The elusive Congo craton margin during Gondwana breakup: Insights from lithospheric mantle structure and heat-flow beneath the Xaudum kimberlite province, NW Botswana

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ABSTRACT

The continental lithospheric mantle (CLM) beneath the southern margin of the Congo craton has remained elusive mainly because of thick Phanerozoic sedimentary cover concealing possible kimberlite and lamproite diatremes. In this study, we explore this lithospheric mantle section by using major and trace element compositions of mantle-derived clinopyroxene and garnet xenocrysts from kimberlites of the ca. 84 Ma Nxau Nxau cluster in NW Botswana, which is part of the poorly known Xaudum kimberlite province extending into northern Namibia. We utilize these data to better understand the thermal and compositional evolution of the lithospheric mantle at the southern margin of the Congo craton. The clinopyroxene population (83 individual grains) comprises Cr-rich and Cr-poor diopsides with variable major (Al_2O_3, Na_2O, Mg#) and incompatible trace element (U, Th, Zr, Hf, Nb, Ta, REEs) compositions. The large garnet population studied (496 individual grains) is dominated by lherzolitic G9 (38%) and `megacrystic´ G1 (41%) compositions, with minor contributions from Ti-metasomatized G11 (7%) and eclogitic G3 (6%) cratonic mantle sources. Harzburgitic G10 garnet is very rare (two grains only), consistent with a lherzolite-dominated CLM section in a craton margin position. The eclogitic garnet population has compositions akin to garnet from high-Mg cratonic mantle eclogite xenoliths, and such compositions have recently been interpreted as metasomatic in origin within the mantle xenoliths literature.

Pressure–temperature calculations using the single-grain clinopyroxene technique reveal a relatively cold cratonic geotherm of 37-38 mW/m^2 for the study region during the Late Mesozoic. For peridotitic garnets, projections of calculated Ni-in-garnet temperatures onto the independently constrained regional conductive geotherm suggest that lherzolite dominates at <145 km depths, whereas high-Ti lherzolitic G11 garnets and `megacrystic´ G1 garnets originate mostly from greater depths, down to the lithosphere base at 150 to 210 km depth. The apparent confinement of ´megacrystic´ G1 garnet to the bottom of the lithosphere suggests formation from infiltrating asthenosphere-derived proto-kimberlite liquids during melt–rock interactions. In general, the data suggest that the CLM beneath NW Botswana is depleted to about 145 km depth, and between 145-210 km depths a thick metasomatized layer is identified, representing the transition into the

underlying asthenosphere. A relatively thin lithosphere beneath NW Botswana is consistent with the proposed craton margin setting, especially when compared to the thicker cratonic roots beneath the central regions of the Congo and Kalahari cratons in Angola and South Africa, respectively, reaching down to 250 km depth and possibly even deeper.

The compositional dissimilarity between the deepest-derived garnets from kimberlites in NW Botswana (i.e., from the diamond stability field) and garnets that occur as inclusions in diamond from cratons worldwide suggests extensive overprinting of the lowermost cratonic lithosphere by oxidative melt-related metasomatism. This finding, together with the very low diamond grades of the Xaudum kimberlites, points to a diminished diamond potential of the large and mostly unexposed 'cratonic' region (e.g., covered by thick desert sand) located between the major diamond mining districts of the Congo craton to the north (e.g., Catoca) and the Kalahari craton to the south (e.g., Orapa and Jwaneng).

Keywords: Congo craton, continental mantle lithosphere, garnet and clinopyroxene xenocrysts, cratonic geotherm, Cretaceous kimberlites, thermobarometry, asthenospherization

INTRODUCTION

Although kimberlites represent only small volumes of mantle-derived magma, they play an essential role in our understanding of the nature and origin of the continental lithospheric mantle (CLM) because they carry materials from otherwise inaccessible parts of Earth's interior. The abundance of xenoliths and xenocrysts in kimberlites and related rocks is strongly biased, with xenolith-rich and xenolith-poor occurrences typically existing within a single volcanic field or cluster (O'Reilly and Griffin, 2006), also mirrored by the skewed diamond grades of individual magmatic bodies within a given kimberlite field (Tappe et al., 2018a). Significant sampling bias is also known from the discrete magmatic pulses that led to different kimberlite units within a single diatreme structure (Moss et al., 2018). For most xenolith-deficient kimberlites, mantle-derived xenocrysts can be used as a proxy for mantle rocks to determine their compositions and pressure–temperature evolution (e.g., Griffin et al., 2004; Grütter et al., 2004). Using this approach, a large dataset for mantle-derived xenocrysts and micro-xenoliths from cratons worldwide has been produced and improved our understanding of ancient CLM and its diamond potential (e.g., Schulze, 1989; Griffin and Ryan, 1995; Griffin et al., 2004; Ashchepkov et al., 2012; Hunt et al., 2012; Nimis et al., 2020).

The 84 ± 4 Ma Nxau Nxau kimberlites in NW Botswana are located at the southern margin of the Congo craton near the Damara mobile belt (Farr et al., 2018; Tappe et al., 2020), which formed during Pan-African crustal reworking as part of the Gondwana assembly between 600 and 500 Ma (Jelsma et al., 2018). However, the present-day position of the southern craton boundary has remained elusive. The 31 known diatremes and sills of the Nxau Nxau kimberlite cluster contain only small amounts of mantle-derived xenoliths, which are mostly altered (Farr et al., 2018). Diamond exploration activity between 2010 and 2015 produced significant quantities of mantle-derived xenocrysts from Nxau Nxau kimberlite drill cores, including garnet and clinopyroxene, which are the subject of this study.

We report a large major and trace element dataset for mantle-derived garnet and clinopyroxene xenocrysts (~550 grains) to obtain first insights into the lithospheric mantle architecture beneath this poorly explored cratonic region of southern Africa. Our investigation of the deeper lithosphere beneath the Nxau Nxau kimberlite cluster includes an assessment of the compositional and thermal evolution of the regional CLM, including its diamond potential. The results are compared with the petrological information available for key mantle sections on the Congo craton to the northwest and the Kalahari craton to the southeast. The findings reported herein provide essential pieces in the puzzle of the cratonic assembly of former Gondwanaland in sub-Saharan Africa, because it is shown for the first time that the mantle lithosphere beneath NW Botswana has a cratonic origin, albeit strongly overprinted during tectonomagmatic processes in the ´buffer zone` between the Congo and Kalahari cratons.

GEOLOGICAL SETTING

Congo craton

The Congo craton (also known as 'Congo Shield', 'Central African Shield', 'Kasai Shield' and 'Angola–Kasai Craton') comprises vast regions in central and southwest Africa. In the south, it is separated from the Kalahari craton (including the Zimbabwe and Kaapvaal cratons plus the intervening Limpopo belt) by the Late Proterozoic Damara orogenic belt, and in the east from the Tanzania craton by the Early Proterozoic Kibaran orogenic belt (Fig. 1). Other orogenic belts flanking the Congo craton include the Lufilian Arc in the southeast, the Kaoko and West Congolian belts in the west, and the Central African fold belt in the north (e.g., Key and Ayres, 2000).

