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The petrogenesis of the Neoarchean Kukuluma Intrusive Complex, NW Tanzania

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Abstract

The Kukuluma Intrusive Complex (KIC) is a late Archean igneous complex, dominated by monzonite and diorite with subordinated granodiorite. The monzonite and the diorite suites have low silica content ($SiO_2 \leq$ 62 wt%), moderate Mg# ($Mg\#_{average}$ = 49), high Sr/Y ($Sr/Y_{average}$ = 79) and high La/Yb ($La/Yb_{average}$ = 56) ratios, and strongly fractionated (La_n/Yb_n = 9 to 69) REE patterns. Their moderate Ni ($Ni_{average}$= 50 ppm), Cr ($Cr_{average}$= 85 ppm), variable Cr/Ni ratio (0.65-3.56) and low TiO_2 ($TiO_{2average}$= 0.5 wt%) indicate little to no interaction with the peridotitic mantle. For most major elements (Al_2O_3, FeO_t, Na_2O, TiO_2 and P_2O_5) the monzonite and the diorite suites display subparallel trends for the same SiO_2 content indicating they represent distinct melts. Intrusions belonging to the diorite suite have high Na_2O ($Na_2O_{average}$ = 4.2 wt %), Dy/Yb_n ($Dy/Yb_{n\text{-}average}$ = 1.6), a positive Sr anomaly and uncorrelated Nb/La and Zr/Sm ratios suggesting derivation from partial melting of garnet-bearing amphibolite. Intrusions belonging to the monzonite suite have higher Na_2O

($Na2O_{average}$ = 5.61 wt %), Dy/Yb_n ($Dy/Yb_{n\text{-}average}$ = 2.21), a negative Sr anomaly and correlated Nb/La and Zr/Sm ratios consistent with derivation from partial melting of eclogite with residual rutile. Small variations in the Th/U ratio and near chondritic/MORB average values ($Th/U_{monzonite}$ = 3.65; $Th/U_{diorite}$ = 2.92) are inconsistent with a subducting slab signature, and it is proposed that the monzonite and the diorite suites of the KIC formed by partial melting of garnet-bearing, lower mafic crust of an oceanic plateau. The granodiorite suite has lower Mg# ($Mg\#_{average}$ = 41), moderately fractionated REE, low Sr/Y ($Sr/Y_{average}$ = 20), La/Yb ($La/Yb_{average}$ = 15), Dy/Yb_n ($DyYb_{n\text{-}average}$= 1.24) and small negative Eu anomalies suggesting derivation from partial melting of amphibolite and plagioclase fractionation. Near-MORB Th/U ($Th/U_{average}$ = 2.92) and Zr/Sm ($Zr/Sm_{average}$ = 30.21) ratios are consistent with intracrustal melting of amphibolite.

Archean rocks with an “adakitic” geochemical signature have been used to argue in favour of a plate tectonics regime in the Archean. The results presented here suggest that tectonic models for the Tanzania Craton, which invoke a subduction-related setting for all greenstone belts may need revision.

Introduction

The geochemical signature of intermediate to felsic rocks with fractionated REE patterns and high Sr/Y and La/Yb ratios has been interpreted to indicate melt derivation from a subducted slab at amphibolite to eclogite facies conditions (Defant and Drummond, 1990; Drummond and Defant, 1990). Their particular geochemical signature, including a high Mg# (molecular (Mg/Mg+Fe) x100) and enriched large-ion lithophile elements (LILE) were interpreted to represent different degrees of interaction between slab melts and mantle peridotite in the mantle wedge (e.g. Kay, 1978; Tatsumi and Ishizaka, 1981; Shirey and

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Hanson, 1984; Stern et al., 1989; Defant and Drummond, 1990; Drummond and Defant, 1990; Tatsumi, 2006; Moyen, 2009; Castillo, 2012). Arc rocks with similar geochemical signatures, including andesite, dacite, sodic rhyolite and their plutonic equivalents, were grouped under the term "adakites" by Defant and Drummond (1990) implying they share a specific petrogenetic history, namely, melting of the subducted slab. Another class of rocks sharing similar petrogenetic processes (e.g. melting of mantle peridotite metasomatised by subduction fluids/melts) and geochemically similar to adakites includes the high-Mg andesites or sanukitoids (e.g. Tatsumi and Ishizaka, 1981; Shirey and Hanson, 1984; Tatsumi, 2006; Tatsumi, 2008), and the crustal contaminated sanukitoids of South India described as "Closepet-type" granites (Jayanada et al., 1995).

Both adakites and sanukitoids are derived from melting of a metamorphosed, garnet-bearing, mafic igneous rock protolith (e.g. Thorkelson and Breitsprecher, 2005). In the case of sanukitoids this probably involved melting of a metasomatised mantle wedge, and in the case of adakites the subducting slab, but the important message is that both suites are generally interpreted as imparting a subduction signature. Martin et al. (2005) subdivided adakites into two groups, low-silica adakites (LSA) and high-silica adakites (HSA), corresponding to distinct petrogenetic processes. In this subdivision the petrogenesis of LSA involves melting of subduction modified peridotite as originally proposed by Defant and Drummond (1990). In contrast, the HSA are proposed to be analogues of the late Archean tonalite-trondhjemite-granodiorite (TTG) magmas and derived from partial melts of subducted basaltic crust in the garnet stability field, which reacted with peridotite during ascent (Martin et al., 2005). Archean rocks with an "adakitic" geochemical signature have been used to argue in favour of a plate tectonics regime in the Archean (e.g. Martin, 1999; Polat and Kerrich, 2001; Manya et al., 2007; Manykyamba et al., 2007; Mohan et al., 2013; Kwelwa et al., 2013).

Alternative models for rocks with an “adakitic’ signature have been proposed, and involve fractional crystallization in the garnet stability field (e.g. Kamber et al., 2002; Macpherson et al., 2006; Richard and Kerrich, 2007; Rooney et al., 2011), by melting of thickened mafic lower crust (e.g. Atherton and Petford, 1993; Rudnick, 1995; Wang et al., 2005), or through the interaction of delaminated eclogitic lower crust with the underlying mantle (e.g. Bedard et al., 2003; Tuloch and Kimbrough, 2003; Gao et al., 2004; Wang et al., 2005; Goss et al. 2011). Since most of the continental crust was formed in the Archean (e.g. Taylor and Mclennan, 1995; Tatsumi, 2008; Hacker et al., 2015) and the Archean rock record is dominated by rocks with an adakite-like (TTG’s; e.g. Condie, 2005; Moyen, 2011; Moyen and Martin, 2012) geochemical signature, understanding the petrogenetic processes that resulted in the formation of rocks with an adakitic signature in the Archean is essential. This is particularly important for the late Archean period when major shifts in the composition of the TTG suites are interpreted to reflect fundamental changes in global tectonics (e.g. Condie, 2005; Martin et al., 2010; Moyen and Martin, 2012; Condie, 2014). In this contribution we present major and trace element geochemical data from the Kukuluma Intrusive Complex (KIC) that intruded the Neoarchean Geita Greenstone Belt of NW Tanzania and discuss the petrogenesis of the KIC, and the implications of this for the tectonic evolution of the Geita Greenstone Belt.

Regional geology

From a geological perspective, the Tanzania Craton was initially divided in three major litho-stratigraphic units: the Dodoman, the Nyanzian and the Kavirondian Supergroups (e.g. Stockley, 1936; Quennel et al., 1956; Harpum, 1970; Gabert, 1990). The Dodoman was interpreted to represent the basement to the Nyanzian, while the Kavirondian unconformably overlays the Nyanzian. The Dodoman consists of high-grade mafic and felsic granulite with subordinate lower-grade schist and thin slivers of greenstone; the Nyanzian consists of mafic and felsic volcanics, ironstone, tuff and epiclastic sediments, while the Kavirondian consists of conglomerate, quartzite, grit, sandstone and siltstone. Borg and Krogh (1999) have shown that migmatitic gneisses, dated at 2680±3 Ma, that occur in the northern part of the Tanzania Craton are much younger than the Dodoman age units (interpreted to be ≥ 3000 Ma in age), and, therefore, cannot represent basement units. This was later confirmed by Chamberlain and Tosdal (2017), Kabete et al (2012) and Sanislav et al. (2014), who reviewed the existing geochronological data for the entire Tanzania Craton and concluded that there is no evidence of Dodoman age rock units in the northern half of the Tanzania Craton. Kabete et al. (2012), based on geophysical interpretation and limited field observations, divided the Tanzania Craton into a series of NW trending, shear-zone bounded accretionary terranes; they subdivided northern Tanzania (Fig. 1) into the Lake Nyanza Superterrane, Mwanza-Lake Eyasi Superterrane and the East Lake Victoria Superterrane.

The geology of the northern half of the Tanzania Craton is dominated by granite, gneiss and greenstone belts. Borg and Shackleton (1997) identified six greenstone belts: the Musoma-Mara, Kilimafedha, Iramba-Sekenke, Shinyanga-Malita, Nzega and Sukumaland greenstone belts. Although these greenstone belts share some common geological features, differences in age and geochemical signature between the individual greenstone belts indicate that their stratigraphy and geological evolution must be treated separately (e.g. Manya et al., 2007; Manya and Maboko, 2008).

The Sukumaland Greenstone Belt comprises a series of individual greenstone fragments separated by shear zones and granitoid intrusions. These fragments appear to share common stratigraphic features (e.g. Borg et al., 1990; Borg, 1994) similar to the Nyanzian and Kavirondian Supergroups (Manya and Maboko, 2003), but each fragment is large enough to be defined as a greenstone belt in its own right (Cook et al., 2015). The Nyanzian

Supergroup in the Sukumaland Greenstone Belt was subdivided into Lower and Upper Nyanzian (Manya and Maboko, 2003). The Lower Nyanzian is dominated by tholeiitic mafic volcanics with minor felsic volcanics and shale. Sm-Nd whole rock model ages (e.g. Manya and Maboko, 2003; Manya and Maboko, 2008; Cook et al., 2015) indicate that the mafic volcanics of the Lower Nyanzian in the Sukumaland Greenstone Belt were erupted at ca. 2820 Ma. Based on their similar eruption ages, geochemistry and Nd isotopic signature the mafic volcanics of the lower Nyanzian have been grouped into the Kiziba Formation (Cook et al., 2015) that was interpreted to have formed in an oceanic plateau-like environment. The Kiziba Formation has been metamorphosed to lower amphibolite facies (Cook et al., 2015). The Lower Nyanzian is overlain by the Upper Nyanzian, which is composed of shale, volcanoclastics, ironstone, chert, sandstone, siltstone and mudstone (Borg et al., 1990; Borg, 1994). In the Geita Greenstone Belt (GGB, Sanislav et al., 2014), which forms the northern part of the Sukumaland Greenstone Belt (Fig. 1), the contact between the mafic volcanics of the Kiziba Formation and the Upper Nyanzian is marked by a major shear zone (Cook et al., 2015). The entire sequence is complexly deformed with eight deformation events identified in the Upper Nyanzian sequence that hosts the Nyankanga, Geita Hill and Matandani-Kukuluma gold deposits in the GGB (Sanislav et al., 2015, 2017; Figs 1, 2). The Kavirondian Supergroup (Manya and Maboko, 2003) occurs as isolated outcrops that unconformably overly the Upper Nyanzian and is composed of conglomerate, grit, quartzite and sandstone (e.g. Borg et al., 1990; Borg, 1992, 1994).

