the Geita Shear Zone. In all three scenarios, the melt gets
829 contaminated with old zircons during ascent. It is worth noting that, so far, old zircon grains

830 were found only within the vicinity of the Geita Shear Zone. a) Old basement underlying both
831 the TTG and the greenstone. If that is the case old zircon grains could be found further south.
832 b) Old basement underlying only the TTG and GSZ dips north at lower depths. If this is the
833 case, GSZ could be a major tectonic boundary separating two distinct terrains. It would also
834 suggest that the shear zones initiated before 2700 Ma. Old zircon grains could be found further
835 south. c) A fragment of rifted old basement underlies part of the greenstone belt. If that is the
836 case old zircon grains would occur only in certain domains.

837 **Figure 12.** Diagrams showing the distribution of zircon ages for central Tanzania (a) and for
838 northern Tanzania (b). Data from Sanislav et al., (2014) updated with new age data from Many
839 (2016a) and present study.

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