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

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Abstract

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

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

1. Introduction

Large scale shear zones are well documented features in Archean terrains worldwide (e.g. Bédard et al., 2003; Cassidy et al., 2006; Chardon et al., 2008; Jelsma and Dirks, 2002; Dirks et al., 2013) and their structural style has implications for the tectonic history and crustal growth models of Archaean cratons. One of the most common Archean structural styles is the dome and keel geometry (e.g. Pilbara, Kaapvaal, Western Dharwar), linked to gravity-driven tectonics and redistribution of rock masses through vertical processes (e.g. Choukroune et al., 1995; Chardon et al., 1996; Collins et al., 1998; Bédard et al., 2003; Van Kranendonk, 2011). The majority of these types of structures are characterised by gneiss domes with radial elongation lineations and normal shear sense indicators suggesting diapiric ascent through the crust. Between the gneiss domes, synformal-shaped greenstone belts are preserved. In general, the geometry of the greenstone belts follows the contours of the gneiss domes resulting in cusplike shapes. Typical examples include the Barberton Greenstone Belt in South Africa (e.g. Dirks et al., 2013; Van Kranendonk et al., 2014; Brown, 2015) and the greenstone belts in the Pilbara Craton of Western Australia (e.g. Collins et al., 1998; Van Kranendonk et al., 2004). Alternative interpretations for dome and keel geometries include core complexes exposed by extensional unroofing (Kloppenburg et al., 2001) or antiformal culminations exposed by cross-folding (Blewett et al., 2004).

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

The structural and tectonic evolution of the Archean Tanzania Craton is poorly understood (Kabete et al., 2012a; Kabete et al., 2012b) and in many instances stratigraphic correlations based on sparse geochronological data are contradictory (e.g. Borg and Krogh, 1999; Manya et al., 2006; Sanislav et al., 2014). A first attempt to define tectonic and structural boundaries within the Tanzania Craton was made by Kabete et al. (2012a) based on existing geological maps and the interpretation of geophysical datasets. A series of NW trending shear zones were proposed and interpreted to delineate superterrane boundaries (Fig. 1). However, the existence of these large-scale shear zones, their continuity, style, age and kinematic sense was never confirmed with field studies, and the terrane-accretion model, therefore, remains entirely speculative.

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

2. Regional geology

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

The Nyanzian and the Kavirondian Supergroups are concentrated in the northern half of the Tanzania Craton and were placed stratigraphically above the Dodoman Supergroup (e.g. Quenell et al 1956; Gabert, 1990). The Nyanzian Supergroup is overlain unconformably (Gabert, 1990) by the Kavirondian Supergroup. The lowermost unit of the Nyanzian is dominated by mafic volcanics and volcanoclastics with minor felsic volcanics erupted at ~ 2820 Ma (e.g. Borg and Krogh, 1999; Manya and Maboko, 2003; Cook et al., 2016). These are overlain by felsic volcanic and pyroclastic units inter-bedded with banded ironstone, volcanoclastic sequences and immature turbiditic sediments (e.g. Borg, 1992; Borg and Shackleton, 1997; Sanislav et al., 2015). All greenstone sequences from the northern half of Tanzania were assigned to the Nyanzian Supergroup. Six major greenstone belts were described (Fig. 1): the Sukumaland Greenstone Belt (SGB), The Nzega Greenstone Belt, the Iramba-Sekenke Greenstone Belt, the Shynianga-Malita Greenstone Belt, the Kilimafedha Greenstone Belt and the Musoma-Mara Greenstone Belt (Borg and Shackleton, 1997). The greenstone belts are fragmented by shear zones and intruded by ~2800 to 2600 Ma granitoids (Kabete et al., 2012b; Sanislav et al., 2014). Most of the greenstone belts are oriented along ~ E-W lineaments (Fig. 1), one of these lineaments forming the southern margin of the Geita Greenstone Belt. The Kavirondian Supergroup consists mainly of coarse-grained conglomerate, grit and quartzite (Gabert, 1990) and its areal extent is poorly constrained.

