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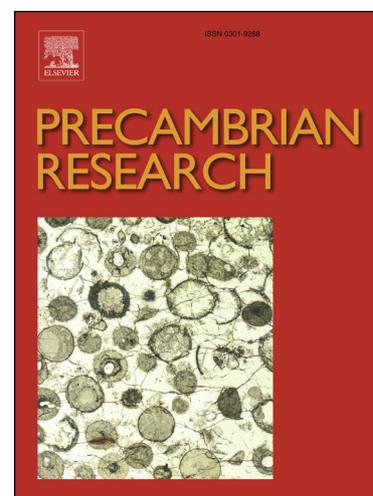
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**The Neoproterozoic Upper Ruvubu Alkaline Plutonic Complex (Burundi) revisited:
large-scale syntectonic emplacement, magmatic differentiation and
late-stage circulations of fluids**

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Abstract

The Upper Ruvubu Alkaline Plutonic Complex (URAPC) in Burundi consists of three separate intrusions, each with a specific emplacement age and petrological composition. Three main units are recognized: an outer unit with silica-saturated plutonic rocks (from gabbro to granite), an inner unit with silica-undersaturated plutonic rocks (feldspathoidal syenite with subordinate feldspathoidal monzonite and ijolite) and a carbonatitic body in the subsoil, known by drilling. The URAPC is quite large in size (~24 km long and up to 10 km wide). It is considered to have been intruded syntectonically in an overall extensional context, thanks to the kilometric shear zones that accommodated its emplacement. Radiometric ages from literature range from 748 to 705 Ma and point to structurally-controlled magmatic differentiation followed by long-lived circulations of late-stage fluids postdating the

emplacement of a part of the undersaturated rocks and the carbonatites. In the north-western part of the outer unit, gabbro likely has been emplaced at a deeper structural level than the granite, which represents a more apical structural level of emplacement.

This petrological, geochemical and isotopic (Sr-Nd-Hf) study concentrates on the processes that generated the URAPC: (i) fractional crystallization, evidenced by the chemical evolution trends of the major and trace elements, and by marked P, Ti and Ba anomalies in the trace element patterns; (ii) crustal assimilation/contamination, as shown by the wide range of Nd isotope compositions and the general increase of the Sr isotope ratios with increasing SiO₂ contents, and (iii) late-magmatic/hydrothermal alteration inducing an increase of the Sr isotope composition without changing significantly the Nd isotope composition.

The isotopic data are consistent with an asthenospheric mantle source, though less depleted than the Depleted Mantle (DM), contaminated by the Subcontinental Lithospheric Mantle (SCLM). The silicate and carbonate magmatic series are cogenetic. The outer unit is clearly more contaminated than the inner unit, whereas the carbonatitic body could have evolved by liquid immiscibility.

The URAPC lies within East Africa's Western Rift Valley, which is marked by 23 alkaline plutonic complexes. Their emplacement has been ascribed to reactivation of Proterozoic lithospheric weakness zones resulting from the breakup of the Neoproterozoic supercontinent Rodinia supercontinent.

Keywords: Neoproterozoic, carbonate-silicate melt immiscibility, fractional crystallization, crustal contamination/assimilation, SCLM, Sr-Nd-Hf isotopes.

1. Introduction

More than 850 alkaline complexes with or without associated carbonatites have been identified in Africa (Woolley, 2001). They represent about 40% of all occurrences worldwide. Their ages range from 2 billion years (Ga) (Phalaborwa, South Africa) to the present day (Oldoinyo Lengai, Tanzania), with a marked abundance peak pointing to the Neoproterozoic and the Lower Paleozoic (roughly 850-500 millions years (Ma); Midende *et al.*, 2014). The tectonic context at the time of emplacement of the Upper Ruvubu Alkaline Plutonic Complex (URAPC; around 700Ma; Midende *et al.*, 2014) is likely related to rifting during Rodinia breakup; this setting also explains fairly well the peak of age emplacements between 850 and 500 Ma.