The northern part of the GGB is intruded by 2620-2660 Ma (Sanislav et al., 2014) high-K granite batholiths while the southern part of the GGB is bordered by gneiss from which it is separated by a ductile shear zone. The eruption of the mafic volcanics forming the Kiziba Formation in the GGB was dated at ~ 2820 Ma (Manya and Maboko, 2008; Cook et al., 2015), whilst the maximum depositional age for the Upper Nyanzian sediments has been variably given as 2771 (Chamberlain and Tosdal., 2007) and 2702 Ma (Sanislav et al., 2014). Borg and Krogh (1999) dated a diorite sill (Sanislav et al., 2015) that intruded the ironstones in the Geita Hill deposit at 2699±9 Ma confirming that the sedimentation of the Upper Nyanzian in the Geita Greenstone Belt probably ceased by ~ 2700 Ma. Detailed mapping (Sanislav et al., 2015, 2016) around the Nyankanga and Geita Hill gold deposits indicate that the Upper Nyanzian sediments experienced an early extensional shearing event (D_1) followed by four compressional folding events (D_2-D_5) and three transpressional to transtensional brittle-ductile shearing events along discrete shear zones (D_6-D_8). Zircon ages from intrusive porphyries within the greenstone belt and the surrounding granite constrain all tectonic activity between 2820 and 2620 Ma (Manya and Maboko, 2008; Sanislav et al., 2014).

Petrographic description of the Kukuluma Intrusive Complex

The Kukuluma Intrusive Complex (KIC) consists of a series of NW trending intermediate igneous rocks (Fig. 2) that intruded the folded sequence of the Upper Nyanzian sediments, during the D_2-D_3 compressional stages affecting the GGB, sometime between 2680-2700 Ma (Kwelwa, 2017). Three major gold deposits, Matandani, Kukuluma and Area 3 W, occur along the contact between the KIC and the sediments (Fig. 2). The KIC is dominated by equigranular, fine- to medium-grained and locally porphyritic, intermediate intrusives (Fig. 3) and subordinate felsic (Fig. 3) porphyritic dykes.

The intermediate intrusive bodies are weakly to moderately foliated, indicating syn-D_3 emplacement (using the deformation scheme of Sanislav et al., 2015). The mineralogy is dominated by plagioclase (30-45 %), amphibole (30-40 %), alkali-feldspar (5-25 %), biotite (5-15%) pyroxene (5-10%) and quartz (5-20%). Based on the mineralogical composition the intermediate intrusives of the KIC can be separated into a diorite suite (gabbro-diorite and diorite; Figs. 3a and b) and a monzonite suite (monzodiorite and monzonite; Figs. 3c and d).

The feldspars are only partly sericitized (Figs. 3b and d) while some of the mafic minerals are partly replaced by metamorphic actinolite. Accessory minerals include apatite, magnetite, and rutile. Minor chlorite and carbonate are present as disseminated minerals, partly replacing the mafic minerals or along veins.

The felsic porphyritic dykes (Figs. 3e and f) show a narrower variation in their mineralogical composition. Their mineralogy is dominated by quartz (15-40 %), plagioclase (50-70%), K-feldspar (5-40%), biotite (5-15%) and amphibole (1-10%). The main accessory minerals are apatite and zircon. Based on their mineralogical composition the felsic phase of the KIC varies between granodiorite and tonalite. The plagioclase is partly sericitized and the mafic minerals are partly replaced by chlorite.

Methodology for major and trace element analyses

Whole rock geochemical analyses were performed at the Advanced Analytical Centre at James Cook University (JCU) on samples collected from drill core. All samples were collected away from the mineralized zones to minimize the effect of alteration. All samples were studied under the petrographic microscope and only samples that showed the minimum alteration were selected for further analyses. Approximately 1 kg of material was milled from each sample to a fine powder in a tungsten carbide ring mill. Major elements were analysed by conventional X-ray fluorescence (XRF) using a Bruker-AXS S4 Pioneer XRF Spectrometer on fused beads. The fused beads were prepared from rock powders mixed with 12:22 borate flux (XRF Scientific Limited, Perth, Australia) at 1:8 sample to flux ratio that were fused to glass after heating to 1050 °C in a F-M 4 Fusion Bead Casting Machine (Willunga, Australia). Chips of the fused beads were mounted into a standard epoxy puck and analysed for a range of trace elements using a Geolas Pro 193 nm ArF Excimer laser ablation unit (Coherent) coupled to a Varian 820 quadrupole ICP-MS. Helium was used as the carrier gas (0.8 l/min), which was subsequently mixed with Ar via a mixing bulb between the ablation cell and the ICP-MS to smooth the ablation signal. Laser energy density was set to 6 J/cm^2, and the laser spot size and repetition rate were set to 120 µm and 10 Hz, respectively. Each fused bead was analysed 3 times and average values are reported. The ICP-MS was tuned to ensure robust plasma conditions and low oxide production levels (≤ 0.5% ThO) with the plasma power set at 1.25 kW. NIST SRM 610 glass was used as a bracketed external standard using the standard reference values of Spandler et al. (2011). Data were quantified using Ca (as previously determined by XRF on the same fused bead) as the internal standard, and data were processed using the Glitter software (Van Achterbergh et al., 2001). To monitor precision and accuracy of the analyses, we analysed Hawaiian basalt reference glass (KL2-glass; n=21) as a secondary standard (Jochum et al., 2006). The precision for REE analyses by LA-ICP-MS is better than 5% (mostly <3%), and the accuracy is often <2%. The standard reference material NIST612 (n=11) was analysed as a ternary standard. The precision for all the elements, besides Zn (3.5%) and Ge (~8.3%,) is <2%, and <1% for REE. The accuracy for all the elements (standard reference concentrations taken from Spandler et al., 2011) is <3%. The only exceptions are Tb (6.5%), Ge (~8.3%), Sb (~9%), and Zn (~5.3%) where relatively large uncertainties in the NIST612 glass have to be taken into consideration.

Alteration and element mobility

The KIC rocks are deformed, metamorphosed and locally overprinted by hydrothermal alteration related to gold mineralisation concentrated along its margins. The top 100 meters of the intrusive complex is highly weathered so that all samples were collected from diamond drill holes that intercepted the intrusive complex at more than 400 meters below the surface, and away from mineralised zones. Petrographic examination of the samples revealed minor carbonate and chlorite alteration indicating that the samples have been hydrated and carbonated. The loss of ignition (LOI) values of up to 5.5% confirms the petrographic observations and requires that all samples be screened for element mobility. On the alteration boxplot of Large et al. (2001), which combines the alteration index of Ishikawa ($100(K_2O+MgO)/(K_2O+MgO+Na_2O+CaO)$) and the chlorite-carbonate-pyrite alteration index (($100(MgO+FeO)/(MgO+FeO+Na_2O+K_2O)$), all samples plot (Fig. 4a) into the field of least altered rocks. However, to further test the element mobility for the KIC samples we used only the monzonite and diorite suites, because the higher number of samples allows compositional variations induced by post-magmatic alteration to be identified more easily. Na, K, Rb and Sr are all easily mobilised during low-temperature hydrothermal alteration and metamorphism. The post-magmatic disturbance of Na, K, Rb and Sr by hydrothermal alteration and metamorphism can be tested by plotting their concentration against the LOI values. A lack of correlation indicates little or no significant disturbance while well-correlated trends indicate significant disturbance and mobility (e.g. Papoulis et al., 2004; Harvey et al., 2014). The lack of any correlation between these elements and the LOI values (Fig. 4b, c, d and e) combined with their coherent behaviour on other geochemical plots suggests that the post-magmatic alteration did not significantly mobilise these elements, and they can, therefore, be used for petrogenetic interpretations. In general REE and HFSE are considered immobile during hydrothermal alteration and greenschist facies metamorphism, but situations where the REE were mobile have been documented (e.g. Wood et al., 1976; Condie et al., 1977); with the LREE considered to be more mobile than the HREE (Sun and Nesbit, 1978). To test the mobility of the REE from the KIC rocks, we plotted the concentration of La against Zr (Fig. 4f), and to test the mobility of the HFSE we plotted the concentration of Nb against Zr (Fig. 4g). The strong positive correlation that exists between these elements in combination with the coherent behaviour of the REE and HFSE on chondrite and primitive mantle normalised plots indicate that these elements most probably retained their original concentrations. The ratio of highly incompatible elements such as Th and U should be near chondritic ($Th/U_{chondrite}$= 3.63; Sun and McDonough, 1989) unless disturbed by alteration processes when U is mobile under oxidizing conditions. The average Th/U ratio of all KIC rocks ($Th/U_{monzonite}$=3.65; $Th/U_{diorite}$= 2.92; $Th/U_{granodiorite}$= 2.92) is near chondritic suggesting little to no mobility of these elements during hydrothermal alteration and metamorphism.

The geochemistry of the KIC

The geochemical composition of the KIC (Table 1 and Fig. 5) is characterised by: intermediate SiO_2 (59.17 wt%), moderate #mg (0.47), high Al_2O_3 and FeO (15.83 wt% and 5.66 wt% respectively) and moderate MgO (2.78 wt%). The K_2O/Na_2O ratio is less than 1 (0.70) and $CaO+Na_2O$ is more than 8 (8.82%). The Y content is low (14.5 ppm) and Sr is high (765 ppm) with an average Sr/Y ratio of 59. The HREE are depleted relative to the LREE with an average La/Yb ratio of 42 and the Cr content is moderately high (62 ppm). The chondrite normalized REE pattern (Fig. 6) show fractionated patterns while the primitive mantle normalized multi-element patterns show negative Nb and Ti anomalies and a general enrichment in the large ion lithophile elements (LILE).