The Geita Greenstone Belt (GGB) is a large greenstone fragment forming most of the northern part of what has previously been defined as the SGB (Fig. 1). It contains elements of the Nyanzian and the Kavirondian stratigraphy (Fig. 2). The base of the greenstone belt is dominated by lower-amphibolite facies metamorphosed mafic volcanics forming the Kiziba Formation (Cook et al., 2016). Whole rock Sm-Nd dating of the mafic volcanics yield model ages of ~2820 Ma (Manya and Maboko, 2008; Cook et al., 2016). The mafic volcanics of the Kiziba Formation are positioned below a complexly deformed sequence of intercalated volcanoclastics, ironstones and immature turbidites, which experienced greenschist facies metamorphism (Cook et al., 2016; Sanislav et al., 2015, 2017). The greenstone sequence was intruded syntectonically by diorite (D_2 to D_4) and by late-syntectonic felsic porphyries (D_6 to D_7). The northern, eastern and western margins of the greenstone are intruded by 2660 to 2620

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

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

3. Geita Shear Zone

The Geita Shear Zone (GSZ) is a mylonitic shear zone that defines the southern margin of the GGB and can be mapped along strike for at least 50 km (Fig. 2). The general orientation of the GSZ is ~ E-W (Fig. 2) and the overall straight map pattern and gravity data for the greenstone belt suggest that the shear zone is steeply dipping. Regional aeromagnetic data suggest that the GSZ extends further to the east and west (Fig. 1) outside of the area mapped in Figure 2. In the Geita area the shear zone marks the boundary between mafic volcanics of the Kiziba Formation to the north and TTG granitoid-dominated gneiss and granites to the south (Fig. 2). The shear zone is generally poorly exposed, and is characterised on aeromagnetic data sets by a 700-900 m wide zone with an intensely developed planar fabric (Fig. 2), and in the field by scattered outcrops of partly recrystallised mylonite and striped gneiss (Fig. 3a and b) Aeromagnetic datasets further indicate the presence of well-foliated gneiss across a width of at least 0.8 km to the south of the main shear zone, whilst to the north, the shear zone appears to transect well-foliated meta basalt with the northern boundary of the GSZ defined by a low-angle truncation zone within deformed meta-basalt. The contact between the Kiziba Formation and the TTG gneisses is not exposed anywhere in the Geita Greenstone Belt, but the two units crop out within 50 meters of each other SW of Geita Town (GR 411520-9679010 - all grid references in WGS84-36S), which represents one of the few areas with excellent outcrop in the shear zone.

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

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

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

Along the entire length of the GGB, the GSZ has been truncated and displaced by NW-trending, near-vertical, D_7 faults (Fig. 2; Sanislav et al., 2015; 2017). These faults contain near-horizontal lineations defined by quartz fibres and striations and brittle-ductile deformation textures including well preserved S-C fabrics indicative of a dextral sense of shear. The D_7 shear zones are characterised by numerous, planar milky quartz veins that intruded parallel to the strike of the D_7 faults. Networks of conjugate pairs of NW- and NE-trending quartz veins are also common in the mylonitic gneisses of the GSZ within the vicinity of D_7 faults. Dextral displacements of the GSZ along the principle D_7 faults are in the order of 2-4 km.

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

Away from the core of the shear zone, some of the truncating granite dykes show only weak fabrics or no deformation features at all. Some granite dykes did not develop a foliation, but preserve a stretching lineation parallel to the dominant lineation direction in the shear zone, defined by quartz aggregates, aligned biotite grains and weakly deformed feldspar phenocrysts. The TTG gneisses are medium- to coarse-grained and their mineralogy is dominated by quartz and plagioclase with minor amounts of k-feldspar, biotite and hornblende. The cross cutting granite dykes and bodies intruding the gneisses of the GSZ during the shearing events are quartz-rich and have approximately equal amounts of plagioclase and k-feldspar with only minor amounts of biotite and hornblende. Based on their mineralogy the granites have been classified as monzogranites. It is common for these granites to contain large k-feldspar phenocrysts (Figs. 3h and i).

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

4. Geochronology

Nine samples were collected for zircon dating along and across the GSZ (Fig. 2) to constrain the time of shearing. Seven samples come from the granitoids south of the GSZ and two samples (K81 and IH3) come from small tonalite bodies that intruded into the Kiziba Formation. These tonalite bodies are part of the greenstone sequence affected by the GSZ and their time of emplacement is interpreted to provide an upper age limit for the GSZ. From south of the GSZ two samples were collected from the TTG gneisses (SM02 and SM25), which may provide a maximum time for shearing. Three samples were collected from the deformed granites that intruded into the TTG gneisses and that retain the shear fabric, i.e. these granites may have been intruded during shearing (SM07, SM23 and SM27). Two samples were collected from the undeformed Nyankumbu granite (SM56 and SM57) that crosscuts the GSZ and constrain the cessation of shearing along the GSZ.