As a rule, the geometry of the many alkaline complexes in East Central Africa is subcircular, with an average size not exceeding ~10 km x ~8 km (although some complexes are around 20-25 km across), and all these complexes are spatially associated with the East African Rift (Woolley, 2001). Along the western branch of this rift system, 23 Neoproterozoic alkaline plutonic complexes, including the Upper Ruvubu Alkaline Plutonic Complex (URAPC), are distributed over a distance of ~1700 km (Tack *et al.*, 1984; Kampunzu *et al.*, 1985).

The URAPC consists of three major units (Fig. 1): an outer unit, predominantly composed of quartz-bearing syenite, granite and gabbro-diorite, an inner unit of feldspathoidal syenite, and a small carbonatitic body (called Matongo carbonatite) that presently is not outcropping but only known by drilling. The URAPC differs from many other alkaline complexes by its outstanding dimensions (~24 km long x 10 km wide) and elliptical shape (Tack *et al.*, 1984; Midende *et al.*, 2014). These geometric attributes have been related to syntectonic emplacement conditions accommodated by the development of large shear zones (Boulvais *et al.*, 2015; Decrée *et al.*, 2015). This occurred in the overall extensional tectonic regime that affected a broad region of Central Africa during Rodinia breakup (Burke *et al.*, 2003; Tack *et al.*, 1996). This model is consistent with the general context of emplacement considered for

alkaline complexes (distensive regions associated or not with rifts; Black *et al.*, 1985). The URAPC was emplaced between ~748 Ma and ~705 Ma (Tack *et al.*, 1984; 1996; Demaiffe, 2008; Midende *et al.*, 2014) and is ascribed to reactivation of Proterozoic lithospheric weakness zones during the breakup of the supercontinent Rodinia (Tack *et al.*, 1984; 1996; Kampunzu *et al.*, 1985; Midende *et al.*, 2014).

The present paper complements recent studies dealing with various aspects of the URAPC (Midende *et al.*, 2014; Boulvais *et al.*, 2015; Decrée *et al.*, 2015 and 2016). It presents a comprehensive petrographical, mineralogical and geochemical (major and trace elements, and radiogenic isotopes) study of the complex. Sr, Nd and Hf isotope compositions of the three units mentioned above were determined in order to identify the source of the parental magma(s), to test the cogenetic nature of the whole suite of rocks, and to evaluate the extent of crustal contamination (if any) that took place during the emplacement and the differentiation processes. In the light of the syntectonic emplacement of the complex, the impact of circulation of late-stage magmatic fluids is discussed too.

2. Geological setting

2.1. Regional geology

The Neoproterozoic Upper Ruvubu Alkaline Plutonic Complex (Fig. 1) is located in NW Burundi along the Western Rift. Field mapping shows that it intruded the Western Domain of the Karagwe-Ankole Belt, known to consist of Mesoproterozoic metasedimentary and metavolcanic rocks (Akanyaru Supergroup, Gikoro and Pindura groups; Fernandez-Alonso *et al.*, 2012). The rocks attributed to these groups were intruded by S-type granites (e.g. Kayanza granite) linked with the ~1375 Ma intraplate bimodal magmatic extensional “Kibaran event” associated with asthenospheric upwelling (Tack *et al.*, 2010). At ~1000 Ma, peraluminous

granites intruded the Karagwe-Ankole Belt in a regional collisional context (Debruyne *et al.*, 2015). In the Western Domain, intrusion of the tin granites is dated at ~ 986 Ma (Tack *et al.* 2010). The tin mineralization developed in a quiescent setting following the paroxysm of deformation (Dewaele *et al.*, 2011; Melcher *et al.*, 2015). Both the magmatism and the fold-and-thrust belt geometry of the Karagwe-Ankole Belt have been ascribed to a far-field effect of Rodinia amalgamation. They illustrate the convergence between the Congo and Tanzania-Bangweulu Cratons, regionally marked by the ~1.0 Ga Southern Irumides collisional orogeny (Fernandez-Alonso *et al.*, 2012). For the Kibara Belt, which is separated along trend from the Karagwe-Ankole Belt over a distance of 300 km by older terranes (Palaeoproterozoic Rusizi belt), geochemical and isotopic data point to a subduction setting (Debruyne *et al.*, 2015).

