

Classification and petrogenesis of the Tongo dike-01 from the Tongo-Tongoma cluster, Sierra Leone: constraints from bulk-rock geochemistry

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Abstract

The Man Craton in West Africa, like the Kaapvaal Craton in southern Africa, hosts diamondiferous kimberlites. However, West African kimberlites are commonly micaceous and unusual relative to archetypal South African kimberlites. Petrographically, they appear more similar to orangeites (aka Group II kimberlites), which represent a type of olivine-lamproite. A suite of 14 representative samples from the Cretaceous Tongo dike-01, Sierra Leone have been analysed for their bulk-rock major and trace elements as well as Sr-Nd-Pb isotope geochemistry. The primary objectives of this study are: 1) provide detailed petrographic observations of the dike, 2) classify the dike relative to kimberlites worldwide, 3) constrain the geochemical effects of secondary processes on bulk-rock analyses, 4) provide an estimate of the close-to-primary parent magma composition, and 5) constrain the petrogenesis of these diamondiferous rocks.

The major element chemistry of the Tongo dike-01 reflects concentrations that are similar to both archetypal kimberlites and orangeites. Major elements such as SiO_2 ($\sim 28.20 \pm 3.90$ wt. %) and CaO ($\sim 12.50 \pm 1.80$ wt. %) display archetypal kimberlite concentrations whereas Cr_2O_3 ($\sim 0.20 \pm 0.01$ wt. %) and P_2O_5 ($\sim 1.65 \pm 0.60$ wt. %) resemble those that define orangeites. The high abundance of phlogopite in this dike is illustrated by the high bulk-rock concentrations of K_2O ($\sim 3.03 \pm 0.50$ wt. %) and Al_2O_3 ($\sim 4.08 \pm 1.00$ wt. %), similar to those of orangeites. Like the major element geochemistry, the trace element geochemistry of the Tongo dike-01 also displays mixed archetypal kimberlite and orangeite traits. Trace elements such as Nb ($\sim 365.0 \pm 50.4$ ppm) and Y ($\sim 18.77 \pm 6.60$ ppm) possess concentrations that are similar to kimberlites whereas Rb ($\sim 160.0 \pm 14.8$

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Prior to interpretation of primary processes and parent magma composition estimates, the effects of secondary processes were first evaluated. These secondary processes include crustal contamination, ilmenite contamination and olivine entrainment/fractionation. Samples that had experienced these secondary processes were excluded and a suite of unaltered/least contaminated samples was compiled in order to constrain the close-to-primary magma composition of the Tongo dike-01 and interpret primary petrogenetic processes effecting the kimberlites. To determine a representative parent magma composition, a total of six out of the fourteen samples were excluded from consideration. The estimated close-to-primary magma composition for the Tongo dike-01 is suggested to be SiO₂ $\sim 28.20 \pm 3.90$ wt. %, Fe₂O₃ $\sim 10.20 \pm 1.70$ wt. %, TiO₂ $\sim 1.70 \pm 0.30$ wt. %, Al₂O₃ $\sim 4.08 \pm 1.00$ wt. %, K₂O $\sim 3.03 \pm 0.50$ wt. %, La $\sim 363.12 \pm 25.43$ ppm, Gd $\sim 15.87 \pm 4.20$ ppm, Yb $\sim 1.020 \pm 0.06$ ppm, ($^{87}\text{Sr}/^{86}\text{Sr}$)_i ~ 0.7039 , ($^{206}\text{Pb}/^{204}\text{Pb}$)_i ~ 18.88 , ($^{208}\text{Pb}/^{204}\text{Pb}$)_i ~ 40.02 and ($^{143}\text{Nd}/^{144}\text{Nd}$)_i $\sim 0.51253 \pm 0.00001$.

Although the petrography and major element concentrations are similar to orangeites found in South Africa, the trace element and Sr-Nd-Pb isotope geochemistry of the Tongo dike-01 reflects a kimberlite composition. Thus, the Tongo dike-01 is more consistent with being classified as a relatively rare type of 'mica-rich' kimberlite rather than orangeite. Kimberlites from around the world

derive from the same asthenospheric mantle reservoir and their major element chemistry is controlled by the compositions/mineralogy of the lithospheric mantle assimilated during kimberlite evolution. The similarity of the trace element ratios and Sr-Nd-Pb isotopes of the Tongo dike-01 in this study relative to archetypal kimberlites worldwide strongly implies that the Tongo dike-01 derives from the same asthenospheric reservoir as these kimberlites, although mineralogically the Tongo dike-01 is different and has a different parent melt major element composition. This is interpreted to reflect the contribution of lithospheric mantle material that is mineralogically different to that assimilated by archetypal kimberlites during the ascent of the Tongo dike-01 parent magma through the sub-continental lithospheric mantle (SCLM). In the case of the Tongo dike-01, its primary melt is K₂O-rich and must have assimilated more K₂O-rich material in the SCLM. Such material is typically present as metasomatic products, e.g., Phlogopite-Ilmenite-Clinopyroxene (PIC) xenoliths observed in South Africa kimberlites. These xenoliths tend to possess abundant phlogopite. Thus, the main difference between the Tongo kimberlite and archetypal SA kimberlites is the fact that Tongo kimberlite assimilated more K₂O-rich metasomatised material in the SCLM during its evolution.

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1. Introduction

Kimberlites are ultramafic igneous rocks that originate from the Earth's asthenosphere and can be found in cratonic environments worldwide (e.g., Giuliani and Pearson, 2019 Elements). These rocks are mostly known for containing diamonds and mantle xenoliths derived from the sub-continental lithospheric mantle (SCLM) (Mitchell, 1986). Another type of magma that comes from the Earth's interior that are diamondiferous is lamproite. The key difference in the petrogenesis between kimberlites and lamproites is that lamproites originate from the Earth's lithospheric mantle whereas kimberlites are derived from the asthenosphere (Mitchell, 1986). The classification of kimberlites versus lamproite, however, is not straightforward and a key research objective of this study is to address classification of unusually micaceous-rich diamond-bearing rocks from Sierra Leone, West Africa, as either lamproite or kimberlite.

In the past, kimberlites in South Africa were classified into two groups: "basaltic" (non-micaceous) Group I kimberlites and "micaceous" Group II kimberlites (Smith, 1983). The main petrographic and geochemical differences between Group I kimberlites and Group II kimberlites would be Group II kimberlites are more enriched mica and potassium (Smith, 1983; Mitchell, 1995; le Roex et al., 2003; Becker and le Roux, 2006). Group I kimberlites have been shown to have high $^{143}\text{Nd}/^{144}\text{Nd}$ isotope ratios and low $^{87}\text{Sr}/^{86}\text{Sr}$ compared to present-day Bulk Silicate Earth and shows similarity to OIB's (Oceanic Island Basalts) (Smith, 1983; Mitchell, 1995; le Roex et al., 2003; Becker and le Roux, 2006). Group II kimberlites have low $^{143}\text{Nd}/^{144}\text{Nd}$ and high $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratios compared to present-day Bulk Silicate Earth composition, and unlike Group I kimberlites, they are derived from the SCLM (Smith, 1983; Mitchell, 1995; Becker and Roux, 2006; Coe et al., 2008; Howarth et al., 2011). More recently, the terminology has changed and Group I kimberlites are now referred to as archetypal kimberlite or simply

kimberlite whereas Group II kimberlites are termed orangeite (Mitchell, 1995). Furthermore, orangeites have been classified as a sub-type of lamproites, termed carbonate-rich olivine lamproite (Pearson et al., 2019). Hereafter, for simplicity, I use kimberlite for true archetypal kimberlites, orangeite for the rocks formerly termed Group II kimberlites, and lamproite for diamondiferous lamproites outside of South Africa.

Southern African kimberlites and orangeites have for a long time influenced the knowledge that we have about the petrogenesis and geochemistry of diamondiferous rock types worldwide. Diamondiferous rocks from West Africa are characteristically micaceous and possess transitional geochemical, petrographic, and mineralogical characteristics between orangeites and kimberlites (Taylor et al., 1994; Howarth and Giuliani, 2020). These rocks are micaceous containing high concentrations of phlogopite relative to kimberlites and possess similar petrographic features to orangeites (e.g., Taylor et al., 1994; Skinner et al., 2004). However, the bulk-rock geochemistry of the Koidu kimberlite found in Sierra Leone is intermediate between kimberlite and the Aries lamproite (Taylor et al., 1994) and the bulk-rock geochemistry of Weasua (Liberia) is mostly similar to kimberlites (Skinner et al., 2004). Based on these previous studies it appears that West Africa possesses kimberlites, lamproites and maybe transitional rocks. And in all these studies very little isotope data has been collected for West African kimberlites and no geochemical research has been conducted on the Tongo dikes. The key objective of this study will include collecting the first robust Sr-Nd-Pb isotope dataset for the region which is crucial in constraining the classification and petrogenesis of the Tongo dike-01.

1.1 Objectives

A suite of 14 representative samples have been obtained from Tongo dike-01 in collaboration with the mining company formerly known as Stellar diamonds,

which has now been incorporated in Newfield ltd. (Figure 1). These samples were obtained from several texturally distinct phases within the dike and are analysed to: (1) produce detailed petrographic observations, (2) provide analyses of the major and trace elements, taking into account the geochemical effects of peridotite entrainment, crustal contamination, fractional crystallisation and alteration and (3) provide amongst the first Sr-Nd-Pb isotope data for these rocks. The objective would then be to: (4) explain their petrogenesis to gain understanding of the processes involved in the formation of these unusual rocks and the evolution experienced *en route* to the surface and, (5) classify these rocks within the standard classification of kimberlites, transitional kimberlites and orangeites worldwide.

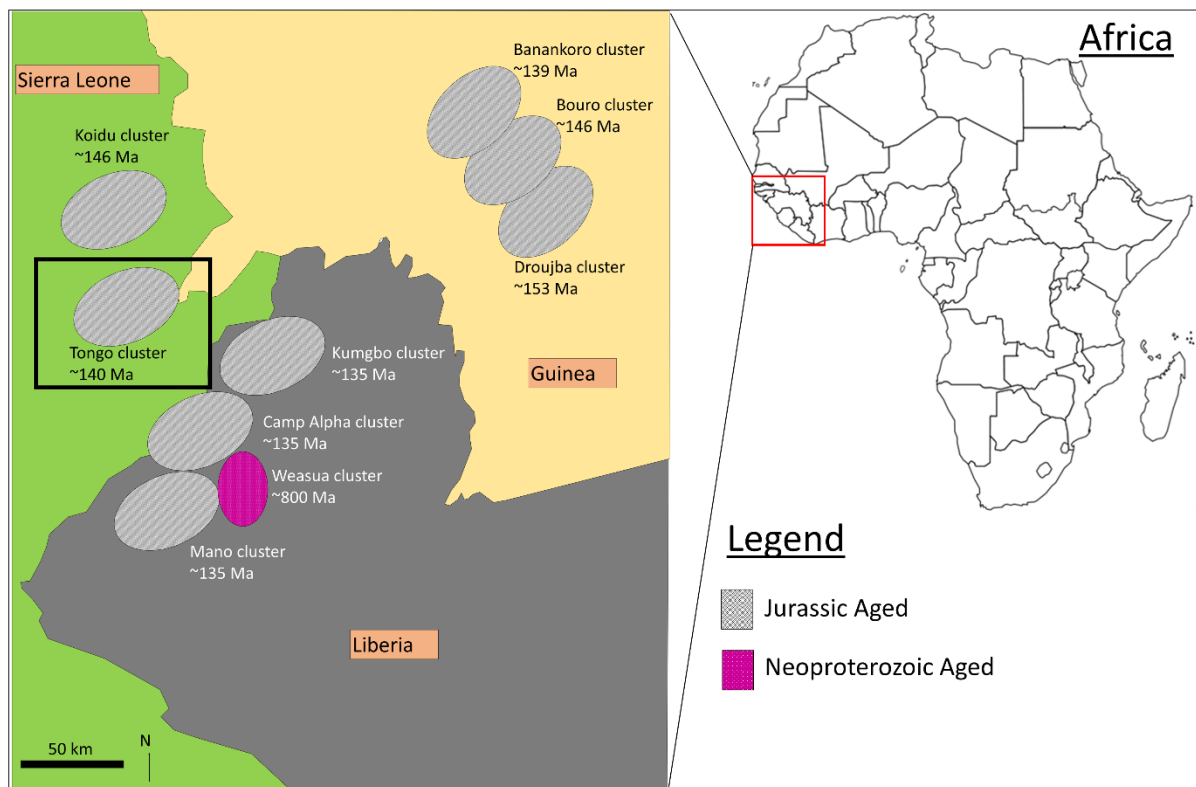


Fig. 1. Map of the localities of West African kimberlite clusters from Sierra Leone, Liberia, and Guinea. The rocks analysed in this paper are from the Tongo dike-01 in the Tongo cluster demarcated by the black box. This map has been modified from Skinner et al. (2004).

2. Literature Review

In this chapter two major sections are discussed in detail with regards to what features define diamondiferous rocks. The first section is comparing kimberlite, orangeite and lamproite characteristics. This section is divided between the mineralogy, bulk-rock geochemistry, and isotope chemistry that define archetypal kimberlites, orangeites and lamproites. The second section of the chapter specifically looks at the mineralogy and chemistry of West African diamondiferous rocks, which is where the Tongo dike-01 samples come from. Although they may be all diamondiferous rocks, West African rocks have shown to be different from archetypal kimberlites, orangeites and lamproites worldwide and this chapter will be discussing the features that distinguish these rocks from one another and what relationships exist between them.

2.1 Comparing kimberlite, orangeite and lamproite characteristics

2.1.1 Mineralogy

Kimberlites are a clan of potassic ultrabasic, volatile-rich (mostly CO₂) rocks that exhibit a distinguishing inequigranular texture that comes from containing megacrysts (1- 20cm in size) and macrocrysts (0.5 to 10mm in size) within a fine-grained (<0.5mm) matrix (e.g., Clement and Skinner, 1979; Clement, 1982; Clement et al., 1984; Smith et al., 1985; Mitchell, 1995; Maitre et al., 2005; Mitchell, 2008; Mitchell et al., 2019). These mega/macrocrysts are xenocrystic and are derived from the mantle and are not related to kimberlite groundmass crystallisation. The macrocryst assemblage is predominantly made up of anhedral olivine, garnet, Cr-diopside, phlogopite, enstatite, Ti-poor chromite and magnesian ilmenite (e.g., Smith et al., 1985; Mitchell, 1995; Maitre et al., 2005). Olivine mega/macrocrysts are the most abundant phase comprising ~25 vol.% in typical kimberlites that have not undergone severe fractionation *en route* to the surface (Clement and Skinner, 1979; Clement, 1982; Clement et al., 1984; Smith

et al., 1985; Mitchell, 1995; Woolley et al., 1996). The groundmass in kimberlites is made up of a second-generation of phenocrystic olivine (euhedral to subhedral in shape) together with the following primary minerals: spinel, ilmenite, phlogopite, perovskite, monticellite, serpentine, carbonate, and apatite (e.g., Clement and Skinner, 1979; Clement, 1982; Clement et al., 1984; Smith et al., 1985; Mitchell, 1995; Woolley et al., 1996; Maitre et al., 2005; Mitchell, 2008; Mitchell et al., 2019). Most kimberlites possess a late forming poikilitic mica i.e., barian phlogopite, which is of the kinoshitalite series (Mitchell, 1995; Woolley et al., 1996). A common feature within kimberlite is the replacement of earlier formed, phlogopite, olivine, apatite and monticellite by calcite and deuteritic serpentine (Skinner and Marsh, 2004; Stripp et al., 2006; Mitchell, 2008). Macrocryst-poor (<10 vol.% macrocrysts) varieties are termed aphanitic kimberlites, and are mostly made up of second-generation olivine, serpentine, spinel and calcite, with also minor perovskite, ilmenite, apatite and phlogopite (Mitchell, 1995; Woolley et al., 1996; le Roux et al., 2003; Harris et al., 2004).

The presence of the following primary phases characterizes lamproites: clinopyroxene, amphibole, leucite, forsteritic olivine, iron-rich leucite, poikilitic tetraferriphlogopite, iron-rich sanidine, aluminium-poor phlogopite, and glass (Bergman, 1987; Mitchell and Bergman, 1991). Not all of the above-mentioned minerals have to be present in order for the rock to be classified as a lamproite (Woolley et al., 1996). The minor and accessory minerals found in lamproites include perovskite, wadeite, magnesiochromite, and apatite, and then there are the less common minor minerals such as ilmenite, shcherbakovite, jeppeite, enstatite and armalcolite (Bergman, 1987; Mitchell and Bergman, 1991; Woolley et al., 1996). Importantly, lamproites generally do not contain carbonate, a key phase aiding classification of lamproite relative to orangeite.

Orangeites are a group of volatile-rich (mostly H₂O-rich) ultrapotassic peralkaline rocks, predominantly composed of phlogopite in several textural associations,

including phenocrysts (0.05mm to 0.5mm), macrocrysts (0.5mm to 10mm), and groundmass mica (<0.05mm) (Mitchell, 1995; Coe et al., 2008). The phlogopite found in orangeites has been shown to vary in composition from phlogopite to tetraferriphlogopite and can constitute up to 40 vol.% of bulk samples (Mitchell, 1995; Coe et al., 2008; Howarth et al., 2011). Olivine can occur as three types in orangeites, megacrysts (1cm - 5cm), macrocrysts (0.5mm to 10mm) and phenocrysts (0.05mm to 0.5mm). These three populations can occupy up to 30 vol.% of the total volume in orangeites (Coe et al., 2008; Howarth et al., 2011). The groundmass is characterized by phases such as: spinels that have compositions that range from Mg-bearing chromite to Ti-bearing magnetite, diopside that may be enveloped by titanian aegirine; monazite that is rich in REE; REE-rich and Sr-rich perovskite; Sr-rich apatite; Mn-bearing ilmenite; and Nb-bearing rutile (Mitchell, 1995; Coe et al., 2008; Howarth et al., 2011). Orangeites possess a much smaller size and lower abundance of spinel and perovskite compared to kimberlites (Mitchell, 1995; Coe et al., 2008; Howarth et al., 2011). In addition to the above phases, more evolved orangeites contain potassium richterite, aegirine, leucite and sanidine and appear more similar to typical lamproites (Mitchell, 1995; Howarth et al., 2011).

The primary silicate phases of kimberlites and orangeites differ, although it is not always possible to tell the difference between a kimberlite and an orangeite based on petrographic examination alone (Mitchell, 1995; Taylor and Kingdom, 1999). Mica-rich kimberlites can share mineralogical and petrographical features with orangeites such as the presence of high-Sr apatite, diopside in the groundmass, accessory mangiferous ilmenite and tetraferriphlogopite rims around the groundmass; examples of such kimberlites would be the Jagersfontein kimberlite, the Liaoning kimberlites of North East China and the Kuusamo cluster of the Karelian craton in Finland (Taylor and Kingdom, 1999; Tomkins et al., 1999; Dalton et al., 2020). In addition, there is mineralogical overlap between orangeites and lamproites. The evolved orangeites have a closer mineralogical relationship

to lamproites, however, unlike lamproites they generally contain carbonate in the groundmass (Bergman, 1987; Mitchell and Bergman, 1991; Mitchell, 1995). Certain minerals allow clear identification of kimberlite vs. orangeite vs. lamproite. The presence of diopside or any other potassium-rich phases like K-richrichterite and leucite are only present in lamproites/orangeites and not in kimberlite, however, orangeites don't always contain diopside (Mitchell, 1995; Howarth et al., 2011).

2.1.2 Geochemistry

Kimberlites

The MgO and SiO₂ contents of kimberlites display a broad positive correlation (Figure 2a). The FeO contents for kimberlites vary from 7.60 wt. % to 15.40 wt. % and display a broad and negative correlation against MgO (Figure 2b). The K₂O vs TiO₂ is an important variation diagram as it reflects phlogopite abundance in diamondiferous rocks (Figure 2c). Kimberlites have lower K₂O and Al₂O₃ concentrations than orangeites and low K₂O and Al₂O₃ reflects the low amounts of phlogopite in kimberlites (Figure 2c). High Field Strength Elements (HFSE) accompanied by other incompatible elements usually cover a wide range of concentrations in kimberlites (Becker and le Roux, 2006). Kimberlite data shows that they follow a similar patterns for Rare Earth Elements (REE) (Figure 5) and are significantly more enriched in light REE (e.g., La = 300-1300 x chondrite) compared to heavy REE (e.g., Lu = 3-8 x chondrite).

Primitive mantle-normalized plots show that kimberlites have a depletion in K, Ti, Sr, Rb, K and Hf, but a high enrichment in the more incompatible elements (e.g., Ba = 23-327 and Nb = 156-605 x primitive mantle) (Figure 6a). The negative anomalies (K, Sr, Ti) seen in the primitive mantle-normalized plot are a characteristic feature used to define archetypal kimberlites (Figure 6a) (le Roux

et al., 2003; Harris et al., 2004; Becker and le Roux, 2006). Potassium and Rb tend to have a larger negative anomaly compared to Ti anomalies (Becker and le Roux, 2006). Incorporation of zircon megacrysts can lead to strong positive Zr-Hf anomalies in kimberlites (Figure 6a) (Becker and le Roux, 2006). Kimberlite initial Sr-Nd-Pb isotope ratios are $(^{206}\text{Pb}/^{204}\text{Pb})_i \sim 19.547$ to 22.075 , $(^{207}\text{Pb}/^{204}\text{Pb})_i \sim 15.693$ to 15.774 , $(^{208}\text{Pb}/^{204}\text{Pb})_i \sim 40.061$ to 40.123 , $(^{87}\text{Sr}/^{86}\text{Sr})_i \sim 0.70328$ to 0.70537 , and $(^{143}\text{Nd}/^{144}\text{Nd})_i \sim 0.512491$ to 0.512711 (Figure 4) (Smith, 1983; Smith et al., 1985; le Roex et al., 2003; Becker and le Roex, 2006; Collerson et al., 2010).

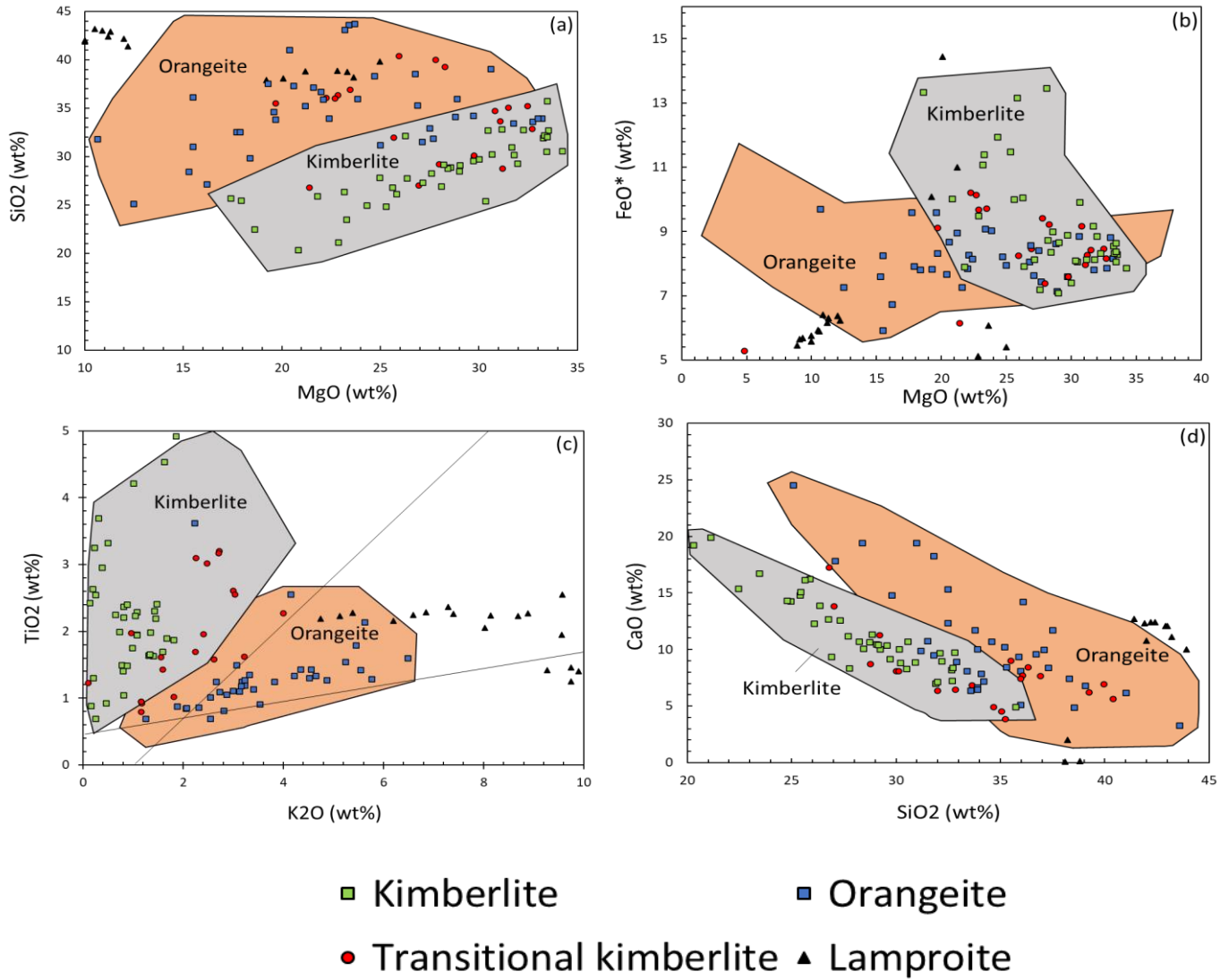


Fig. 2. Variation of (a) MgO vs SiO₂, (b) MgO* vs FeO*, (c) K₂O vs Ni, (d) SiO₂ vs CaO for kimberlites (le Roex et al., 2003 and Becker and le Roux, 2006), orangeites (Becker and le Roux, 2006 and Coe et al., 2008), transitional kimberlites (Becker et al., 2007) and Leucite Hills and Aries lamproites (Kornprobst, 1982 and Mirnejad and Bell, 2006). The light grey field represents kimberlite compiled using le Roex et al. (2003) and Becker and le Roux (2006) data. The tan field represents orangeite compiled using Becker and le Roux (2006) Coe et al. (2008) and Howarth et al. (2011) data.