The monzonite suite

The geochemical composition of rocks that belong to the monzonite suite (Fig. 5; Table 1) from the KIC is characterized by intermediate SiO_2 (51.7- 62.1 wt%), FeO (3-9 wt%), MgO (2.5-5.6wt%) and CaO (3.8-7 wt%), moderate K_2O (0.9-3.6 wt%), high Al_2O_3 (14.4-16.5 wt%) and Na_2O (4.5-6.6 wt%) and low TiO_2 (0.4-0.6 wt%). They have high $CaO+Na_2O$ (8.6-12.9), high Sr (537-1563 ppm) and high LREE (La_n = 241-777 ppm; the subscript “n” refers to chondrite normalized). These features combined with a low K_2O/Na_2O

ratio (0.1-0.8), low Y (11-30 ppm), low HREE (Yb_n = 4-11) and high Sr/Y and La/Yb ratios (30-119 and 67-102 respectively) indicate that the monzonite suite has geochemical characteristics similar to adakites, sanukitoids and Closepet-type granite. Martin et al (2005) suggested that less differentiated sanukitoids (SiO_2<62 wt%) are similar to LSA and Closepet-type granite. However, rocks that belong to the monzonite suite from the KIC have, on average, higher Y, Yb and La/Yb than the LSA, higher #mg, Sr, Cr, Sr/Y and La/Yb than the Closepet-type granite, and higher La/Yb than the average sanukitoid. At the same time the monzonite suite has lower TiO_2, #mg, Sr, Cr, Ni and Sr/Y than the LSA, lower TiO_2, Y, Yb than the Closepet-type granite and lower Cr compared to the average sanukitoid. The chondrite normalized REE pattern (Fig. 6a) of monzonite suite rocks is subparallel to the pattern from average LSA, sanukitoid and Closepet-type granite and shows the strong LREE enrichment characteristic for these type of rocks. When plotted on a primitive mantle normalized multielement diagram (Fig. 6b) the monzonite suite shows strong negative anomalies for Nb and Ti, and moderate negative anomalies for Zr and Sr. Their pattern is subparallel to that of the LSA, sanukitoids and Closepet-type granite. Notable differences are the positive Sr anomaly for the LSA, the lack of a Sr anomaly in sanukitoids and the lack of a negative Zr anomaly in LSA, sanukitoids and Closepet-type granite.

The diorite suite

Rocks that belong to the diorite suite (Fig. 5; Table 1) have similar SiO_2 (53-63 wt%) contents compared to monzonite suite rocks, but slightly higher Al_2O3 (14.4-17.1 wt%), FeO (2.8-7.8 wt%), and TiO_2 (0.3-0.7 wt%), and slightly lower Na_2O (3.1-5.6 wt%), K_2O (1.2-3.2 wt%) and P_2O_5 (0.1-0.3 wt%). These values combined with K_2O/Na_2O ~ 0.54, $CaO+Na_2O$ ~ 8, low Y (8-12 ppm), low HREE (Yb_n = 3-7 ppm) and high Sr (572-1062 ppm), Cr (49-99 ppm) and LREE (La_n= 45-161 ppm) suggest that the diorite suite also shares geochemical features with sanukitoid, adakite and Closepet-type granite. The average composition of rocks from the diorite suite is similar to the average composition of HSA except for lower SiO_2 (58.6 vs 64.8), higher FeO (6.39 vs. 4.27), higher Cr (75 vs 41) and higher Sr/Y (85 vs 56). The chondrite normalized REE pattern (Fig. 6c) is similar to the average HSA and subparallel to, but at lower concentrations than the average LSA, sanukitoid and Closepet-type granite. On a primitive mantle normalized multi-element diagram (Fig. 6d) diorite suite rocks show pronounced negative Nb anomalies, moderately negative Ti anomalies and moderately positive Sr anomalies, all of which are also typical for LSA. The overall pattern is similar to that of HSA, except for the positive Sr anomaly, and is subparallel to the average pattern of LSA, sanukitoid and Closepet-type granite, but at lower concentrations.

The granodiorite suite

Four samples from the KIC were classified as granodiorites. Although the samples were collected a few hundred meters apart their major and trace element composition is almost identical (Fig. 5; Table 1). They have moderate SiO_2 (av. 62.5 wt%), low FeO (av. 4.9 wt%), MgO (av. 1.9 wt%), CaO (av. 3.7 wt%) and high K_2O (av. 3.67 wt%) when compared to rocks from the monzonite and the diorite suite. Their K_2O/Na_2O ratio is high (av. 1.1), $CaO+Na_2O$ is low (av. 7.4), Y is low (16 ppm), HREE are low (Yb_n = 10), LREE are moderately high (La_n= 100 ppm), and Sr and Cr content are relatively high (332 and 21 ppm respectively). The lower Sr content (< 400 ppm), a Sr/Y ratio of less than 40 and a La/Yb ratio of less than 20 suggest that the granodiorites cannot be considered to have an adakite-like signature sensu Defant and Drummont (1990). However, as pointed out by Moyen (2009), HSA can have a Sr/Y ratio as low as 20. On a chondrite normalized REE diagram (Fig. 6e) the granodiorites display a subparallel trend to that of the HSA but they plot at higher concentrations. They also show a weak negative Eu anomaly indicative of plagioclase fractionation. On a primitive mantle normalized multi-element diagram (Fig. 6f) their pattern is similar to that of the HSA, except that they have a weak negative Sr anomaly while the HSA have a weak positive Sr anomaly.

Petrogenesis of the KIC

Relative timing of emplacement

Rocks of the granodiorite suite have been dated at 2651 Ma to 2667 Ma (Kwelwa, 2017), but there are no direct age data available for rocks forming the monzonite and diorite suites. However, field relationships help constrain their relative timing of emplacement. The intrusives of the monzonite and diorite suites occur as a series of intrusive bodies subparallel to the NW-SE trending regional fabric (Fig. 2). The rocks contain a weakly to well-developed foliation that is subparallel to the axial planar surface of regional D_3 folds (Sanislav et al., 2015, 2017) indicating coeval and syntectonic emplacement. The granodiorite suite rocks are not foliated indicating that their emplacement postdates the emplacement of the diorite and the monzonite suite. Felsic dykes similar in composition to the granodiorite suite outcrop in the Kukuluma and Matandani deposits where they crosscut the folded sequence and are crosscut by brittle ductile shear zones.

In the Nyankanga and Geita Hill areas, monzonite and diorite dykes and sills intrude during D_2 and D_3, i.e. at a time relative to deformation that is near identical to the relative timing observed in Kukuluma (Sanislav et al., 2015, 2017). Borg and Krogh (1999) provide an age of 2699±9 Ma for a diorite dyke from Geita Hill, and Chamberlain and Tosdal (2007) report an age of 2698±14 Ma for diorite in the Nyankanga pit where it has been cross cut by several generations of felsic dykes dated at 2685-2696 Ma (Chamberlain and Tosdal, 2007). Therefore, by comparison, the monzonite and diorite suites of the KIC are interpreted to have been emplaced between 2685-2700 Ma.

Depth and source of melts

Fractionated REE patterns (Fig. 6), high Sr/Y and La/Yb ratios (Figs. 7a and b), and low Y and Yb contents suggest that garnet was present as a fractionating or residual phase in the melt (e.g. Martin et al., 2005; Moyen, 2009; Castillo, 2012). However, high Sr/Y and La/Yb ratios can also reflect an enriched source (Moyen, 2009), can be produced by the fractionation of amphibole, and by the delayed crystallization of plagioclase in hydrous mafic magmas (Castillo, 2012), while fractionated REE patterns may result from amphibole fractionation (e.g. Romick et al., 1992; Richards and Kerrich 2007). Continental crust has high Sr/Y and La/Yb ratios, therefore, melting of continental crust and/or mixing with continental crust may impart high Sr/Y and La/Yb to their derivative melts. The low SiO_2 and moderate Mg# of the KIC rocks suggest a mafic to ultramafic source, and preclude any significant contribution from felsic rocks. In mafic melts fractionation of amphibole may increase the La/Yb ratio of the residual melt, but the REE pattern will not develop a strongly concave shape. As magma becomes more dacitic the hornblende REE distribution coefficients increase and magmas develop concave REE patterns and high La/Yb ratios (Romick et al., 1992). So the net effect of amphibole and plagioclase fractionation is an increase in La/Yb and decrease in Dy/Yb (Moyen, 2009), whereas garnet fractionation or partial melting with residual garnet will increase the Dy/Yb ratio in the melt (e.g. Macpherson et al., 2006; Davidson et al., 2007). Kelemen et al. (2003) proposed that melts with a clear garnet (eclogite) signature should have Dy/Yb_n ratios $\geq$ 1.5. All samples of the monzonite suite and the majority of the diorite suite samples have $Dy/Yb_n > 1.5$ (Fig. 7c) suggesting that their high Sr/Y and La/Yb ratios are related to deep melting. Eclogite melts reacting with the mantle (Kelemen et al., 2003) would decrease both the Dy/Yb and the La/Yb ratios (Fig. 7d) of the initial melt. Therefore, a lack of eclogite melting signature in some of the samples (i.e. the granodiorite suite) does not automatically rule out their derivation from eclogite/garnet-bearing melts. Moreover, plagioclase crystallization can decrease Sr/Y ratios and increase Y concentrations. Thus, a deep melting signature (based on this ratio) can be erased by large degrees of plagioclase fractionation (e.g. Richard and Kerrich, 2007).

The major element variation diagrams show that for the same SiO_2 content (Fig. 5), the diorite and the monzonite suites display subparallel trends for most of the elements. It is particularly obvious for Al_2O_3, FeO_t, Na_2O, TiO_2 and P_2O_5. Assuming that the two suites were derived from rocks having a similar composition this subparallel evolution of the major elements cannot be explained by magma mixing or by fractional crystallization alone and requires melting at different pressures. For example, the Al_2O_3 content of melts becomes depleted with increasing pressure at the same degree of partial melting (e.g. Hirose and Kushiro, 1993; Spandler et al., 2008). The negative correlation between SiO_2 and Al_2O_3 (Fig. 5a) in the diorite suite may indicate garnet fractionation or residual garnet, which will effectively reduce Al_2O_3 with increasing SiO_2 in the melt (Macpherson et al., 2006; Davidson et al., 2007). The positive correlation observed in the monzonite suite may indicate that garnet was partly consumed during melting. The monzonite suite rocks tend to have higher Na_2O at the same CaO (Fig. 8a) compared to the diorite suite rocks. This can also indicate a higher pressure during melting as Na_2O becomes more compatible in clinopyroxene at higher pressure (e.g. Kogiso et al., 2004).

On chondrite normalized diagrams (Fig. 6) the REE patterns for the two suites are subparallel, but the LREE elements are more fractionated for the monzonite suite than the diorite suite. This is also illustrated by much higher La/Yb_n and Dy/Yb_n ratios (Fig. 7) suggesting that rocks belonging to the monzonite suite may represent deeper melts compared to rocks from the diorite suite. Their primitive mantle normalized trace element patterns (Fig. 6) are also sub-parallel, with the notable difference that the diorite suite rocks have a positive

Sr anomaly while the monzonite suite rocks have negative Sr and Zr anomalies. The presence of a significant positive Sr anomaly in the diorite suite cannot be explained by melting or crystallization unless plagioclase is involved. The lack of any correlation between the Sr/Sr* and the MgO (Fig. 8b) excludes fractionation. Thus, a plagioclase-rich component is required in the melt source region. Alternatively, interaction of the melt with a plagioclase-rich region (assimilation) will produce a similar effect. However, assimilation will result in a large decrease in FeO_t and a large increase in Al_2O_3 with decreasing MgO (e.g. Peterson et al., 2014), which is not the case here. Therefore assimilation can be excluded.