At ~800-750 Ma, the emplacement of alkaline complexes in the western branch of the East African Rift is overall related to the Rodinia breakup and continental rifting (e.g., Burke *et al.*, 2003; Kampunzu *et al.*, 1997; Tack *et al.*, 1996), with the possible involvement of a superplume present below Rodinia (Bell, 2001; Ernst and Buchan, 2003; Li *et al.*, 2008). Nevertheless, the breakup appears to be diachronous with a succession of extensional regimes (e.g., Delpomdor *et al.*, 2013).

As a result of the ~550 Ma Gondwana amalgamation, N-S trending discrete thrust sheets and imbricate structures have overprinted large areas of the Karagwe-Ankole Belt, amongst others in NW Burundi (Fernandez-Alonso *et al.*, 2012).

The URAPC has a roughly NW-SE oriented elliptic shape (Fig.1A). Three magmatic units have been distinguished: an “outer” silica-saturated unit, an “inner” silica-undersaturated unit and a rather small carbonatite body in the subsoil, referred to as Matongo carbonatite (Tack *et al.*, 1984 and references therein; Fig.1B). The NW-SE trending elongate outer unit is an intimate association of plutonic rocks ranging from olivine-bearing gabbro and diorite through

quartz-bearing syenite to granite (Figs.2A-D). The spatially more limited N-S trending ellipsoidal “inner” unit consists of feldspathoidal syenite and a subordinate amount of feldspathoidal monzonite (Figs.2E-G). It cross-cuts the outer unit and forms the core of the URAPC complex. Concentric subvertical magmatic layering suggests rapid emplacement conditions (Fig.2E) with very little (if any) contamination from (and interaction with) rocks from the outer unit when the inner feldspathoidal syenite intruded. The outer unit is outcropping in two parts of the URAPC, i.e. in the SE (Buraniro granite, displaying a large crescent-shaped external zone) and in the NW (Maranga granite, extending over a more limited external area). In the field, no cross-cut limits are observed between the gabbro, diorite, quartz-bearing syenite and granite, but rather a progressive transition (Tack *et al.*, 1984). Xenoliths deriving from the outer unit occur in the inner unit, whereas numerous dykes of feldspathoidal syenite (with sharp/clear-cut contacts) intrude the outer unit, especially in its NW part. When observable, the contacts of outer unit with the surrounding metasedimentary parent rocks (in particular to the west of the massif) are often tectonized with more or less well-expressed foliation. The extent of the elongate N10E trending Matongo carbonatite body, originally discovered by geophysics, has been revealed by drilling campaigns showing a ~2750 m long and 250 m wide, dyke-like intrusive body with subvertical magmatic layering. Fine- and coarse-grained calciocarbonatites and a ferrocyanatite have been identified in the Matongo carbonatite body (Midende, 1984; Demaiffe, 2008; Midende *et al.*, 2014). About 3 km to the north of the Matongo carbonatite (Fig. 1B), a magmatic breccia, surrounded by feldspathoidal syenites of the inner unit, contains carbonatitic veinlets suggesting that the Matongo carbonatites postdate the emplacement of the inner unit.

Quartz-bearing syenites of the outer unit, kaolinized by hydrothermalism, are exposed at the surface and locally capped by a supergene lateritic crust (Inamvumvu laterite; Fig.1B). The area of kaolinization coincides with a metasomatic aureole (finitisation) appearing on top of

and around the Matongo carbonatite. Kaolonization seems associated with the location of deformed zones. These are vertical shear zones, several kilometres long, tens of metres wide (Fig.1B; Decrée *et al.*, 2015). They controlled the emplacement of the URAPC and hence its present-day geometry. Kaolinization was likely enhanced by fluid circulation along these zones of high permeability. Secondary phosphates occur on top of the carbonatite body. Isolated megacrysts of zircon and ilmenite (up to 6 cm in size) have been found in laterite and kaolin in the Inamvumvu area (~1 km north of the Matongo carbonatite and ~ 1km south of Kiziba; Fig. 1B). Although referred to as “Matongo zircons” (Fransolet and Tack, 1992; Burke, 1998; Midende *et al.*, 2014), these megacrysts do not come from the Matongo carbonatite body itself but are likely linked with pegmatoid veinlets intruding the kaolinized quartz-bearing syenite, as observed during carbonatite exploration works in the 1980s (British Sulphur Company, unpublished internal report). A genetic relationship with the carbonatite body is corroborated by the nature of their primary carbonate inclusions (Burke, 1998) pointing to a relatively late-stage circulation of fluids originating from the nearby Matongo carbonatite body.