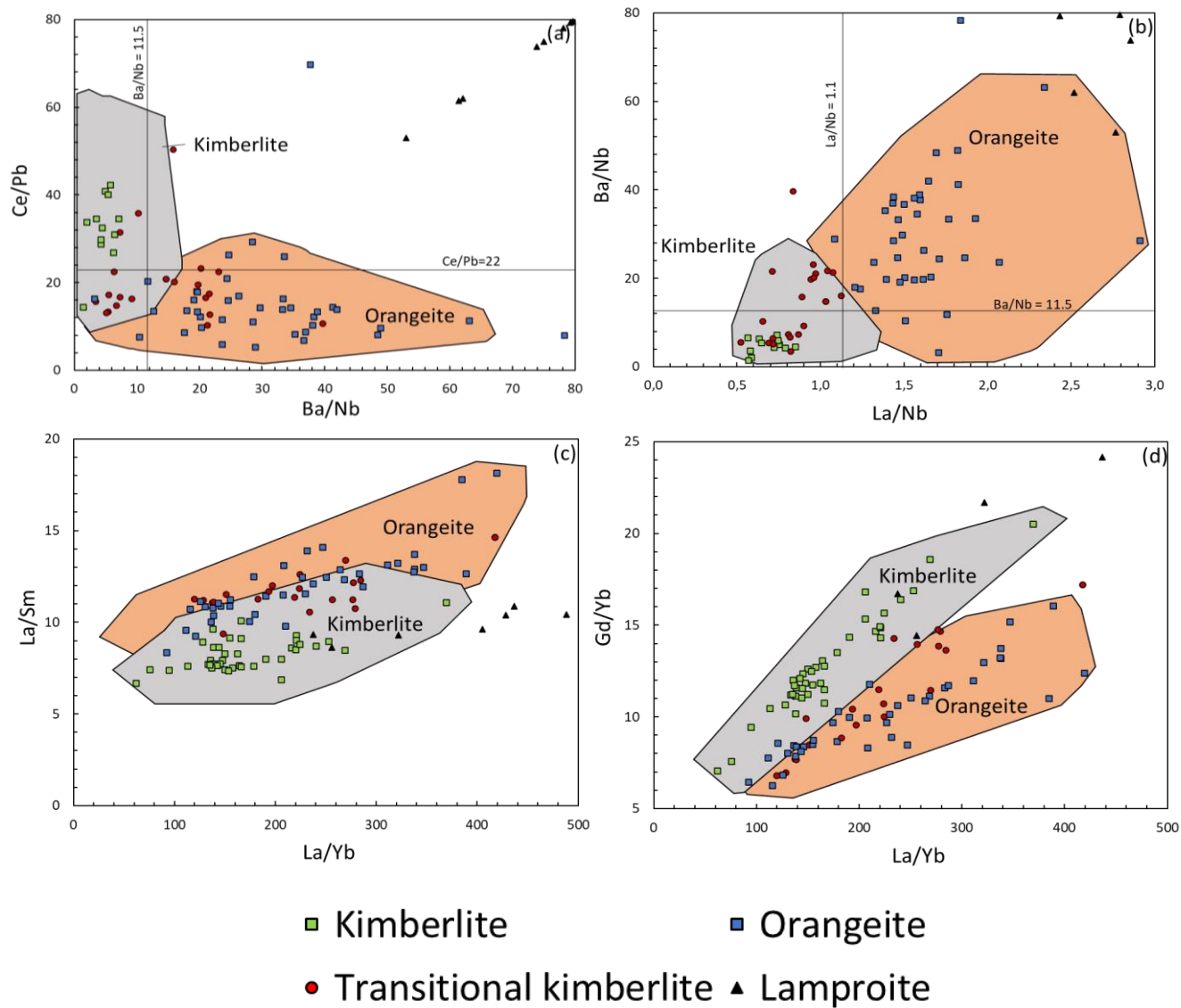


Fig. 3. Variation of (a) Ba/Nb vs Ce/Pb (b) La/Nb vs Ba/Nb (c) La/Yb vs La/Sm (d) La/Yb vs Gd/Yb CaO for kimberlites (le Roex et al., 2003 and Becker and le Roux, 2006), orangeites (Becker and le Roux, 2006 and Coe et al., 2008), transitional kimberlites (Becker et al., 2007) and Leucite Hills and Aries lamproites (Kornprobst, 1982 and Mirnejad and Bell, 2006). The light grey field represents kimberlite compiled using le Roex et al. (2003) and Becker and le Roux (2006) data. The tan field represents orangeite compiled using Becker and le Roux (2006) Coe et al. (2008) and Howarth et al. (2011) data.

Kimberlite isotope ratios show that the mantle source of kimberlites worldwide appears to have evolved along a uniform isotopic trajectory over a period of 2 billion years indicating that kimberlites globally derive from the same homogeneous reservoir (Woodhead et al., 2019; Giuliani et al., 2021). Kimberlites originate at depths greater than 140 km through the process of partial melting of mantle peridotite (Smith, 1983; Mitchell, 1986; Mitchell, 1995; Maitre et al., 2005; Mitchell, 2008; Stamm and Schmidt, 2017; Mitchell et al., 2019). During kimberlite formation, kimberlite magmas entrain lithospheric material *en route* to the surface, altering the composition of the primary kimberlite magma resulting in kimberlites with varying mineralogy but similar isotopic ratios (Woodhead et al., 2019; Giuliani et al., 2020).

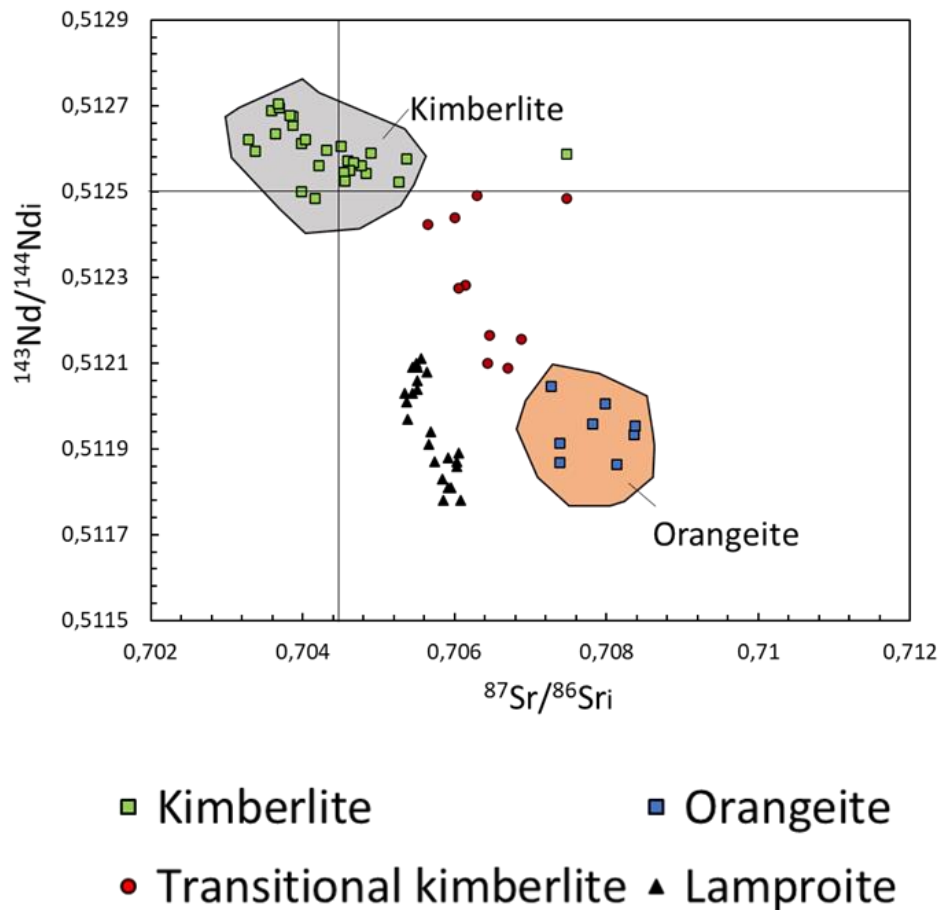


Fig. 4. Nd-Sr diagram showing isotopic concentration of kimberlites (le Roex et al., 2003 and Becker and le Roux, 2006), orangeites (Becker and le Roux, 2006 and Coe et al., 2008), transitional kimberlites (Becker et al., 2007) and Leucite Hills and Aries lamproites (Kornprobst, 1982 and Mirnejad and Bell, 2006). The light grey field represents kimberlite compiled using Becker and le Roux (2006) data and the tan field represents orangeite compiled using Becker and le Roux (2006) data.

Lamproites

Lamproites in general have a higher concentration in SiO₂, K₂O and TiO₂ and lower FeO and CaO concentrations compared to both kimberlites and orangeites (Bergman, 1987; Lustrino et al., 2016). When lamproites are compared to kimberlites and orangeites, lamproites have lower concentrations in mantle compatible elements (e.g., Sc, Ni, Cu, Cr, Zn and Co), but higher concentrations in incompatible elements (e.g., REE, Zr, Rb, Ba, U and Ba) (Bergman, 1987; Lustrino et al., 2016). Although the average REE concentrations of lamproites overlap with those of kimberlites and orangeites, it must be noted that REE concentrations of lamproites exceeds those of kimberlites and orangeites (Bergman, 1987; Lustrino et al., 2016). Phlogopite and madutipic lamproites are characterized by a large range in isotopic characteristics with Sr-Nd isotopes that have an initial value of $(^{87}\text{Sr}/^{86}\text{Sr})_i \sim 0.70874$ to 0.70961 and $(^{143}\text{Nd}/^{144}\text{Nd})_i \sim 0.51223$ to 0.51222 $(^{206}\text{Pb}/^{204}\text{Pb})_i \sim 18.54$ to 18.60 , $(^{207}\text{Pb}/^{204}\text{Pb})_i \sim 15.62$ to 15.70 , $(^{208}\text{Pb}/^{204}\text{Pb})_i \sim 38.88$ to 19.16 (Lustrino et al., 2016).

Depletion of Ca, Na, Al and the enrichment of Ti, Mg and K of primitive lamproites indicates that lamproites come from partial melting of enriched harzburgite sources under reduced water-rich conditions (Bergman, 1987; Mitchell and Bergman, 1991; Lustrino et al., 2016). The melt that forms these rocks comes from SCLM sources (Bergman, 1987; Mitchell and Bergman, 1991; Lustrino et al., 2016). Ancient subduction zones and many ancient “mantle metasomatic” events are required for the long-term production of this source (Bergman, 1987; Mitchell and Bergman, 1991; Lustrino et al., 2016).

Orangeites

Orangeites compared to kimberlites have higher SiO₂ concentrations (29.0 wt. % - 43.1 wt. %) (Becker and le Roux, 2006). Unlike kimberlites, MgO and SiO₂ in orangeites don't show any form of correlation (Mitchell, 1995; Becker and le Roux, 2006; Coe et al., 2008; Howarth et al., 2011). There is a large variation in MgO concentration and Mg number (Mg number = 0.80-0.91; MgO = 17.1 wt. % - 36.3 wt. %) amongst macrocrystic orangeites, which reflect variable olivine modal abundances (Becker and le Roux, 2006). This rock is characterized by low concentrations of TiO₂ (0.56 wt. % - 1.40 wt. %) relative to kimberlites and higher concentrations of K₂O (2 wt. % - 7 wt. %) and Al₂O₃ (3 wt. % - 5 wt. %); a characteristic used to help distinguish the two rock types (Becker and le Roux, 2006). There is a good correlation between SiO₂ and CaO (4.8 wt. % - 21.3 wt. %) and for any given SiO₂ concentration, orangeites tend to have higher CaO concentrations than kimberlites (Figure 2d).

Transition metals such as Cr (485-2326 ppm) and Ni (570-2062 ppm) correlates well with MgO which is again a reflection of varying olivine modal abundance (Becker and le Roux, 2006). What is more significant is that for any given MgO concentration, orangeites tend to have a higher Ni concentration compared to kimberlites (Becker and le Roux, 2006). Lead and Large Ion Lithophile Elements (LILE) concentrations of orangeites display a poor correlation against the more immobile incompatible elements (e.g., La) but are notably higher than those in kimberlites (Becker and le Roux, 2006). Incompatible trace element ratios have proven to be useful for distinguishing kimberlites from orangeites (Becker and le Roux, 2006; Coe et al., 2008; Howarth et al., 2011). Orangeites display higher ratios of Th/Nb (0.21 ± 0.06), Ba/Nb (24 ± 10) and La/Nb (1.6 ± 0.5) and lower ratios of Ce/Pb (16 ± 12) compared to kimberlites (Figure 3a and 3b). Orangeite REE patterns are sub-parallel on the chondrite normalized REE plots and are extremely enriched by Light Rare Earth Elements (LREE) (La; 1600 x chondrite)

compared to Heavy Rare Earth Elements (HREE) (Lu; 2-10 x chondrite) (Figure 5b). Orangeites have notably higher La/Yb and La/Sm ratios, but kimberlites have a higher Gd/Yb ratio (Figure 3c and 3d). Orangeites initial Sr-Nd-Pb isotope ratios are $(^{87}\text{Sr}/^{86}\text{Sr})_i \sim 0.70728$ to 0.70837 , $(^{143}\text{Nd}/^{144}\text{Nd})_i \sim 0.511865$ to 0.512055 , $(^{206}\text{Pb}/^{204}\text{Pb})_i \sim 17.638$ to 17.878 , $(^{207}\text{Pb}/^{204}\text{Pb})_i \sim 15.528$ to 15.637 and $(^{208}\text{Pb}/^{204}\text{Pb})_i \sim 37.089$ to 38.826 (Smith, 1983; Smith et al., 1985; Becker and le Roex, 2006).

Orangeite magmas come from sources that possess low U/Pb and Nd/Sm and high Rb-Sr ratios (Smith, 1983). This source is interpreted to be the geochemically enriched SCLM including highly metasomatised lithologies such as those represented by the mica-amphibole-rutile-ilmenite-diopside (MARID) xenoliths (Giuliani et al., 2015). There are four principal stages that describe the formation of orangeites: 1) mantle source development; 2) source melting; 3) ascent of melt into the crust and along the way, melt evolution and contamination; 4) crystallisation of melt once emplaced on Earth's surface (Mitchell, 1995).

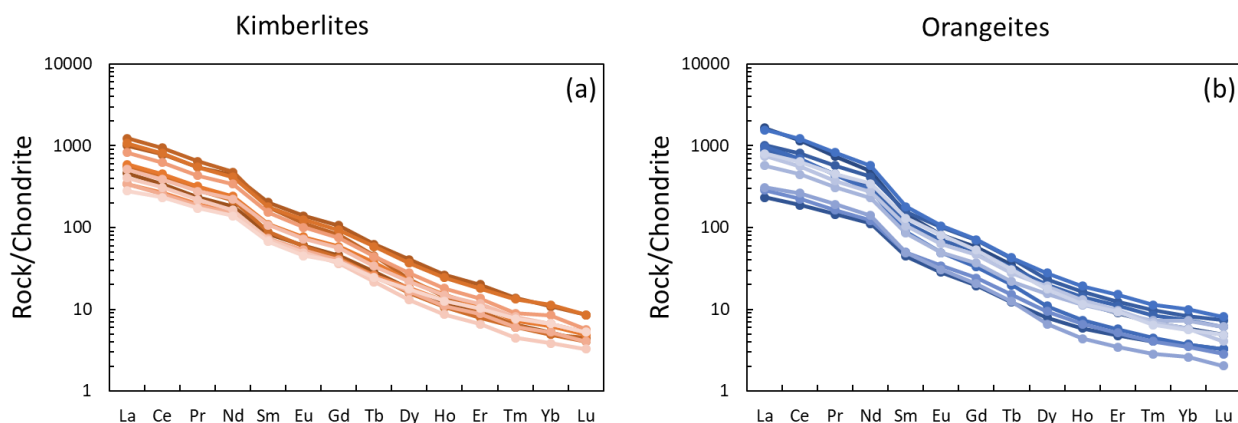


Fig. 5. Chondrite normalized REE plot for (a) kimberlite (Becker and le Roux, 2006) and (b) orangeite (Becker and le Roux, 2006).

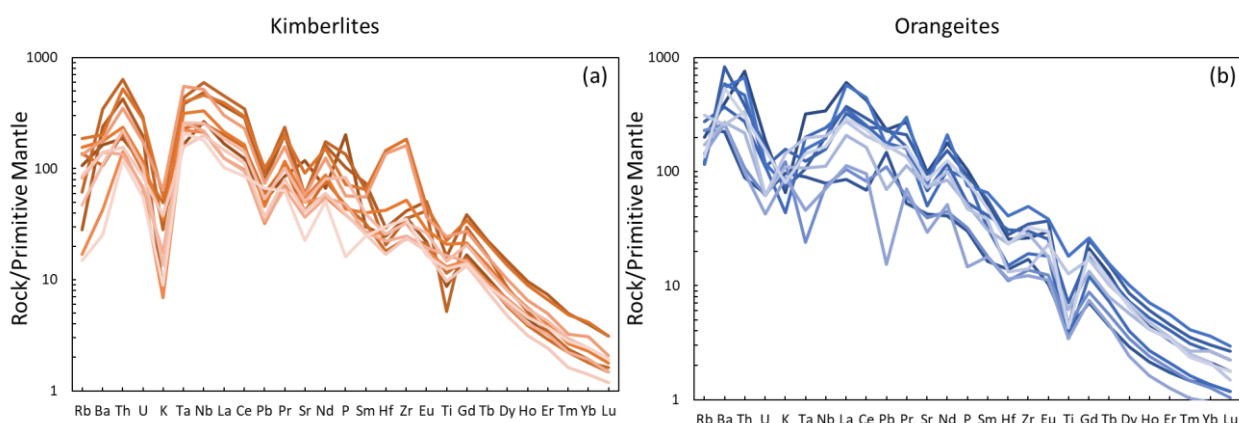


Fig. 6. Primitive mantle normalized trace element patterns for (a) kimberlite (Becker and le Roux, 2006) and (b) orangeite (Becker and le Roux, 2006).

Transitional kimberlites

Transitional kimberlite is a descriptive term used to describe rocks that display intermediate isotopic and petrographic characteristics between kimberlite and orangeite (Becker et al., 2007). Transitional kimberlites that are Mg-rich tend to be richer in olivine macrocrysts (Becker et al., 2007). These rocks can have concentrations of Al_2O_3 that range from 1.76 wt. % to 6.85 wt. % and FeO from 8.7 wt. % to 9.2 wt. % (Becker et al., 2007). A scattered relationship between K_2O (0.01 wt. % - 4.40 wt. %) and TiO_2 (0.79 wt. % - 3.20 wt. %) is observed, but in contrast SiO_2 (26.6 wt. % - 41.3 wt. %) and CaO (3.9 wt. % - 17.2 wt. %) show a broad negative correlation (Figure 2c and 2d).

Transitional kimberlites have a large range in transition metal concentrations (e.g., Co = 58-203 ppm, Ni = 628-1609 ppm, Cr = 1069-2417 ppm). These rocks tend to show a significant overlap with kimberlite and orangeite LILE (Ba = 612-3663 ppm, Sr = 467-1617 ppm, Rb = 5-200 ppm, Pb 7-26 ppm) and HFSE (Nb = 92-434 ppm, Ta = 4-18 ppm, Zr = 84-475 ppm). The inter-element ratios are also intermediate between kimberlites and orangeites (e.g., Ba/Nb = 14 ± 7 , La/Nb = 0.87 ± 0.15 , Ce/Pb = 20.0 ± 6.8 , Th/Nb = 0.12 ± 0.03 , K/Rb = 156 ± 20 (Becker et al., 2007). Transitional REE patterns are sub-parallel on chondrite normalized REE plots and are extremely enriched by Light Rare Earth Elements (LREE) (Figure 7). Transitional kimberlites are characterized by Sr-Nd isotopes that have an initial value of $(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.70575\text{-}0.70732$ and $(^{143}\text{Nd}/^{144}\text{Nd})_i = 0.512125\text{-}0.512532$ (Becker et al., 2007).

A possible theory that could explain the formation of transitional kimberlite is the mixing of sources of orangeite and kimberlite magmas in the mantle, by sharing a common plumbing system during emplacement (Smith, 1983; Becker et al., 2007). Another possible theory for the formation of transitional kimberlites is explained by Chalapathi Rao et al. (2009). They proposed a subduction event could have caused the overlapping of geochemical and petrological characteristics forming transitional kimberlites.

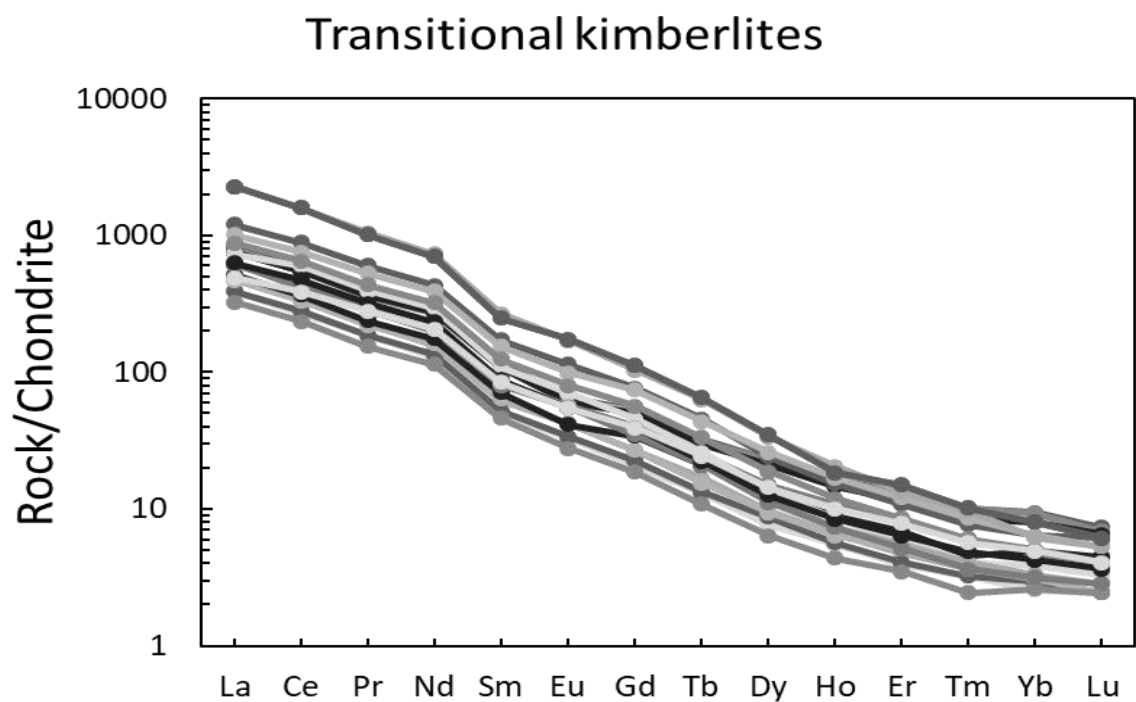


Fig. 7. Chondrite normalized REE plot for transitional kimberlite (Becker et al., 2007).