The Archean crust of the Congo craton (3.4–2.6 Ga) comprises granulites, gneisses, granites and amphibolites, and some gabbro–charnockite and migmatite complexes (e.g., Walraven and Rumvegeri, 1993). A large portion of the Congo craton is covered by Phanerozoic strata, with metamorphic and igneous basement rocks being exposed principally at four craton edges, known as the Angolan, Kasai, Mbomou and Ntem blocks (Jelsma et al., 2018). Additional information about the architecture of the Congo craton and its margins can be obtained from the study of kimberlite-borne xenoliths and xenocrysts, with three principal study regions: *(1)* the Kasai block (Mbuji Mayi – Tshibwe kimberlite clusters), *(2)* the central Angola block (Catoca and Lunda kimberlite clusters), and *(3)* the Bangweulu block (Kundelungu kimberlite and Kapamba lamproite clusters) (Batumike et al., 2009; Ashchepkov et al., 2012; Robles-Cruz et al., 2012; Nikitina et al., 2014; Kosman et al., 2016; Korolev et al., 2021; Ngwenya and Tappe, 2021; Tappe et al., 2023). These relatively few studies provide the current knowledge base for our understanding of the nature and evolution of the CLM beneath the Congo craton and its margins.

Demarcation of the southern boundary of the Congo craton is especially complicated because thick sedimentary cover sequences, including Damara-age thrust sheets, conceal this portion of the craton in southern Angola and northern Namibia and Botswana (e.g., Key and Ayers, 2000). Although geophysical (aeromagnetic and gravity) and borehole data suggest that the southern margin of the Congo craton is located in NW Botswana (Khoza et al., 2013), supporting evidence from the petrology of lithospheric mantle-derived rocks and minerals from beneath this key region is lacking thus far. We obtained such insightful materials from several small kimberlite bodies of the Nxau Nxau volcanic cluster in NW Botswana, and they were investigated to further constrain the elusive southern margin of the Congo craton.

The Nxau Nxau kimberlites of the Xaudum province

The Xaudum kimberlite province is located on the inferred southernmost extent of the Congo craton. It comprises four kimberlite clusters, namely Sikereti, Gura, Kaudom and Nxau Nxau (de Wit, 2013) (Fig. 1). Whereas the Nxau Nxau kimberlites are located in Botswana, the other three nearby clusters occur in Namibia. The Nxau Nxau cluster contains at least 31 discrete kimberlite bodies out of the 45 known magmatic bodies for the entire Xaudum kimberlite province (Fig. 2). The Xaudum kimberlite province is generally poorly explored by comparison to other kimberlite fields in southern and central Africa (de Wit et al., 2016).

In the Nxau Nxau cluster, kimberlite magmas were emplaced in the form of maar-diatremes (i.e., volcanic pipes), dykes and sills. They intruded the Phanerozoic sedimentary rocks and dolerites of the Karoo Supergroup (Smith, 1984), the Late Proterozoic sedimentary rock sequence of the Damara Supergroup, and Late Archean basement granitoids (Batumike et al., 2009). The emplacement age of the Nxau Nxau kimberlites (ca. 84 Ma; Farr et al., 2018) overlaps with the prominent Late Cretaceous peak of kimberlite magmatic activity across southern Africa and Brazil (Griffin et al., 2014; Tappe et al., 2018b), probably in response to plate reorganization during the late phase of West Gondwana breakup. The Nxau Nxau kimberlites typically consist of olivine macrocrysts (mostly altered) set in a fine-grained groundmass of olivine, spinel, perovskite, apatite, phlogopite, calcite and serpentine (Farr et al., 2018), i.e., a typical hypabyssal Group-1 kimberlite mineral assemblage and texture (Mitchell, 2008). In agreement with the petrography, the major and trace element compositions and Sr-Nd-Hf isotopic signatures of the Nxau Nxau kimberlites reflect a typical Group-1 kimberlite petrogenesis, and in the absence of anomalous ^{182}W an asthenospheric upper mantle source for this volcanism appears most likely (Tappe et al., 2020).

SAMPLES AND ANALYTICAL TECHNIQUES

Garnet (496 grains) and clinopyroxene (83 grains) xenocrysts recovered from heavy mineral concentrates of the K20 and K21 kimberlite pipes in the Nxau Nxau cluster were mounted in epoxy resin and polished for *in-situ* major and trace element analysis at the University of Johannesburg. The major and minor element compositions were measured using a CAMECA SX100 electron microprobe (EPMA), equipped with four WDS spectrometers. The measurements were performed with a beam current of 20 nA and an accelerating voltage of 20 kV at an electron beam size of 1 µm. Trace element concentrations were measured for 443 garnet grains and for a selection of 59 clinopyroxene xenocrysts by LA-ICP-MS using a Thermo Scientific iCAP RQ mass spectrometer attached to a 193 nm ArF RESOlution SE155 excimer laser. The NIST612 glass was the primary calibration standard, and ^{29}Si was used for internal standardization. Two in-house matrix-matched standards (GHR1 garnet and KBY17 clinopyroxene; Tappe et al., 2021) and several USGS basaltic glasses (BCR2G, BHVO2, BIR1G) were analyzed repeatedly alongside our samples to monitor data quality. A slightly higher variance of up to 30% was observed between the recommended values for KBY17 clinopyroxene and the trace element concentrations determined for this material in this study. However, the concentrations of most elements in GHR1 garnet and USGS basaltic glasses are in excellent agreement (i.e., <15% variability) with the recommended values (Supp. Table 1). Trace element data for clinopyroxene xenocrysts with >1 µg/g Ba are rejected because of a possible imprint from the trace element enriched host kimberlite melt (Shaikh et al., 2020). Full details about the analytical techniques are provided in Appendix 1, and the complete dataset is given in Supp. Table 1.

RESULTS

Major and trace element compositions

Clinopyroxene

Clinopyroxene xenocrysts collected for this study are mostly diopside ($Wo_{41.1–49.9}En_{46.5–53}Fs_{2.7–8}$) with variable Cr_2O_3 (0.03–3.48 wt.%) and Al_2O_3 (0.30–5.87 wt.%) contents and a wide range of Mg# (=Mg/(Mg+Fe^{2+})) (0.86–0.95) (Fig. 3a–c; Supp. Table 1). Among these xenocrysts, Cr-rich clinopyroxenes (1.03–2.85 wt.% Cr_2O_3) that are eligible for P–T calculations (see the thermobarometry section) have moderate Na_2O (1.06–2.98 wt.%) and Al_2O_3 (1.23–3.97 wt.%) contents. The trace element concentrations show a wide range: Zr (1–144 µg/g), Y (0.1–8.9 µg/g), Hf (0.1–6.8 µg/g) and Ce (6.2–40.1 µg/g) (Supp. Table 1). The normalized extended trace element patterns (using the primitive mantle values recommended by Palme and O'Neill, 2003) for clinopyroxenes have a humped shape with negative anomalies at Nb–Ta, Zr, Pb, Ti, and Y with respect to REEs (Fig. 4a, b). In contrast, the clinopyroxene xenocrysts that were excluded from P–T calculations because of their low Cr contents show a wide range of incompatible trace element concentrations, with humped trace element patterns exhibiting positive Zr–Hf anomalies (Fig. 4a, b).