The only viable explanation is that the positive Sr anomaly is related to the source rock. We propose that the diorite suite was formed by melting of garnet-bearing amphibolite and plagioclase was completely transferred into the melt, leaving behind a Sr-depleted (relative to Ce and Nd) residue of garnet-clinopyroxene-rutile eclogite. Further melting of the eclogite with residual rutile produced the monzonite suite with negative Sr and Zr anomalies. Zr and Hf have similar chemical properties and should not fractionate from each other in geological processes; i.e. their ratio should be chondritic in all earth materials (e.g. Zr/Hf=~36.3; Sun and McDounough, 1989). The diorite suite has an average Zr/Hf ratio of 36.8 (Fig. 8c), which is similar to the chondritic value, but the monzonite suite has an average Zr/Hf ratio of 42.7 (Fig. 8c), which exceeds the chondritic value suggesting that these elements were fractionated from each other. Experimental data on amphibole/melt partition coefficients (e.g. Foley et al., 2002; Tiepolo et al., 2007) have shown that amphibole can fractionate most HFSE causing negative Ti and Nb anomalies, but only high-Mg amphibole can fractionate Zr from Hf. The ability of garnet to fractionate Zr from Hf is dependent on pressure and MgO content (e.g. Green et al., 2000; van Westrenen et al., 1999). The only mineral able to effectively fractionate HFSE from each other is rutile (Stalder et al., 1998; Foley et al., 2000). If rutile was the residual phase, the Nb/La and Zr/Sm ratios of the melt will correlate positively (Münker et al., 2004), but if high-Mg amphibole was the residual phase the melt ratios of these elements will correlate negatively. The monzonite suite shows a clear positive correlation between Nb/La and Zr/Sm (Fig. 8d) implying residual rutile. However, the diorite suite shows no correlation between these two ratios. Rutile cannot coexist with basaltic melts arising from the partial melting of peridotite (e.g. Ryerson and Watson, 1987; Woodhead et al., 1993; Thirlwall et al., 1994), because it reacts with the olivine to form orthopyroxene and ilmenite. Thus, the most likely source for the monzonite suite is rutile-bearing, garnet-clinopyroxene eclogite.

Melt mantle interaction

The low SiO_2 content, average Mg numbers, and relatively high Ni and Cr concentrations indicate that the source rocks for the KIC must be mafic or ultramafic. Their intermediate composition ($SiO_2 \leq 62$ wt%) suggests that fractionation played a minor role in their petrogenesis and they are close to primary magmas. From this point of view the rocks of the KIC resemble LSA and sanukitoid. However, there are some important differences. Firstly, at the same SiO_2 content the rocks of the KIC have much lower MgO compared to LSA (Fig. 5b). Secondly, the rocks of the KIC overlap the field of mafic experimental melts (Figs 3a,b and f), whereas LSA rocks plot above it, and the sanukitoids overlap with it, but only for low MgO concentrations. Because of their low SiO_2, high Mg numbers and high Cr and Ni concentrations, the LSA and the sanukitoids are commonly interpreted to have formed by partial melting of mantle peridotite metasomatised by felsic melts (e.g. Shirey and Hanson, 1984; Rapp et al., 1999; Martin et al., 2005). The rocks of the KIC have lower Mg numbers (at the same SiO_2 content; Fig. 5b) and much lower Cr, Ni, Sr, K, Rb and Nb concentrations compared to LSA rocks and the sanukitoids (Fig. 9 and Table 1). The difference between KIC rocks and LSA rocks is clearly illustrated in Figure 9 where the composition of KIC rocks overlaps the composition of the HSA, and closely resembles the composition of experimental melts of basalt (Figs. 7a and b).

On the K vs Rb diagram (Fig. 9a) some of the LSA rocks plot subparallel to the Y-axis suggesting high K/Rb ratios, which were interpreted to reflect Rb depletion by selective melting of metasomatic amphibole in a peridotite source (e.g. Martin et a., 2005 and references therein). However, in the absence of metasomatism, both peridotite and basaltic melts result in K/Rb ratios much lower than average oceanic basalt (K/Rb=1071; Sun and McDonough, 1989). High Sr contents can reflect deep melting at pressures above the plagioclase stability field, melting of a source that was already high in Sr, and/or melt interaction with high-Sr geological materials (e.g. Moyen, 2009). Given the low SiO_2 ($\leq$ 62 wt%) of the KIC samples, their source rock must have been mafic or ultramafic. Experimental melting of basalt produced liquids with up to 1000 ppm Sr (Fig. 9b), but to achieve the high Sr observed in LSA, interaction with mantle peridotite is required (e.g. Martin et al., 2005). Rocks from the diorite and granodiorite suites plot within the fields of basaltic melts and HSA (Fig. 9b), while some of the samples from the monzonite suite plot at higher Sr values (~1500 ppm) within the field of LSA, which may indicate some level of interaction with mantle peridotite. However, if the source of the KIC rocks was mafic lower crust (Sr = 348 ppm; Rudnick and Gao, 2003; Sr = 289 ppm; Hacker et al., 2015) rather than an average oceanic basalt (Sr = 90 ppm; Sun and Mcdonough, 1989), than high Sr values observed in the monzonite suite do not necessarily require interaction with peridotite mantle. Maybe the most useful ratio to use when separating melts derived from partial melting of metasomatised mantle peridotite (LSA) and melts derived by partial melting of mafic rocks (HSA) is the Cr/Ni ratio (Fig. 9c; Martin et al., 2005). The Cr/Ni ratio for KIC samples (Fig. 9c) is clearly distinct from that of the LSA and overlaps the field of the HSA suggesting a mafic source and minimum interaction with the mantle. The lack of correlation between the

Cr/Ni ratio and the Mg# (Fig. 9d) suggests that the Cr/Ni ratio of the KIC samples is a source characteristic and not dependent on fractionation.

Tectonic setting

The rocks of the KIC have major and trace element signatures similar to "adakitic' rocks. Although the original description of adakites (e.g. Defant and Drumkond, 1990) specifically indicates that their geochemical signature is derived from partial melting of a subducted slab, it is clear now that high Sr/Y and La/Yb ratios alone cannot be used to unequivocally indicate a subduction setting (e.g. Moyen, 2009; Castillo, 2012), and rocks with an "adakitic" signature can form in different tectonic settings as well. The KIC was emplaced syn-tectonically along axial planar surfaces of upright regional folds suggesting a period of crustal thickening between 2685-2700 Ma. The age data from the northern half of the Tanzania Craton (e.g. Kabete et al., 2012; Sanislav et al., 2014) suggest that growth of this part of the craton started at ~ 2820 Ma with extensive tholeiitic mafic volcanism (e.g. Manya and Maboko, 2003, 2008; Cook et al., 2015) followed by a period dominated by the intrusion of diorite and TTG and completed with the intrusion of the 2620-2660 Ma high-K granites (Sanislav et al., 2014). Thus, the KIC was emplaced during the transition period from higher depth TTG magmas to shallower depth high-K magmas.

Cook et al. (2015) proposed that the ~ 2820 Ma mafic volcanics (Kiziba Formation) that form the base of the stratigraphic sequence (Lower Nyanzian) in the Sukumaland Greenstone Belt were emplaced in an oceanic plateau like setting. However, it is unclear at the moment whether or not the Upper Nyanzian sediments (intruded by the KIC) were deposited on top of the Kiziba Formation or the two units were tectonically juxtaposed. In the Geita region the contact between the two units is structural (Cook et al., 2015), but there appears to be evidence that the Geita Greenstone Belt is underlain by the mafic rocks of the Kiziba Formation suggesting that crustal growth in this part of Tanzania occurred by partial melting and maturation of an oceanic plateau by lower crustal delamination/modification (e.g. Vlaar et al., 1994; Zegers and van Keken, 2001; Bedard, 2006; Bedard et al., 2013; Cook et al., 2015). Chiaradia (2015) showed that there is a strong correlation between the Sr/Y ratio and MgO content of recent arc magmatism, and the upper plate thickness indicating that source processes (slab melting, slab melt-mantle interactions) do not play a major role in the generation of high Sr/Y signatures. This implies that high Sr/Y ratios occur at lower MgO content, suggesting that thicker crust favours magma evolution at deeper levels, thus Sr/Y increases steadily with magmatic differentiation (Fig. 10a). In contrast, Archean rocks show a sudden increase of the Sr/Y ratio between ~2.5 and ~0.5 wt% MgO (Fig. 10a) suggesting that, as opposed to modern arc lava, source processes control the Sr/Y ratio of Archean rocks (Chiaradia, 2015). Source processes may include partial melting in the garnet stability field of subducted mafic crust or partial melting of delaminated lower mafic crust.

To investigate a possible subduction component in the generation of the KIC rocks we use Th/U vs Zr/Hf ratios. Partial melting in the mantle wedge can be excluded, as detailed above, based on the presence of Nb and Ti depletion due to residual rutile, which cannot coexist with basaltic melts arising from the partial melting of peridotite (e.g. Ryerson and Watson, 1987; Woodhead et al., 1993; Thirlwall et al., 1994) because it reacts with the olivine to form orthopyroxene and ilmenite. Given the highly incompatible behaviour of Th and U, normal magmatic processes cannot significantly fractionate these elements from each other. U and Th are easily fractionated during surface processes, because of the higher mobility of U during weathering and under oxidizing conditions. Seafloor alteration and addition of slab fluids will lower the Th/U ratio while dehydration and addition of sediment melts will increase the Th/U ratio (e.g. Bebout, 2007). Figure 10b shows that the diorite and the granodiorite suites have almost MORB-like Th/U and Zr/Hf ratios, thus precluding a subduction component. The monzonite suite has higher Zr/Hf ratios due to residual rutile (see above), and the Th/U ratio varies between the values for the lower mafic crust end-member of Hacker et al. (2015) and the values for the average lower continental crust of Rudnick and Gao (2003). Overall, the KIC shows a very narrow variation in Th/U ratios, which is more consistent with partial melting of the mafic lower crust than partial melting of a subducted oceanic crust.

Conclusions

In general Archean igneous rocks with adakite-like signature are interpreted to indicate a subduction setting. We have shown that although the rocks of the KIC can be easily classified as “adakites”, detail screening of their composition revealed important differences. Given the recognition that rocks with adakite-like signature can form in a variety of tectonic settings from non-unique petrogenetic processes requires a re-examination of the existing Archean datasets. Higher Archean geothermal gradients would have favoured the development of thicker lithospheric roots and partial to complete eclogitization of the mafic lower crust; the removal of the eclogitised crust by delamination would favour partial melting of the thickened lower crust to generate adakite-like rocks. This scenario is similar to the interpretation of the post-tectonic adakite-like rocks from the Tibetan Plateau with the main difference that the KIC is syn-tectonic. Alternatively, the KIC formed by partial melting of eclogitised mafic lower crust of an Archean oceanic plateau.

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Figure captions

Figure 1

Simplified geological map of northern Tanzania (a) and the geological map of Geita Greenstone Belt (b). SGB – Sukumalanad 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. Inset map of Africa showing the location of Archean blocks.

Figure 2

Geological map of the eastern part of the Geita Greenstone Belt showing the location of the Kukuluma Intrusive Complex.