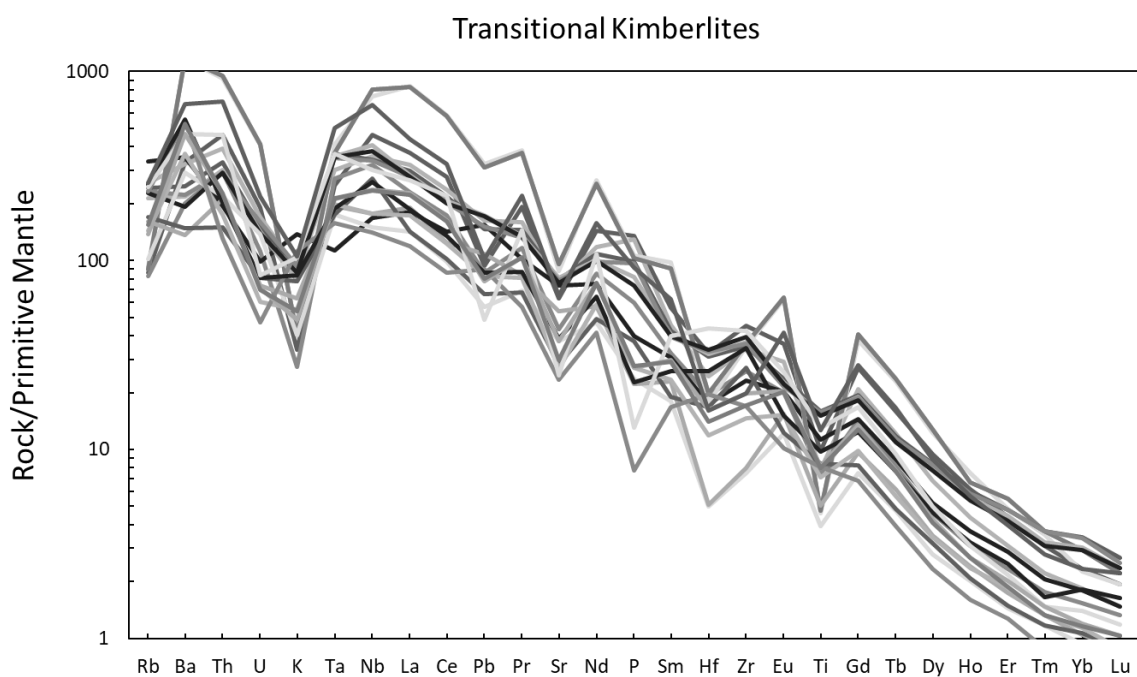


Fig. 8. Primitive mantle normalized trace element patterns plot for transitional kimberlite (Becker et al., 2007).

2.2 West African diamondiferous rocks

West African diamondiferous rocks from the Man craton occur both as small pipes and prolific dike systems and possess features that belong to kimberlite and orangeite (Howarth and Giuliani, 2020; Haggerty 2017; Skinner et al., 2004; Taylor et al., 1994). These rocks occur in numerous clusters of Jurassic age (135 – 155 Ma) in Liberia, Sierra Leone, and Guinea and one of Neoproterozoic age (~800 Ma) in Liberia (Weasua cluster) (Figure 1).

Kimberlite formation can be controlled by tectonic activity (Haggerty, 2017). Camp Alpha kimberlites sit in a well-defined regional trend along paleo fracture zones (Precambrian aged) and known as the ‘Liberian trend’ whereas the Koidu kimberlites dike trends fall under the ‘Sierra Leone trend’ which are traced to Mesozoic transform faults. The Koidu kimberlite, from Sierra Leone, are micaceous (mineralogically similar to orangeites) containing high modal abundances of phlogopite relative to typical South African kimberlites (e.g., Taylor et al., 1994; Skinner et al., 2004). However, the bulk-rock geochemistry of the Jurassic Koidu kimberlite displays intermediate characteristics between typical kimberlite and the Aries (phlogopite) lamproite (Taylor et al., 1994). In contrast, the Jurassic-aged Camp Alpha diamondiferous rocks in NW Liberia are remarkably similar in bulk-rock geochemistry to archetypal kimberlite, both in major and trace elements (Haggerty, 2017).

The Tongo-Tongoma cluster of Sierra Leone contains phlogopite and olivine macrocrysts in a groundmass of spinel, olivine, abundant phlogopite, perovskite, serpentine and carbonate (Howarth and Giuliani, 2020). The phlogopite in the Tongo rocks is characterised by simultaneous enrichment of Al_2O_3 and FeO, which is a typical feature for kimberlites and not orangeites (Howarth and Giuliani, 2020). These features and their kimberlite-like spinel compositions have led the Tongo rocks to be classified as mica-rich kimberlite (Howarth and Giuliani, 2020).

Rocks from the Neoproterozoic Weasua cluster are made up of olivine macrocrysts that sit in a groundmass consists of olivine, diopside, phlogopite, perovskite, spinel, apatite, calcite, clinopyroxene and serpentine (Howarth and Giuliani, 2020). The geochemistry of spinel and the presence of primary diopside (which is not a kimberlite mineral) in the Weasua cluster strongly suggests that these rocks classify as lamproites (Howarth and Giuliani, 2020). The incompatible trace element ratios of these rocks similarly suggests that the Weasua cluster contains lamproites (Haidula, 2010). However, samples analysed by Skinner et al. (2004) from the Weasua cluster did not show evidence of primary diopside and the bulk-rock geochemistry is consistent with kimberlite classification, and thus were classified as archetypal kimberlites. Howarth and Giuliani (2020) suggested that it is likely that both lamproites and kimberlites are present in the Weasua cluster.

Diamondiferous rocks from West Africa are not easily classified and a key piece of the puzzle missing is that of Sr-Nd-Pb isotopes. Only two samples from the Koidu kimberlite have been analysed for Sr-Pb isotopes $(^{87}\text{Sr}/^{86}\text{Sr})_i = 0.70448\text{-}0.70481$ and $(^{143}\text{Nd}/^{144}\text{Nd})_i = 0.512454\text{-}0.512687$ and appear similar to Cretaceous southern African kimberlites. Therefore, one of the key objectives of this study will entail collecting and analysing the newly collected Sr-Nd-Pb isotope data from the Tongo dike-01 to bring more enlightenment on the behaviour of kimberlites on the Man craton.

2.2.1 Geochemistry of West African kimberlites

A total number of 40 samples from 3 different localities were used to compile variations diagrams that look at the geochemical behaviour of West African diamondiferous rocks. Fifteen samples (kimberlite with transitional features similar to the Aries olivine lamproite of Kimberly Australia) were from the Koidu region in Sierra Leone (sample data from Taylor et., 1994), 18 samples

(kimberlite) from the Camp Alpha region in north west Liberia (sample data from Haggerty, 2017) and 10 samples (kimberlite and olivine lamproite hybrid) were from the Weasua region in Liberia (sample data from Skinner et al., 2004). This represents all geochemical analyses on the West African kimberlites to date that are available in the published literature.

The geochemistry of West African kimberlites has been shown to be influenced by crustal contamination and contamination by ilmenite macrocrysts (e.g., Taylor et al., 1994). To measure the degree of contamination and the extent of weathering, a contamination index (C.I.) was developed by Roger Clement for Kimberley kimberlites and applied to the West African kimberlites by Taylor et al. (1994):
$$C.I. = \frac{SiO_2 + Al_2O_3 + Na_2O}{2K_2O + MgO}$$
. Samples that reach values higher than C.I. = 1.5 are considered as “highly contaminated” samples (Clement, 1982). The degree of Fe-Ti-oxide (ilmenite) macrocryst accumulation can also be constrained using an Ilmenite Index:
$$Ilm.I = \frac{FeO^* + TiO_2}{2K_2O + MgO}$$
. Samples with concentrations greater than Ilm.I = 0.8 are considered contaminated by ilmenite macrocrysts (Taylor et al., 1994).

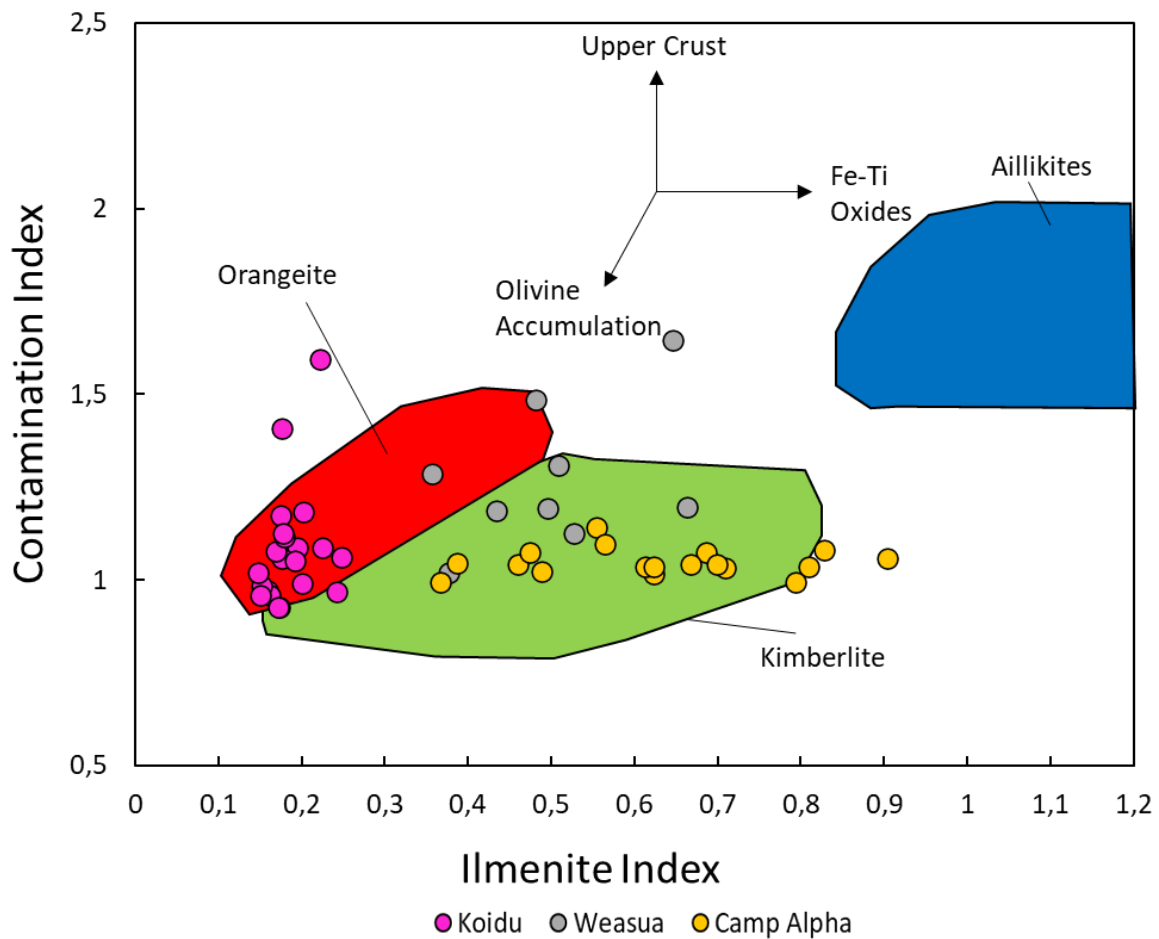


Fig. 9. Contamination index (Clement, 1982) versus Ilmenite contamination index (Taylor et al., 1994) diagram. The green field represents kimberlite that are similar to South African kimberlites and the red orangeite field represents that of South African orangeites. The yellow dots represent samples from Camp Alpha (Liberia) (Haggerty 2017), the pink dots are samples from the Koidu region (Sierra Leone) (Taylor et al., 1994) and grey dots are samples from Weasua (Liberia) (Skinner et al., 2004). The vectors represent the directions of the different processes that can change the chemistry of the rock.

The majority of the reported samples from the Koidu region are uncontaminated and unaltered ($\text{Ilm.I} < 0.52$ and $\text{C.I.} < 1.5$) and plot in the orangeite field, however, there are 2 samples that are outliers as a result of abundant granitic fragments (Figure 9). Samples from Liberia show traces of both crustal and ilmenite contamination but are mostly affected by xenocrystic ilmenite contamination resulting in low $\text{Mg\#}$ and variable ilmenite index values (Figure 9). Liberian samples unlike those from Sierra Leone are non-micaceous similar to archetypal kimberlites and lie in the kimberlite (i.e., archetypal kimberlite) field (Figure 9).

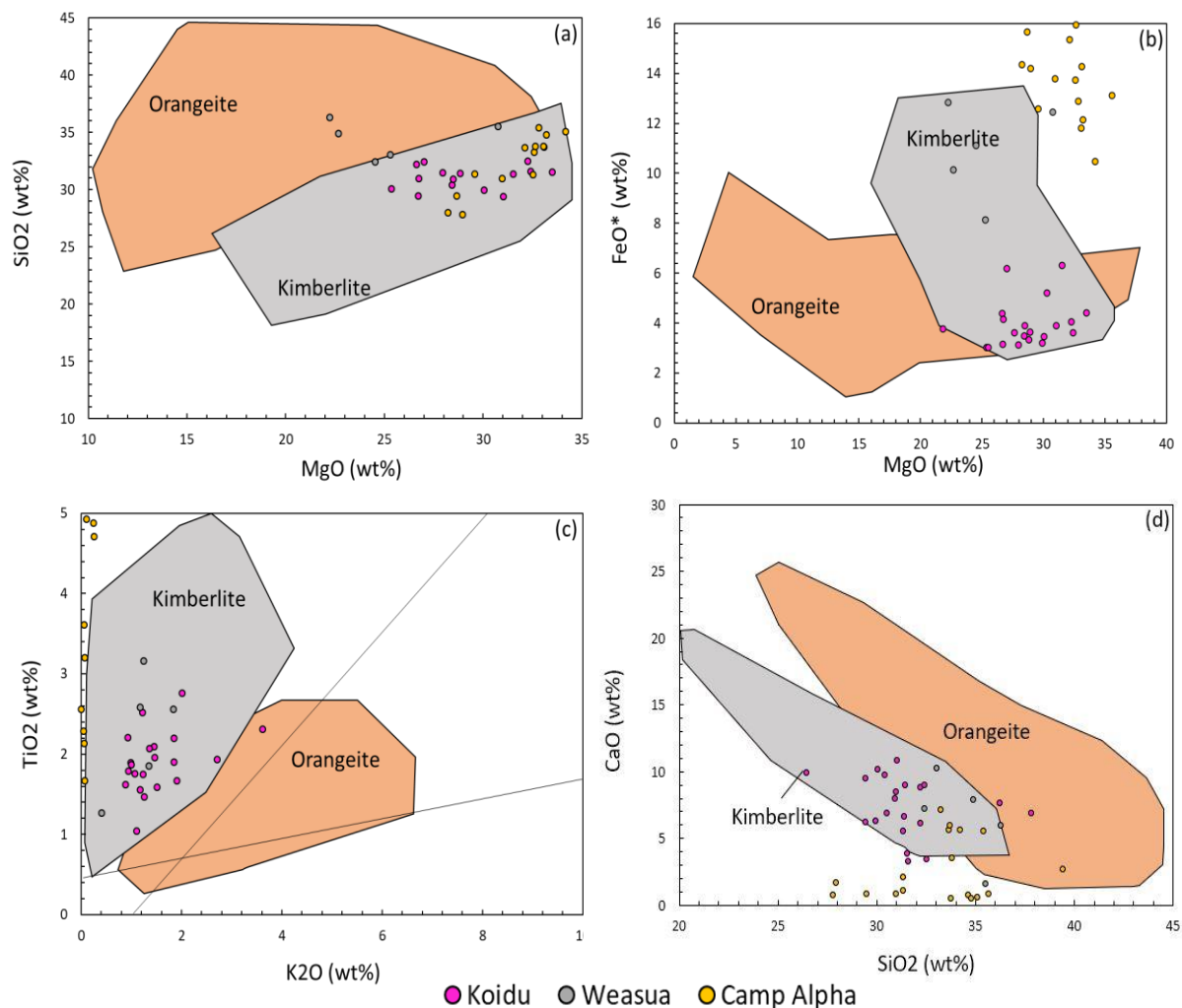


Fig. 10. Variation of (a) MgO vs SiO₂, (b) MgO* vs FeO*, (c) K₂O vs Ni, (d) SiO₂ vs CaO for West African diamondiferous rock. The yellow dots represent samples from Camp Alpha (Liberia) (Haggerty 2017), the pink dots are samples from the Koidu region (Sierra Leone) (Taylor et al., 1994) and grey dots are samples from Weasua (Liberia) (Skinner et al., 2004). The light grey field represents kimberlite compiled using le Roex et al. (2003) and Becker and le Roux (2006) data. The tan field represents orangeite compiled using Becker and le Roux (2006) Coe et al. (2008) and Howarth et al. (2011) data.

Diamondiferous rocks found in West Africa display a small compositional range for a majority of the major elements and do not seem to show any affiliation to one group, they display both kimberlite and orangeite characteristics (Figure 10). The Mg# (avg. Mg# = 85) are high due to high modal abundances of olivine in the groundmass (Taylor et al. 1994). The West African diamondiferous rocks, in particular Koidu, contain abundant phlogopite and hence are K₂O-rich but yet these rocks are not as rich in K₂O as orangeites (Figure 10c).

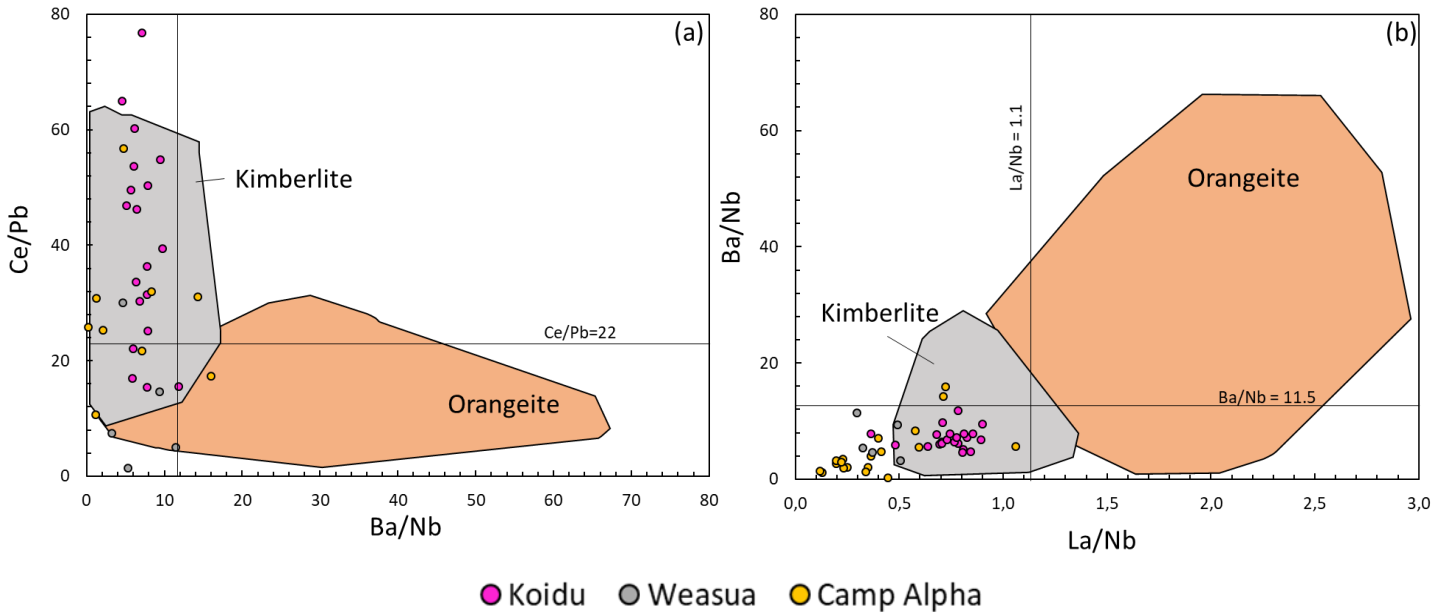


Fig. 11. Variation diagrams of (a) Ba/Nb vs Ce/Pb (b) La/Nb vs Ba/Nb for West African diamondiferous rock. The yellow dots represent samples from Camp Alpha (Liberia) (Haggerty 2017), the pink dots are samples from the Koidu region (Sierra Leone) (Taylor et al., 1994) and grey dots are samples from Weasua (Liberia) (Skinner et al., 2004). The light grey field represents kimberlite compiled using le Roex et al. (2003) and Becker and le Roux (2006) data. The tan field represents orangeite compiled using Becker and le Roux (2006) Coe et al. (2008) and Howarth et al. (2011) data.

Becker and le Roex (2006) showed that the one of the most robust ways to classify kimberlite/orangeite is using trace elements ratios (e.g., Ce/Pb, Ba/Nb, La/Nb and La/Sm). The trace element ratio data of West African samples are similar to those that define archetypal kimberlite (Figure 11a), but in figure 11b most of the Liberian samples (Weasua and Camp Alpha) have very low La/Nb ratios

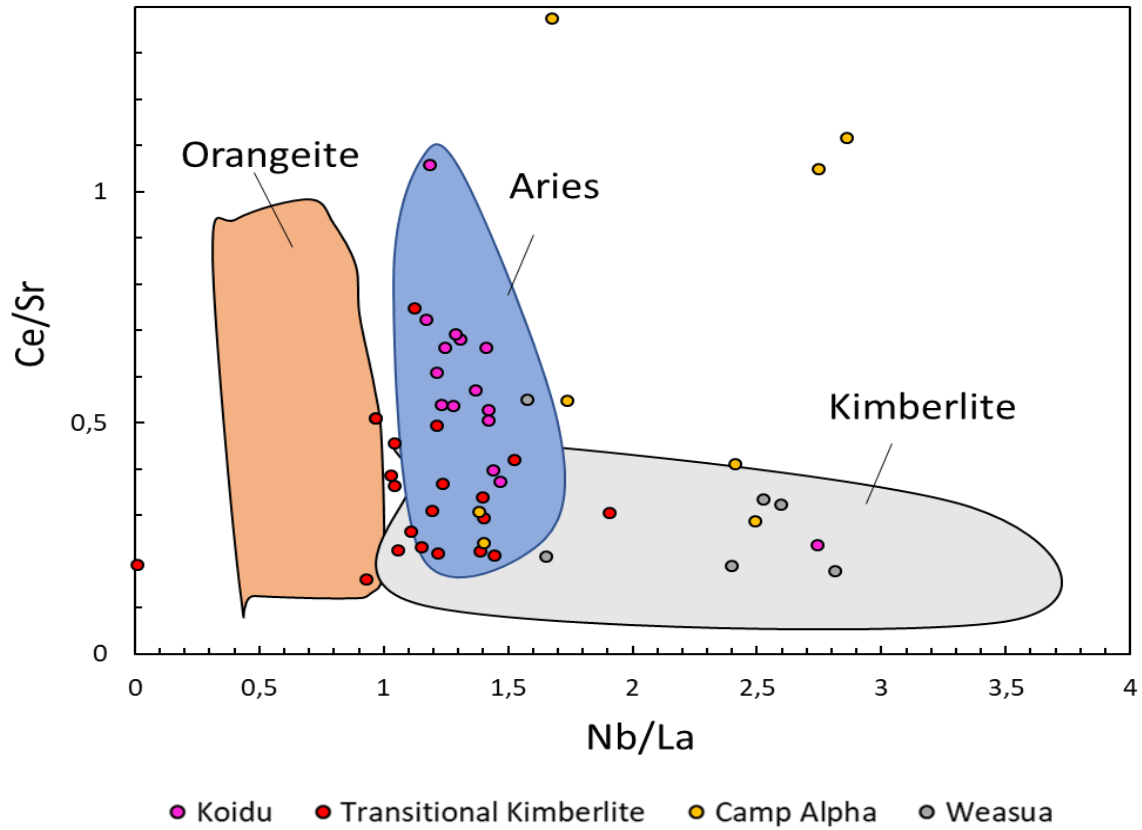


Fig. 12. Trace element covariation plot for Ce/Sr vs. Nb/La of West African diamondiferous rock. The yellow dots represent samples from Camp Alpha (Liberia) (Haggerty 2017), the pink dots are samples from the Koidu region (Sierra Leone) (Taylor et al., 1994) and grey dots are samples from Weasua (Liberia) (Skinner et al., 2004). The light grey field represents kimberlite compiled using le Roex et al. (2003) and Becker and le Roux (2006) data. The tan field represents orangeite compiled using Becker and le Roux (2006) and Howarth et al. (2011) data. The blue field represents Koidu-Aries kimberlites compiled using Taylor et al. (1994) data.

The majority of the Koidu samples share Nb/La and Ce/Sr trace element ratios that define the Aries lamproite whereas Camp Alpha and Weasua samples fall between the kimberlite and the Aries field (Figure 12). What all of these data shows us is that West African kimberlites cannot be simply classified as either kimberlite or orangeite. They are of an intermediate nature that display both kimberlite and orangeite/lamproite characteristics but leaning more towards kimberlite. To get a better understanding of these group of rocks the production and analysis of radiogenic isotopic data would be necessary, which currently does not exist.

3. Sample preparation and analytical techniques

Bulk-rock geochemical analyses for all 14 samples, which were obtained from Stellar diamonds now known as Newfield ltd, were prepared and analysed at the University of Cape Town. To avoid contamination, a hydraulic splitter was used to break down the samples so that weathered surfaces could be removed and broken down into smaller fragments. The smaller fragments produced, were then put into a jaw crusher to form small chips of rock. From these small chips of rocks, fresh kimberlite fragments that did not contain calcite veins, mantle or crustal xenoliths, were selectively handpicked in order to be powdered. The fresh samples were turned into powder using a Siebtechnik carbon steel swing mill, producing between 50 to 75g of powder.

3.1 X-Ray Fluorescence (XRF) analysis

The major element concentrations were measured by using an X-Ray Fluorescence spectrometer (XRF). In order to measure the major elements, the powdered samples had to be prepared into fused disks. Preparation went as follows; 2g of powdered sample was placed into a ceramic crucible and dried overnight in an oven at 110° C. Once dry, the samples were weighed again in order to calculate the amount of water content adsorbed (wt. % H₂O), and once calculated, the dried powders were then placed back into the oven at 900° C for a minimum of four hours in order to extract all extractable volatiles such as CO₂ and H₂O, and also oxidize ferrous iron to ferric iron. Once four hours had passed, the samples were placed into a desiccator for cooling and once cooled the powders were weighed again to determine the amount of weight percentage lost on ignition.