Garnet

Garnets analyzed in this study (n=496) show a wide range of compositions with Mg# between 0.68–0.86 and Ca# (= Ca/(Ca+Mg+Fe^{2+}+Mn)) between 0.01–0.30 (Supp. Table 1). The wide range of compositions is also reflected in the G-type classification scheme of Grütter et al. (2004), which indicates the presence of at least eight different garnet types (Fig. 5a). The garnet assemblage studied is dominated by megacrystic G1 (41%) and lherzolitic G9 (38%) types, along with a minor proportion of high-Ti lherzolitic (metasomatized peridotite) G11 garnets (7%), eclogitic G3 garnets (6%), high-Cr pyroxenite garnets (G5) and pyroxenite/low-Ca eclogite garnets (G4) (3%), as well as unclassified G0 garnets (4%) (Fig. 5b). Only three garnet grains were identified as wehrlitic G12 and two grains as harzburgitic G10 varieties, which amounts to <2% of the total garnet xenocryst population studied.

Lherzolitic G9 garnets are relatively depleted in Zr (mostly <50 μg/g) and Y (mostly between 0–20 μg/g) compared to megacrystic G1 and Ti-metasomatized G11 garnets (mostly between 20–80 μg/g Zr, 10–25 μg/g Y) (Supp. Table 1). Megacrystic G1 and Ti-metasomatized G11 garnets also show elevated Zr/Y ratios. Similarly, Ni concentrations in lherzolitic G9 garnets are lower (<50 μg/g Ni) than in megacrystic G1 and Ti-metasomatized G11 garnets (mostly 40–100 μg/g Ni). Besides these similarities in trace element concentrations between megacrystic and metasomatized garnets, some subtle differences were observed. For example, the positive Ti-anomaly is observed only in megacrystic G1 and a few lherzolitic G9 garnets. Also, the garnet 'megacrysts' show relatively strong enrichment in Zr-Hf and lack the positive Sc anomaly (relative to the HREE and V) compared to their Ti-metasomatized garnet counterparts (Fig. 7a, d).

Nearly half of the lherzolitic G9 garnets and all of the Ti-metasomatized G11 garnets, along with the megacrystic G1 and harzburgitic G10 garnets, show 'normal' REE_N patterns with strongly depleted LREEs relative to the flat MREEs and HREEs (Fig. 6a–d). The remaining lherzolitic G9 garnets plus wehrlitic G12 garnets have 'sinusoidal' REE_N patterns with a weaker depletion in LREEs in addition to enriched Nd-Sm-Eu and progressive depletion from Gd-Tb-Dy to Dy-Ho-Er (Fig. 6c). Overall, the REE_N patterns (both normal and sinusoidal) and extended trace element patterns of the lherzolitic G9 garnets exhibit a wide range of shapes (Fig. 6c, d), indicating gradational changes in the compositions of these garnets. In contrast, the REE_N and extended trace element patterns of the other garnet types exhibit less scatter compared to the G9 garnets. In extended trace element diagrams, all garnet types show prominent negative Sr, La and Th anomalies and positive U anomalies (Fig. 7a–d). Positive Nb, Ta, Zr and Hf anomalies are also common. Extended trace element diagrams of wehrlitic G12 garnets and a few lherzolitic G9 garnets are marked by depletion in Zr, Hf, Ti, Y and the MREEs to LREEs compared to other garnet types (Fig. 7c–d). Eclogitic G3, pyroxenitic/low-Ca eclogitic G4 and 'unclassified' G0 garnets show 'normal' REE_N patterns and negative Th, La, Sr and Sm anomalies in extended trace element diagrams (Fig. 8a–f). Their REE_N and extended trace element patterns are very similar to those of the megacrystic G1 garnets (Fig. 7, 8).

Thermobarometry

Clinopyroxene

We applied the single-pyroxene thermobarometer of Nimis and Taylor (2000) to our samples, but out of 84 individual clinopyroxene xenocrysts, only 27 passed the compositional filter for suitability implemented by Nimis and Grütter (2010) and refined further by Ziberna et al. (2016). After additional pressure corrections on clinopyroxene grains that had equilibrated at >50 kbar (Nimis et al., 2020), our dataset yielded a range of temperatures and pressures between 525–1100°C and 20–52 kbar (Fig. 9), which is consistent with a conductive geotherm of 38 mW/m^2 using the Hasterok and Chapman (2011) model.

Recently, Sudholz et al. (2021a) recalibrated the Cr-in-clinopyroxene barometer of Nimis and Taylor (2000) over an extended pressure range of up to 70 kbar. We calculated equilibrium pressures for our clinopyroxene xenocrysts applying this new calibration, and we obtained pressures in the range of 25 to 57 kbar, i.e., 3 to 4 kbar higher pressures relative to the original calibration. Data treatment using the Sudholz et al. (2021a) calibration yielded a slightly colder geotherm of 37 mW/m^2 compared with 38 mW/m^2 using the Nimis and Taylor (2000) method. Given that the difference between calibrations is very minor, we will work with the 38 mW/m^2

geotherm to enable more robust comparisons with the many literature datasets that used the original Cr-in-clinopyroxene barometer (e.g., Ashchepkov et al., 2012; Hunt et al., 2012; Ziberna et al., 2016; Tappe et al., 2021; Nkere et al., 2021; Shaikh et al., 2020).

Garnet

We calculated Ni-in-garnet temperatures (T_{Ni}) for 386 peridotitic garnets using three different thermometers (Ryan et al., 1996; Canil, 1999; Sudholz et al., 2021b). The calibrations of Ryan et al. (1996) and Sudholz et al. (2021b) tend to give a broader range of temperatures than the calibration by Canil (1999), with high-temperature outliers beyond the mantle adiabat that we consider unrealistic (Supp. Fig. 1). Furthermore, the Sudholz et al. (2021b) calibration tends to give higher temperatures for garnet grains with elevated Cr#, often resulting in a similar range of temperatures for different Ni contents (Supp. Fig. 2). In contrast, the algorithm by Canil (1999) does not produce such high-temperature outliers, and the results appear to be geologically meaningful (Shaikh et al., 2020). Therefore, we only discuss T_{Ni} results obtained with the Canil (1999) method. This Ni-based thermometer gave a temperature range of 793–1329 ºC for our garnet xenocrysts. The frequency distribution of T_{Ni} shows two prominent peaks at around 975 ºC and 1100 ºC (Fig. 10a). Whereas lherzolitic (G9) garnets dominate the low-T peak, 'megacrystic' (G1) and Ti-metasomatized (G11) garnets are more prominent at the high-T peak. Our T_{Ni} results have been projected onto the clinopyroxene-derived 38 mW/m^2 Nxau Nxau geotherm (Fig. 10b), using the method discussed in Shaikh et al. (2019). The T_{Ni} projections suggest that these garnets were sampled from the lithospheric mantle by ascending kimberlite magmas between 210 and 105 km depths, with a possible thermal lithosphere-asthenosphere boundary at around 210 km depth (based on the intersection between the conductive geotherm and the mantle adiabat). However, geochemical proxies such as Y- and Ti-depletion in garnet (Griffin and Ryan, 1995) suggest that the base of the lithospheric mantle (i.e., the geochemical LAB) is located at shallower depth of around 145 km (Fig. 11a, d), which is consistent with geophysical data that cover the study region (e.g., Evans et al., 2019; White-Gaynor et al., 2020). A cratonic geotherm is assumed to be close to the conductive model up to the temperatures calculated for the lithosphere base, however, it may show a kinked nature at higher temperatures (Finnerty and Boyd, 1987; Griffin et al., 2003). This assumption may cause an underestimation of the depths of origin for the high-temperature garnets, but this minor caveat does not alter the here developed model for the CLM beneath the southernmost portion of the Congo craton in a significant way.