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

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

5. Results

5.1. The TTG samples predating the GSZ

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

The cathodoluminescence images of the zircon grains from sample SM25 revealed the presence of grains with complex internal growth domains (Fig. 6d). The majority of the zircons grains contain a low luminescence core with a weakly developed concentric zoning surrounded by a very thin, high luminescence rim (Fig. 6d). Two concordant and nine nearly concordant analyses from this domain have a $^{207}Pb/^{206}Pb$ weighted average of 2706±13 Ma (Fig. 6e) and an identical concordia age of 2706±13 Ma (Fig. 6f), which are interpreted to represent the emplacement age. One large zircon grain (Fig. 6d) contains an elongated and sub-rounded inherited core with low luminescence and without any internal zoning surrounded by a thick median overgrowth with higher luminescence and concentric zoning. The median overgrowth has a thin and weakly developed high luminosity rim. Two U-Pb analyses from the inherited core yielded a concordant analysis with a $^{207}Pb/^{206}Pb$ age of 2840±22 Ma and one nearly concordant analysis with a $^{207}Pb/^{206}Pb$ age of 2826±21 Ma. One nearly concordant analysis from the median overgrowth gave a $^{207}Pb/^{206}Pb$ age of 2710±22 Ma, which is similar to the emplacement age. Three zircon grains contained large, low luminescence cores surrounded by well-developed, high luminescence rims (Fig. 6d). Only one core analysis returned a nearly concordant $^{207}Pb/^{206}Pb$ age of 2702±20 Ma, similar to the emplacement age, while the rims returned nearly concordant analyses with a $^{207}Pb/^{206}Pb$ weighted average age of 2661±24 Ma (Fig. 6e) and a concordia age of 2665±41 Ma (Fig. 6f). The rim age is interpreted to represent zircon overgrowth during a later thermal event related to the emplacement of the granites that intruded the TTG's during shearing along the GSZ.

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

Samples IH3 and K81 were collected from two small tonalite bodies that intruded into the mafic rocks of the Kiziba Formation (Fig. 2). Both tonalite outcrops contain a well-developed foliation having a similar orientation to the foliation found in the surrounding mafic volcanics, which occur to the north of the GSZ. Sample IH3 was collected near the contact with the TTG gneisses where the foliation is sup-parallel to the trend of GSZ. Sample K81 was collected further north of the GSZ boundary where the foliation forms a small angle to the trend of the GSZ. The tonalite bodies are interpreted to have been emplaced before development of the GSZ. The zircon grains separated from these samples are generally medium- to small-sized, prismatic, and mostly euhedral in shape. Cathodoluminescence images of zircon grains from sample IH3 (Fig. 7a) show that most of the zircon grains have low luminescence and poorly developed, concentric zoning. It is common for these zircon grains to display a thin, high-luminescence band, which only partly surrounds the grain, and which, in turn, is overgrown by a narrow rim (Fig. 7a). Ten concordant and near concordant analyses from these zircons gave a $^{207}Pb/^{206}Pb$ weighted average age of 2708±15 Ma (Fig. 7b) and a similar concordia age of 2708±25 Ma (Fig. 7c). These ages are interpreted to indicate the emplacement age of the tonalite bodies. The rim overgrowth was too narrow to be analysed. Sample IH3 also contained a few higher luminescence zircon grains (Fig. 7a) with some of them having a narrow rim overgrowth. Four analyses of these higher luminescence zircon grains returned concordant or near concordant ages. Their $^{207}Pb/^{206}Pb$ weighted average age is 2831±21 Ma (Fig. 7b) while the concordia age is 2831±22 Ma (Fig. 7c). These ages are similar to the ages of the inherited zircon cores found in sample SM25 suggesting that these two intrusions may have sampled a similar reservoir.