Figure 3

Photomicrographs showing the three main rock types found in the KIC. a and b) Medium grained diorite; the mineralogy is dominated by amphibole (mostly actinolite) and plagioclase with minor quartz. c and d) Medium grained monzonite; the mineralogy is dominated plagioclase, k-feldspar, biotite and amphibole with minor quartz. The diorite and monzonite have been deformed and metamorphosed to greenschist facies. As a result amphiboles and pyroxenes have been partly replaced by actinolite. Note that the feldspars in both rock types are not altered to sericite and appear fresh under microscope suggesting that the samples have not been significantly affected by hydrothermal alteration. e and f) Photographs of a porphyritic granodiorite dyke. Note that the feldspars from the granodiorite have been partly replaced by sericite but appear mostly fresh under microscope. Small amounts of carbonate and chlorite, disseminated or as microveins are present in all samples.

Figure 4

Series of diagrams showing that although the rocks of the KIC have been hydrated and carbonated, as indicated by the petrography and LOI values, their major and trace element composition was very little disturbed. In the alteration boxplot of Large et al. (2001) all samples plot in the field of least altered rocks (a). Diagrams b and c show that there is no correlation between the concentration of two of the most mobile major elements, Na_2O and K_2O, and the LOI values. The same pattern is observed in d and e where the concentration of two of the most mobile trace elements, Rb and Sr, are plotted against the LOI values suggesting that most likely the concentration of these elements is close to their initial values. The mobility of REE and HFSE was tested by plotting the values of La (f) and Nb (g) against Zr, a highly immobile element. The good correlations suggest that these elements were most likely immobile during post-magmatic alteration and metamorphism.

Figure 5

Major elements variation diagrams for the KIC. The grey area shows the field of sanukitoids from Martin et al., (2010). The field of LSA (continuous line in Figure 5b) is from Castillo (2012) and the field of basaltic experimental melts is from Rollison (1997) and Martin et al., (2005).

Figure 6

Chondrite normalized REE diagrams (a, c and e) and primitive mantle normalized trace element diagrams (b, d and f) for the KIC rocks. Also shown is the average of LSA, HSA, sanukitoids and Closepet-type granite from Martin et al., (2005).

Figure 7

Sr/Y vs Y (a) and La/Yb_n vs Yb_n (b) diagrams for the KIC samples. The field of LSA (ligh grey) and HSA (darker grey) in (a) is from Castillo (2012) and the field of sanukitoids (dashed line) is from Martin et al., (2005). The field of adakites in (b) is from Moyen and Martin (2012). The diagrams in (c) and (d) show that the samples that have high Sr/Y (c) and La/Yb_n ratios (d) also have high Dy/Yb_n ratio indicative of high pressure melting. The large square (diorite suite) and the large circle (monzonite suite) show the samples with the highest Mg# which also have the Sr/Y, La/Yb_n and Dy/Yb_n ratios. The line with arrow in (d) shows that the interaction of eclogitic melts with the mantle peridotite leads to a decrease in the La/Yb and Dy/Yb ratios in the melt (Kelemen et al., 2003).

Figure 8

Diagram (a) showing that at similar CaO values the monzonite suite has higher Na_2O which may reflect clinopyroxene in the source and melting at higher pressure. The lack of correlation (b) between the MgO and the Sr/Sr* suggest that the Sr anomaly is not the result of plagioclase fractionation. (c) shows that Zr and Hf are fractionated from each other in the monzonite suite which we attribute to residual rutile. The effect of residual rutile is shown in (d) where the positive correlation between Nb/La and Zr/Sm ratios is indicative of residual rutile (Münker et al., 2004). The dashed lines in (c) and (d) shows the chondritic ratios for the respective elements while the arrows in (d) show the effect of residual rutile (positive correlation) vs the effect of residual high-Mg amphibole (negative correlation).

Figure 9

Diagrams showing the compositional differences between LSA and HSA on K vs Rb (a), Sr vs CaO+Na_2O (b) and Cr/Ni vs TiO_2 (c) compiled by Martin et al., (2005). In (a) the continuous line shows the average K/Rb ratio in MORB (Sun and Mcdonough, 1989) while the arrows show the effect of metasomatism, which increases the ratio and the effect of partial melting, which decreases the ratio. In all three diagrams the KIC samples resemble more the HSA than the LSA and mostly overlap the field of experimental basaltic melts. The diagram in (d) show that there is no correlation between the Cr/Ni ratio and the Mg# indicating that the Cr/Ni ratio in the KIC samples is a source characteristic rather than the result of fractionation.

Figure 10

Diagram showing the variation of the Sr/Y ratio with MgO in modern arcs, Archean adakites and experimental melts from Chiaradia (2015). Chiaradia (2015) showed that in the modern arcs the Sr/Y ratio is a function of crustal thickness and the gradual increase of the Sr/Y ratio with increased crustal thickness also correlates with decreasing MgO suggesting that the Sr/Y ratio in modern arcs better reflects intracrustal processes than source characteristics. The sudden increase in Sr/Y ratios at low MgO in Archean adakites is similar to the data obtained for experimental basaltic melts and is consistent with partial melting of the lower mafic crust in the Archean. The KIC samples have Sr/Y and MgO values similar to the experimental basaltic melts and the Archean adakites suggesting lower mafic crust melting. The diagram in (b) shows that the diorite and the granodiorite suite have near MORB/chondritic Zr/Hf and Th/U ratios while the monzonite suite has Zr/Hf ratios similar to the mafic end member (mlc) of the lower crust (Hacker et al, 2015) and the Th/U ratio varies between the mafic end member of the lower crust and the lower continental crust (lcc) values of Rudnick and Gao (2003). However, the Th/U ratios in all samples are near chondritic suggesting that Th and U were not fractionated from each other as required by a subduction environment.

Element	LSA	HSA	Sanukitoid	Closepet	Monzonite suite	Diorite suite	Granodiorite	KIC (average)
SiO_2 (%)	56.25	64.80	58.76	56.39	56.41	58.61	62.50	57.51
TiO_2 (%)	1.49	0.56	0.74	1.20	0.49	0.51	0.55	0.50
Al_2O_3 (%)	15.69	16.64	15.80	15.79	15.70	16.09	15.69	15.90
FeO (%)	5.82	4.27	5.28	6.60	5.71	6.39	4.89	6.05
MnO (%)	0.09	0.08	0.09	0.13	0.07	0.08	0.08	0.07
MgO (%)	5.15	2.18	3.90	3.38	3.54	2.90	1.89	3.22
CaO (%)	7.69	4.63	5.57	5.45	4.99	4.58	3.71	4.79
Na_2O (%)	4.11	4.19	4.42	3.94	5.61	4.20	3.37	4.90
K_2O (%)	2.37	1.97	2.78	3.17	2.59	2.16	3.67	2.38
P_2O_5 (%)	0.66	0.20	0.39	0.72	0.50	0.22	0.16	0.36
K_2O/Na_2O	0.58	0.47	0.63	0.80	0.49	0.51	1.10	0.50
$CaO+Na_2O$	11.80	8.82	9.99	9.39	10.59	8.78	7.08	9.69
$FeO+MgO+MnO+TiO_2$	12.55	7.09	10.01	11.31	9.81	9.88	7.42	9.85
Y (ppm)	13.00	10.00	18.00	37.00	17.24	9.794	16.41	13.52
Yb (ppm)	0.93	0.88	1.32	2.05	1.18	0.85	1.57	1.02
Sr (ppm)	2051.00	565.00	1170.00	978.00	1128.38	834.34	331.96	981.36
Cr (ppm)	157.00	41.00	128.00	50.00	86.91	75.10	21.24	81.00
Ni (ppm)	103.00	20.00	72.00	38.00	46.65	53.00	13.43	49.83
Sr/Y	162.21	55.65	63.98	26.43	71.39	85.19	20.25	78.29
La/Yb	44.19	21.82	45.38	44.34	82.76	26.68	15.20	54.72
Mg#	0.61	0.48	0.57	0.48	0.53	0.46	0.41	0.49

Table 1. Table showing the average composition of the adakites, sanukitoids and Closepet-type granite (Martin et al., 2005) and the average composition of the Kukuluma Intrusive Complex rocks.