Data treatment shows that all Ti-metasomatized G11 garnets and most of the ‘megacrystic’ G1 garnets were likely sampled from the deepest levels of the CLM (Fig. 10a, b). The accuracy of our temperature estimates for the 'megacrystic' garnet population relies on the assumption that these grains equilibrated with ordinary mantle olivine containing approximately 3000 µg/g Ni (Canil, 1999). Because all of the olivine macrocrysts in the Nxau Nxau kimberlites studied are altered (Farr et al., 2018), it is impossible to test a megacrystic origin using mineral compositions (e.g., Moore and Costin, 2016). However, olivine is an integral part of the discrete low-Cr megacryst suite in kimberlites (Nixon and Boyd, 1973) so that the assumption of coexistence between garnet and olivine (among other phases) in the megacryst source is well founded. We note further that genetic links between low-Cr megacrysts and sheared peridotites from the lowermost CLM have been suggested, and in these plastically deformed peridotites ultra-coarse garnet clasts and other megacryst phases are typically embedded in an olivine-dominated matrix containing between 3000 and 4000 µg/g Ni (e.g., Tappe et al., 2021). Nkere et al. (2021) showed that the Ni-based garnet thermometers can be applied to megacrysts, because independent constraints yielded similar equilibration temperatures.

Our dataset reveals some prominent depth-specific compositional features. For example, there is a higher variance in the TiO_2 and Zr concentrations, with more garnets having higher Mg# in the deeper CLM portions compared to the shallower mantle (Fig. 11a, b, c). In the $(Nd/Dy)_N$ versus $(Nd/Ce)_N$ diagram (Supp. Fig. 3a), the 'megacrystic' G1, Ti-metasomatized G11 and harzburgitic G10 garnets fall into the NW-quadrant with high $(Nd/Ce)_N$ and low $(Nd/Dy)_N$, which is consistent with their strongly LREE depleted 'normal' REE patterns. Lherzolitic G9 garnets show a wide compositional range in the $(Nd/Dy)_N$ versus $(Nd/Ce)_N$ diagram consistent with their 'normal' and 'sinusoidal' REE_N patterns. When projected down to depth, there is a substantial increase in $(Nd/Dy)_N$ at ~140 km with low to moderate $(Nd/Ce)_N$ ratios (Supp. Fig. 3b, c). This indicates that garnets from this depth exhibit the strongest sinuosity in their REE patterns.

DISCUSSION

Architecture and thermal state of the lithospheric mantle beneath NW Botswana

Clinopyroxene and garnet xenocrysts studied here define a relatively cold conductive geotherm of ~38 mW/m^2 for the CLM beneath NW Botswana, with a lithospheric thickness of ~145 km (Fig. 9; 10b). Our data indicate some distinct lithological and compositional variations within the studied lithospheric mantle section. HREE depletion in clinopyroxene xenocrysts indicates their derivation from a garnet-bearing lithology (Fig. 4a). Most garnet grains studied were derived from lherzolites (G9, 38%) and metasomatized lherzolites (G11, 41%), which testifies to a relatively fertile nature of the CLM beneath NW Botswana. The reconstructed vertical distribution of garnet xenocrysts shows compositionally distinctive CLM layers, with a shallow less metasomatized layer (<145 km depth) dominated by lherzolites, and a deeper more extensively metasomatized layer (145–210 km depths) dominated by metasomatized lherzolites and megacrystic garnets (Fig. 10a, b). The shallow CLM layer also contains rare harzburgites and wehrlites, two lithologies that are seemingly absent from the deeper layer, implying an original difference between CLM layers, or lack of sampling, or metasomatic overprinting. An increasing metasomatic overprint with depth is supported by the observed increase in metasomatic alteration of the rare harzburgite and wehrlite derived crystals from within the shallow CLM layer.

The extensive metasomatism of the deeper layer, possibly representing the lithosphere-asthenosphere transition, is evident from the larger variance in TiO_2 contents of garnets with increasing depth (Fig. 11a). This observation is consistent with bottom-up refertilization of the CLM noticed for many cratons worldwide (e.g., Griffin and Ryan, 1995; Griffin et al., 1999a; Hunt et al., 2012; Ziberna et al., 2013; Aulbach et al., 2017a; Shaikh et al., 2020; de Freitas Rodrigues et al., 2023). Garnet Mg# is increasing with depth (Fig. 11b), which may be a function of increasing temperature (O'Neill and Wood, 1979), or an increasing degree of depletion in basaltic melt component over the course of lithosphere evolution (e.g., Lehtonen and O'Brien, 2009). The latter option is less likely because these high-Mg# garnets near the lithosphere base show trace element evidence for melt/fluid related fertilization. Furthermore, there is no clear trend in the garnet Cr# versus depth systematics (Supp. Fig. 4a), which argues against compositionally induced changes of garnet Mg#. A pronounced metasomatic character of the lower CLM layer is also evident from the V/Sc, Zr/Y and Ti/Eu systematics of garnet (Fig. 12b, c, d; Supp. Fig. 4b), with an apparent increase in these trace element ratios with depth. This is also consistent with the common occurrence of peridotitic garnet with ‘normal’ REE_N patterns in the lower layer compared to the upper CLM layer (Supp. Fig. 3b, c). It is also important to note that many garnet xenocrysts from the lower layer show transitional compositions between 'normal' and 'sinusoidal' end-member REE patterns (Fig. 6a–d), suggesting gradational trace element enrichment through the interaction of continuously evolving melt with the CLM wall rocks (e.g., Ziberna et al., 2013), or increasing levels of interaction between mantle peridotite and infiltrating melt. The nearly constant Mg# of 'megacrystic' G1 garnets suggests that the metasomatic melt was buffered with respect to its Fe–Mg content. The apparent confinement of 'megacrystic' G1 garnets to the lower CLM layer (Fig. 10a, b) stands in contrast to the distribution of megacrysts in other cratonic mantle sections worldwide. For example, in the Slave (Kopylova et al., 2009) and Eastern Dharwar cratons (Shaikh et al., 2020), 'megacrystic' G1 garnets and other members of the megacryst suite appear to be dispersed throughout the CLM column, suggesting their crystallization from kimberlitic precursor melts at various depth levels (e.g., Shaikh et al., 2021). Unlike these examples, the relatively restricted distribution of 'megacrystic' G1 garnets beneath NW Botswana may be explained by their formation as part of stockwork-like bodies of proto-kimberlite melt that stalled at the lithosphere base (Nkere et al., 2021). The strongly metasomatized and megacryst-rich lower CLM layer identified beneath NW Botswana resembles what has been recently described as a thick lithosphere-asthenosphere transition zone beneath the central Kaapvaal craton, a layer that is dominated by deformed fertile lherzolites and associated megacrysts (Tappe et al., 2021). Similarly, the fertilized deeper layer between 145 and 210 km depth beneath NW Botswana may represent a lithosphere-asthenosphere transition zone at the southern margin of the Congo craton. An alternative explanation for the relatively high temperatures recorded by the majority of 'megacrystic' G1 garnets (and the Ti-metasomatized G11 garnets) could be heating during interactions with 'hot' metasomatic melt, as has been suggested to explain the abundance of Ti-rich CLM lithologies beneath the Chidliak kimberlite field in NE Canada (Kopylova et al., 2019).