In order to prepare the fused disks, 0.7g of prepared powdered sample was added to 6g of Claisse XRF flux. This powder mixture underwent melting and was homogenised in the Claisse M4 Fluxer. The Claisse M4 Fluxer had to melt and mix the powder mixture fourteen times in order to make the required homogenous melt, which had to be poured into a hot platinum mould. Once the melt had solidified and cooled it was then removed from the mould producing fused disks ready to be placed in the XRF.

The Panalytical Axios XRF spectrometer, was calibrated using international certified standards and were prepared in the same manner as the samples being analysed.

3.2 Inductively Coupled Plasma-Mass Spectrometry (ICP-MS) Analysis

The 14 samples plus the SARM-39 standard, were analysed for a suite of trace element (e.g., Li, Sc, V, Cr, Co, Ni, Cu, etc.) concentrations using the Thermo-Fisher X-Series II quadrupole ICP-MS within the geological sciences department of the University of Cape Town. To prepare the samples to be analysed, 0.05g of crushed sample powder is weighed out and placed into a 3ml Teflon vials in which 2ml of concentrated hydrofluoric acid (HF) and nitric acid (HNO₃) are added with a mixture of 3:1 was added. The Teflon vial was closed and moved into Teflon capsule of a steel Parr digestion vessel. In order to decrease the pressure difference between the outer Teflon and inner Teflon, 2ml of very pure water was poured into the outer vessel. The vessels were firmly sealed and put into a 200° C oven for a period of 24 hours to allow for complete digestion. After 24 hours, the 3ml ml Teflon vial was expelled from the capsule and the cap was cautiously removed and was put on a hot plate of 100° C to allow the acid mixture to evaporate, followed by the addition of 2ml of nitric acid (HNO₃). Once acid had

completely evaporated, the samples were mixed with a nitric acid solution (5%), then poured for analysis into 15ml centrifuge tubes.

3.3 Sr-Nd-Pb isotope analysis

3.3.1 Separation Chemistry

4ml of conc. HF:HNO₃ (4:1 ratio) was added to 50mg of sample and placed in 7ml Teflon beakers. The beakers were then closed and left to digest for 48 hours on a hot plate at 140° C. Once 48 hours had passed, the beakers were opened and left to dry at the same heating temperature (140° C). The left-over residue in the beaker was then mixed with 2ml of 65% HNO₃ and left to dry. Once cooled, the newly formed solution was dissolved in 1.7ml of 2M HNO₃ for Sr and Nd elemental separation chemistry.

For Pb preparation, samples were mixed with 5% HNO₃ into 50ml centrifuge tubes to about 25g. 500µL of the prepared solution was then pipetted into a set of 15ml centrifuge tubes followed by the addition of 9.5ml of 5% HNO₃ for further dilution.

To separate Sr and Nd, the chemistry of Mikova and Denkova (2007) was applied. In order to run Sr.Spec and TransUranic-element Specific (Tru.Spec) a rack of vials were set up for each Spec. Both vial racks had waste beakers placed underneath each column. 2 x 1ml of 2M HNO₃ was used to condition the columns before use. Then the Sr.Spec rack was placed above the Tru.Spec rack followed by the addition of 1ml of 2M HNO₃ to condition the new set up. Once conditioned, 1.45ml of prepared sample was loaded. After loading the samples, 0.5ml of 2M HNO₃ was added to rinse the columns and then 4 x 0.5ml of 2M HNO₃ to wash the columns. After separation, a further 2 x 0.5ml of 2M HNO₃ was used to wash the

columns again. Waste from the Sr.Spec was collected in the waste beakers that sat underneath the columns and once collected, the beakers were replaced with clean marked 7ml Teflon beakers in which Sr was collected with 6 x 0.5ml of MQ water. After separation, the collected Sr was left to dry and cool and once cooled 2ml of 0.2% HNO_3 is added to newly formed product before being placed under ultrasonication for 30 minutes.

0.25ml of 0.05M HNO_3 was used to rinse the Tru.Spec columns. To set up for Ln.Spec, a new set vials were set up. 2 x 1ml of 0.05M HNO_3 was used to condition the new rack of vials. The newly formed set up was then washed with 6 x 0.5ml of 0.05M HNO_3 . Once the Ln.Spec columns were ready for collection, the columns were rinsed with 2 x 0.5ml of 0.05M HNO_3 and then further rinsed with 8ml of 0.25M HCl. Waste from the Ln.Spec was collected in the waste beakers that sat underneath the columns and once collected, the beakers were replaced with clean marked 15ml Teflon beakers in which Nd was collected with 7ml of 0.25M HCl. After separation, the collected Nd was left to dry and then converted to nitrate using 1ml 2B HNO_3 . This process was repeated, and the collected fraction was left to cool before the addition of 2ml of 2% HNO_3 . The newly formed product was then placed under ultrasonication for 30 minutes.

For Pb separation, Savillex Teflon columns were utilised. These columns were first cut into 2cm lengths before fitted with frits that had a thickness of 2.5mm and a porosity of 30 μm . To optimise flow the frits were made wet with methanol. After filling the columns with 0.05M HNO_3 , they were left to drip out and then the bed of the column was filled with Sr.Spec resin (Eichrom) mixed with a few millilitres of 0.05M HNO_3 . To clean and condition the columns three cycles of 10ml of 6M HCl and 10 ml of 0.05M HNO_3 were ran through the column. Two millilitres of each sample were then mixed with 1M HNO_3 and then centrifuged at 4000rpm for

twenty minutes to remove any solid residues. This newly formed solution was then loaded in two 0.9ml aliquots into the columns. After setting up the columns, the columns were rinsed with 0.5ml 1M HNO₃ followed by 2 x 1ml of 7M HNO₃ to get rid of any unwanted elements such as barium. The Pb fractions were then collected into a Teflon beaker with 4ml of 6M HCl. Once collected the Pb fractions were set on a 140°C hotplate to dry. Once dry, Pb fraction underwent two nitric conversions. A few drops of nitric acid were added into the beakers and left to dry on hotplate, this process was repeated twice. After nitric conversion, 1ml of 2% HNO₃ was added and ultrasonicated for thirty minutes. The Pb fractions were then ready to be placed in the MC-ICP-MS lab for analysis.

3.3.2 Multicollector Inductively Coupled Plasma-Mass Spectrometry (MC-ICP-MS)

Analysis

Isotope analysis was conducted using the NuPlasma HR Multicollector Inductively Coupled Plasma-Mass Spectrometer (HR MC-ICP-MS) housed within the Department of Geological Sciences of the University of Cape Town. Sr analysis takes place in a 200ppb 0.2% HNO₃ solution with NIST SRM987 used as the reference standard. The normalizing value of the standard is (⁸⁷Sr/⁸⁶Sr) = 0.710255. Rb interference was corrected using the natural (⁸⁵Rb/⁸⁷Rb) ratio and the measured signal for ⁸⁵Rb, and instrumental mass fractionation that uses exponential law and an (⁸⁶Sr/⁸⁸Sr) value of 0.1194. Nd analysis takes place in a 50ppb 2% HNO₃ solution with JNdi-1 used as the reference standard in a Nu Instruments DSN-100 desolvating nebuliser. The normalizing value of the standard is (¹⁴⁴Nd/¹⁴⁶Nd) = 0.512115. Ce and Sm interferences were corrected using natural Ce and Sm isotope concentrations and the measured signal from ¹⁴⁷Sm and ¹⁴⁰Ce, and instrumental mass fractionation was corrected using the exponential law and a (¹⁴⁶Nd/¹⁴⁴Nd) value of 0.7219.

Lead isotope ratios are measured in a 50 ppb 2% HNO₃ solution, using a NIST SRM 981 standard with reference values for (²⁰⁶Pb/²⁰⁴Pb) and (²⁰⁷Pb/²⁰⁴Pb) of 16.9405 and 15.4963 (Galer and Abouchami, 1998). To produce a 10:1 Pb:Ti ratio NIST SR M997 (the standard) was added to all samples and standards. To correct isobaric Hg interference, every hour an on-peak background measurement of 30s in the running acid solution was performed and each analysis was made up of sixty 10s integrations. To correct the Pb isotope ratios for instrumental mass fractionation, ²⁰⁵Tl/²⁰³Tl valued at 2.3889 and the exponential law were used.

4. Results

4.1 Petrography

Table 1: Tongo dike-01 samples divided by mineralogical texture.

Macrocrystic phlogopite kimberlite	Aphanitic kimberlite	Aphanitic phlogopite kimberlite
TP0202	TP0101	TP0203
TP0306	TP0204	TP0305
TP0509	TP0408	TP0307
TP0510		TP1115
TP0812		
TP1013		
TP1014		

The Tongo dike-01 samples can be divided into two main textural rock types, macrocrystic and aphanitic, but the aphanitic samples can be further broken down into olivine-rich aphanitic and phlogopite-rich aphanitic based on the mineralogy of the groundmass (Figures 13, 14 and 15). Macrocrystic samples are samples that predominantly possess large anhedral olivine grains ($>0.5\text{mm}$) that originate from mantle xenoliths and if the observed specimen has >10 vol.% macrocrysts then it is classified as macrocrystic and aphanitic samples would-be those that possess <10 vol.% of macrocrysts (e.g., Clement et al., 1985; le Roex et al., 2003).

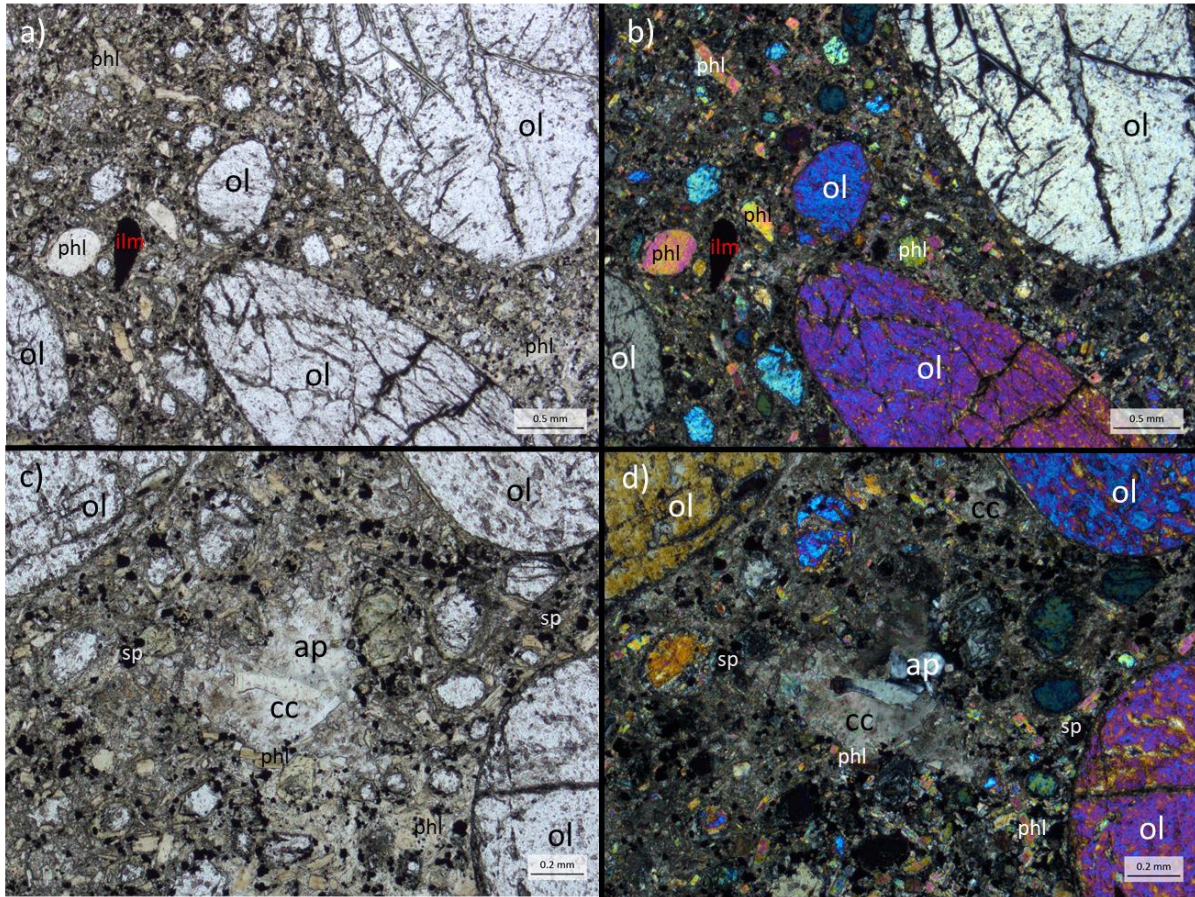


Fig. 13. Photomicrographs of representative samples from the Tongo dike-01. Figures a), b), c) and d) display a macrocrystic sample (TP0202). a) Plain Polarized Light (PPL) photomicrograph of TP0202 displaying the size and abundance of the olivine macrocrysts and b) Cross Polarized Light (XPL) photomicrograph of the same image as figure a). Figures c) (PPL) and d) (XPL) are a close up of the macrocrystic sample displaying the groundmass phases present. Key: ol - olivine; phl – phlogopite; ilm – ilmenite; sp – spinel; ap – apatite; cc – calcium carbonate (calcite); sep – serpentine; pry– perovskite.

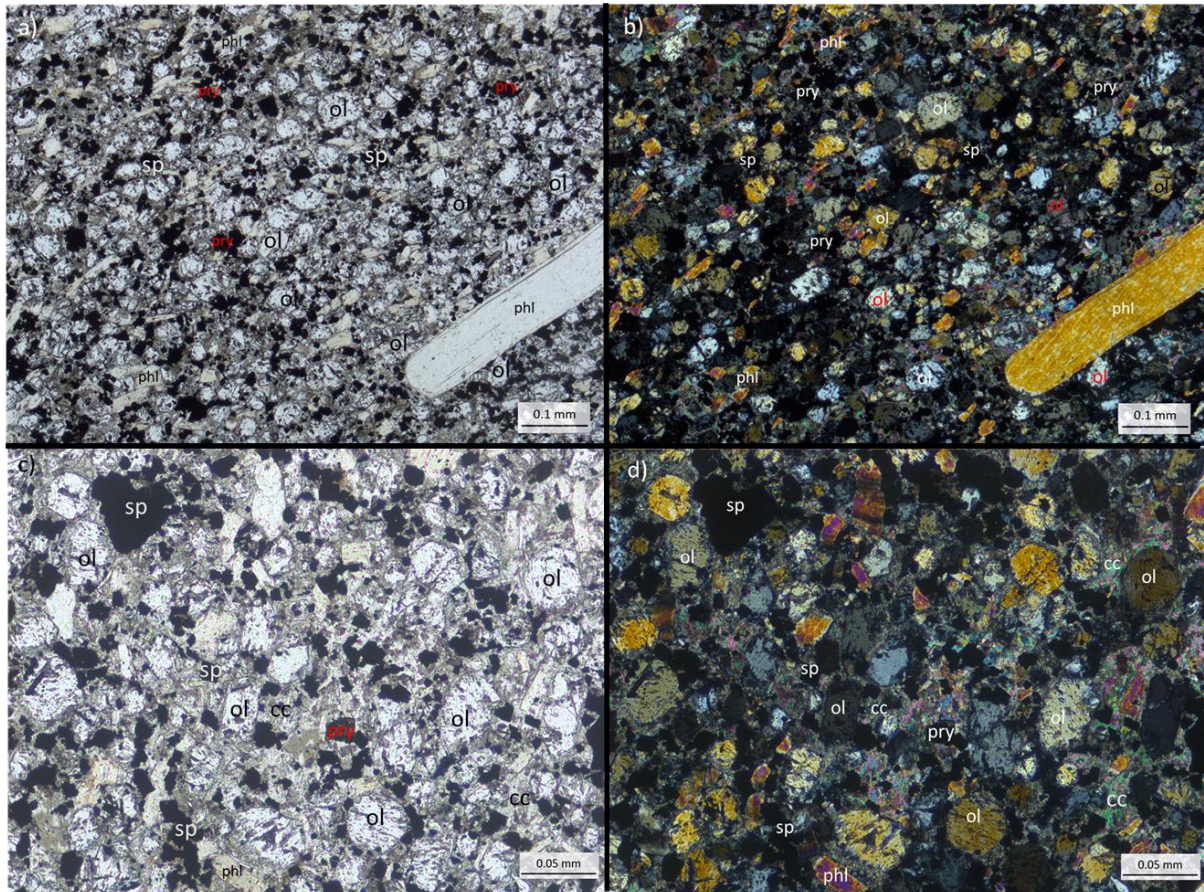


Fig. 14. Photomicrographs of representative samples from the Tongo dike-01. Figures a), b), c) and d) display an olivine-rich aphanitic sample (TP0101). Figure a) (PPL) and b) (XPL) display the abundance of the olivine and phlogopite microcrysts, whereas figures c) (PPL) and d) (XPL) are a close up of the olivine-rich aphanitic sample displaying the groundmass phases present. Key: ol - olivine; phl – phlogopite; ilm – ilmenite; sp – spinel; ap – apatite; cc – calcium carbonate (calcite); sep – serpentine; pry – perovskite; PPL - Plain Polarized Light; XPL - Cross Polarized Light.

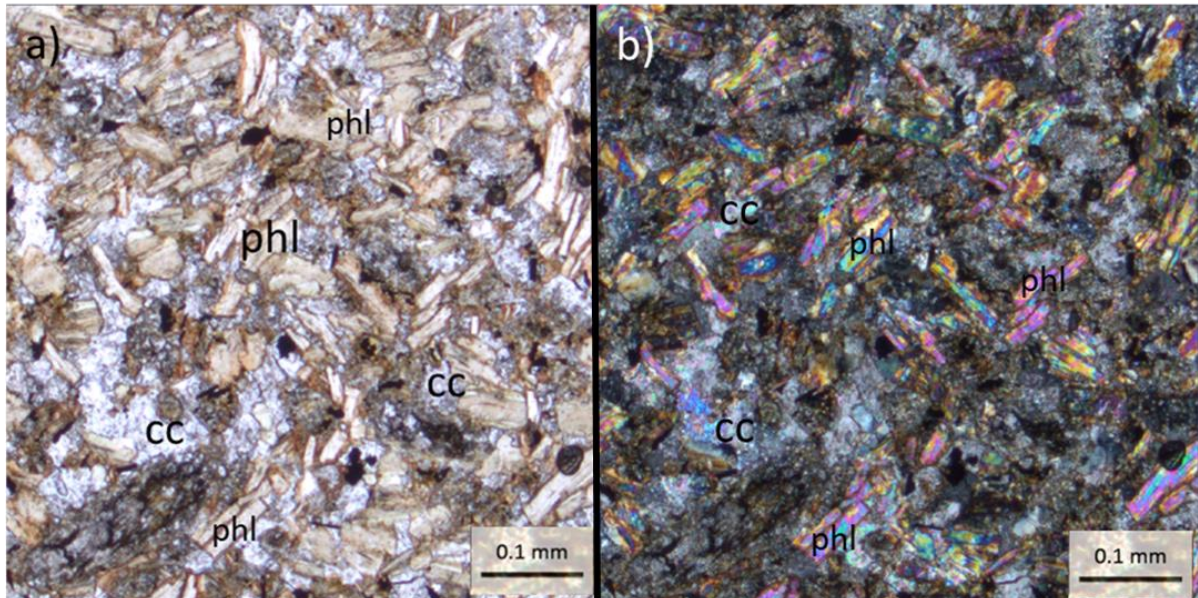


Fig. 15. Photomicrographs of representative samples from the Tongo dike-01. Figures a) and b) display a phlogopite-rich aphanitic sample (TP0101). Figure a) (PPL) and b) (XPL) display the abundance and size of the phlogopite microcrysts and the interstitial calcium carbonate that holds the phlogopite in place. Key: ol – olivine; phl – phlogopite; ilm – ilmenite; sp – spinel; ap – apatite; cc – calcium carbonate (calcite); sep – serpentine; pry – perovskite; PPL – Plain Polarized Light; XPL – Cross Polarized Light.

4.1.1 Macrocrystic samples (Figure 13)

A total of 14 samples were observed and seven of those samples were macrocrystic (Table 1). Sample TP0202 is a good representative of a macrocrystic (>10 vol.% macrocrysts) sample from the Tongo dike-01 and is described in detail here (Figure 13a, b, c and d). The specimen contains rounded to sub-rounded macrocrysts (>0.5mm) of olivine up to 10mm in size, set in a phlogopite-rich groundmass. The olivine macrocrysts occupy ~35 vol.%, display undulose extinction and also show minor evidence of serpentinisation (low temperature alteration) but the majority remain unaltered. Phlogopite macrocrysts are also present with a modal percentage of ~5%; these grains are predominantly lath shaped with rounded to sub-rounded edges, they display undulose extinction and their size ranges from 5mm to 10mm. Trace amounts of rounded ilmenite macrocrysts (~0.5mm) are observed. The groundmass (~30-40 vol.%) is composed of subhedral olivine microcrysts (<0.5mm), lath shaped phlogopite (<0.1mm), subhedral to euhedral spinel (<0.1mm), apatite laths (<0.4mm) and trace amounts of perovskite (<0.1mm). Interstitial to all of these phases are calcite and serpentine.

4.1.2 Olivine-rich aphanitic samples (Figure 14)

Three of the 14 samples observed are olivine-rich aphanitic samples (Table 1). TP0101 is a good representative of an olivine-rich aphanitic (<10 vol.% macrocrysts) sample and described here in detail (Figure 14a, b, c and d). The olivine-rich aphanitic specimens share the same features as the macrocrystic samples; the only key difference is the low portion (<10 vol%) of olivine and phlogopite macrocrystic grains, but rather have subhedral olivine microcrysts that range from 0.1mm to 0.5mm in size and occupy ~50% of the total volume. The groundmass (~40 vol%) is composed of lath shaped phlogopite (<0.1mm), subhedral spinel (<0.1mm) and trace amounts of perovskite (<0.1mm). Very little to no apatite can be seen, no signs of olivine alteration (serpentinisation)

and phlogopite is less abundant (~5-8 vol%) in the groundmass compared to the macrocrystic samples even though they share similar characteristics.

4.1.3 Phlogopite-rich aphanitic samples (Figure 15)

Four of the fourteen samples observed were phlogopite-rich aphanitic samples (Table 1). TP0305 is a good representative of a phlogopite-rich aphanitic (<10 vol.% macrocrysts) specimen (Figure 15a and b). Phlogopite grains are lath shaped ~0.1 mm in size and occupy 50% of the total volume. These crystals display undulose extinction and also possess a red/brownish rim of tetraferriphlogopite. Olivine microcrysts (<0.2mm) are observed as heavily altered (serpentinized) pseudomorphs and occupy ~5 % of the total volume. Unidentified needle shaped opaque's (0.05-0.1mm) are observed and are found in trace amounts. There are subhedral shaped spinel (<0.05mm) inclusions observed within the phlogopite and are only found in trace amounts. These minerals sit in a fine-grained interstitial material (~40 vol.%) composed of mostly calcite and lesser serpentine which encloses the phlogopite and other minerals.

4.2 Bulk-Rock Geochemistry

A total of 14 samples from the Tongo dike-01 of the Tongo-Tongoma cluster in Sierra Leone, West Africa, were selected for geochemical analysis. The major and trace element data of these samples are recorded in Table 2 and 3 and Sr-Nd-Pb isotope data is recorded in table 4. By using this data, a number of variation diagrams were plotted to show the geochemistry of the Tongo rocks.

Table 2: Bulk-rock major elements from the Tongo dike-01, Sierra Leone. All concentrations are reported in (wt. %).