Sample K81 yielded a large number of zircon grains. Cathodoluminescence images revealed that the majority of zircon grains are similar to the zircon grains separated from sample IH3; in general, they display low luminescence, with poorly developed concentric zoning, with or without a thin outer high-luminescence rim and a further, narrow rim overgrowth (Fig. 7d). Twenty-three concordant and nearly concordant analyses from these zircons gave a $^{207}Pb/^{206}Pb$ weighted average age of 2711±8 Ma (Fig. 7e) and an identical concordia age of 2711±8 Ma (Fig. 7f), which are interpreted to indicate the emplacement age. However, many of these zircon grains contain irregularly shaped, inherited cores with very low luminescence and no internal zoning (Fig. 7d). These cores always contain a thin and highly luminescent outer band that may or may not completely surround the core and have a narrow rim overgrowth. A few of these cores were large enough to be analysed. One nearly concordant analysis from a low luminescence core yielded a $^{207}Pb/^{206}Pb$ age of 3231±19 Ma. Six other analyses performed on inherited cores yielded almost identical ages so that they were treated together. The $^{207}Pb/^{206}Pb$ weighted average age of these six analyses is 3004±21 Ma (Fig. 7e) while their concordia age is 3005±26 Ma (Fig. 7f).

5.3. The granite samples emplaced during movement on the GSZ

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

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

5.4.The Nyankumbu granite emplaced into the GSZ

Two samples (SM56 and SM57) were collected from the Nyankumbu granite to constrain the minimum age of the GSZ. Sample SM57 yielded a large number of zircon grains while sample SM56 yielded only few zircon grains. Both samples contain a mixed zircon population with inherited cores and xenocrysts. Cathodoluminescence images show that in sample SM56 the majority of the grains are in general high luminescence with well-defined growth zoning (Fig. 10a). The inherited cores have low luminescence and in many instances are surrounded by rim overgrowths with a higher luminescence. A few zircon grains with low luminescence and a weakly defined growth zoning were interpreted as xenocrysts. Five high luminescence zircon grains yielded nearly concordant results with a $^{207}Pb/^{206}Pb$ weighted average age of 2638±16 Ma (Fig. 10b) and an identical concordia age (Fig. 10c). Two low luminescence zircon xenocrysts yielded a combined $^{207}Pb/^{206}Pb$ age of 2655±25 Ma (Fig. 10b). Three zircon grains contained inherited cores that were large enough to be analysed (Fig. 10a). Their $^{207}Pb/^{206}Pb$ ages are similar so that they were treated together and their $^{207}Pb/^{206}Pb$ weighted average age is 2709±22 Ma (Fig. 10b). The 2638±16 Ma age is interpreted to indicate the emplacement age.

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

6. Discussion

6.1. Structural history of the Geita Shear Zone

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

The TTG gneisses and the deformed granites, south of the shear zone, and the mafic volcanics of the Kiziba Formation north of the shear zone, have similarly trending foliations (Figs. 4a and c), but differently oriented mineral lineations (Figs. 4b and d) and associated kinematic history, further supporting a different origin for the two deformation fabrics. The Kiziba Formation is characterised by moderately NE plunging lineations with top to the south shear-sense indicators, consistent with SW directed sinistral reverse movement while the shallowly W plunging lineations and the lack of a dominant shear sense direction along the mylonitic margin of the GSZ are consistent with N-S directed flattening strain. The kinematics of the Kiziba formation is similar to the main period of deformation of the greenstone belt, which produced the dominant D_2-D_3 folding with SW-directed vergence and the sinistral-reverse D_6 shears present in the Geita Hill and Nyankanga deposits (Sanislav et al., 2015; 2017). Both the GSZ and the Kiziba Formation are affected and displaced by a series of dextral-NW trending brittle-ductile shear zones (Fig. 2). These NW trending shear zones are of D_7