Such high-temperature mantle xenoliths appear to be common in craton margin positions and have also been reported from off-craton kimberlites along the Proterozoic mobile belts that border against the western and southern Kaapvaal craton margins (Janney et al., 2010). In the context of the southern margin of the Congo craton examined here, the deeper and hotter garnets could also stem from underthrust CLM portions of the Damara orogen. A similar notion was put forward by Griffin et al. (2009) for high-temperature garnets recovered from kimberlites in southern India, speculating on the presence of a large-scale tectonic suture zone at mantle depth. However, mantle-derived crystal cargo from many more kimberlite localities, spread along and across the southern Congo craton margin, would need to be studied to substantiate this idea in the African context.

The trace element concentrations of hypothetical melts in equilibrium with garnet and clinopyroxene xenocrysts were calculated using the partition coefficients for carbonated silicate melts at deep upper mantle conditions between 6 and 12 GPa (Girnis et al., 2013). The reconstructed trace element compositions were compared with those of the bulk hypabyssal kimberlites from Nxau Nxau as reported in Tappe et al. (2020). In general, we note similarities between the bulk kimberlite and metasomatized garnet trace element compositions. However, the strong depletions in Th, U and La for the melts that crystallized the megacrystic G1 and Ti-metasomatized G11 garnets at depth, relative to bulk kimberlite at surface, suggest fractionation of these elements by other common megacryst phases such as clinopyroxene and ilmenite, or even zircon (Fig. 13). Overall, the trace element modelling results for garnet are consistent with the notion of megacryst formation through the interaction of carbonate-rich proto-kimberlite melts with cratonic mantle lithosphere (Kopylova et al., 2009; Tappe et al., 2021). In contrast, the trace element patterns of the bulk kimberlites and the inverted melts using clinopyroxene xenocrysts do not match, which suggests a lack of a direct genetic relationship or distinctly different evolutionary paths of the kimberlitic liquids involved.

Diamond prospectivity at the southern margin of the Congo craton

Whereas the eclogitic diamond potential of kimberlites cannot be assessed with reasonable confidence using xenocrystic garnet compositions, it is common practice to utilize such mineralogical data to evaluate the peridotitic diamond potential of kimberlite bodies and clusters (e.g., Gurney and Moore, 1993). Our data suggest that the CLM beneath NW Botswana exhibits a suitable thickness and thermal state that would promote diamond formation and preservation within a thin interval between ~120 and 145 km depth (i.e., 25 km thick; Fig. 10a, b). This narrow diamond window is significantly thinner than beneath major diamond mining districts in sub-Saharan Africa, from where >50 km thick diamond windows have been reported (e.g., Griffin et al., 1999a, 2004; Ashchepkov et al., 2012; Nimis et al., 2020; Tappe et al., 2021).

The majority of peridotitic diamonds in a global database are linked to harzburgites and depleted lherzolites (Stachel et al., 2004; Stachel and Harris, 2008); however, at Nxau Nxau the major and trace element compositions of the most deeply derived garnet xenocrysts indicate that harzburgitic and depleted lherzolitic potential diamond source rocks are absent or scant below 145 km depth (Fig. 14a-c). The deeper garnet populations are dominated by Ti-metasomatized G11 and 'megacrystic' G1 compositions, which fall outside the compositional field for garnets found as inclusions in diamonds from cratons worldwide (Stachel and Harris, 2008). Mineral compositions dissimilar from those of inclusions in diamond, as recorded in the global database, suggest a low diamond prospectivity in NW Botswana. This is consistent with first previous exploration results that reported a lack of macro-diamonds and a general paucity of micro-diamonds for the Xaudum kimberlite province (de Wit et al., 2017).

The diamond potential of continental shields may be affected by poor diamond preservation in their mantle source rocks, including the complete destruction of the diamond endowment, possibly triggered by heating, decompression and oxidation (e.g., Stagno et al., 2015; Howarth et al., 2023). In addition, kimberlites and related carbonated magma types are known to be able to resorb diamond (Fedortchouk et al., 2019). For example, it has been argued that diamonds beneath the metacratonic Bangweulu Block in the D.R. Congo (Fig. 1) were destroyed by melt interaction in the mantle source, explaining the barren nature of the ca. 30 Ma Kundelungu kimberlite field (Batumike et al., 2009). Importantly, the same block – inferred to be an extension of the Congo craton – contained lithospheric diamonds at ca. 180 Ma as evident from the diamondiferous nature of Karoo-age lamproites that erupted on its southern margin (Tappe et al., 2023). Also, a barren suite of mantle-derived eclogite xenoliths from the highly diamondiferous Orapa kimberlites in NE Botswana (Kalahari craton) was interpreted to have interacted with proto-kimberlite melt to the extent of complete diamond destruction (Aulbach et al., 2017b; 2020a). For the southern margin of the Congo craton in NW Botswana, our data reveal the extensive nature of mantle metasomatism that overprinted the lower CLM and may have diminished the diamond potential of this poorly exposed cratonic region.

Comparisons of lithospheric mantle sections across sub-Saharan Africa

Southernmost (Nxau Nxau) versus central Congo craton (Mbuji Mayi and Catoca)

Lithospheric mantle sections from different kimberlite locations on or around cratons provide valuable insights into CLM architecture and thermal evolution (e.g., Griffin et al., 2004; Sand et al., 2009; Ashchepkov et al., 2012; Hunt et al., 2012; Kopylova et al., 2019; Shaikh et al., 2020; de Freitas Rodrigues et al., 2023). Recently, Ozaydin et al. (2021) presented magnetotelluric evidence showing that lithospheric mantle sections reconstructed from kimberlite-borne mantle cargo appear to be representative of CLM architecture. Such a combined geophysical and petrological approach is lending new confidence to the interpretations of lithosphere-scale structures based on mantle xenolith/xenocryst studies alone. However, studies based only on petrology have the additional advantage of exploring craton architecture and thermal state through time.