Beside fenitisation, phosphate mineralization and pegmatoid veinlets containing megacrysts evidence of late-stage fluid circulations and alterations is obvious in the field (Fig.1B), as: (i) local enrichment in blue sodalite and yellow or pinkish cancrinite, and in garnet and associated clinozoisite-epidote in feldspathoidal syenites (in the southern portion of inner unit); and (ii) carbonatite veinlets and related breccias with their hydrothermal REE-Sr mineralization (with monazite, ancylite and strontianite) in the feldspathoidal syenites (north of Kiziba).

2.2. Geochronology and isotope geochemistry

The emplacement ages of the outer and inner units of the URAPC have been determined in the 1980-1990s. These ages, although not recent, can still provide a useful indication of the emplacement age of the complex. The inner unit gave a U-Pb bulk zircon age of 739 ± 7 Ma (Tack *et al.*, 1984) and the outer unit an age of 748 ± 2 Ma (Buraniro granite, SE part of the outer unit; unpublished data, and Tack *et al.*, 1995). A fission-track age of 762 ± 33 Ma obtained on titanite from the inner unit (Van den haute, 1986) overlaps with the 739 ± 7 Ma age, whereas an unpublished zircon Pb-evaporation age of 741 ± 2 Ma obtained for the Buraniro granite (Tack *et al.*, 1995) is in line with the 748 ± 2 Ma age. The Rb-Sr whole rock isochrons for the inner and outer units of the URAPC converge to ~ 700 Ma, respectively 699 ± 13 Ma (MSWD=5.2) and 707 ± 17 Ma (MSWD=1.6) (Tack *et al.*, 1984).

The “Matongo zircon” megacrysts (see 2.1) yielded a LA-ICP-MS U-Th-Pb age of 705.5 ± 4.5 Ma (Midende *et al.*, 2014), which is inferred (see 2.1.) to be a minimum age for the crystallization and emplacement of the Matongo carbonatite body. This age is consistent with the Pb-Pb whole rock isochron age of 690 ± 32 Ma obtained on the carbonatites (DemaiFFE, 2008) and matches with the Rb-Sr whole rock ages of the inner and outer units (699 ± 13 Ma and 707 ± 17 Ma, respectively). An unpublished zircon Pb-evaporation age of 738 ± 4 Ma obtained on similar megacrysts (Tack *et al.*, 1995), although less precise, could suggest a slightly older age for the carbonatite body.

If the bulk zircon U-Pb ages obtained for the inner and outer units (respectively 739 ± 7 Ma and 748 ± 2 Ma) are considered to be reliable emplacement ages, the significantly younger Rb-Sr “ages” of the same units could then correspond to the resetting of the primary (= magmatic) Rb-Sr ages as a result of the emplacement, at ~ 705 Ma, of the Matongo carbonatite body itself or, alternatively, of late-stage circulation of fluids originating from it. Emplacement ages spanning over a few tens of million years (between ~ 748 and ~ 705 Ma for

the URAPC) are not exceptional for alkaline complexes including carbonatites; see amongst others some African Cenozoic provinces, e.g. Eastern Rift, Virunga, Hoggar, Tibesti.

The Sr and Nd isotope compositions provided by the URAPC rocks suggest a cogenetic relation between the carbonatites and the feldspathoidal syenites (Tack *et al.*, 1996). Positive initial ϵNd values (+0.7 to +5.2) and low initial $^{86}\text{Sr}/^{87}\text{Sr}$ ratios (0.7025-0.7030) support a mantle-derived origin without evidence for significant assimilation of crustal material (Tack *et al.*, 1984). The inferred source appears to be less depleted in incompatible elements than the typical Depleted Mantle (DM) (Demaiffe, 2008).

2.3. Syntectonic emplacement

The URAPC differs from many other alkaline massifs by its relatively large size and elliptic shape (Fig.1A) (Tack *et al.*, 1984; Midende *et al.*, 2014). Boulvais *et al.* (2015) and Decrée *et al.* (2015) argue that these properties relate to the syntectonic emplacement of the complex in the overall long-lived extensional tectonic regime associated with the Rodinia breakup (Burke *et al.*, 2003; Hanson, 2003; Tack *et al.*, 1996).