Major Oxide	TP0101	TP0202	TP0203	TP0204	TP0305	TP0306	TP0307
SiO ₂	25.73	30.82	25.92	27.88	31.80	32.00	32.84
TiO ₂	3.65	1.71	2.06	1.88	1.64	1.50	1.50
Al ₂ O ₃	1.67	2.53	4.81	4.75	3.92	3.81	4.83
Fe ₂ O ₃ *	15.71	10.71	11.02	11.15	9.41	9.81	8.62
MnO	0.26	0.17	0.22	0.21	0.21	0.20	0.20
MgO	31.94	33.86	20.99	20.31	17.50	23.37	18.73
CaO	7.92	7.03	14.06	13.11	14.61	8.61	10.85
Na ₂ O	0.11	0.11	0.15	0.15	0.35	0.23	0.37
K ₂ O	0.78	1.58	3.53	3.30	3.99	3.22	4.38
P ₂ O ₅	1.02	0.83	2.63	1.34	1.58	0.96	1.17
SO ₃	0.10	0.12	0.20	0.20	0.10	0.14	0.23
Cr ₂ O ₃	0.47	0.24	0.21	0.21	0.19	0.18	0.18
NiO	0.12	0.33	0.06	0.09	0.08	0.11	0.06
H ₂ O-	0.20	0.21	0.74	1.39	1.41	1.69	2.91
LOI	9.81	9.93	12.76	13.17	12.41	13.42	12.30
Sum	99.49	100.17	99.37	99.14	99.20	99.26	99.15

Table 2: Continued

Major Oxide	TP0408	TP0509	TP0510	TP0812	TP1013	TP1014	TP1115
SiO ₂	24.38	28.28	34.54	28.93	31.04	18.34	24.65
TiO ₂	1.31	1.91	1.37	1.62	1.43	3.41	1.62
Al ₂ O ₃	3.53	4.64	2.66	2.84	4.51	2.93	4.61
Fe ₂ O ₃ *	10.82	11.07	7.11	10.07	7.59	14.38	7.16
MnO	0.17	0.20	0.13	0.17	0.16	0.35	0.22
MgO	20.06	21.18	22.53	31.25	16.82	17.55	12.54
CaO	13.23	12.47	9.50	8.09	15.76	17.70	23.93
Na ₂ O	0.22	0.15	0.25	0.14	0.26	0.14	0.17
K ₂ O	3.06	3.01	2.27	1.89	3.41	1.77	3.31
P ₂ O ₅	1.79	1.31	1.19	1.07	1.86	1.53	1.24
SO ₃	0.27	0.13	0.13	0.11	0.19	0.12	0.27
Cr ₂ O ₃	0.15	0.22	0.20	0.23	0.16	0.35	0.19
NiO	0.08	0.10	0.11	0.14	0.06	0.09	0.08
H ₂ O-	1.68	1.58	1.93	0.33	1.57	0.88	0.43
LOI	18.60	13.09	15.21	12.53	14.77	19.43	18.98
Sum	99.36	99.32	99.13	99.40	99.60	98.97	99.41

Table 3: Trace element concentrations from the Tongo dike-01, Sierra Leone. All concentrations are reported in ppm.

Trace elements	TP 0101	TP 0202	TP 0203	TP 0204	TP 0305	TP 0306	TP 0307	TP 0408
Li	2.08	3.21	3.04	6.86	30.18	13.67	43.23	27.65
Sc	23.68	18.08	20.04	30.38	24.31	21.79	24.03	19.79
V	181.0	118.3	135.9	151.3	129.4	102.0	147.8	176.1
Cr	3154	1627	1350	1525	1185	1259	1242	941
Co	97.43	96.54	57.85	76.43	55.11	72.37	62.22	54.02
Ni	975	1367	488	780	535	833	540	576
Cu	81.18	66.27	51.82	37.72	66.09	49.98	54.63	75.91
Zn	117.8	71.0	80.0	91.0	79.6	77.6	82.2	67.1
Rb	46.0	84.9	142.7	155.5	174.5	130.2	158.1	122.9
Sr	1403	1241	1699	1430	1813	1394	1888	2195
Y	13.92	10.38	31.14	27.55	22.53	15.18	14.16	19.07
Zr	372	200	448	372	294	234	223	209
Nb	522	287	490	470	439	363	367	373
Cs	3.51	3.91	4.06	4.47	4.10	3.25	3.15	1.52
Ba	1570	2098	3114	2763	2181	1979	2083	2009
La	340	187	588	389	297	292	340	344
Ce	624	310	922	683	472	505	581	577
Pr	72.1	35.3	107.2	78.0	53.2	56.3	63.5	61.2
Nd	241	117	363	266	180	189	210	204
Sm	27.96	14.00	45.67	32.49	22.77	21.88	23.94	24.46
Eu	6.25	3.36	11.07	7.82	5.77	5.10	5.56	6.01
Gd	13.96	7.53	25.62	18.34	13.65	11.72	12.18	13.74
Tb	1.39	0.78	2.61	1.91	1.46	1.21	1.18	1.45
Dy	4.61	2.82	9.07	7.16	5.61	4.16	3.88	5.05
Ho	0.583	0.385	1.209	1.004	0.794	0.561	0.526	0.691
Er	1.25	0.83	2.49	2.26	1.81	1.24	1.23	1.47
Tm	0.123	0.099	0.286	0.245	0.193	0.127	0.146	0.169
Yb	0.650	0.438	1.365	1.277	0.966	0.626	0.799	0.970
Lu	0.081	0.062	0.177	0.155	0.120	0.084	0.108	0.140
Hf	9.31	4.60	10.57	10.35	6.98	5.86	6.08	5.73
Ta	27.47	12.63	20.50	19.22	15.76	15.28	17.11	13.54
Pb	7.97	6.55	10.02	8.72	31.07	10.33	19.70	11.38
Th	66.13	26.75	51.66	57.15	29.04	35.22	40.57	52.48
U	8.97	5.74	9.98	8.54	8.38	6.44	6.72	12.34
La/Sm	12.14	13.38	12.87	11.98	13.04	13.36	14.22	14.07
La/Yb	522.39	427.80	430.60	304.78	307.35	466.53	426.04	354.90
La/Nb	0.65	0.65	1.20	0.83	0.68	0.80	0.93	0.92
Ba/Nb	3.01	7.30	6.35	5.88	4.97	5.44	5.67	5.39
Ce/Pb	78.29	47.25	91.98	78.32	15.20	48.88	29.49	50.66
Nb/La	1.54	1.53	0.83	1.21	1.48	1.24	1.08	1.08
Ce/Sr	0.44	0.25	0.54	0.48	0.26	0.36	0.31	0.26

Table 3: Continued

Trace Elements	TP 0509	TP 0510	TP 0812	TP 1013	TP 1014	TP 1115
Li	6.69	85.72	11.10	64.80	49.45	35.38
Sc	23.34	18.23	17.74	19.12	30.04	28.33
V	159.1	103.0	103.8	133.0	205.3	135.6
Cr	1421	1527	1439	1018	2515	1352
Co	72.53	72.20	82.43	51.34	84.39	54.91
Ni	754	921	1143	560	805	484
Cu	38.14	76.63	75.33	68.71	76.10	61.58
Zn	92.6	56.1	63.4	72.2	120.1	81.1
Rb	124.1	116.4	101.0	153.9	94.9	167.4
Sr	1388	1637	1775	1369	1479	1637
Y	16.59	13.51	10.38	20.64	19.55	26.68
Zr	297	188	219	356	435	235
Nb	396	356	287	428	548	431
Cs	1.57	1.17	1.23	1.32	1.33	1.50
Ba	2563	1305	1848	1990	1493	2675
La	321	239	196	267	348	344
Ce	581	404	332	459	614	636
Pr	64.5	44.0	36.6	50.2	66.1	68.1
Nd	218	149	122	169	223	231
Sm	25.64	17.99	14.33	20.80	26.93	28.26
Eu	5.91	4.39	3.35	5.13	6.28	6.68
Gd	13.67	10.19	7.70	11.90	14.30	15.48
Tb	1.40	1.07	0.80	1.33	1.56	1.72
Dy	4.61	3.57	2.88	4.89	5.41	6.28
Ho	0.632	0.486	0.395	0.722	0.740	0.920
Er	1.35	1.11	0.88	1.55	1.55	1.96
Tm	0.145	0.124	0.089	0.185	0.168	0.230
Yb	0.801	0.718	0.470	0.942	0.876	1.195
Lu	0.098	0.097	0.060	0.125	0.104	0.163
Hf	7.14	4.45	5.06	9.77	10.98	7.70
Ta	19.24	13.37	12.60	14.59	23.02	17.56
Pb	11.71	6.89	8.13	8.21	14.59	13.63
Th	50.00	70.36	25.25	66.22	77.39	71.04
U	5.07	10.53	5.39	6.15	16.67	5.13
La/Sm	12.50	13.27	13.65	12.82	12.91	12.18
La/Yb	400.20	332.32	416.60	283.20	396.95	288.05
La/Nb	0.81	0.67	0.68	0.62	0.63	0.80
Ba/Nb	6.48	3.67	6.45	4.65	2.73	6.20
Ce/Pb	49.62	58.65	40.80	55.95	42.06	46.68
Nb/La	1.24	1.49	1.46	1.61	1.58	1.25
Ce/Sr	0.42	0.25	0.19	0.34	0.42	0.39

Table 4 (Rb-Sr): Isotopic compositional data for measured (m) and initial (i) Sr-Nd-Pb isotopes for the Tongo dike-01.

Sample	Rb	Sr	(⁸⁷ Sr/ ⁸⁶ Sr) _m	(⁸⁷ Sr/ ⁸⁶ Sr) _i	SD
TP0101	46.0	1403	0.70398	0.70377	0.000009
TP0202	142.7	1699	0.70442	0.70398	0.000011
TP0203	155.5	1430	0.70451	0.70397	0.000013
TP0204	122.9	2195	0.70504	0.70434	0.000011
TP0306	84.9	1241	0.70635	0.70575	0.000013
TP0408	130.2	1394	0.70483	0.70447	0.000011
TP0509	124.1	1388	0.70495	0.70438	0.000012
TP0510	116.4	1637	0.70540	0.70495	0.000014
TP0812	101.0	1775	0.70588	0.70552	0.000009

Table 4 (Nd-Sm): Continued

Sample	Nd	Sm	(¹⁴³ Nd/ ¹⁴⁴ Nd) _m	(¹⁴³ Nd/ ¹⁴⁴ Nd) _i	εNd(t)	SD
TP0101	241	27.96	0.512632	0.512573	2.2	0.000016
TP0202	363	45.67	0.512621	0.512549	1.8	0.000012
TP0203	266	32.49	0.512612	0.512556	1.9	0.000013
TP0204	204	24.46	0.512605	0.512549	1.8	0.000009
TP0306	117	14.00	0.512610	0.512540	1.6	0.000016
TP0408	189	21.88	0.512599	0.512544	1.7	0.000012
TP0509	218	25.64	0.512613	0.512553	1.8	0.000012
TP0510	149	17.99	0.512612	0.512550	1.8	0.000013
TP0812	122	14.33	0.512633	0.512573	2.2	0.000018

Table 4 (U-Pb-Th): Continued

Sample	U	Pb	Th	(²⁰⁶ Pb/ ²⁰⁴ Pb) _m	(²⁰⁶ Pb/ ²⁰⁴ Pb) _i	SD
TP0101	8.97	7.97	66.13	20.19	19.27	0.0010
TP0202	9.98	10.02	51.66	19.93	18.71	0.0009
TP0203	8.54	8.72	57.15	19.83	19.12	0.0010
TP0204	12.34	11.38	52.48	19.76	19.03	0.0012
TP0306	5.74	6.55	26.75	19.42	19.62	0.0010
TP0408	6.44	10.33	35.22	20.13	18.88	0.0012
TP0509	5.07	11.71	50.00	19.51	19.16	0.0010
TP0510	10.53	6.89	70.36	20.40	19.16	0.0009
TP0812	5.39	8.13	25.25	19.93	19.39	0.0012

Table 4: Continued

Sample	$(^{207}\text{Pb}/^{204}\text{Pb})_m$	$(^{207}\text{Pb}/^{204}\text{Pb})_i$	SD	$(^{208}\text{Pb}/^{204}\text{Pb})_m$	$(^{208}\text{Pb}/^{204}\text{Pb})_i$	SD
TP0101	15.67	15.25	0.0010	41.30	40.88	0.0033
TP0202	15.65	15.27	0.0008	39.90	39.48	0.0025
TP0203	15.63	15.28	0.0010	40.25	39.84	0.0029
TP0204	15.62	15.26	0.0011	40.47	40.05	0.0034
TP0306	15.60	15.41	0.0009	39.59	39.17	0.0031
TP0408	15.65	15.22	0.0011	40.71	40.29	0.0033
TP0509	15.61	15.45	0.0010	40.36	39.94	0.0034
TP0510	15.67	15.10	0.0008	42.31	41.90	0.0030
TP0812	15.65	15.40	0.0011	39.98	39.57	0.0035

4.2.1 Major element geochemistry

The MgO and SiO₂ concentrations of both macrocrystic and aphanitic samples do not show a correlation (Figure 16a), with MgO content varying from 13 wt. % to 34 wt. %, whereas SiO₂ ranges from 18 wt. % to 35 wt. %. The FeO contents range from 7 wt. % to 16 wt. % and also show no correlation with MgO (Figure 16b). The aphanitic samples have a higher concentration in K₂O (3.1 wt. % - 4.9 wt. %) and Al₂O₃ (2.53 wt. % - 4.83 wt. %) compared to the macrocrystic samples (1.3 wt. % - 3.8 wt. %) but have the same range in TiO₂ concentration (1.1 wt. % - 3.9 wt. %). The CaO (6 wt. % - 24 wt. %) concentrations display a wide range with no distinct correlation with SiO₂ (Figure 16d).

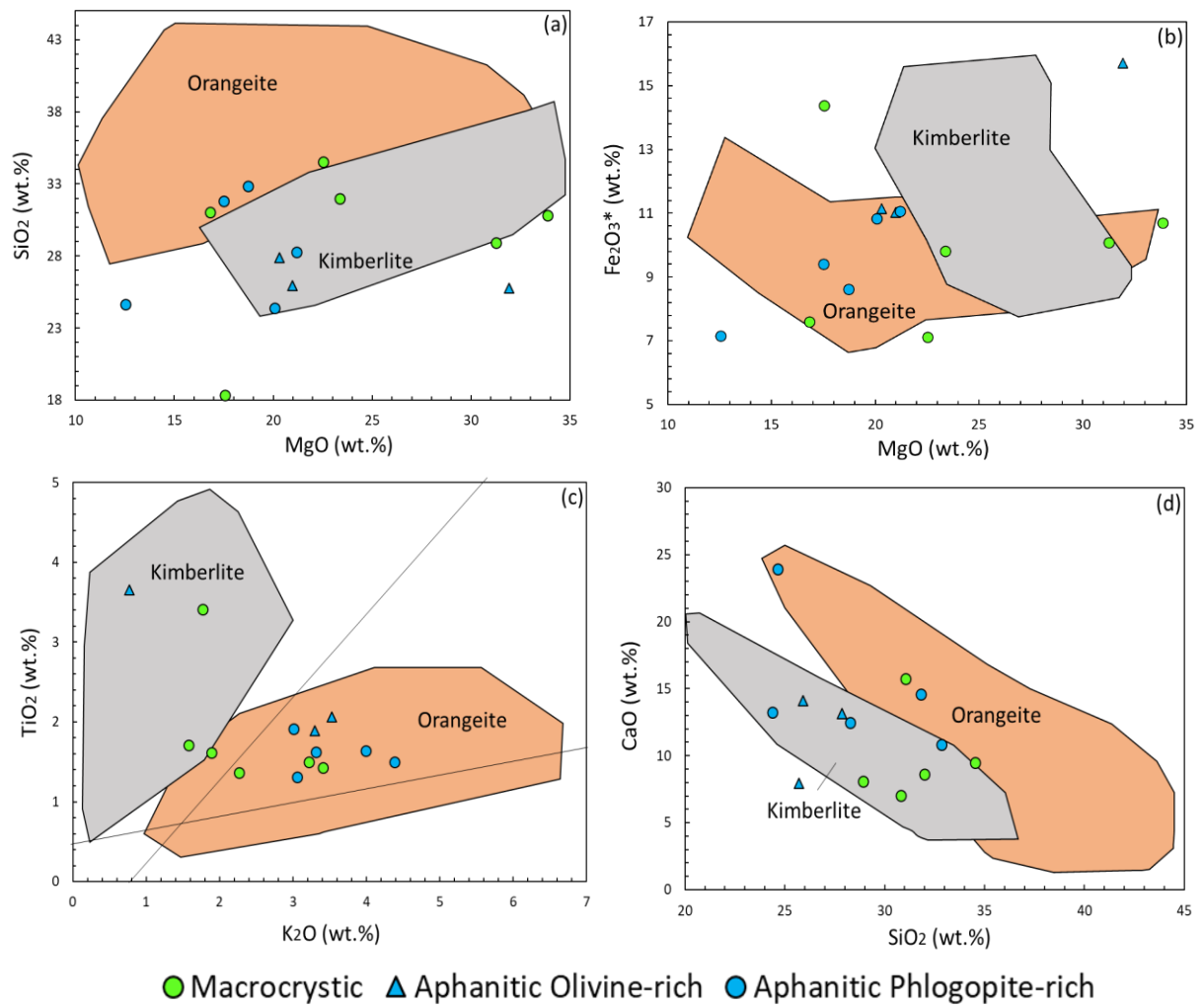


Fig. 16. Variation of (a) MgO vs SiO₂, (b) MgO* vs FeO*, (c) K₂O vs Ni, (d) SiO₂ vs CaO for Tongo dike-01. The light grey field represents kimberlite compiled using le Roex et al. (2003) and Becker and le Roex (2006) data. The tan field represents orangeites compiled using Becker and le Roex (2006) Coe et al. (2008) and Howarth et al. (2011) data.

4.2.2 Trace element geochemistry

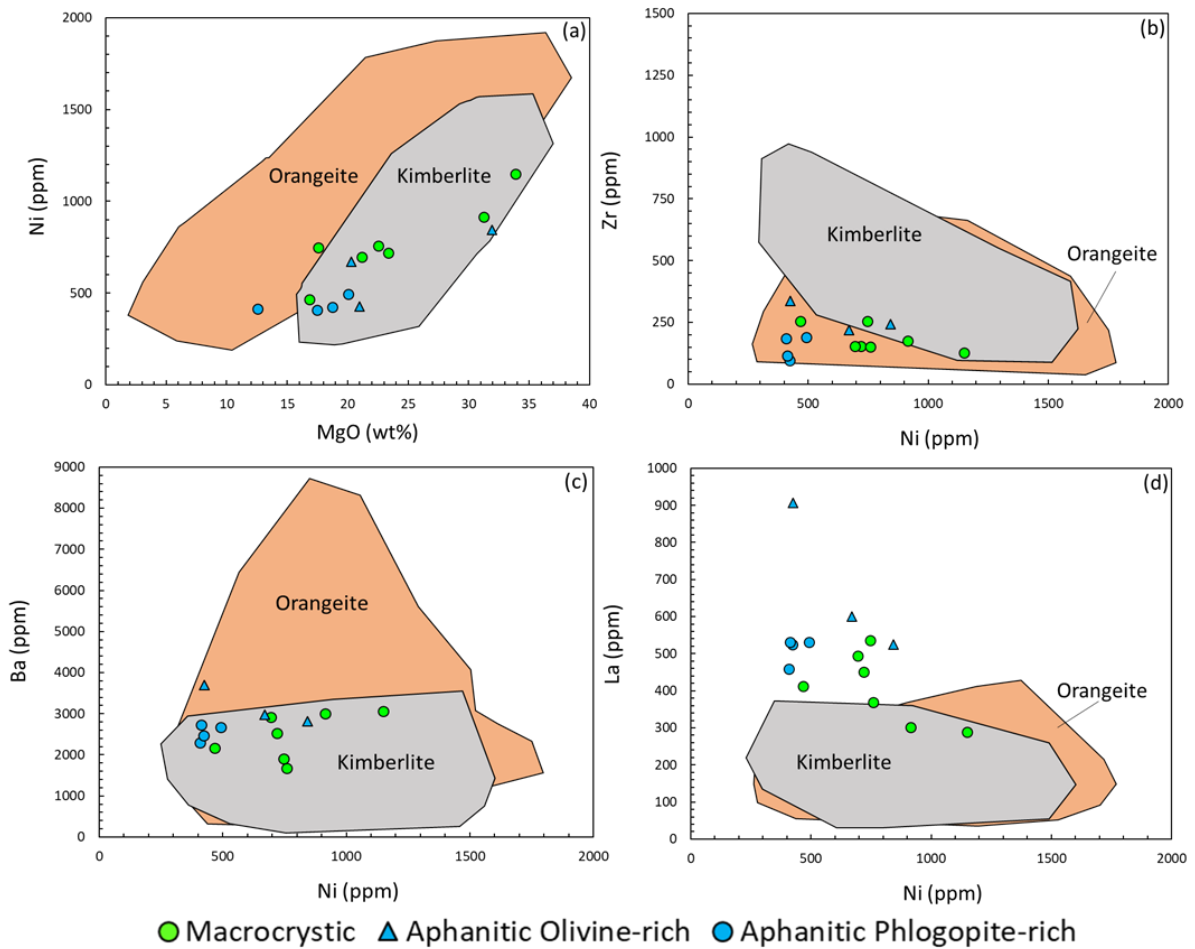


Fig. 17. Variation of (a) MgO vs Ni, (b) Ni vs Zr, (c) Ni vs Ba, (d) Ni vs La for Tongo dike-01. The light grey field represents kimberlite compiled using le Roex et al. (2003) and Becker and le Roex (2006) data. The tan field represents orangeites compiled using Becker and le Roex (2006) Coe et al. (2008) and Howarth et al. (2011) data.

Transition metals such as Ni show a positive correlation against MgO with Ni varying in concentration ranging from 340 ppm to 1200 ppm (Figure 17a). Zirconium and Ba are relatively constant at varying Ni concentrations (Figures 17b and 17c) and La has no relationship with Ni (Figure 17d). The contents of Zr vary between 100 ppm and 350 ppm, Ba 1600 ppm and 3700 ppm and La 270 ppm and 900 ppm.

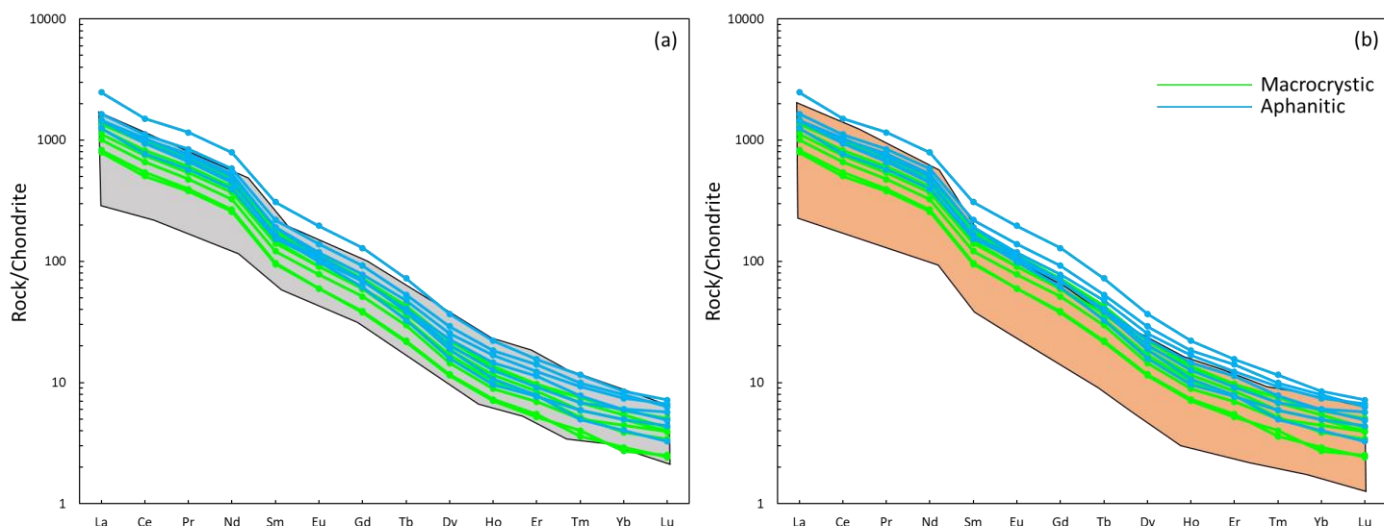


Fig. 18. Chondrite normalized rare earth elements plot for Tongo dike-01 rock samples. For clarity macrocrystic samples (blue lines) are distinguished from aphanitic samples (green lines). The light grey field in figure (a) represents kimberlite compiled using le Roex et al. (2003) and Becker and le Roux (2006) data. The tan field in figure (b) represents orangeite compiled using Becker and le Roux (2006), Coe et al. (2008) and Howarth et al. (2011) data.

The Tongo dike-01 samples are enriched in light rare earth elements (LREE) relative to heavy rare earth elements (HREE), with La concentrations ranging from ~700 to ~1500 times chondrite whereas Lu ranges from ~2 to ~7 times chondrite (Figure 18). The chondrite normalized plot shows that, the aphanitic samples are more enriched in REE compared to macrocrystic samples but show a REE pattern that is sub-parallel to the macrocrystic samples (Figure 18).