Below, we compare lithospheric mantle sections reported from the more central parts of the Congo craton at Mbuji Mayi (and Tshibwe) in the D.R. Congo and Catoca in Angola with the CLM architecture at the southern margin of the Congo craton in NW Botswana (this study). We note that these reconstructed CLM sections were probed by relatively young kimberlite magmatism during the Cretaceous between ca. 120 and 70 Ma (Robles-Cruz et al., 2012; Farr et al., 2018; Tappe et al., 2018b). Whereas the lithosphere was in excess of 200 km thick beneath the central Congo craton, with extensive metasomatic overprinting at its base (Batumike et al., 2009; Ashchepkov et al., 2012; Kosman et al., 2016; Korolev et al., 2021), our findings from the southern craton margin suggest the presence of a 145 km thick lithosphere, with a thick and heavily overprinted lithosphere–asthenosphere transition zone down to 210 km depth (Fig. 9, 10). The obtained 37-38 mW/m^2 geotherm for the southern margin of the Congo craton is similar to those reported for the locations at the centre of the craton, ranging between 35 and 40 mW/m^2 (Ashchepkov et al., 2012; Batumike et al., 2009). It must be noted, however, that Kosman et al. (2016) reported geotherm estimates of >40 mW/m^2 for the Kasai block, obtained by garnet and clinopyroxene thermobarometry on inclusions in alluvial diamonds. Given that the geological meaning of such information from alluvial diamonds is far from straightforward, our discussion will focus on more conventional approaches to heat-flux estimations for continental shields. These thermal constraints suggest that the CLM beneath NW Botswana is truly cratonic, albeit compositionally strongly overprinted, possibly due to tectonomagmatic activity during Gondwana assembly and breakup since around 600 Ma. Although metasomatic overprinting is also prominent for the deep mantle root beneath the central regions of the Congo craton (150-240 km depths; Batumike et al., 2009; Ashchepkov et al., 2012), it was apparently less intensive compared to the southern margin, where no harzburgitic components appear to have 'survived' at depth (Fig. 10). Congo craton architecture, as broadly reconstructed here, is consistent with the global model of craton formation and evolution, which identifies thick lithospheric roots beneath the central regions of cratons surrounded by thinning margins that accommodate most of the tectonic stresses and act as pathways to plume-related magmatism (e.g., intensive mantle metasomatism) over the course of Earth history (Shirey et al., 2002; Begg et al., 2009; Pearson et al., 2021). More recent mantle plume activity across sub-Saharan Africa appears to have further complicated the distribution of thick versus thinned and overprinted cratonic mantle lithosphere (e.g., Evans et al., 2019; Celli et al., 2020), but these geophysical models cannot be tested using mantle petrology studies because of the extremely rare occurrence of Cenozoic kimberlite magmatism (Tappe et al., 2018b).

Southern Congo craton margin versus Kalahari craton

The Kalahari craton experienced episodic kimberlite magmatism since at least Early Proterozoic times (e.g., Tappe et al., 2018b), which has allowed petrologists to constrain the evolution of the CLM beneath southern Africa. The highly refractory peridotites of the Kaapvaal mantle lithosphere were formed by extensive Archean melt depletion (e.g., Pearson et al., 1995), with numerous subsequent events of metasomatic enrichment involving different styles and melt/fluid compositions (Griffin et al., 2003; Grégoire et al., 2003; Bell et al., 2005; Simon et al., 2007; Rehfeldt et al., 2008; Kobussen et al., 2009). Nevertheless, as evidenced by the main kimberlite magmatic phase during the Mesozoic between 160 and 80 Ma, the Kaapvaal lithosphere has remained relatively thick (~190–220 km) and slightly warmer (~40 mW/m^2) (e.g., Kobussen et al., 2009; Bell et al., 2003 Muller et al., 2013; Ozaydin et al., 2021) compared to the Nxau Nxau lithospheric mantle section in NW Botswana (~145 km thick lithosphere at 38 mW/m^2; this study).

Three distinct layers have been recognized beneath the Kaapvaal craton: (1) a fertile lherzolite-dominated shallow layer (~90–125 km depth) with high orthopyroxene/olivine ratios (~0.5), (2) a middle layer dominated by depleted lherzolite and harzburgite (125–170 km depth) with lower orthopyroxene/olivine ratios (~0.25), and (3) an extensively metasomatized lower layer at >170 km depth (e.g., Kobussen et al., 2009; Griffin et al.,1999b, 2003; Özaydin et al., 2021). The shallower layer beneath the Kaapvaal craton was strongly affected by potassic metasomatism resulting in phlogopite enrichment (e.g., Fitzpayne et al., 2018) that coincides with a seismically detected mid-lithospheric discontinuity (Smart et al., 2021). Such phlogopite-rich mantle metasomatism is not known from the southern margin of the Congo craton (e.g., Ashchepkov et al., 2012). Instead, a distinctive basaltic melt-related metasomatic style, possibly linked to Karoo-age LIP magmatism (i.e., the Okavango mafic dyke swarm), is prevalent beneath the Nxau Nxau kimberlite cluster in NW Botswana (this study). This metasomatized layer beneath the Nxau Nxau kimberlite cluster is thicker than the metasomatic portion of the Kaapvaal lithosphere (~145–210 km versus 180-200 km depth). In general, the metasomatic history of the Kaapvaal craton is better studied and thus better understood, and several major magmatic events can be held responsible for reworking of the wider Kalahari CLM; for example, the Bushveld LIP at ca. 2 Ga (Richardson and Shirey, 2008; Zhang et al., 2022), the Umkondo LIP at ca. 1.1 Ga (Koornneef et al., 2017), the Karoo LIP at ca. 180 Ma (Rehfeldt et al., 2008; Burness et al., 2020), and Mesozoic orangeite/lamproite and subsequent kimberlite magmatism (e.g., Giuliani et al., 2015; Tappe et al., 2018b). Importantly, the off-craton mantle lithosphere beneath the Proterozoic mobile belts that surround the Kaapvaal craton is relatively thin and notably warmer than the central portion of the craton (e.g., Janney et al., 2010; Muller et al., 2013). An example for this pattern is the Gibeon kimberlite field on the Paleoproterozoic Rehoboth block in Namibia (Fig. 1), where the lithosphere is ~180 km thick and characterized by a warm cratonic geotherm of 40–45 mW/m^2 (Bell et al., 2003; Muller et al., 2013). Furthermore, a variably thinned (as thin as 180 km) and warmer off-craton lithosphere has been identified beneath the Mesoproterozoic Namaqua–Natal mobile belt (Janney et al., 2010). Mantle-derived peridotite xenoliths from these off-craton regions are relatively depleted and have experienced at least two stages of melt extraction, followed by variable metasomatic re-enrichment. These off-craton CLM sections also experienced a Mesozoic heating episode possibly related to continental breakup that might have caused up to 30 km of lithosphere thinning due to thermal erosion. Recently, Pearson et al. (2021) extended the boundary of the Kalahari craton to include all Proterozoic domains such as the Rehoboth block as well as the Kheis and Namaqua-Natal belts, because the formation of these mobile belts during the Proterozoic marks the stabilization of thick CLM roots.