The URAPC was emplaced under NW-SE oriented stretching parallel to the long axis of the elliptic shape. This is consistent with the NW-SE trending regional foliation and the NE-SW oriented shortening. Emplacement was controlled by the development under ductile regime of submeridian large shear zones (Fig. 1B), especially in the eastern part of the inner unit. These shear zones host a vertical foliation that bears a stretching lineation plunging at $\sim 30^\circ$ towards the North. Typical C/S structures (described in Decrée *et al.*, 2015) have been recognized along some shear zones, which implies that they were active during magmatic emplacement. Therefore, these shear zones controlled the emplacement of the URAPC from the earliest

stage of the magmatic differentiation and crystallization through magmatic-hydrothermal transition up to circulations of late-stage hydrothermal fluids.

For example, pegmatoid veins (containing nepheline, sodalite and cancrinite) that correspond to tension gashes in zones of high strain are encountered in deformed zones within the feldspathoidal syenites. A comparable mineral assemblage occurs in mylonitic zones (within the nearby quartz-bearing syenite) as infillings of *en échelon* sigmoidal tension gashes. These veins and infillings are indicative of fluid circulations during both magmatism and deformation. Some veins display internal deformation, others not.

Brittle deformation at the scale of the URAPC is ubiquitous and superimposed to ductile deformation. It led to rock fracturing (Fig.2B), locally with slickensides and steps on the fractured planes. The rock fragments are cemented by a neoformed phyllosilicate-rich mineral assemblage suggesting that the brittle deformation took place in the shallow crust. The brittle deformation is observed not only in the URAPC but also on a regional scale, e.g. in Western Burundi (adjacent ~1375 Ma S-type Kayanza granite and (ultra)mafic rocks of the Kabanga-Musongati alignment; ~ 1205 Ma A-type granites of Central Burundi; Tack *et al.*, 2010) and even more to the West across the Western Rift in the Itombwe Syncline of DR Congo (Fernandez-Alonso *et al.*, 2012). This deformation, which is also accompanied by numerous submeridian faults, is attributed to a far-field effect of the distant East Africa Orogen, located east of the Archaean Tanzania Craton, as a result of the ~550 Ma Gondwana amalgamation (Rosetti *et al.*, 2007; Fernandez-Alonso *et al.*, 2012).

3. Material and methods

Fifty-five rock samples of the URAPC have been studied (rock collection Tack, 1977-1982, listed LT "xxx", and field archives, stored at the Royal Museum for Central Africa (RMCA),

Tervuren, Belgium. In addition, eight samples (listed GPS "xxx") were collected in 2014 during field work carried out in the frame of the digitalization program of the geological map of Burundi (partly funded by the Belgian Development Agency, BTC). All investigated samples are presented in Table 1 and their location reported on Fig.1B. Most analytical data, including mineralogical composition, petrographic texture and geochemistry (major and trace elements) come from two unpublished master theses submitted at Ghent University (Vergauwen, 1990, inner unit of URAPC, and Buyschaert, 1993, outer unit of URAPC). Microprobe data (from these theses) were obtained at the Centre d'Analyse par Microsonde pour les Sciences de la Terre at the Université Catholique de Louvain (Belgium). Major and trace elements analyses (Tables 2-3) were performed at the Geology Department of Ghent University (using atomic absorption spectrometry, AAS) and at the RMCA (using X-ray fluorescence, XRF, and ICP-MS).

To complete the geochemical database (Tables 2-3) additional trace element analyses were carried out on 14 rocks (samples from the RMCA collection and from the field) at the Université Libre de Bruxelles (ULB). For major and trace elements concentration measurements, ~50 mg of powdered samples have been melted by alkaline fusion after adding 0.8g of lithium metaborate and 0.2 g of lithium tetraborate in a graphite crucible. After 10 minutes at 1000 °C, the beads were redissolved in a stirring solution of HNO₃ 5% plus traces of HF. Major elements were measured on the ThermoScientific ICAP ICP-AES at ULB, using Y as internal standard. Two USGS standards have been used (BHVO-2 and AGV-2) and the total reproducibility was better than 2% for each element. Loss on ignition was measured on a 0.5 g aliquot after 5 h at 800 °C. Trace elements were measured on an Agilent 7700 ICP-MS, also at ULB, by adding In as internal standard. The same USGS standards were used. The total reproducibility was systematically better than 5%.