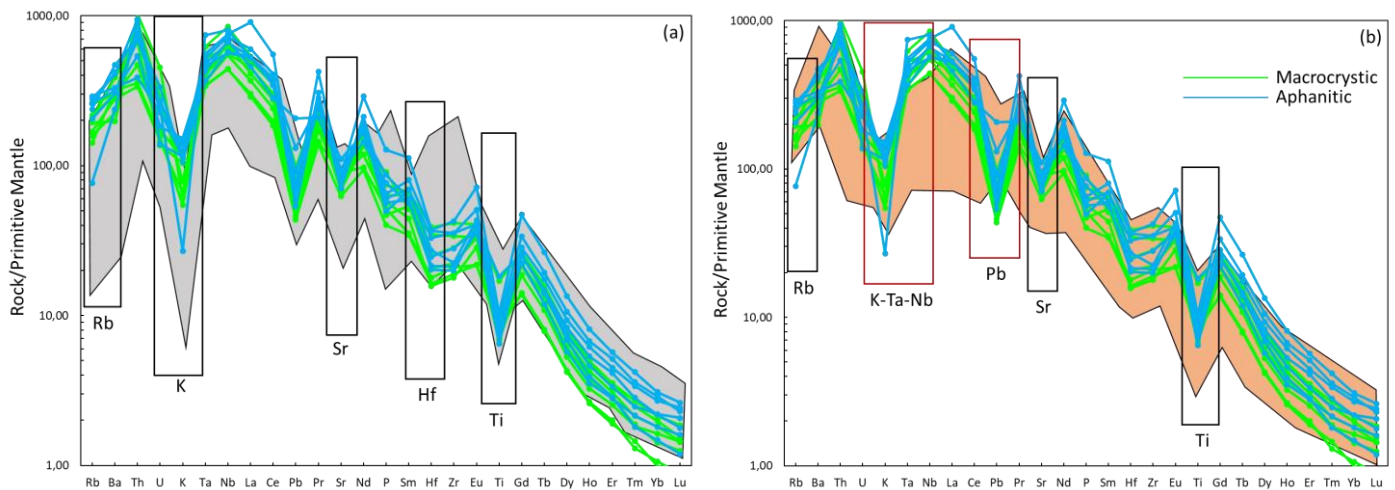


Fig. 19. Primitive mantle normalized trace element diagram for incompatible trace elements from the Tongo dike-01. For clarity macrocrystic samples (blue lines) are distinguished from aphanitic samples (green lines). The light grey field in figure (a) represents kimberlite compiled using le Roex et al. (2003) and Becker and le Roux (2006) data. The tan field in figure (b) represents orangeite compiled using Becker and le Roux (2006), Coe et al. (2008) and Howarth et al. (2011) data.

Trace elements such as Rb, K, Sr, Pb, Hf and Ti show negative anomalies relative to sub-parallel enrichment patterns (Figure 19). The primitive mantle normalized patterns show that the aphanitic samples display strong enrichment of trace elements compared to macrocrystic samples.

Trace element ratios Ce/Pb range from 15 to 79 whereas Ba/Nb ratios range from 1 to 7. The samples are well spread out along the Ce/Pb axis but not along the Ba/Nb axis (Figure 20a). Samples are clustered into a small area with La/Nb ratios ranging from 0.5 to 1.1 whereas Ba/Nb ratios range from 1 to 7 (Figure 20b). The La/Yb vs La/Sm ratios display a positive correlation, with La/Sm ratios ranging from 11 to 14 and La/Yb ratios from 270 to 480 (Figure 20c). The La/Yb vs Gd/Yb plot shows a strong positive correlation (Figure 20d).

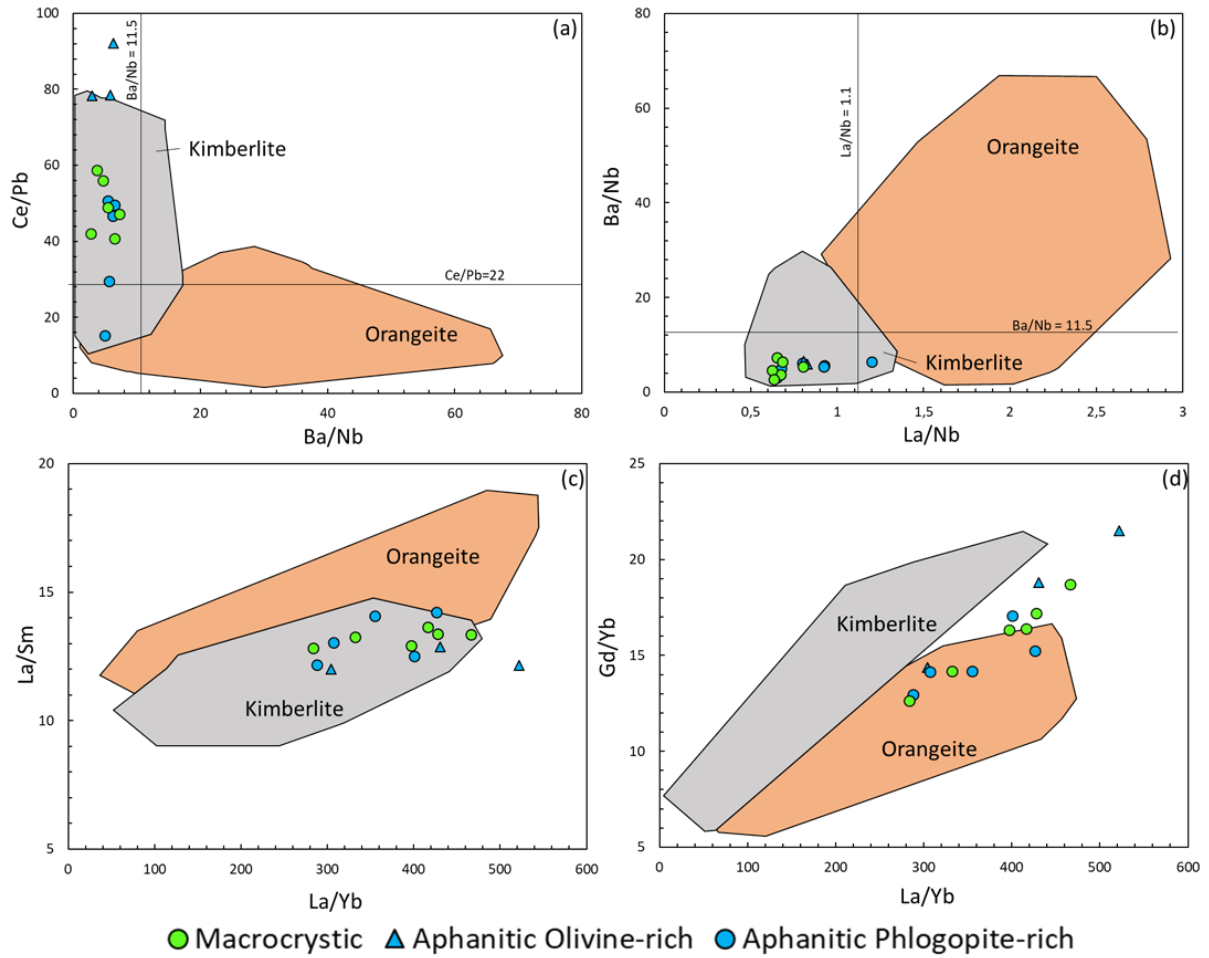


Fig. 20. Variation of (a) MgO vs Ni, (b) Ni vs Zr, (c) Ni vs Ba, (d) Ni vs La for Tongo dike-01. The light grey field represents kimberlite compiled using le Roex et al. (2003) and Becker and le Roex (2006) data. The tan field represents orangeites compiled using Becker and le Roex (2006), Coe et al. (2008) and Howarth et al. (2011) data.

4.2.3 Sr-Nd-Pb Isotope geochemistry

A total of nine out of the 14 samples were analysed for Sr-Nd-Pb isotope geochemistry. The measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for the Tongo dike-01 have a range of $0.70391\text{--}0.70443 \pm 0.000009\text{--}0.000014$ (Table 4). The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios for the Tongo dike-01 were calculated using the following equation

$$\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_m = \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_i + \left(\frac{^{87}\text{Rb}}{^{86}\text{Sr}}\right)_m (e^{\lambda t} - 1)$$

where m - is the measured ratios, i - is the

calculated initial ratios, t - the age of the Tongo dike-01 (140 Ma; Skinner et al., 2004) and $\lambda = 1.42 \times 10^{-11}/\text{yr}$. The calculated initial ratios have a range of $0.70377\text{--}0.70572$, (Figure 21a and d). The measured $^{143}\text{Nd}/^{144}\text{Nd}$ ratios for the Tongo dike-01 have a range of $0.512606\text{--}0.512638 \pm 0.00009\text{--}0.00018$ (Table 4).

The calculated initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios for the Tongo dike-01 were calculated using the isochron equation $\left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}}\right)_m = \left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}}\right)_i + \left(\frac{^{147}\text{Sm}}{^{144}\text{Nd}}\right)_m (e^{\lambda t} - 1)$ where m - is the measured ratios, i - is the calculated initial ratios, t - the age of the Tongo dike-01 (140 Ma; Skinner et al., 2004) and $\lambda = 6.54 \times 10^{-12}/\text{yr}$. The calculated initial ratios have a range of 0.512532-0.512572. The measured $^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios for the Tongo dike-01 range from 19.423-20.4001 $\pm$ 0.0009-0.012 for $^{206}\text{Pb}/^{204}\text{Pb}$, 15.608-15.667 $\pm$ 0.0008-0.0011 for $^{207}\text{Pb}/^{204}\text{Pb}$ and 39.585-40.705 $\pm$ 0.0025-0.0035 for $^{208}\text{Pb}/^{204}\text{Pb}$ (Table 4). The calculated initial $^{206}\text{Pb}/^{204}\text{Pb}$ ratios for the Tongo dike-01 were calculated using the following equation $\left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)_m = \left(\frac{^{206}\text{Pb}}{^{204}\text{Pb}}\right)_i + \left(\frac{^{238}\text{U}}{^{204}\text{Pb}}\right)_m (e^{\lambda_{238}t} - 1)$ where m - is the measured ratios, i - is the calculated initial ratios, t - the age of the Tongo dike-01 (140 Ma; Skinner et al., 2004) and $\lambda_{238} = 1.551 \times 10^{-10}/\text{yr}$. The calculated initial ratios have a range of from 18.725 to 19.486 (Figure 21b, c and d). For $^{207}\text{Pb}/^{204}\text{Pb}$ the initial ratios were calculated using the following equation $\left(\frac{^{207}\text{Pb}}{^{204}\text{Pb}}\right)_m = \left(\frac{^{207}\text{Pb}}{^{204}\text{Pb}}\right)_i + \left(\frac{^{235}\text{U}}{^{204}\text{Pb}}\right)_m (e^{\lambda_{235}t} - 1)$ where m - is the measured ratios, i - is the calculated initial ratios, t - the age of the Tongo dike-01 (140 Ma; Skinner et al., 2004) and $\lambda_{235} = 9.849 \times 10^{-10}/\text{yr}$. The calculated initial ratios range from 15.097 to 15.412 (Figure 21c). $^{208}\text{Pb}/^{204}\text{Pb}$ initial ratios were calculated using $\left(\frac{^{208}\text{Pb}}{^{204}\text{Pb}}\right)_m = \left(\frac{^{208}\text{Pb}}{^{204}\text{Pb}}\right)_i + \left(\frac{^{232}\text{Th}}{^{204}\text{Pb}}\right)_m (e^{\lambda_{232}t} - 1)$ where m - is the measured ratios, i - is the calculated initial ratios, t - the age of the Tongo dike-01 (140 Ma; Skinner et al., 2004) and $\lambda_{232} = 4.948 \times 10^{-11}/\text{yr}$. The calculated initial ratios range from 39.173 to 41.896.

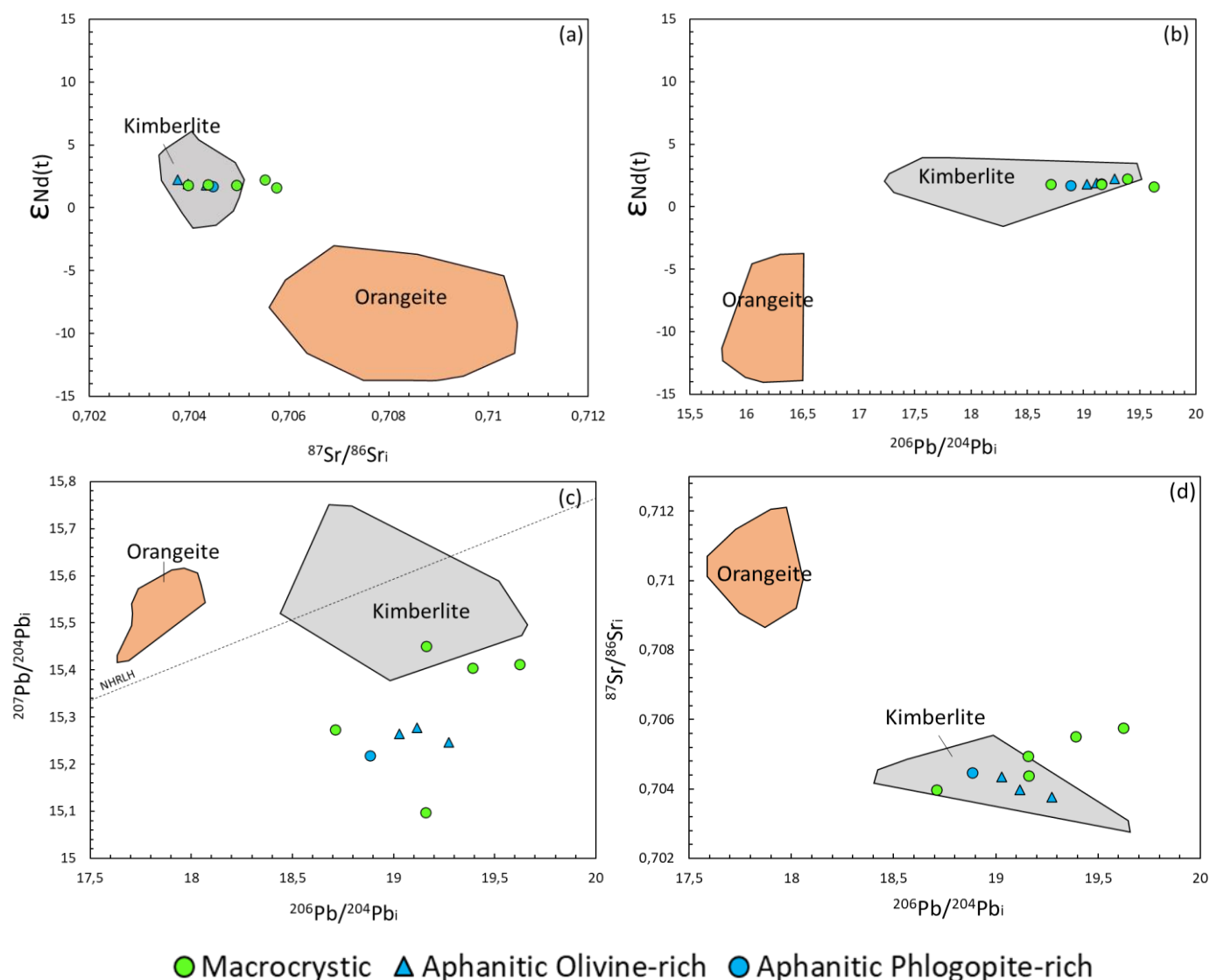


Fig. 21. Isotope ratio plots for Tongo dike-01 samples: (a) $(^{87}\text{Sr}/^{86}\text{Sr})_i$ vs $\epsilon_{\text{Nd}}(t)$, (b) $(^{206}\text{Pb}/^{204}\text{Pb})_i$ vs $\epsilon_{\text{Nd}}(t)$, (c) $(^{206}\text{Pb}/^{204}\text{Pb})_i$ vs $(^{207}\text{Pb}/^{204}\text{Pb})_i$, and (d) $(^{206}\text{Pb}/^{204}\text{Pb})_i$ vs $(^{87}\text{Sr}/^{86}\text{Sr})_i$. The light grey field represents kimberlite compiled using Nowell et al. (2004) and Becker and le Roux (2006) data. The tan field represents orangeite compiled using Fraser and Hawkesworth (1992) and Becker and le Roux (2006) data. NHRHL (Northern Hemisphere Reference Line of Hart, 1984).

5. Discussion

5.1 Alteration and crustal contamination

Due to the permeable and brecciated nature of diamondiferous rocks they are susceptible to alteration (Berg and Allsopp, 1972). These rocks experience alteration through various processes such as deuteric alteration of primary minerals (primary process), contamination by crustal xenoliths (secondary process) and alteration through movement of meteoric waters (secondary process) (Berg & Allsopp, 1972; Berg, 1994; Price et al., 2000; le Roex et al., 2003; Sparks, 2013). In addition, West African diamondiferous rocks are known for containing abundant ilmenite xenocrysts leading to contamination of bulk-rock analyses (e.g., Taylor et al., 1994; Skinner et al., 2004). Therefore, constraining the geochemical effects of ilmenite and crustal contamination is of great importance before interpretation and classification of the rocks can be made.

Crustal contamination by xenoliths during sample preparation can have a strong influence on the composition of these diamondiferous rocks (Clement, 1982). The following section deals with assessing the effects of xenolith contamination during sample preparation. To assess the degree of crustal and ilmenite xenocryst contamination and the extent of weathering, a contamination index ($C.I. = \frac{SiO_2 + Al_2O_3 + Na_2O}{2K_2O + MgO}$) and an ilmenite index ($Ilm.I = \frac{FeO^* + TiO_2}{2K_2O + MgO}$) were used. Samples that reached values higher than $C.I. = 1.5$ and $Ilm.I = 0.8$ were considered as “highly contaminated” samples (Clement, 1982; Taylor et al., 1994). Samples reported from the Camp Alpha kimberlites by Haggerty (2017), for example, have a low C.I. index indicating they are not severely crustally contaminated, however, they do have a large range in Ilm.I index suggesting variable contributions from ilmenite xenocrysts. In this case, the likely close-to-primary magma is the one with lowest Ilm.I index with the least geochemical effect of ilmenite xenocryst contamination (Figure 22).

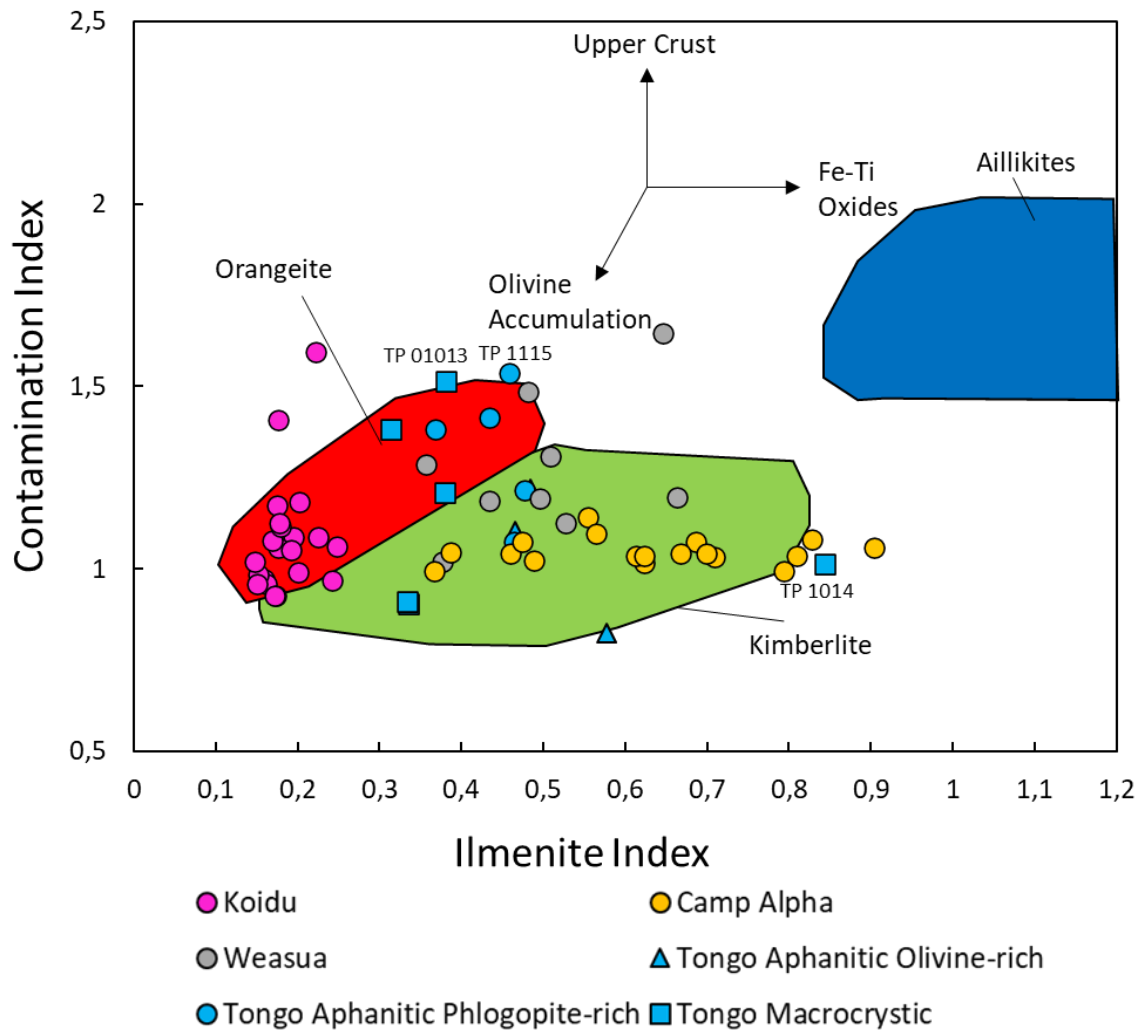


Fig. 22. Contamination index (Clement, 1982) versus Ilmenite contamination index (Taylor et al., 1994) diagram. The green field represents archetypal kimberlite, and the red field represents that of South African orangeites. The yellow dots represent samples from Camp Alpha (Liberia) (Haggerty 2017), the pink dots are samples from the Koidu region (Sierra Leone) (Taylor et al., 1994) and grey dots are samples from Weasua (Liberia) (Skinner et al., 2004). The vectors represent the directions of the different processes that can change the chemistry of the rock.

The majority of samples from the Tongo dike-01 analysed in this study are uncontaminated and unaltered ($\text{Ilm.I} < 0.55$ and $\text{C.I.} < 1.5$) and plot in the orangeite and kimberlite fields (Figure 22). The aphanitic samples have a range of $\text{Ilm.I} = 0.38\text{--}0.59$ and $\text{C.I.} = 0.70\text{--}1.45$ whereas the macrocrystic samples have a range of $\text{Ilm.I} = 0.28\text{--}0.39$ and $\text{C.I.} = 0.8\text{--}1.41$. However, there are three samples that are outliers and are likely a result of crustal and ilmenite contamination; sample TP1013 ($\text{C.I.} = 1.514$), TP 1115 ($\text{C.I.} = 1.536$) and TP1014 ($\text{Ilm.I} = 0.8433$). These samples thus do not represent true kimberlite melt compositions but

rather reflect the addition of chemical components from an external source (Figure 22).

Sample TP1013, TP1115 and TP1014 were not included in Sr-Nd-Pb isotope analyses because they contained some crustal xenoliths that were visible in the samples during preparation, consistent with their high contamination index discussed above. Thus, they were excluded from isotope analyses and are not present in figures 23 and 24 illustrating the correlation of contamination index and Sr-Nd-Pb isotopes.

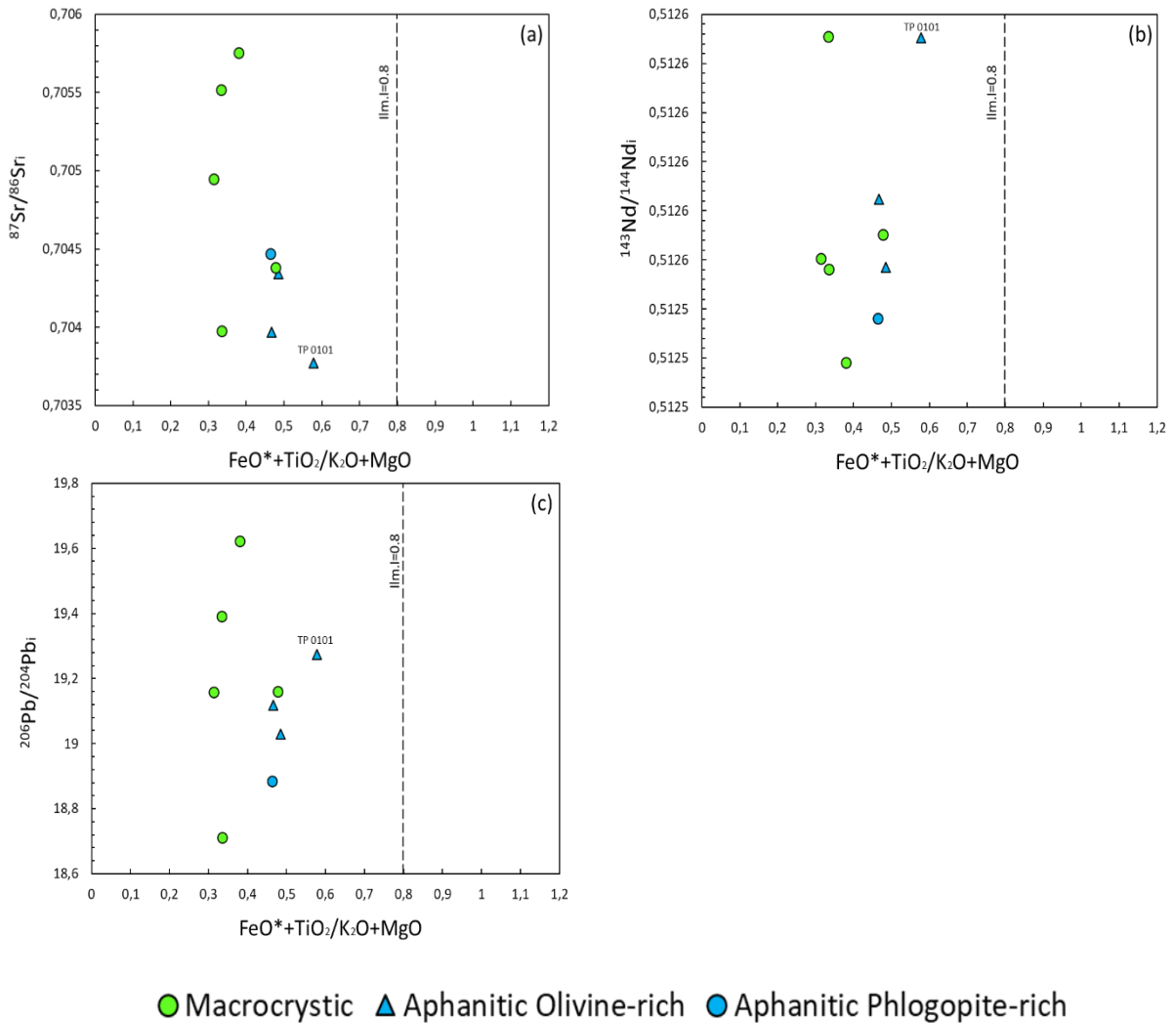


Fig. 23. Ilmenite contamination index vs isotope ratios for Tongo dike-01 samples: (a) (⁸⁷Sr/⁸⁶Sr)_i vs Ilmenite index, (b) (¹⁴³Nd/¹⁴⁴Nd)_i vs Ilmenite index and (c) (²⁰⁶Pb/²⁰⁴Pb)_i vs Ilmenite index. The dotted line represents the maximum ilmenite value a sample can have before its considered contaminated.