The above scenario is different from the southern margin of the Congo craton, where cold cratonic geotherms prevail despite strong metasomatic overprinting (Fig. 9, 10). It is therefore permissible to speculate that the Congo and Kalahari cratons may be contiguous at lithospheric mantle depth (see also the interpretation of magnetotelluric data in Evans et al., 2019), and that

Neoproterozoic to Early Paleozoic tectonism in the region (i.e., Gondwana assembly) was largely confined to the upper continental crust (i.e., thin-skinned tectonics). Similar ideas have been expressed for Late Proterozoic tectonic processes that operated along the eastern margins of the Congo craton, where contiguous cratonic mantle lithosphere has been proposed to link the central African cratonic assemblage to the Tanzania craton in the east (Link et al., 2010) and to the Kalahari craton (Zimbabwe cratonic block) in the southeast (Ngwenya and Tappe, 2021). This could potentially mean that a significant part of the Gondwana assembly was shaped by intraplate deformation, with crustal shortening and strike-slip movements hundreds of kilometres away from plate margins and *bona fide* collision zones.

Remarks on non-peridotitic mantle-derived garnets

Major and minor element contents of eclogitic G3 and pyroxenitic/low-Ca eclogitic G4 garnets can be used to identify their source rock characteristics. For example, Na_2O contents of garnets were used to classify kimberlite-borne eclogite xenoliths from Roberts Victor mine in South Africa into Group I (with large-sized rounded garnets that have on average >0.09 wt.% Na_2O that are set in a clinopyroxene matrix) and Group II (with Na_2O <0.09 wt.% in garnets interlocking with clinopyroxene) varieties (McCandless and Gurney, 1989). Similarly, garnet Na_2O contents of 0.04 and 0.07 wt.% were arbitrarily taken as cut-off values for diamond-indicator eclogitic garnets by Gurney (1984). Using pyrope-almandine-grossular proportions of garnet, mantle-derived eclogite xenoliths can be divided into Type A, Type B and Type C (Taylor and Neal, 1989). More recently, trace element signatures such as Eu/Eu* anomalies, $LREE_N$ depletion patterns, and Sr-Zr-Hf anomalies were included into the classification of eclogites (Jacob et al., 2009; Viljoen et al., 2005; Aulbach and Jacob, 2016; Smart et al., 2017a; Radu et al., 2019). Based on mineral and reconstructed bulk compositions, Aulbach and Jacob (2016) described eclogites as three main types – eclogites, pyroxenites and gabbroic counterparts. The eclogites were divided further into high-Ca (Ca# for garnet = molar Ca/(Ca + Fe + Mg + Mn) > 0.2), high-Mg (Ca# ≤0.2 and Mg# = molar Mg/(Mg + Fe_{total}) > 0.6) and low-Mg (Ca# ≤ 0.2 and Mg# ≤0.6) types by these authors. We apply some of these major and trace element proxies to eclogitic (G3) and pyroxenitic/low-Ca eclogitic (G4) garnets (40 grains) recovered from the Nxau Nxau kimberlites to constrain their protoliths in more detail.

Most of the eclogitic (G3) and pyroxenitic/low-Ca eclogitic (G4) garnet grains are marked by Na_2O contents below 0.1 wt.% with high Mg# values (>0.6), which is indicative of a high pyrope content typical for Type A bimineralic eclogites (Taylor and Neal, 1989). These authors attributed Type A garnet compositions with high REE contents to fractional crystallization processes. Furthermore, Na_2O and TiO_2 (0.2–0.6 wt.%) contents of eclogitic (G3) and pyroxenitic/low-Ca eclogitic (G4) garnet from the Nxau Nxau kimberlites are comparable with the Group II eclogites identified by McCandless and Gurney (1989).

Such low Na_2O contents were considered indicative of last equilibration at relatively low pressures (e.g., Sobolev and Lavrent'ev, 1971; Gurney and Moore, 1993). However, Grütter and Quadling (1999) demonstrated that elevated Na in eclogitic garnet is not an ideal indicator of an association with diamond because garnet compositions from many graphite-bearing eclogite xenoliths overlap with those of garnets found as inclusions in diamond. Accordingly, high-Na eclogitic garnets can also occur at pressures lower than those that define the graphite-diamond phase transition. It is possible that high Na_2O contents (>0.07 wt.%) in eclogitic garnets reflect derivation from diamond-facies eclogites, or eclogite assemblages that equilibrated at graphite-facies P-T conditions but with high bulk Na contents, permitting elevated solubility of Na in the garnet structure at relatively low pressures (Grütter and Quadling, 1999). Conversely, those eclogitic garnets with <0.07 wt.% Na_2O may represent graphite-facies eclogites, or diamond-facies eclogites with low bulk Na contents. Our eclogitic garnet data do not allow a clear assignment to either graphite- or diamond-facies equilibration conditions of their eclogitic source rocks. The eclogitic (G3) and pyroxenitic/low-Ca eclogitic (G4) garnet xenocrysts studied do not show Eu/Eu* anomalies (Fig. 8a–f), which suggests a non-gabbroic (basaltic) nature of the protolith (Aulbach and Jacob, 2016; Smart et al., 2017b). The high Mg# values coupled with low Ca# values for these garnets suggest that they are derived from disaggregated high-Mg eclogite xenoliths (Fig. 15b), commonly interpreted to have metasomatic imprints from their protracted journey through Earth's crust and mantle (De Stefano et al., 2009; Smart et al., 2009; Tappe et al., 2011; Smart et al., 2012; Aulbach et al., 2017b). The eclogitic garnets studied have very similar trace element systematics to those from Mg-rich metasomatized eclogite xenoliths from the Orapa kimberlite field in NE Botswana (Aulbach et al., 2017b) and the Catoca kimberlite cluster in NE Angola (Korolev et al., 2021), with the exception of slightly higher Zr-Hf concentrations compared to peridotitic garnets (Fig. 8d). However, the extended trace element patterns of eclogitic (G3) and pyroxenitic/low-Ca eclogitic (G4) garnets from Nxau Nxau in NW Botswana do not show negative Zr-Hf anomalies relative to neighbouring MREEs (Fig. 8b, d). Very similar trace element patterns are shown by garnets from Type IIA bimineralic eclogites from Roberts Victor (Hardman et al., 2001; Radu et al., 2019). We suggest that the subtle Zr-Hf enrichment in garnet relative to the REEs may be indicative of mantle metasomatism, especially when coupled to notably higher Ti contents relative to the global eclogite database (Fig. 15a). This observation has been corroborated frequently by other proxies for metasomatic overprinting of mantle-derived eclogite xenolith suites, such as the presence of hydrous phases and incompatible trace element enrichment in clinopyroxene (e.g., Aulbach et al., 2020b; Smart et al., 2021).

SUMMARY AND CONCLUSIONS

This study investigated the major and trace element geochemistry of mantle-derived garnet and clinopyroxene xenocrysts from the Nxau Nxau kimberlites in NW Botswana. The EPMA and LA-ICP-MS datasets for these lithospheric mantle materials are the first of their kind for this poorly exposed region at the southern margin of the Congo craton. The following conclusions can be drawn.