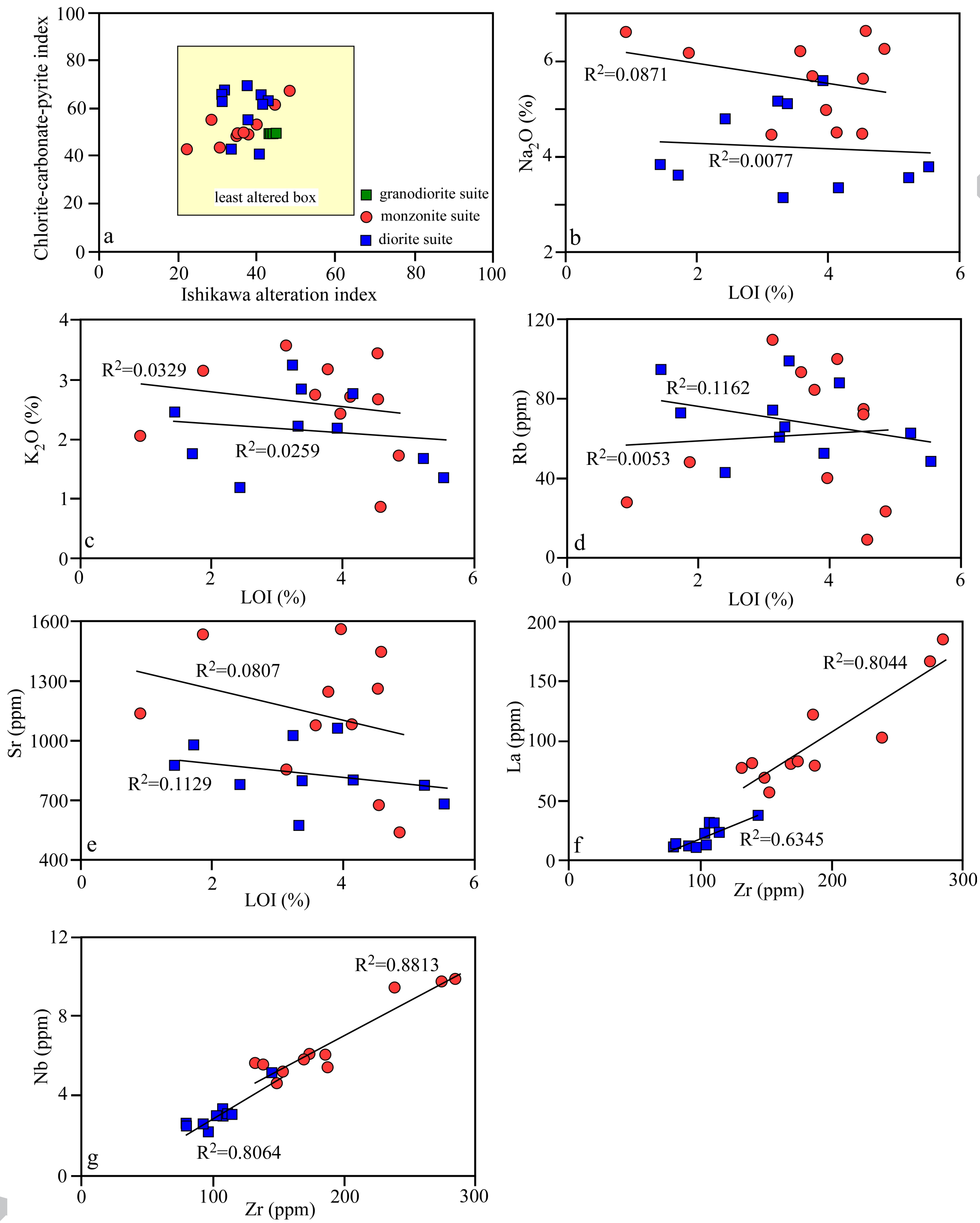
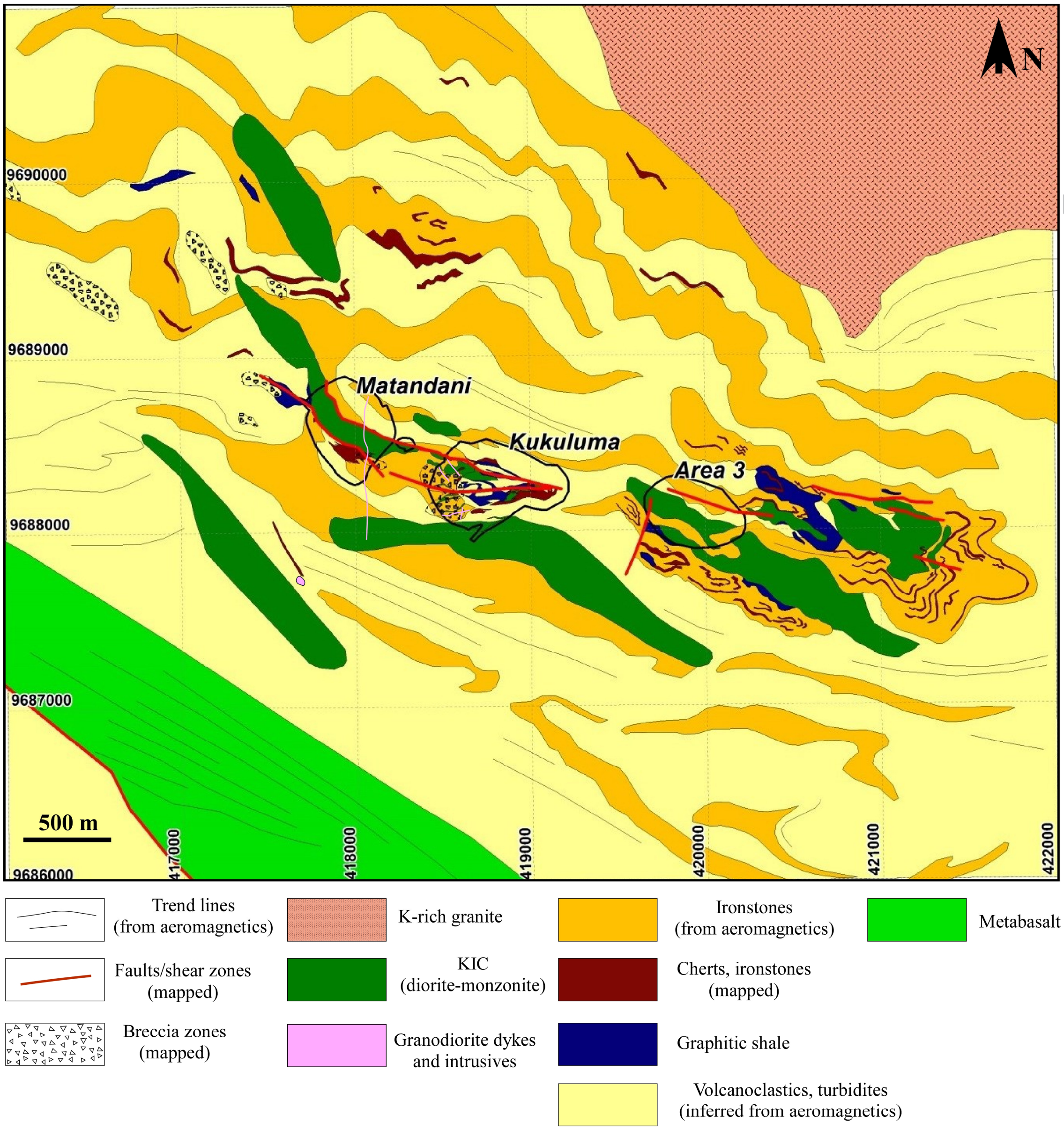

Chlorite-carbonate-pyrite index
Ishikawa alteration index
least altered box
granodiorite suite
monzonite suite
diorite suite
a
b
Na_2O (%)
LOI (%)
R^2=0.0871
R^2=0.0077
c
K_2O (%)
R^2=0.0329
R^2=0.0259
d
Rb (ppm)
R^2=0.1162
R^2=0.0053
e
Sr (ppm)
R^2=0.0807
R^2=0.1129
f
La (ppm)
Zr (ppm)
R^2=0.8044
R^2=0.6345
g
Nb (ppm)
R^2=0.8813
R^2=0.8064

Figure 4



902 **Figure 2**

903

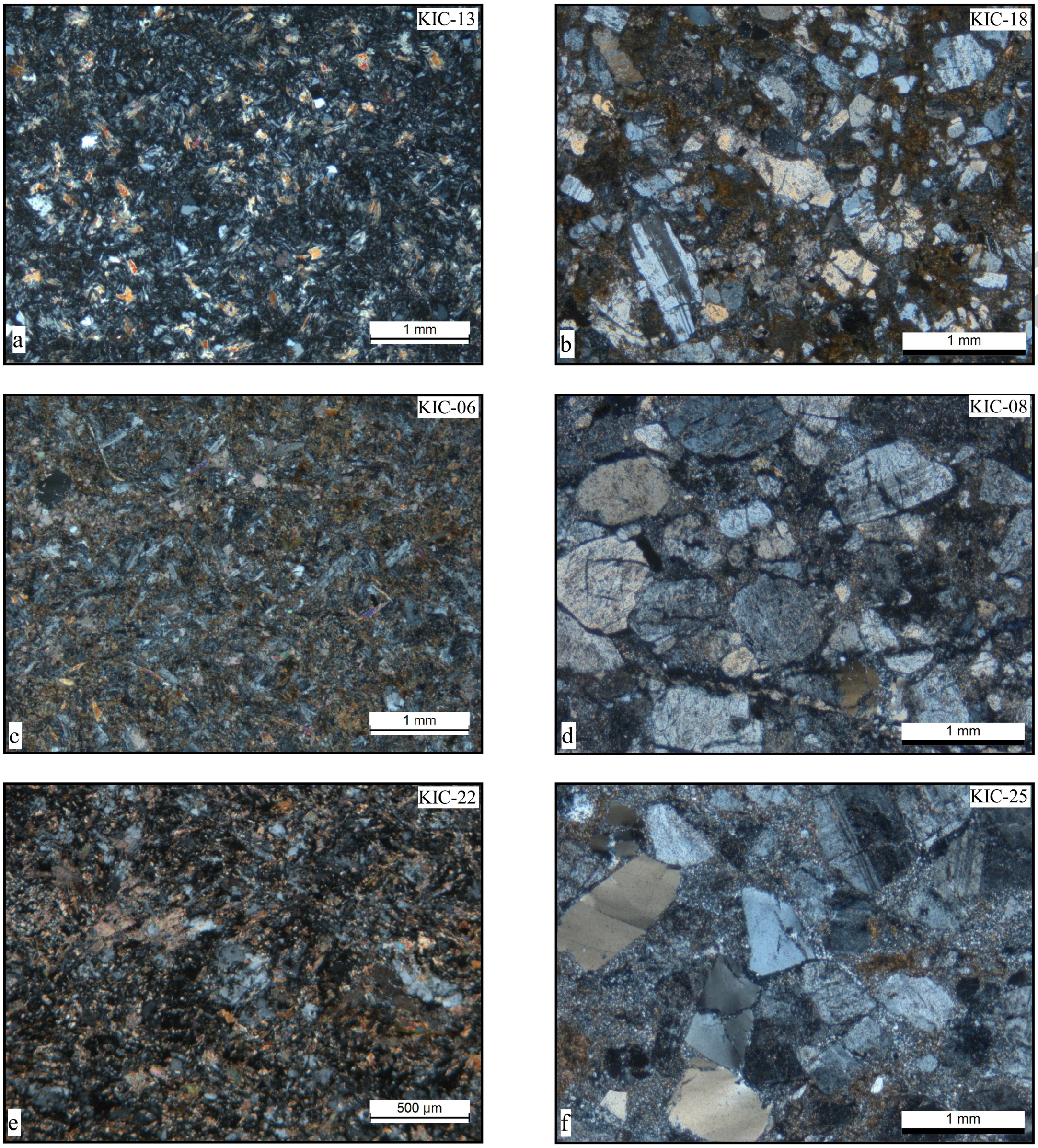

KIC-13
1 mm
a
KIC-18
1 mm
b
KIC-06
1 mm
c
KIC-08
1 mm
d
KIC-22
500 µm
e
KIC-25
1 mm
f