origin (Sanislav et al., 2015; 2017) meaning that the GSZ must predate D_7 deformation. In the Geita Hill deposit, 2699 Ma (Borg and Krogh, 1999) diorite dykes are folded by the D_3 folds suggesting that the main period of folding occurred after 2700 Ma. The high-K granites that intruded the eastern, northern and western part of the greenstone belt between 2660 Ma and 2620 Ma (Sanislav et al., 2014) truncate all the deformation fabric in the greenstone including the D_7 structures indicating a minimum age of 2660 Ma for the D_7 deformation. The ~ 2700 Ma (Fig. 6) tonalite bodies that intruded the Kiziba Formation preserve the same deformation fabric with the host metabasalts indicating that the deformation of the Kiziba Formation occurred most probably after 2700 Ma. Taking into account the sinistral-reverse kinematics with top to the SW shearing of the Kiziba Formation it is likely that the deformation of the Kiziba Formation also occurred between 2700 Ma and 2660 Ma. The two tonalite samples collected from the TTG gneisses have similar ages (Fig. 6) with the tonalite samples that intruded the Kiziba Formation, thus, constraining the maximum age for the GSZ to ~2700 Ma. The three granite samples that intruded the TTG's and contain variably developed deformation fabrics, and were emplacement during movement on the GSZ, thus, their emplacement ages of 2667 and 2680 Ma (Figs. 8 and 9) mark the main activity on the shear zone. The undeformed Nyankumbu granite constrains the minimum age movement on the GSZ at ~2640 Ma (Fig. 10). It is worth noting that the NW-trending D_7 shears do not appear to offset the contact of the Nyankumbu granite, indicating that shearing on the GSZ ceased before the emplacement of the Nyankumbu granite. Since, the D_7 shears may predate the emplacement of the 2660-2620 Ma high-K granites the activity of the GSZ can be constrained to between 2690 Ma and 2660 Ma.

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

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

6.3.The significance of the inherited cores

Tonalite samples collected from both sides of the GSZ not only yield similar emplacement ages, but also contain similarly aged zircon xenocrysts and inherited cores. This suggests that the TTG gneisses that crop out south of the GSZ most probably extend northward beneath the GGB. The TTG gneisses south of the GSZ were initially interpreted to represent Dodoman age basement rocks on which the greenstone sequences were deposited (Bell and Dodson, 1981; Gabert, 1990; Kabete et al., 2012a). However, zircon dating of a migmatitic gneiss from the southern end of the SGB returned an age on 2680±3 Ma (Borg and Krogh, 1999) indicating that at least some of the gneisses in the SGB are not of Dodoman age and do not constitute basement rocks to the greenstone sequences. The zircon ages presented in this study show that the age of emplacement of the TTG gneisses along the GSZ is very similar, at ~ 2710 Ma, confirming previous interpretations that they do not represent Dodoman age rocks and cannot be the basement to the greenstone sequence. For example, whole rock Sm-Nd model ages from the mafic volcanics that form the base of the greenstone sequence indicate eruption at ~2820 Ma (Manya and Maboko, 2003; Manya and Maboko; 2008; Cook et al., 2016) making the deposition of the mafic volcanics at least 100 Ma older than the emplacement of the TTG's. The eruption age of the mafic volcanics is corroborated by zircon dating of interlayered rhyolite and felsic tuffs dated at 2808±3 Ma and 2821±30 Ma respectively (Borg and Krogh, 1999; Sanislav et al., 2014). It is worth noting that zircon xenocrysts and inherited cores from samples collected from both sides of the GSZ yielded ages between 2826 and 2840 Ma which are similar to the eruption age of the mafic and the interlayered felsic volcanics indicating the possibility that partial melting of the base of the greenstone sequence may have contributed to the petrogenesis of the TTG's. However, zircon cores as old as 3000 Ma or 3200 Ma found in samples SM07 and K81 cannot be correlated, at the moment, with any of the rock units from the northern half of the Tanzania Craton and raises the possibility of Dodoman age basement rocks being present at depth (Fig. 11). Kabete et al., (2012a) proposed that at least part of the greenstone sequence in the SGB was deposited on rifted Dodoman basement. It is possible that >3Ga old basement rocks underlying the greenstone belt and the TTG gneisses (Fig. 11a) constitute the source of the old zircons. However, >3Ga zircon xenocrysts are found only in the proximity of the southern margin of the greenstone belt suggesting that the basement rocks do not extend further north. A second possibility is that basement rocks underlie only the TTG's (Fig. 11b). If this is the case, the GSZ must be dipping steeply N at depth to explain the presence of >3Ga zircons within tonalite that intruded the Kiziba Formation. It also implies that the GSZ has a much older history and may represent a major tectonic boundary separating two distinct terrains. The trace element and isotope geochemistry of the mafic volcanics that crop out north and south of the GSZ and form the base of the greenstone sequences in the SGB show no evidence of crustal contamination and were most probably erupted on oceanic crust (Manya and Maboko, 2003; Manya and Maboko, 2008; Cook et al., 2016). This data excludes the possibility that the greenstone sequence was deposited on old basement. However, given the fact that the geometry of the basin in which the greenstone sequence was deposited is unknown, the possibility of small fragments of rifted Dodoman age rocks incorporated in the greenstone sequence and underlying parts of the SGB (Fig. 11c) cannot be excluded. Alternatively, these 3000 to 3200 Ma zircon cores could have a detrital origin of an unknown source and were recycled through partial melting of amphibolite and sediment.