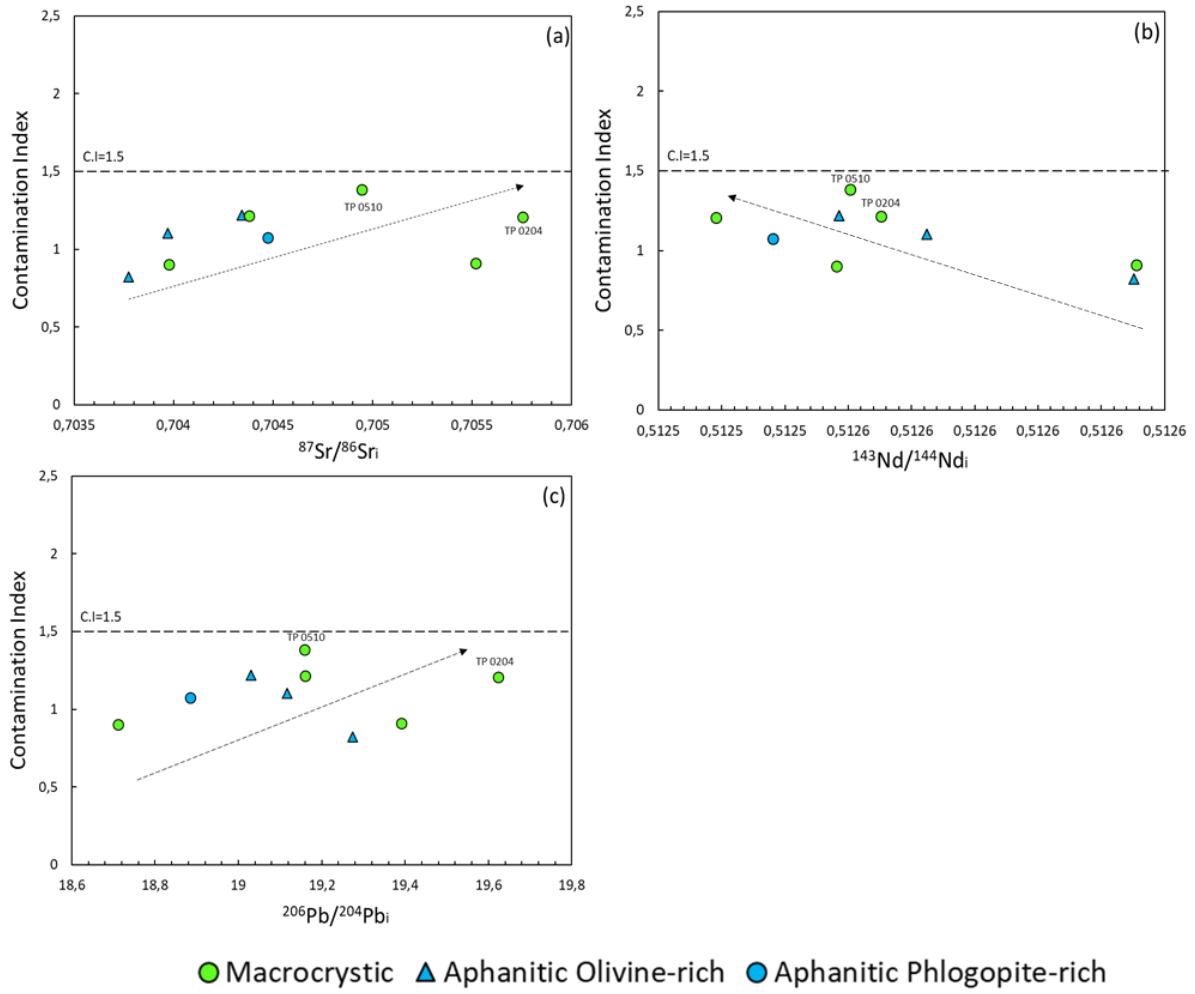


Fig. 24. Crustal contamination index vs isotope ratios for Tongo dike-01 samples: (a) $(^{87}\text{Sr}/^{86}\text{Sr})_i$ vs Crustal contamination index, (b) $(^{143}\text{Nd}/^{144}\text{Nd})_i$ vs Crustal contamination index and (c) $(^{206}\text{Pb}/^{204}\text{Pb})_i$ vs Crustal contamination index. The dotted line represents the maximum contamination value a sample can have before it is considered contaminated.

The crustal contamination index and ilmenite index are further used to evaluate the effects of contamination on radiogenic isotopes measured in this study. The reason for this is that $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ in diamondiferous rocks are commonly affected by secondary processes whereas $^{143}\text{Nd}/^{144}\text{Nd}$ generally retains its primary signature (e.g., Price et al., 2000; le Roex et al., 2003; Becker and le Roux, 2006; Paton et al., 2007). The section below will evaluate the effects of secondary processes on the isotopic composition.

No clear correlation seems to exist between Sr-Nd-Pb isotopes and ilmenite index, suggesting ilmenite contamination does not affect the Sr-Nd-Pb isotope compositions in these rocks (Figure 23). However, a positive correlation can be seen between $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ (where continental crust has a lower $^{143}\text{Nd}/^{144}\text{Nd}$ with increasing C.I.) with the Clement contamination index (Figure 24). This positive trend indicates that samples with higher $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios are likely contaminated by more radiogenic crustal material (samples TP0510 and TP0204) compared to samples with lower $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios. This observation suggests that the samples with the lowest $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios are most likely to reflect the close-to-primary $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ signature of the Tongo dike-01 and this observation is consistent with the studies of Becker and le Roux (2006) and Coe et al. (2008) on the effects of crustal contamination on Sr-Nd isotopes. The $^{143}\text{Nd}/^{144}\text{Nd}$ ratios do not show any correlation against contamination index suggesting crustal contamination does not affect $^{143}\text{Nd}/^{144}\text{Nd}$ ratios in the Tongo dike-01 samples.

5.2 Peridotite entrainment

Macrocrysts that are found in diamondiferous rocks are characteristically anhedral in shape with strained physical properties suggesting that they are xenocrysts that come from broken-up mantle peridotite xenoliths (Clement, 1982; Shee, 1985). Olivine macrocrysts in kimberlites are generally abundant, occupying a volume of $\pm 25\%$, and are xenocrystic (e.g., le Roex et al., 2003; Giuliani, 2018). Therefore, macrocrystic kimberlite samples are expected to have a large proportion of additional olivine chemical components from the entrained olivine xenocrysts, i.e., they will be more Mg-rich and Ni-rich than the primary magma. Thus, it is important to assess the effects of olivine xenocryst chemical components on bulk-rock analyses in order to try constrain the parent magma composition and bring a full understanding to the Tongo parent magma.

Past studies have used aphanitic kimberlites to estimate the composition of primary kimberlite melts as they do not contain significant proportions of olivine macrocrysts (e.g., le Roex et al., 2003; Becker and le Roux, 2006). However, recent studies have shown that even fine-grained olivine microcrysts (<0.5 mm) present in aphanitic kimberlites have xenocrystic core zones indicating contamination of bulk-rock analyses with foreign olivine chemical components (e.g., Giuliani, 2018). Howarth and Giuliani (2020) showed that all olivines in the Tongo dike-01, both macrocrystic and aphanitic samples, contain xenocrystic cores. Thus, aphanitic kimberlites do not truly represent the primary kimberlite, ruling out using aphanitic samples alone to determine primary melt composition.

The entrainment of peridotite mantle xenoliths and in particular olivine xenocrysts derived from the disaggregation of xenoliths is expected to produce a higher concentration of trace elements compatible in olivine such as Co, Cr and Ni and a higher Mg# for macrocrystic rocks compared to aphanitic rock samples (e.g., Coe et al., 2008). The effects of olivine entrainment are best represented by Ni vs MgO and Ni vs K₂O variation diagrams, as shown by Coe et al. (2008). Coe et al. (2008) used Ni vs MgO and Ni vs K₂O variation diagrams with alteration, lherzolite and fractionation (discussed in the next section) vectors representing the directions of different processes that can change the chemistry of the rock. Rocks that possess high concentrations of olivine are displaced towards a lherzolite composition giving evidence that the rock had experienced olivine entrainment.

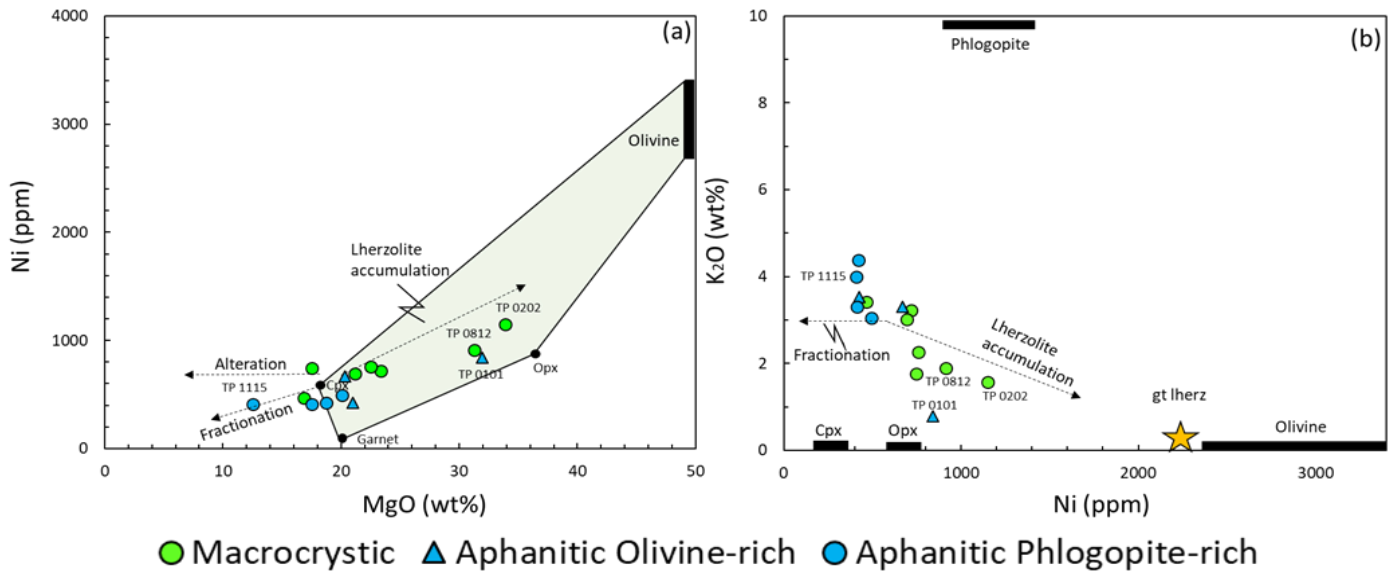


Fig. 25. a) and b) are variation diagrams of Ni vs MgO and K₂O for the Tongo dike-01 showing the effects of peridotite entrainment and olivine fractionation. The light green field represents lherzolite composition. The vectors on the graph represent the directions of the different processes that can change the chemistry of the rock.

The majority of the Tongo dike-01 samples plot in a cluster with MgO concentrations ranging from ~19.0 wt. % to ~25.8 wt. % and Ni concentrations ranging from ~426 ppm to 758 ppm. Samples that are outliers at higher MgO (>30 wt.% MgO) concentrations are likely to reflect addition of olivine (Figure 25a); these are macrocrystic samples and olivine-rich aphanitic samples (TP0202, TP0812 and TP0101). The olivine-rich nature of these samples has resulted in their relatively high concentrations in Ni and MgO, giving evidence to olivine entrainment. This is well illustrated in figure 25a, these samples are displaced towards lherzolite composition. Thus, samples TP0202, TP0812 and TP0101 do not represent a true kimberlite melt but rather have been contaminated by the entrainment of abundant olivine xenocrysts. Furthermore, the olivine-rich aphanitic samples have high MgO, which is consistent with the observation of Howarth and Giuliani (2020) that the olivine microcrysts have xenocrystic cores. This also further suggests that aphanitic kimberlite that contain abundant olivine microcrysts are not representative of parent melts.

5.3 Fractional crystallization and differentiation

The composition of primary kimberlite magma is also affected by crystal fractionation. Previous studies on kimberlites have shown that fractionation of phlogopite, olivine and calcite can take place during shallow emplacement of kimberlites or *en route* to the surface (e.g., Beard et al., 2000; le Roex et al., 2003; Coe, 2004; Harris et al., 2004; Soltys et al., 2020). Crystal fractionation is generally observed in samples by the absence of macrocrystic grains (i.e., aphanitic samples), which will be lost during fractionation. Rocks that possess low concentrations of olivine display low concentrations of MgO and Ni giving evidence of fractionation.

Rocks of an aphanitic texture are believed to have undergone crystal fractionation at some point *en route* to the surface (e.g., le Roex et al., 2003; Harris et al., 2004). Aphanitic kimberlites of the Tongo dike-01 possess higher concentrations of trace elements relative to macrocrystic kimberlites and such an enrichment is more readily achieved through fractionation instead of partial melting (e.g., le Roex et al., 2003; Harris et al., 2004). Macrocrystic samples display similar trace element patterns in primitive-mantle normalised diagrams to the aphanitic samples, suggesting they are geochemically related and that they both derive from the same parent magma (Figures 19). The primary reason for the small difference in concentrations is either due to olivine fractionation or entrainment of abundant xenocrystic olivine by the macrocrystic samples. Since olivine does not contain any trace elements, its presence lowers the trace element concentrations of macrocrystic samples, which explains why these samples have lower trace element concentrations than aphanitic samples (Figures 18 and 19). The different proportions of olivine xenocrysts observed in the Tongo dike-01 indicates that crystal fractionation has had an effect on the parent magma. Other evidence that points to olivine fractionation would be the decrease of Ni with decreasing MgO wt. %; i.e., the opposite trend to olivine entrainment discussed in the previous

section (Figure 25a). The very low concentrations of Ni and MgO of sample TP1115 in figure 25a suggests that significant amounts of olivine have been fractionated making it a good example of what an evolved kimberlite from the Tongo dike-01 would look like. This is also consistent with observation of Howarth and Giuliani (2020) who showed that the phlogopite in this sample is more evolved than other samples from this location. Thus, TP1115 is an end-member sample representing fractionation and evolution of the Tongo dike-01.

Acknowledging that some of the Tongo dike-01 samples have undergone crystal fractionation and olivine entrainment, the primary magma composition of the Tongo dike-01 should lie between these two contrasting processes in Figure 25 (e.g., le Roex et al., 2003).

5.4 Close-to-parent melt composition of the Tongo dike-01

To determine a representative parent magma composition for the Tongo dike-01, a total of six (TP0101, TP0202, TP0812, TP1013, TP1014 and TP1115) out of the original 14 samples were excluded and not considered. These mentioned samples were excluded due to showing one or more of the following processes: crustal contamination, ilmenite xenocryst contamination, olivine entrainment or olivine fractionation and as a result their compositions have been compromised (Table 5). Therefore, the average of eight samples (TP0203, TP0204, TP0305, TP0306, TP0307, TP0408, TP0509 and TP0510) were used to calculate the close-to-parent melt composition.

Table 5: A list of samples that have been excluded from primary magma composition determination and their reasons of exclusion.

Sample No.	Reason of exclusion
TP0101	Olivine entrainment
TP0202	Olivine entrainment
TP0812	Olivine entrainment
TP1013	Crustal contamination.
TP1014	Ilmenite contamination.
TP1115	Crustal contamination. Crystal fractionation.

Major elements such as SiO_2 ($\sim 28.2 \pm 3.9$ wt. %) and CaO ($\sim 12.5 \pm 1.8$ wt. %) display kimberlite like concentrations, however, contrary to this, there are some major elements that share similar concentrations to orangeites due to the highly micaceous nature of the Tongo dike-01. The high abundance of K_2O ($\sim 3.03 \pm 0.50$ wt. %) and Al_2O_3 ($\sim 4.08 \pm 1.00$ wt. %) for the Tongo dike-01 relative to magma composition of archetypal kimberlites ($\text{K}_2\text{O} \sim 0.83$ wt. % and $\text{Al}_2\text{O}_3 \sim 2.76$ wt. %; Becker and le Roex, 2006) tells us that the primary magma is more mica-rich than typical kimberlites and more similar to orangeites ($\text{K}_2\text{O} \sim 3.63$ wt. % and $\text{Al}_2\text{O}_3 \sim 3.76$ wt. %; Becker and le Roux, 2006). The Tongo dike-01 parent magma has a bulk composition that shares characteristics of both orangeite and kimberlite and is possibly unique to the Man Craton.

As suggested by the behaviour of the major elements (previous paragraph), the unusual nature of the Tongo dike-01 has produced trace elements such as Nb ($\sim 365.0 \pm 50.4$ ppm), Y ($\sim 18.77 \pm 6.60$ ppm), Pb ($\sim 12.81 \pm 3.10$ ppm) and Cr ($\sim 1508.9 \pm 101.3$ ppm) that look similar to kimberlites and some that look similar to orangeites (e.g., Rb $\sim 160.0 \pm 14.8$ ppm and Th $\sim 36.22 \pm 5.30$ ppm) (Table 6).

The composition of Sr and Pb isotopes were determined using the lowest $^{87}\text{Sr}/^{86}\text{Sr}$ (TP0202) and $^{206}\text{Pb}/^{204}\text{Pb}$ (TP0408) ratios since they are the least affected by crustal contamination and Nd was determined using an average of the samples (TP0203, TP0204, TP0408, TP0306, TP0510 and TP0509) that were not excluded based on the above-mentioned criteria. Therefore, the inferred Sr-Nd-Pb isotope geochemistry of the close-to-primary magma of the Tongo dike-01 would be $(^{87}\text{Sr}/^{86}\text{Sr})_i \sim 0.70395$, $(^{206}\text{Pb}/^{204}\text{Pb})_i \sim 18.882$, $(^{208}\text{Pb}/^{204}\text{Pb})_i \sim 40.02$ $(^{143}\text{Nd}/^{144}\text{Nd})_i \sim 0.51253 \pm 0.00001$ and $\epsilon\text{Nd} \sim 1.9 \pm 0.23$ (Table 6).

Table 6: Average composition of close to primary magma composition for Tongo dike-01 listed against South African kimberlites and orangeites for comparison.

	Tongo dike-01		S.A Kimberlite		S.A Orangeite	
	Average	SD	Average	SD	Average	SD
SiO ₂ (wt.%)	28.20	3.9	26.15	2.7	33.89	4.8
TiO ₂ (wt.%)	1.71	0.3	2.58	0.7	1.77	1.5
Al ₂ O ₃ (wt.%)	4.08	1.0	2.76	0.45	3.76	2
Fe ₂ O ₃ (wt.%)	10.23	1.7	10.72	1.2	8.76	0.6
MnO (wt.%)	0.19	0.01	0.19	0.03	0.18	0.05
MgO (wt.%)	21.01	1.0	25.2	2.1	23.15	4.2
CaO (wt.%)	12.47	1.8	13.26	2.2	9.96	2
Na ₂ O (wt.%)	0.18	0.02	0.16	0.14	0.25	0.24
K ₂ O (wt.%)	3.03	0.5	0.83	0.46	3.63	1.4
P ₂ O ₅ (wt.%)	1.65	0.6	2.04	1.2	1.85	0.8
SO ₃ (wt.%)	0.19	0.1	0.17	0.15	0.21	0.16
Cr ₂ O ₃ (wt.%)	0.20	0.01	0.18	0.04	0.23	0.09
NiO (wt.%)	0.09	0.0	0.11	0.03	0.14	0.06
H ₂ O- (wt.%)	1.46	0.4	0.66	0.27	1.34	1.7
LOI (wt.%)	14.56	2.5	14.71	1.9	10.75	3.8
Sum (wt.%)	99.27	0.1	99.71	0.2	99.41	0.4
Cu (ppm)	35.20	14.2	95.9	60	42.9	14
Ni (ppm)	608.24	141.7	935	235	977	298
Nb (ppm)	365.03	50.4	239	147	165	88
Zr (ppm)	210.32	76.2	371	98	399	194
Y (ppm)	18.77	6.6	20.5	6	20.9	9
Sr (ppm)	1808.31	381.1	1295	444	1869	671
La (ppm)	363.12	25.43	150	16.5	194.6	17.2
Gd (ppm)	15.87	4.2	11.84	2.2	8.74	3.5
Yb (ppm)	1.016	0.06	1.088	0.01	0.93	0.02
Rb (ppm)	160.91	14.8	47.1	25	151	57
U (ppm)	17.41	6.0	5.76	4	6.14	3
Th (ppm)	36.22	5.3	24.3	18	40.3	27
Pb (ppm)	12.81	3.1	10.4	5	33.3	12
Co (ppm)	78.25	9.2	89.1	9	74.7	9
Cr (ppm)	1508.97	101.3	1508	481	2096	449
V (ppm)	191.71	42.6	155	46	112	46
Ba (ppm)	2787.20	728.9	1380	876	3799	1503
(⁸⁷ Sr/ ⁸⁶ Sr) _i	0.70395	-	0.7044	0.0008	0.7078	5.3611
(¹⁴³ Nd/ ¹⁴⁴ Nd) _i	0.51253	0.00001	0.51263	0.0003	0.5119	0.0001
(²⁰⁶ Pb/ ²⁰⁴ Pb) _i	18.882	-	19.358	n/a	17.458	n/a
(²⁰⁸ Pb/ ²⁰⁴ Pb) _i	40.027	-	40.094	n/a	37.421	n/a
εNd(t)	1.9	0.23	2.2	n/a	-7.2	n/a

5.5 Classification of the Tongo dike-01

Petrography

The petrography of the studied rock indicates that the Tongo dike-01 is a highly micaceous rock resembling orangeites of the Kaapvaal craton. In particular, the common occurrence of dark brick-red margins of tetraferriphlogopite on groundmass phlogopite gives the Tongo dike-01 an orangeite appearance (Figure 15). These phlogopite margins were shown by Howarth and Giuliani (2020) to be more like orangeite phlogopite in chemistry than kimberlitic phlogopite. A feature that is used to discriminate lamproites/orangeites from kimberlites is the presence of diopside as it is a common mineral in lamproites/orangeites but absent in kimberlites (e.g., Mitchell, 1995). There is no evidence of diopside in the Tongo samples, meaning that these samples cannot be easily classified as orangeite. However, that does not mean they can't be classified as orangeites as unevolved orangeites don't have diopside either (Mitchell, 1995; Howarth et al., 2011). Generally, lamproites (e.g., Western Australian diamondiferous lamproites) do not possess calcite in their groundmass, however, orangeites do contain calcite and have been suggested to be re-classified as carbonate-bearing olivine lamproites by Pearson et al. (2019) based on the presence of calcite relative to other lamproites worldwide. Thus, using petrography alone is not enough to classify the Tongo dike-01 as either kimberlite or orangeite, but what can be deduced is that the Tongo dike-01 are unlike lamproites from Western Australia because of the presence of calcite in their groundmass. Furthermore, they are unlike archetypal Kimberley kimberlites based on the high abundance of phlogopite. The mineralogy of these samples suggests either a classification as mica-rich kimberlite or something more similar to orangeites of South Africa.

Major element geochemistry

Major element concentrations of the estimated parent magma for the Tongo dike-01 such as SiO_2 ($\sim 28.2 \pm 3.9$ wt. %), Fe_2O_3 ($\sim 10.23 \pm 1.70$ wt. %), MnO ($\sim 0.19 \pm 0.01$ wt. %), CaO ($\sim 12.47 \pm 1.80$ wt. %) and Na_2O ($\sim 0.18 \pm 0.02$ wt. %) display kimberlite like composition (Table 6). However, not all the major elements display concentrations that are similar to kimberlite, other major elements display orangeite like concentrations such as Cr_2O_3 ($\sim 0.20 \pm 0.01$ wt. %) and P_2O_5 ($\sim 1.65 \pm 0.60$ wt. %). A key feature that is used to identify orangeites relative to kimberlites is the presence of abundant phlogopite, which is reflected in high concentrations of K_2O and Al_2O_3 in the bulk rock geochemistry. Orangeites generally have high K_2O ($\sim 2-7$ wt. %) and Al_2O_3 ($\sim 3-5$ wt. %) whereas kimberlites have low K_2O ($\sim 0-2$ wt. %) and Al_2O_3 ($\sim 0-3$ wt. %). The Tongo dike-01 has a high abundance of phlogopite, which is reflected in the high concentrations of K_2O , which is more similar to orangeites than to kimberlites (Figure 16c). However, other major elements display scatter and plot in both kimberlite and orangeite fields in figures 16a, 16b and 16d, and do not aid in classification. Thus, the high K_2O concentrations are consistent with the highly micaceous nature of the Tongo dike-01 rocks indicating similarities with orangeites in terms of mineralogy and major element chemistry.