Figure 3

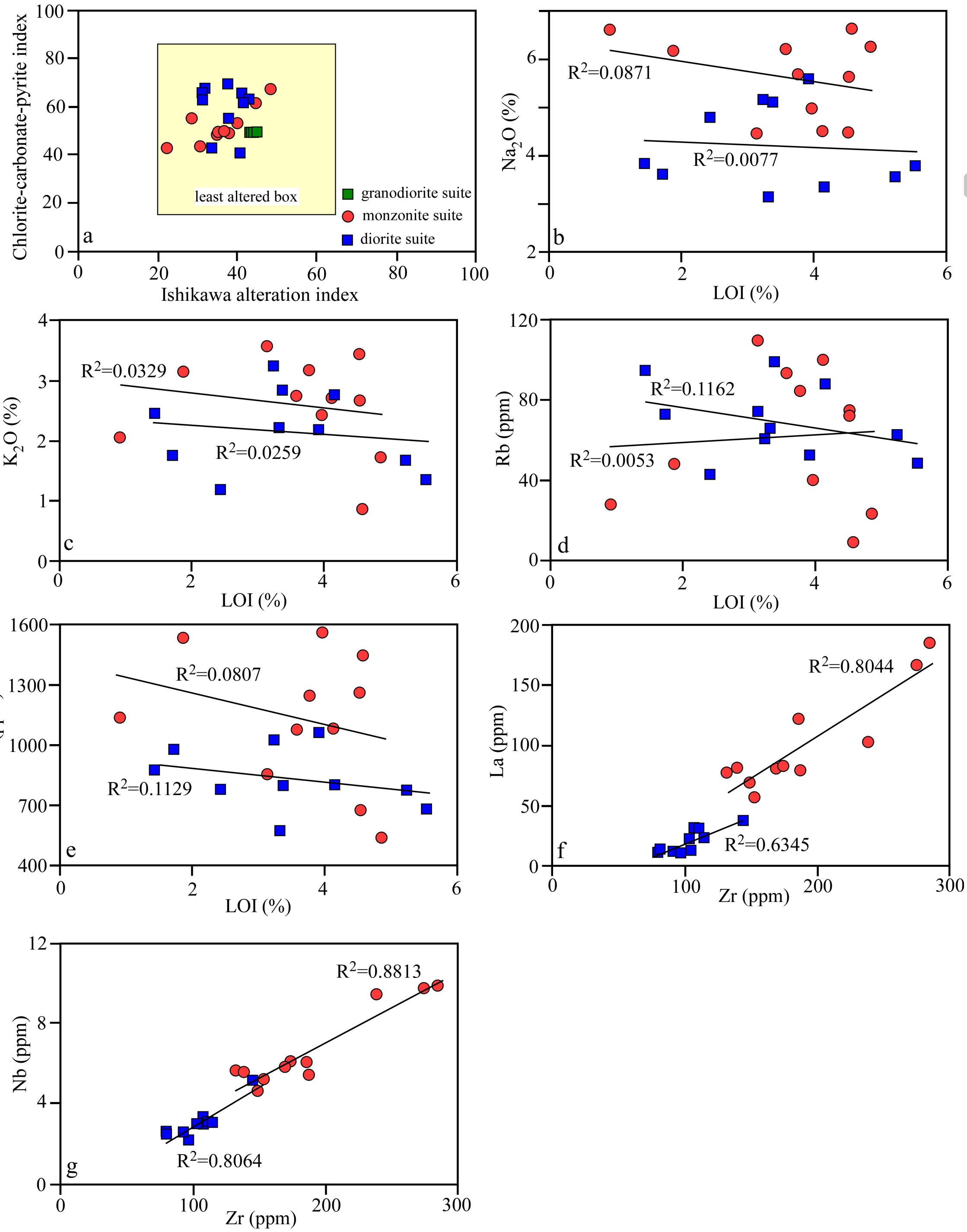
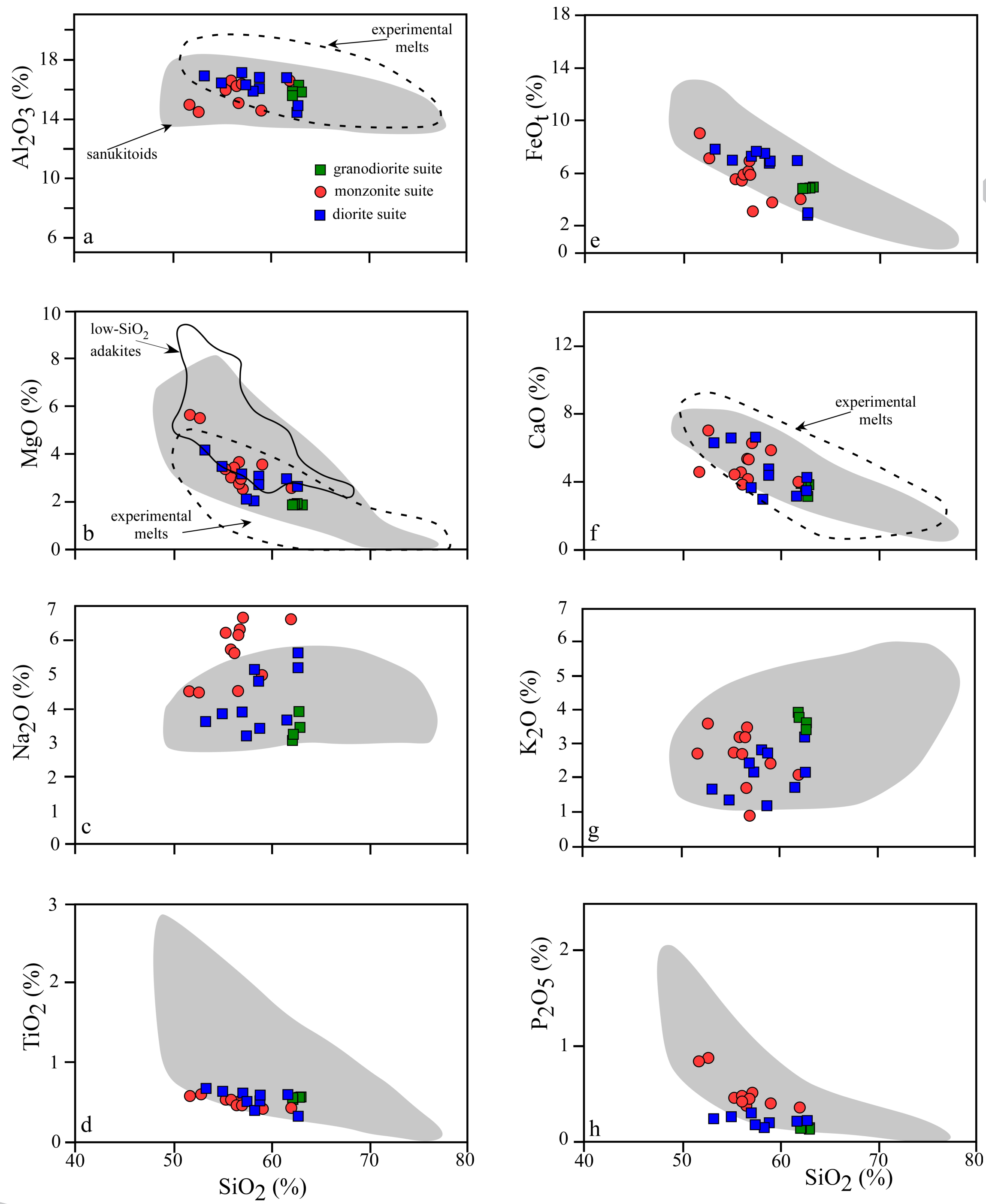

Chlorite-carbonate-pyrite index
Ishikawa alteration index
least altered box
granodiorite suite
monzonite suite
diorite suite
a
Na_2O (%)
LOI (%)
R^2=0.0871
R^2=0.0077
b
K_2O (%)
LOI (%)
R^2=0.0329
R^2=0.0259
c
Rb (ppm)
LOI (%)
R^2=0.1162
R^2=0.0053
d
Sr (ppm)
LOI (%)
R^2=0.0807
R^2=0.1129
e
La (ppm)
Zr (ppm)
R^2=0.8044
R^2=0.6345
f
Nb (ppm)
Zr (ppm)
R^2=0.8813
R^2=0.8064
g

Figure 4



908 Figure 5

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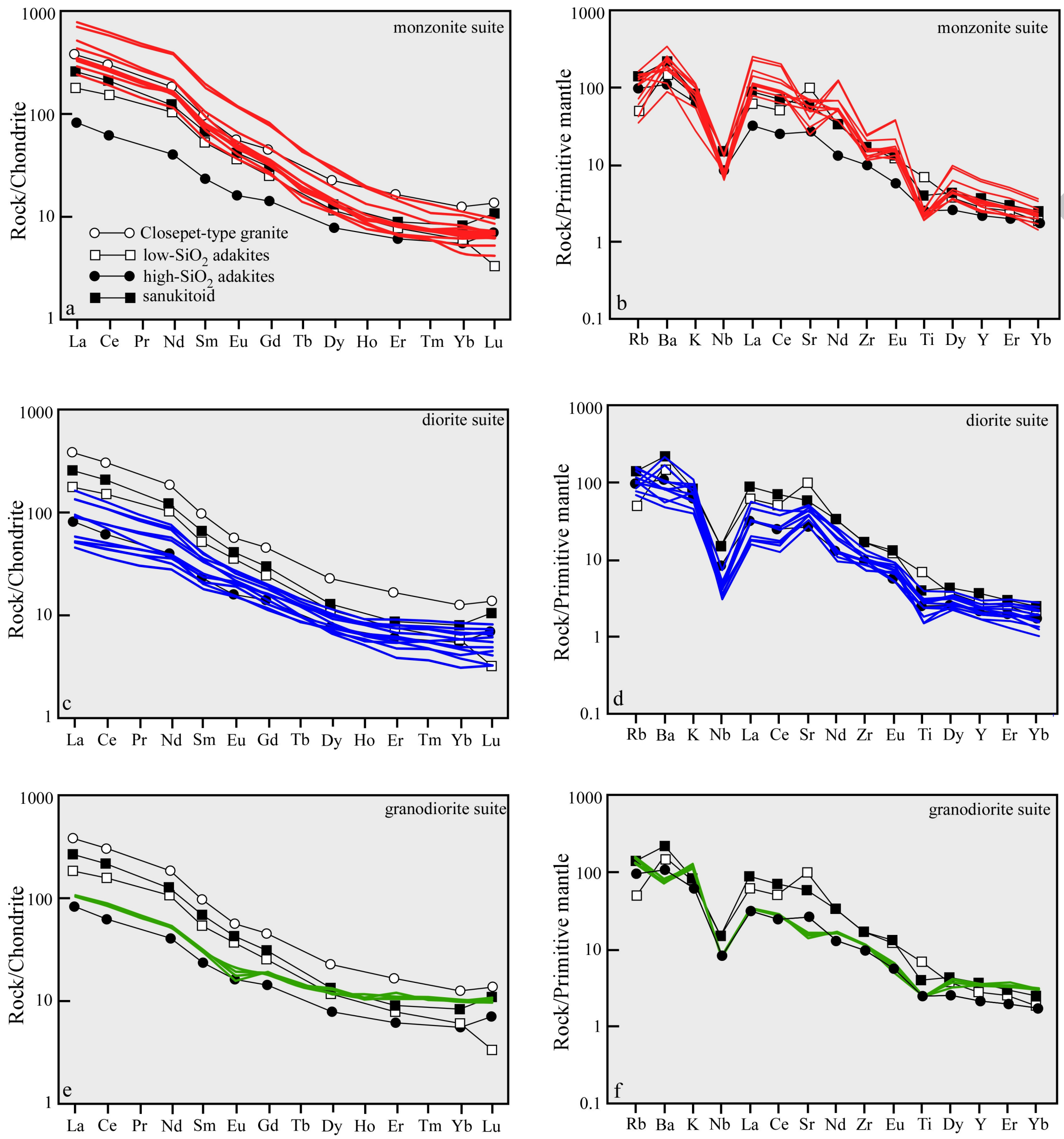