1. Our data reveal the existence of a relatively thin cratonic lithosphere (~145 km thick) beneath the southernmost margin of the Congo craton, characterized by a cold conductive geotherm of ~38 mW/m^2 at ca. 85 Ma.
2. The lithospheric mantle section at the southern margin of the Congo craton is dominated by depleted lherzolites to around 145 km depth, and this unit is underlain by a ~65 km thick strongly overprinted zone dominated by metasomatized lherzolites and megacrysts. This layer possibly represents a lithosphere–asthenosphere transition zone that developed from the bottom third of once ~200 km thick cratonic lithosphere (i.e., progressive asthenospherization). Harzburgite appears to be a very rare component at the southern margin of the Congo craton, probably due to the strong mantle metasomatic imprint.
3. The major and trace element compositions of mantle-derived eclogitic garnet xenocrysts indicate the presence of a high-Mg eclogite component within the cratonic lithosphere. Such a lithology provides yet another record of the protracted metasomatic overprinting of the lithospheric mantle at the southern margin of the Congo craton.

4. The compositional dissimilarity between the deepest-derived garnets from the Nxau Nxau kimberlites and garnet inclusions in diamond from cratons worldwide, together with the very low diamond grades of the kimberlites from the Xaudum province explored so far, suggest a diminished diamond potential of the ‘metacratonic’ lithosphere at the southern margin of the Congo craton.

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FIGURE CAPTIONS

Fig. 1. Map of sub-Saharan Africa showing the location of the Nxau Nxau kimberlite cluster at the inferred southern margin of the Congo craton. Other kimberlite localities and cratonic domains referred to in the main text are shown for orientation (modified after Jelsma et al., 2018). Geographic coordinates and emplacement ages for the kimberlite clusters shown are summarized in Tappe et al. (2018b).

Fig. 2. Location map of the Nxau Nxau kimberlite cluster in NW Botswana. Sample material for this study was recovered from kimberlite pipes K20 and K21, which are shown with the star symbol relative to the other unstudied pipes. Modified after Farr et al. (2018).

Fig. 3. Mg# values of mantle-derived clinopyroxene xenocrysts plotted against their Na_2O (a), Al_2O_3 (b) and Cr_2O_3 (c) contents. Fields for Kalahari megacrysts are from Nkere et al. (2021) and Tappe et al. (2021). Clinopyroxenes suitable for P-T calculations have passed the compositional filters given by Nimis and Grütter (2010) and Ziberna et al. (2016).

Fig. 4. Primitive-mantle normalized trace element patterns (values for normalization after Palme and O'Neill, 2003) for mantle-derived clinopyroxene xenocrysts from the K20 and K21 pipes of the Nxau Nxau kimberlite cluster in NW Botswana at the southern margin of the Congo craton. Fields for Kalahari clinopyroxene megacrysts are after Nkere et al. (2021) and Tappe et al. (2021).

Clinopyroxenes suitable for P-T calculations passed the compositional filters given by Nimis and Grütter (2010) and Ziberna et al. (2016).

Fig. 5. (a) Cr_2O_3 (wt.%) versus CaO (wt.%) G-type classification diagram (Grütter et al., 2004) for mantle-derived garnet xenocrysts from the K20 and K21 pipes of the Nxau Nxau kimberlite cluster in NW Botswana. The radar plot (b) illustrates the relative frequency of each G-type garnet population identified at the southern margin of the Congo craton.

Fig. 6. Chondrite-normalized REE diagrams (normalization after McDonough and Sun, 1995) for megacrystic G1 (a), Ti-metasomatized G11 plus harzburgitic G10 and wehrlitic G12 (b), lherzolitic G9 'sinusoidal' (c) and lherzolitic G9 'normal' garnet xenocrysts from the Nxau Nxau kimberlite cluster in NW Botswana.

Fig. 7. Primitive-mantle-normalized trace element patterns (normalization after Palme and O'Neill, 2003) for megacrystic G1 (a), lherzolitic G9 (b), harzburgitic G10 and wehrlitic G12 (c), and Ti-metasomatized G11 garnets (d) from the Nxau Nxau kimberlite cluster in NW Botswana.

Fig. 8. Chondrite-normalized REE diagrams (normalization after McDonough and Sun, 1995) and primitive-mantle-normalized extended trace element patterns (normalization after Palme and O'Neill, 2003) for 'unclassified' G0, eclogitic G3 and pyroxenitic/low-Ca eclogitic G4 garnets from the Nxau Nxau kimberlite cluster in NW Botswana.

Fig. 9. Pressure-temperature plot for mantle-derived clinopyroxene xenocrysts from the Nxau Nxau kimberlites in NW Botswana. Conductive model geotherms (35 and 38 mW/m^2) are after Hasterok and Chapman (2011), and the diamond–graphite transition curve is after Day (2012).

Fig. 10. (a) Frequency distribution of Ni-in-garnet temperatures (T_{Ni}) (after Canil, 1999) obtained for various classes of mantle-derived garnet xenocrysts (after Grütter et al., 2004) from the Nxau Nxau kimberlites in NW Botswana. (b) Projection of T_{Ni} onto a 38 mW/m^2 geotherm (Hasterok and Chapman, 2011). The diamond–graphite transition curve is after Day (2012).

Fig. 11. Ni-in-garnet temperatures (T_{Ni}) plotted against TiO_2 (a), Mg# (b), and Zr (c) for mantle-derived garnet xenocrysts from the Nxau Nxau kimberlites in NW Botswana. The garnet classes are after Grütter et al. (2004).

Fig. 12. (a) Y (μg/g) versus Zr (μg/g) plot after Griffin et al. (1999b) and (b) Zr/Hf versus Ti/Eu plot after Shu and Brey (2015) for mantle-derived garnet xenocrysts from the Nxau Nxau kimberlites in NW Botswana.

Fig. 13. Primitive-mantle-normalized trace element patterns (normalization values after Palme and O'Neill, 2003) for melts calculated to have beeen in equilibrium with Nxau Nxau clinopyroxene and garnet (G1, G9, G11) under deep upper mantle conditions (6-12 GPa). The whole-rock trace element pattern for hypabyssal kimberlite from Nxau Nxau is taken from Tappe et al. (2020). Partition coefficients for clinopyroxene/melt and garnet/melt are after Girnis et al. (2013).

Fig. 14. (a) Nd/Y versus Sc/Y, (b) Nd/Y versus T_{Ni} (°C), and (c) Mg# versus Cr_2O_3 (wt.%) diagrams for mantle-derived garnet xenocrysts from the Nxau Nxau kimberlites in NW Botswana. Diamond inclusion fields (garnet) after Stachel et al. (2004) and Batumike et al. (2009). The "deeper garnets" mentioned in the figures are derived from >145 km depth.

Fig. 15. (a) TiO_2 (wt.%) versus Na_2O (wt.%) and (b) Ca# versus Mg# diagrams for eclogitic G3 and pyroxenitic/low-Ca eclogitic G4 garnets from the Nxau Nxau kimberlite cluster in NW Botswana. Fields for high-Ca, low-Mg and high-Mg eclogite xenolith suites and the eclogite xenoliths database are after Aulbach and Jacob (2016) and Aulbach et al. (2020b).