Trace element ratios

Becker and le Roex (2006) showed that one of the most robust ways to classify kimberlite vs. orangeite is using trace elements ratios (e.g., Ce/Pb , Ba/Nb , La/Nb and La/Sm). Trace element ratios are generally better for classifying kimberlite/orangeite given they are not affected by crystal fractionation and olivine xenocryst entrainment, plus high concentrations of trace elements in these rocks make these ratios not susceptible to crustal contamination (e.g., le Roex et al., 2003; Becker and le Roex, 2006). The following Tongo dike-01 trace element

ratios; Ce/Pb (16-82), Ba/Nb (1-8), La/Nb (0.6-1.2) and La/Sm (11-13) resemble those of kimberlites (Figure 20a, b and c). The sharing of similar trace element ratios clearly indicates that the Tongo dike-01 is a kimberlite and derived from a similar source to kimberlites worldwide. However, HREE enrichment constrained by using Gd/Yb (11-21) relative to La/Yb (280-520) indicates HREE ratios more similar to orangeites (Figure 20d). Having HREE ratios that are orangeite-like doesn't make the Tongo dike-01 an orangeite but rather demonstrates that it's an unusual kimberlite. This difference in HREE ratios will be discussed more in detail in the section below on partial melting.

The chondrite-normalised REE patterns of the Tongo dike-01, both macrocrystic and aphanitic, have sub-parallel patterns with a large enrichment of LREE (e.g., La = 825-1400 x chondrite) relative to HREE (e.g., Lu = 2-6 x chondrite) and these patterns are the same as those of kimberlites (Figure 18). Kimberlites are defined by relative depletions in Rb, K, Sr, Hf and Ti (le Roex et al., 2003; Harris et al., 2004) in primitive mantle-normalised diagrams whereas orangeites are defined by a depletion in Rb, Ti and Sr, a broad depletion K-Ta-Nb and an enrichment in Pb (Coe, 2004). The Tongo dike-01 is characterised by distinct negative (Rb, K, Sr, Hf and Ti) anomalies that are the same as those that define kimberlites (Figure 19). Orangeite geochemical features such as a broad depletion in K-Ta-Nb and an enrichment in Pb are not observed for the Tongo rocks in the primitive mantle plots but rather there is just a depletion in K and Pb.

Radiogenic isotope geochemistry

Like trace elements, Nd isotope geochemistry is less sensitive to mantle material assimilation compared to major elements; this means that they too can be used to classify kimberlite vs. orangeite rocks (Becker and le Roex, 2006). Typical isotope compositions for southern African kimberlites are, $(^{87}\text{Sr}/^{86}\text{Sr})_i \sim 0.70261\text{--}0.70493$, $(^{206}\text{Pb}/^{204}\text{Pb})_i \sim 18.457\text{--}20.053$, $(^{208}\text{Pb}/^{204}\text{Pb})_i \sim 40.061\text{--}40.121$ and

$(^{143}\text{Nd}/^{144}\text{Nd})_i \sim 0.511952\text{-}0.512775$ whereas for orangeites they are $(^{87}\text{Sr}/^{86}\text{Sr})_i \sim 0.76073\text{-}0.70882$, $(^{206}\text{Pb}/^{204}\text{Pb})_i \sim 17.061\text{-}18.241$, $(^{208}\text{Pb}/^{204}\text{Pb})_i \sim 37.421\text{-}38.822$ and $(^{143}\text{Nd}/^{144}\text{Nd})_i \sim 0.51181\text{-}0.51241$ (Smith, 1983 and Becker and le Roux, 2006). The Tongo dike-01 has $(^{87}\text{Sr}/^{86}\text{Sr})_i \sim 0.70395$, $(^{206}\text{Pb}/^{204}\text{Pb})_i \sim 18.882$, $(^{208}\text{Pb}/^{204}\text{Pb})_i \sim 40.027$ and $(^{143}\text{Nd}/^{144}\text{Nd})_i \sim 0.51253 \pm 0.00001$, which are similar to those that define archetypal kimberlites (Figure 21). Kimberlite magmas worldwide are produced from broadly similar sources and generally share the same isotopic values (Woodhead et al., 2019). Given that the Tongo dike-01 shares the same Sr-Nd-Pb isotope ratios as kimberlites worldwide, this would thus classify the Tongo dike-01 as a kimberlite.

Summary and classification of Tongo dike-01

The petrography and major element concentrations of the Tongo dike-01 are more similar to orangeites, but the trace element and isotope ratios clearly indicate a kimberlite affinity. The exception being HREE depletion relative to kimberlite. Thus, while these rocks display mixed characteristics they are more consistent with classification as a 'mica-rich kimberlite' than orangeite/lamproite and represents a relatively rare type of micaceous kimberlite magmatism not commonly observed on the Kaapvaal craton. An example of a mica-rich kimberlite from the Kaapvaal craton would be the Jagersfontein kimberlite, South Africa (Taylor and Kingdom, 1999). The Jagersfontein mica-rich kimberlite shares mineralogical and petrographical features with orangeites (Taylor and Kingdom, 1999). These features include the presence of high-Sr apatite, diopside in the groundmass, accessory manganiferous ilmenite and tetraferriphlogopite rims around groundmass mica that trend towards those found in orangeites, however, like the Tongo kimberlite, the Jagersfontein kimberlite is isotopically similar to kimberlites (Taylor and Kingdom, 1999). Mica-rich kimberlites are not unique to Africa, as they can also be found in other regions of the world such as the Liaoning

kimberlites of North East China (Tomkins et al., 1999) and the Kuusamo cluster of the Karelian craton in Finland (Dalton et al., 2020).

5.6 Partial melting

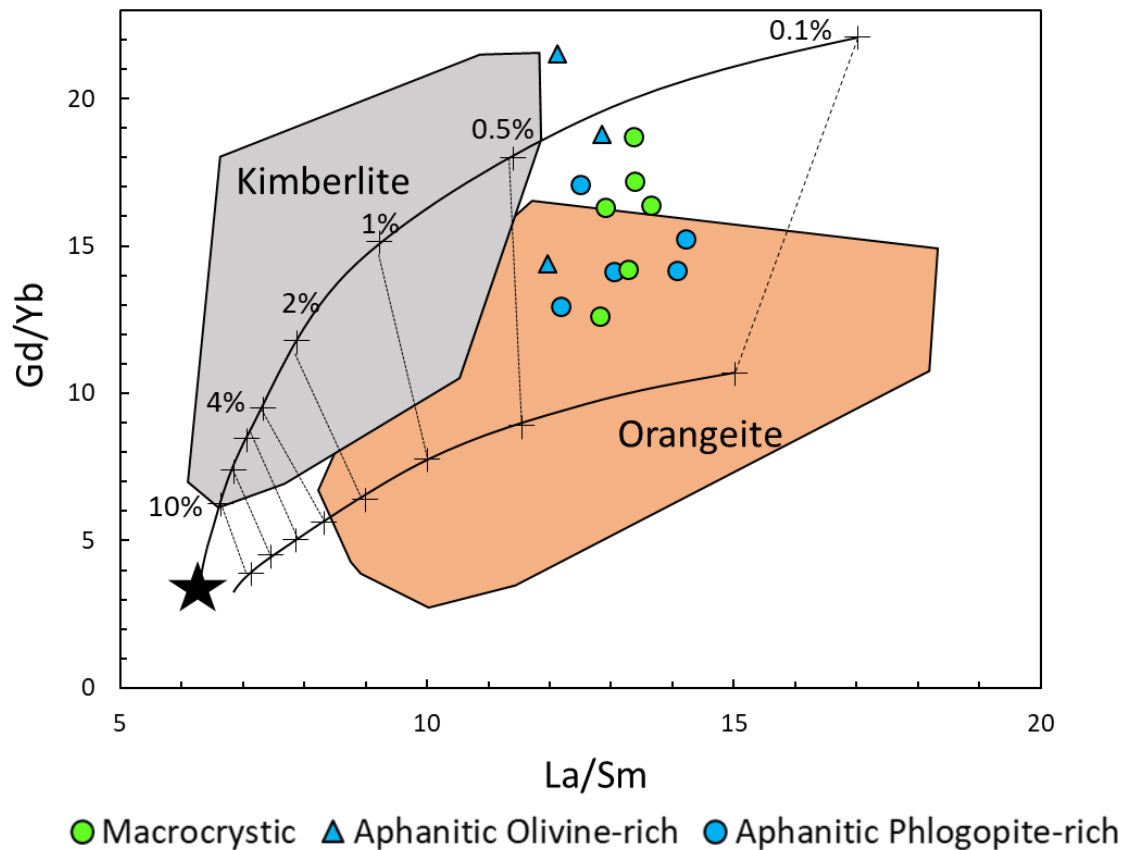


Fig. 26. Variation diagram of Gd/Yb vs La/Sm showing the degrees of partial melting experienced by the Tongo dike-01. The light grey field represents kimberlite compiled using le Roex et al. (2003) and Becker and le Roex (2006) data. The tan field represents orangeites compiled using Becker and le Roex (2006) Coe et al. (2008) and Howarth et al. (2011) data.

One of the key differences between the Tongo dike-01 and that of South African kimberlites are their La/Yb ratios (Figure 20d). Possible reasons for this difference include: (1) Lower degrees of partial melting (0.2 % to 0.4 %; Figure 26) compared to kimberlite (0.5 % - 10.0 %; le Roex et al., 2003 and Becker and le Roex, 2006). The LREE are generally more incompatible than HREE, therefore, at lower degrees of partial melting melts become enriched in LREE over HREE. (2)

Higher proportions of residual garnet in the source. Since HREE are compatible in garnet, the presence of residual garnet in the source during melting will keep HREE in the source resulting in relative depletion in HREE (e.g., Becker and le Roux, 2006 and Coe et al., 2008). (3) Having a different source composition compared to South African kimberlites. Kimberlites worldwide are interpreted to be sourced from the same asthenospheric mantle setting and they assimilate different types of lithospheric material (e.g., Woodhead et al., 2019; Giuliani et al., 2020). Therefore, the different ratios seen in figure 20d may represent different lithospheric mantle material assimilated by the Tongo kimberlite.

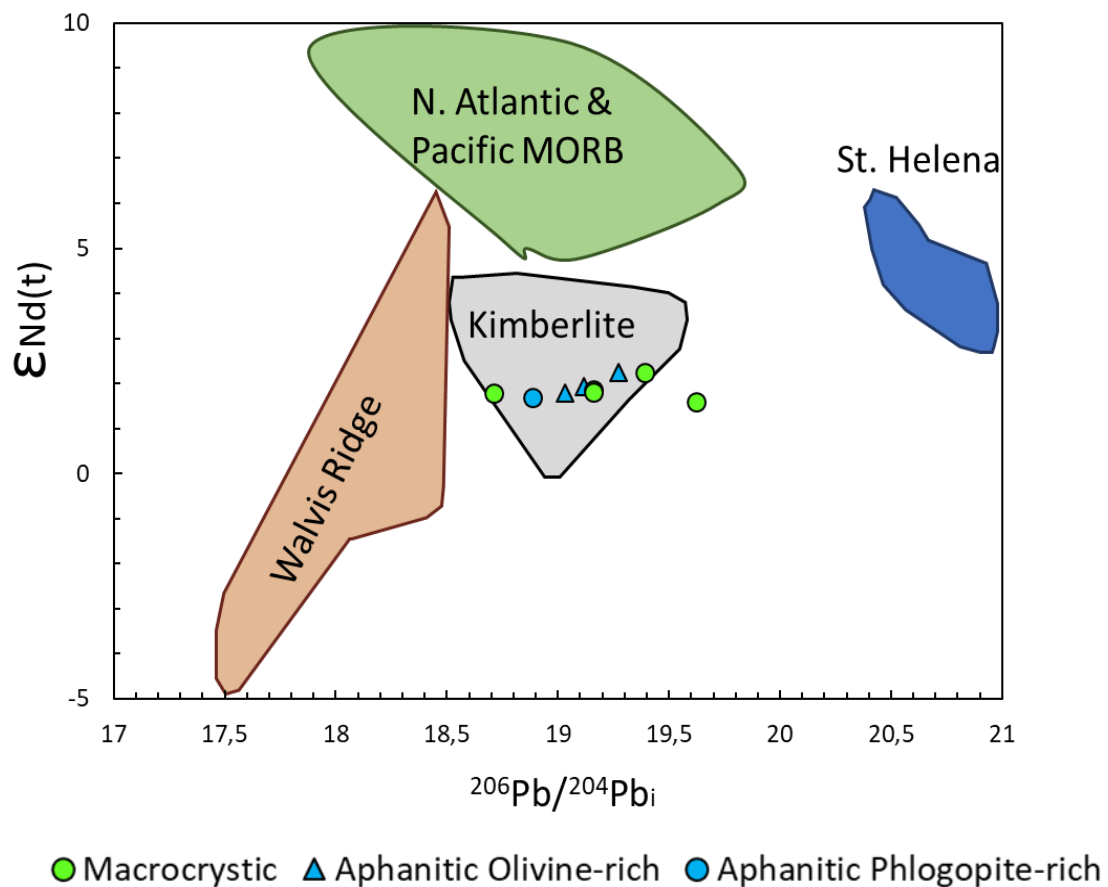


Fig. 27. Isotope ratio plot of $^{206}\text{Pb}/^{204}\text{Pb}_i$ vs $\epsilon\text{Nd}(t)$ for Tongo dike-01. The fields in the diagram represent the rock source areas: Walvis Ridge field (Richardson et al., 1982), Kimberlite field (Nowell et al., 2004 and Becker and le Roux, 2006) North Atlantic and Pacific MORB field (Petrology Database of the Ocean floor website) and St Helena field (Kawabata et al., 2011).

5.7 Implications for the petrogenesis of West African kimberlites

Kimberlites are volcanic rocks that are rich in carbonate and are formed from low degree melting of the asthenospheric mantle (e.g., Haggerty, 1994; le Roex, 2003; Pearson et al., 2014; Stamm and Schmidt, 2017). Kimberlite magmas are known for entraining a large proportion of lithospheric mantle fragments *en route* to the surface, providing insight into the lithospheric composition in different cratonic regions worldwide. Dalton et al. (2019) and Giuliani et al. (2020) showed that during the rise of a kimberlite magma, its interaction with lithospheric material results in a change of its primary melt composition. Entrained lithospheric material tends to control the major element composition of any kimberlite that ascends into the crust (Giuliani et al., 2020).

By removing samples that had been geochemically affected by crustal xenolith and ilmenite xenocryst contamination and olivine entrainment and fractionation, as described in detail in the above sections, the close-to-primary composition of the Tongo dike-01 kimberlite was constrained (Table 6). This close to primary magma composition shows that the Tongo dike-01 kimberlite and South African kimberlites display similar trace element ratios and isotope ratios due to being less sensitive to lithospheric mantle assimilation compared to major elements. This then suggests that the Tongo dike-01 and South African kimberlites are produced from similar asthenospheric sources, consistent with the suggestion of Woodhead et al. (2019) that kimberlites worldwide are derived from the same geochemical source in the asthenospheric mantle.

Kimberlites worldwide, however, range significantly in their major element chemistry and mineralogy, which has been suggested to be related to the influence of lithospheric material assimilated during evolution (Giuliani et al.,

2020). In order to explain the effects of assimilated lithospheric material by kimberlite melt, Giuliani et al. (2020) employed the composition of xenocrystic olivine cores, that provided information on the mantle column that the kimberlite magma traversed, and compared them to the composition of primary kimberlitic growth rims surrounding olivine cores, which represented kimberlite melt after assimilating lithospheric material. Their findings showed that kimberlite melts do change in composition as they assimilate lithospheric material.

The Tongo dike-01 is a micaceous rock and this is reflected in the high K_2O (~ 3.03 wt. %) and Al_2O_3 (~ 4.08 wt. %) contents of the parent melt. The micaceous nature of this kimberlite, thus, indicates that different lithospheric mantle material was assimilated relative to archetypal South African kimberlites. In order for the Tongo kimberlite to be micaceous, the assimilated lithospheric material needs to be K_2O - and Al_2O_3 -rich in order to produce the K_2O - and Al_2O_3 -rich Tongo dike-01 kimberlite magma.

Mantle xenoliths in kimberlites from West Africa are rare and have not been studied in detail other than a few studies on eclogite xenoliths (e.g., Toft, 1989; Rollinson, 1997; Barth, 2002). In contrast to the lack of studies on West African mantle xenoliths, numerous studies have been conducted on South African mantle xenoliths found in kimberlites. Based on the studies of South African mantle xenoliths, two examples of mantle xenoliths have been shown to be highly enriched in K_2O -bearing phases, 1) Mica-Amphibole-Rutile-Ilmenite-Diopside (MARID) xenoliths and 2) Phlogopite-Ilmenite-Clinopyroxene (PIC) xenoliths (e.g., Dawson and Smith, 1977; Gregoire et al., 2002; Giuliani et al., 2015; Fitzpayne et al., 2019). The isotopic features of the MARID rocks are similar to orangeites and in contrast the PIC rocks are similar to kimberlites in isotopic chemistry (e.g., Sweeney et al., 1995; Konzett et al., 1998). The MARID and PIC type xenoliths are a product of metasomatism in the lithospheric mantle, which is the key process

that introduces K₂O into the lithospheric mantle (e.g., Fitzpayne et al., 2018). Given that PIC xenoliths have isotope systematics that are similar to kimberlites, the Tongo kimberlite could have assimilated similar highly metasomatised material leading to a K₂O-rich Tongo kimberlite magma.

However, the Tongo dike-01 is not the only mica-rich kimberlite in West Africa. The Koidu kimberlite which lies 16 kilometres away from the Tongo dike-01 is also a mica-rich kimberlite and also has likely followed a similar petrogenesis process like the Tongo kimberlite given its close proximity. However, if we look at other West African kimberlites such as the Camp Alpha kimberlites from Liberia, these kimberlites are not micaceous like the Tongo and Koidu kimberlites, but all are found on the same Man Craton. Hagerty (2017) talks about different dike trends observed in West Africa and names them “the Sierra Leone trend” and “Liberian trend” referring to the different stages of kimberlite formation. It is likely that each dike trend is formed by different process, where the Sierra Leone trend primarily assimilated a lot of micaceous material in the lithospheric mantle whereas the Liberian trend did not explain why Camp Alpha kimberlites are more similar archetypal kimberlites in South Africa relative to Tongo and Koidu kimberlite.

The assimilation of micaceous metasomatised lithospheric material by kimberlites is not unique to the Man and Kaapvaal Craton but also occurs in other cratons as well, such as Karelian Craton in Finland. Like the Tongo kimberlite, the Kuusamo cluster of the Karelian Craton are highly micaceous, however, their trace element geochemistry is consistent with those that define archetypal kimberlite worldwide; classifying them as mica-rich kimberlites (Dalton et al., 2020). The formation of these mica-rich kimberlites likely follows the same processes that I proposed for the Tongo kimberlite. The Kuusamo kimberlites, like the Tongo kimberlites, assimilated metasomatised K-richrichterite- and phlogopite-bearing

lithospheric material (Dalton et al., 2020) suggesting that its lithospheric mantle may also be highly metasomatised like the Man Craton.

The rarity of mica-rich kimberlites on the Kaapvaal Craton strongly indicates that the mantle section of the Kaapvaal Craton sampled by kimberlites is less metasomatised than the Man Craton. However, the strong overlap in isotopic and trace element chemistry between micaceous and non-micaceous kimberlites e.g., Tongo kimberlite, Koidu kimberlite, South African kimberlite, Kuusamo kimberlites and other kimberlites found worldwide suggests that they all derive from a similar mantle source (Figure 27), which is most likely the asthenospheric mantle, consistent with recent studies of Sun et al. (2014); Woodhead et al. (2019) and Giuliani et al., (2020).

6. Conclusions

The Tongo dike-01 is an unusual type of kimberlite that has major element chemistry that defines both archetypal kimberlite and orangeite. The micaceous nature of the Tongo dike-01 is reflected in its high concentrations of K_2O ($\sim 3.03 \pm 0.50$ wt. %) and Al_2O_3 ($\sim 4.08 \pm 1.00$ wt. %).

Prior to interpretation of primary processes that affected the Tongo dike-01 during formation, the dike was evaluated for the effects of crustal contamination and ilmenite xenocryst contamination during samples preparation. A total of three samples were excluded for showing high indices for contamination. To evaluate the effects of contamination on Sr-Nd-Pb isotopes, crustal contamination and ilmenite indices were put against Sr-Nd-Pb isotopes to identify any possible relationships. The positive correlations between Sr-Pb-Nd and C.I. is indicative that the higher the $(^{87}Sr/^{86}Sr)_i$, $(^{206}Pb/^{204}Pb)_i$ and $(^{143}Nd/^{144}Nd)_i$ (where continental crust has a lower $^{143}Nd/^{144}Nd$ with increasing C.I.) ratios the more radiogenic crustal material had contaminated these analyses. The lowest values of $(^{87}Sr/^{86}Sr)_i$, $(^{206}Pb/^{204}Pb)_i$ and $(^{143}Nd/^{144}Nd)_i$ were used to represent the closest to primary kimberlite parent melt isotope composition. Thus, this study has shown that Sr-Pb-Nd isotopes in kimberlites are affected by crustal contamination. No correlation exists between ilmenite index and Sr-Pb-Nb isotopes suggesting ilmenite contamination does not affect the Sr-Nd-Pb isotope compositions in these kimberlites.

Samples that showed evidence of olivine xenocryst entrainment displayed relatively high concentrations of trace elements compatible in olivine such as Co, Cr and Ni and higher MgO compared to the primary melt, and samples that had experienced olivine fractionation displayed lower concentrations in Ni and MgO

relative to the primary melt. Thus, three samples that displayed traces of olivine entrainment/fractionation were too excluded from further analysis.

Having removed six (TP0101, TP0202, TP0812, TP1013, TP1014 and TP1115) out of the fourteen samples due to showing one or more of the following secondary processes such as crustal contamination, ilmenite xenocryst contamination, olivine entrainment and/or olivine fractionation, a close-to-primary magma composition was constrained, and it looks as follows: $\text{SiO}_2 \sim 28.2 \pm 3.9$ wt. %, $\text{Fe}_2\text{O}_3 \sim 10.2 \pm 1.7$ wt. %, $\text{TiO}_2 \sim 1.70 \pm 0.30$ wt. %, $\text{Al}_2\text{O}_3 \sim 4.08 \pm 1.00$ wt. %, $\text{K}_2\text{O} \sim 3.03 \pm 0.50$ wt. %, $\text{La} \sim 363.12 \pm 25.43$ ppm, $\text{Gd} \sim 15.87 \pm 4.20$ ppm, $\text{Yb} \sim 1.016 \pm 0.060$ ppm, $(^{87}\text{Sr}/^{86}\text{Sr})_i \sim 0.7039$, $(^{206}\text{Pb}/^{204}\text{Pb})_i \sim 18.88$, $(^{208}\text{Pb}/^{204}\text{Pb})_i \sim 40.02$ and $(^{143}\text{Nd}/^{144}\text{Nd})_i \sim 0.51253 \pm 0.00001$. The Tongo dike-01 primary melt composition is K_2O -rich and unusual relative to SA kimberlites.

The similarity of trace elements and Sr-Nd-Pb isotope geochemistry to archetypal kimberlites worldwide strongly suggests that the mica-rich Tongo kimberlite derives from the same asthenospheric reservoir as archetypal kimberlites, even though its major element composition and mineralogy differ from archetypal kimberlites.

Kimberlites worldwide range significantly in their major element chemistry and mineralogy due to the influence of the type of lithospheric material assimilated during formation. Therefore, the introduction of highly enriched K_2O -bearing phases from the SCLM (e.g., Phlogopite-Ilmenite-Clinopyroxene (PIC) xenoliths) via assimilation accounts for the micaceous nature of the Tongo kimberlite. PIC xenoliths are typically a product of metasomatism in the SCLM and possess abundant phlogopite and possibly K-richterite; therefore, assimilating such material during formation is the best explanation why the Tongo kimberlite differs from archetypal kimberlites.

In conclusion, the petrography, bulk-rock major and trace elements, and Sr-Nd-Pb isotope data of the Tongo dike-01 indicates that it represents an unusual type of kimberlite that does not fit within the standard classification of kimberlites, transitional kimberlites or orangeites but rather more similar to micaceous kimberlites. Therefore, an appropriate classification for the Tongo dike-01 would thus be 'mica-rich kimberlite' and unlike archetypal kimberlite, these kimberlites are formed from assimilating K₂O-rich material in the sub-continental lithospheric mantle (SCLM) during their formation.

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