910 **Figure 6**

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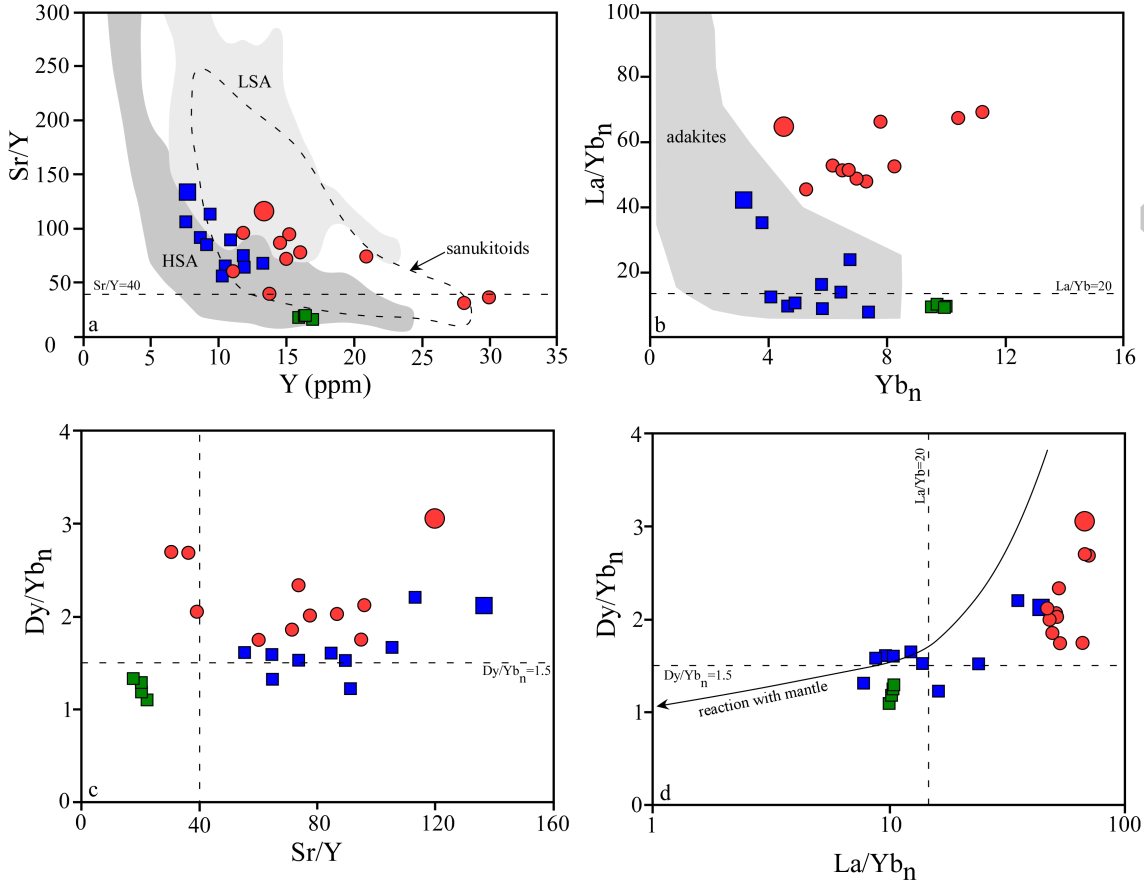

912 **Figure 7**

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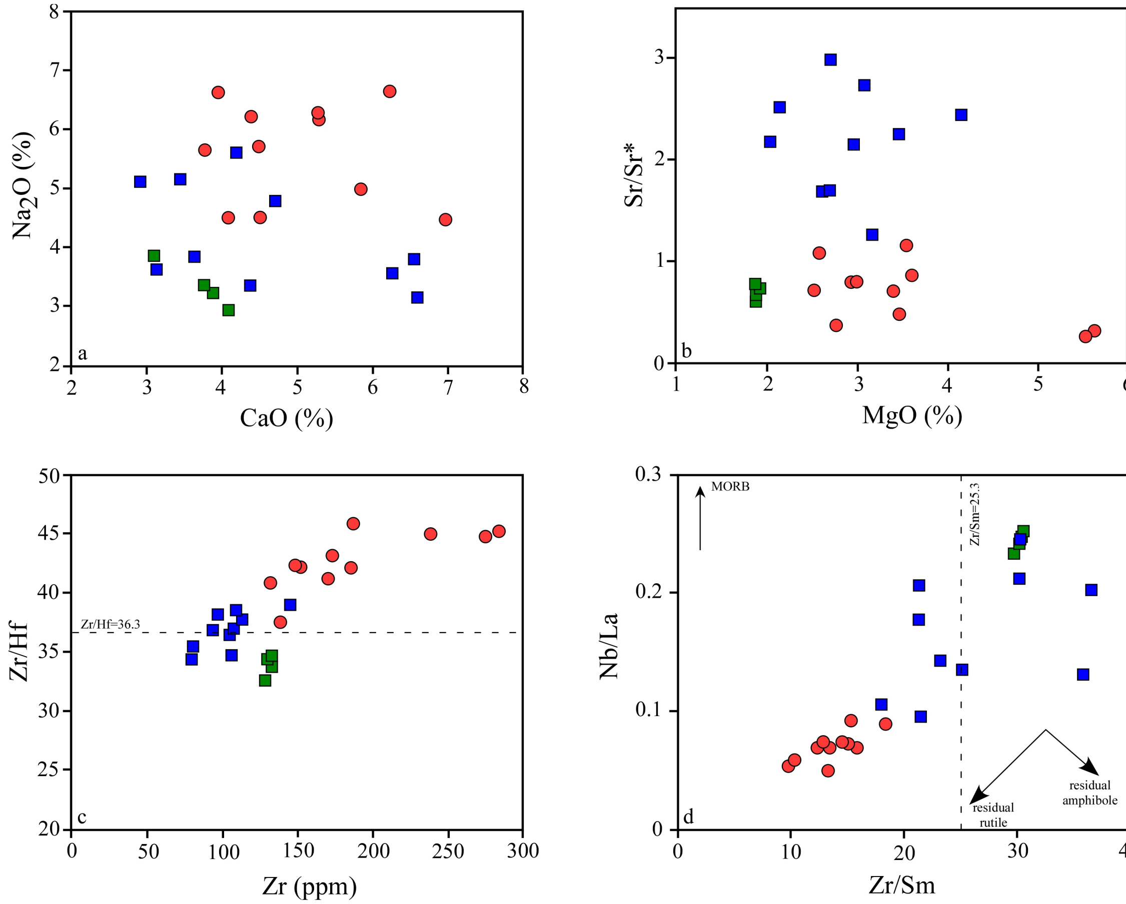
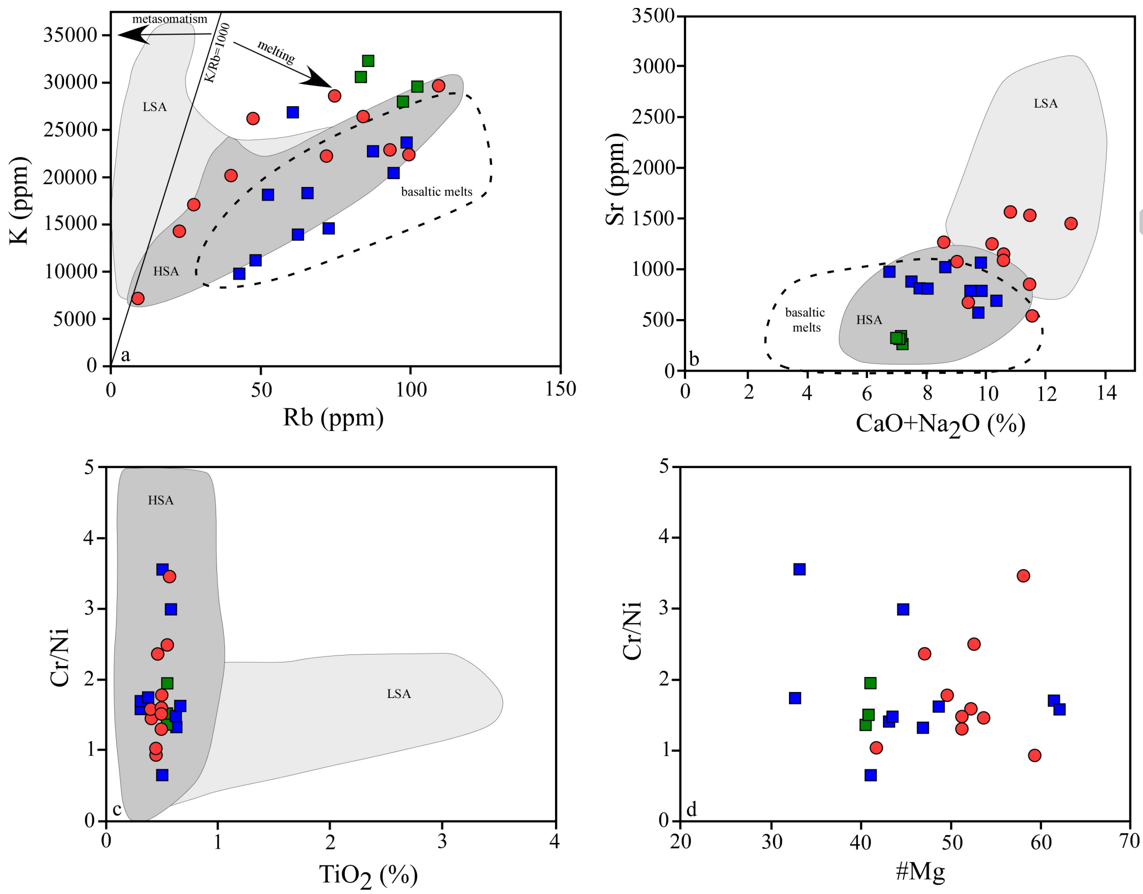


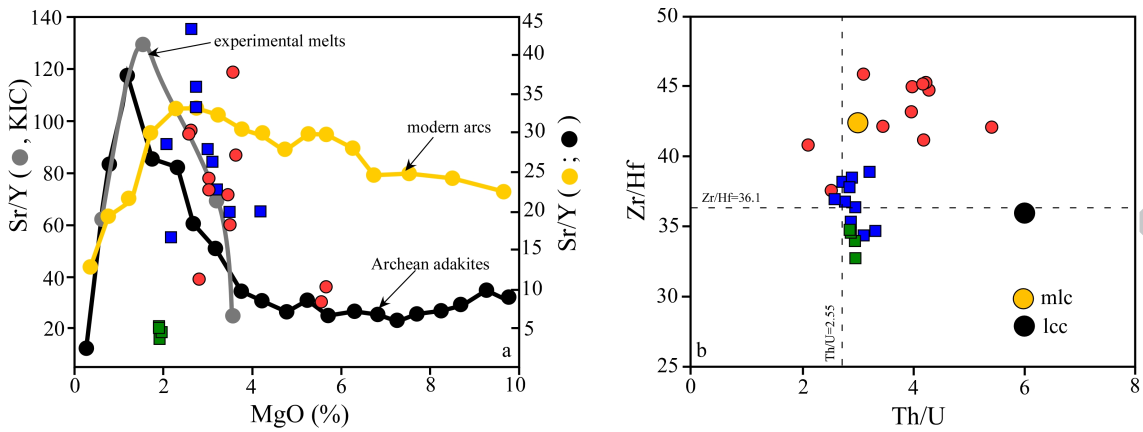
914 **Figure 8**

915



916 **Figure 9**

917



918 **Figure 10**

919

920 **Highlights**

921 • The KIC is a syn-tectonic Neoarchean igneous complex with intermediate composition
922 • The KIC has high Sr/Y and La/Yb ratios indicative of a melting in the garnet stability field
923 • The KIC formed by partial melting of thickened mafic/ultramafic lower crust
924 • The KIC is related to the removal of the eclogitised root of an oceanic plateau

925