For isotopes analyses, performed at ULB, ~200 mg of powdered samples were dissolved by a 3:1 mixture of HNO₃ 14N:HF 29N using high-pressure screwed beakers. After 2 days on a hotplate at 120 °C, the samples were evaporated and redissolved in HCl 6N. After the solutions were clear, several small aliquots were taken for Sr isotopes, Rb/Sr ratio, and Lu-Hf and Sm-Nd chronology. Strontium purification was made using Sr.Spec resin in HNO₃. Some samples rich in Rb necessitated a second pass through the resin to ensure a perfect elimination of Rb. The Rb/Sr ratio was determined without purification on the Agilent 7700 ICP-MS using a calibration curve of gravimetrically prepared mixed Rb and Sr standards. The reproducibility was better than 1%. The Lu-Hf-Sm-Nd cut was spiked with a mixed spike ¹⁷⁶Lu-¹⁷⁹Hf-¹⁵⁰Sm-¹⁴⁸Nd (Debaille *et al.*, 2008). After equilibration on a hot plate for 24h, the purification procedure was the same for both spiked and unspiked samples. In brief, samples were loaded on a cationic column in 1.5 N HCl. Hafnium was immediately collected, while after rinsing, the REE cut was collected in 6N HCl. Neodymium was further purified on a Ln-Spec column using 0.16 N and 0.27 N HCl. For spiked samples, Sm and Lu were also collected by increasing the strength of HCl (0.75 N and 6 N HCl respectively). The Hf cut was loaded on an anionic column using 6N HCl in order to eliminate Fe. Then the Hf cut was further purified from Ti by using a Ln-Spec column and a mixture of 6N HCL and H₂O₂.

The isotopic compositions of Sr, Nd and Hf were measured on a Nu Instrument Nu Plasma 2 multi-collector ICP-MS at ULB. Strontium was measured in wet mode by measuring a standard every 2 samples at 400 ppb in 0.05 HNO₃. Samples were recalculated with the accepted value of the SRM987 (0.710248, Weis *et al.*, 2006), and the reproducibility was better than 20 ppm (n=20). For Nd and Hf, measurements were made using an aridus 2 desolvator and by measuring a standard every 2 samples at 25 ppb in 0.05 N HNO₃ and 0.05N HNO₃:0.05N HF respectively. Samples were respectively recalculated according to the Rennes Nd standard (0.511963, Chauvel and Blichert-Toft, 2001) and the JMC 475 Hf

standard (0.282163, Blichert-Toft, 2001). The total reproducibility was better than 36 ppm for Nd ($n = 38$) and 35 ppm for Hf ($n = 21$). Spiked samples were measured on the Nu Instrument Nu Plasma 1 at ULB. Spike was deconvoluted according to the procedure described in Debaille *et al.*, 2008.

4. Petrography and mineral composition

4.1. Outer unit

The rocks of this unit cannot be considered as strictly alkaline since they contain neither alkali amphibole, alkali pyroxene, nor feldspathoids. The dark plutonic rocks, collectively referred to as “gabbro-diorite” (Tack *et al.*, 1984), comprise gabbro, diorite, quartz monzogabbro and monzodiorite. They are predominantly located in the north-western part of the outer unit where they occupy a relative large area (Fig.1A). This contrasts with the situation in the south-eastern part of the URAPC where these rocks have a more limited and scattered distribution. The gabbros contain either olivine (mainly in the NW part) or clinopyroxene (salite) as main dark minerals. Olivine has an intermediate composition ($\text{Fo}_{42.9-52.4}$) with orthopyroxene (En_{54}) forming sometimes thin reaction rims, locally in association with magnetite. Olivine and clinopyroxene are often surrounded by amphibole (ferro-pargasite or pargasite) and/or biotite (16.09-30.14 % FeO, 6.59-12.99 % MgO), illustrating the discontinuous Bowen series (Figs.3A-B; observed only in the western part of the outer unit). Plagioclase (An_{38-48}) occurs as large (up to 3 mm wide), unzoned or zoned subhedral crystals, frequently saussuritized or albitized. Plagioclase cumulate textures are observed. Alkali feldspar ($\text{Or}_{88-96} \text{Ab}_{4-12}$), being largely subsidiary, and rare quartz are mainly interstitial components. Zircon, apatite and oxides are primary accessory phases while titanite, epidote and leucoxene are of secondary origin.