6.4. Crustal growth and the assembly of Tanzania Craton

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

Moreover, new data suggests that some of these greenstone belts share not only a similar age history but also a similar igneous history with TTG magmatism peaking around 2700 Ma and transitioning into high-K magmatism post 2660 Ma. For example, Manya (2016a) showed that in Nzega Greenstone Belt, TTG magmatism occurred at ~ 2710 Ma and was followed by high-K magmatism at ~2670 Ma. In the SGB, the situation is almost identical, TTG magmatism occurred mainly around 2710 Ma (this study) and transitioned into high-K magmatism post 2660 Ma (Sanislav et al., 2014). Musoma-Mara Grennstone Belt (Fig. 1), located further north, has a similar igneous history, although, there appears to be evidence of felsic magmatism as early as 2840 Ma (Manya et al., 2006; Sanislav et al., 2014). The central part of Tanzania Craton contains ~2815-2691 Ma greenstone fragments which, are more or less coeval with the greenstones from the northern part of the craton but they are embedded in > 3000 Ma granite and gneisses (Kabete et al., 2012b). Thus the terrane boundary between the Lake Nyanza Super Terrane and the Moyowosi-Manyoni Super Terrane (Fig. 1) could be a fundamental boundary separating two distinct cratonic nuclei. In this context the source of the zircon xenocrysts older than 3000 Ma from the TTGs along the Geita Shear Zone becomes important because may hold the key to the timing of the assembly of the Tanzania Craton. If parts of central Tanzania underlie the Sukumaland Greenstone Belt (Figs. 11a and b) and constitute the source of the > 3000 Ma zircon xenocrysts than the assembly of the Tanzania Craton must have occurred prior to the emplacement of the ~2710 Ma TTGs. If that is the case dating of TTG gneisses from the northern half of the craton, particularly near the boundary with the central part of the craton should return an increasing amount of >3000 Ma zircon xenocrysts which is not the case. For example, Manya (2016a) showed that the TTG gneisses from Buzwagi mine in the Nzega Greenstone Belt (Fig. 1), located near the boundary with the central part of the craton, were emplaced at 2713±8 Ma, which is similar to the age of the TTG gneisses found along the Geita Shear Zone. No zircon xenocrysts were reported. Moreover, Borg and Krogh (1999) proposed that prior to 2700 Ma the Nyanzian sedimentation was still active in the Sukumaland Greenstone Belt, a scenario which is at odds with the amalgamation of the craton in the same time. In fact, all shortening occurred after 2700 Ma (Sanislav et al., 2017) and is more likely that this period (<2700 Ma) coincides with the assembly of the Tanzania Craton and the Geita Shear Zone accommodated part of the strain resulted from the amalgamation of the craton.

6.5.Deformation style and Archean tectonics

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

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

7. Conclusions

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

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Figure captions

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

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

Figure 3. Photographs showing outcrops of granitoids found within the Geita Shear Zone. a) mylonitic TTG near the greenstone margin; b) mylonitic TTG near the greenstone margin intruded by syn-deformation granite. Note that the granite cuts across and interfingers with the mylonitic foliation. The granite contains a similar but less intense foliation and mineral lineation with the mylonitic tonalite; c) microphotograph of altered and flattened feldspar porphyroclasts from near the contact with the Kiziba Formation. This sample was also affected by D_7 brittle-ductile shearing and overprinted by epidote and chlorite veins; d) symmetric, highly boudinaged quartz vein within the mylonitic TTG; e) asymmetric plagioclase

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

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

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

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

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

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

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

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

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

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

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

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