Quartz-bearing syenites constitute the greater part of the outer unit. Textures are hypidiomorphic to panidiomorphic-granular, with large grain size variations (fine-grained to coarse-grained). Major constituents include quartz, perthitic alkali feldspar ($Or_{83-67} Ab_{3-6}$), plagioclase ($An_{0.9}$; locally albitized and sericitized), biotite (26.44-33.87 % FeO, 3.02-7.31 % MgO), locally transformed into chlorite and amphibole (ferro-edenite) with inclusions of apatite and zircon (Fig.3C). The modal proportions of quartz and plagioclase vary significantly from one outcrop to another. Myrmekitic texture occurs locally. Apatite, zircon, rutile, ilmenite, magnetite, hematite, titanite, epidote and allanite are accessory constituents. Calcite characterizes rocks affected by late alteration.

The granites of Maranga (in the NW part) and Buraniro (in the SE part) are pinkish, medium- to coarse-grained and differ from the quartz-bearing syenites by a lack of amphibole. Quartz, K-feldspar (often perthitic; $Or_{98} Ab_2$), plagioclase (An_{16}), clinopyroxene (ferrosalite) and biotite (30.02 % FeO, 3.77 % MgO) are the main constituents. Plagioclase is frequently saussuritized and sericitized. Accessory minerals include apatite (often forming inclusions in biotite), zircon, ilmenite, magnetite, titanite, allanite and hematite. Both granite types have subsolvus affinities and are usually hypidiomorphic-granular textured. However, some Maranga granite samples contain idiomorphic, greyish high-temperature quartz phenocrysts whereas in the Buraniro granite, quartz is xenomorphic and sometimes forms vermicular or wormy intergrowths in K-feldspar.

4.2. Inner unit

The silica-undersaturated rocks display various textures and include feldspathoidal syenite (the most abundant rock type) and subordinate feldspathoidal monzonite and ijolite. Most samples are hypidiomorphic-granular textured (Fig.3E) and medium- to coarse-grained, with

perthitic alkali feldspar ($Or_{93-97} Ab_{3-7}$), nepheline, sodalite, cancrinite (after nepheline; Fig.3F), biotite, aegirine-augite and amphibole as main constituents. Accessory minerals include garnet (Fig.3G), titanite, fluorite, apatite, zircon and interstitial calcite. Some samples have a medium- to coarse-grained allotriomorphic-granular texture, consist mainly of perthitic alkali feldspar and feldspathoids and contain less mafic constituents. In the southern part of the complex some feldspathoidal syenites have a porphyritic texture with large (up to several cm long) alkali feldspar crystals, nepheline and biotite, in a fine-grained mesostasis made of feldspathoids, alkali feldspar, biotite, pyroxene, amphibole and accessory minerals. Feldspathoidal syenites with a trachytic (fluidal) texture, outlined by subparallel laths of alkali feldspar, are exclusively observed in the eastern part of the inner unit. They show the same mineral assemblage as the other rock types. Some samples display magmatic layering with an alternation of mm-thick bands rich in perthitic alkali feldspar, alkali feldspar associated with nepheline (Fig.3H) and mafic minerals. Pyrochlore and titanite are common (Fig.3H). Clinopyroxene is Fe-rich with an overall aegirine-hedenbergite composition. The composition of garnet - that often forms a reaction rim around clinopyroxene - is $Pyr_0Alm_{0.3}Gro_{7.65}And_{30.93}$. Amphibole is taramite whereas biotite displays a large compositional range (16.90-32.84 % FeO, 1.07-10.86 % MgO). Nepheline is the dominant feldspathoid, with low amounts of normative quartz and very low CaO and FeO contents ($0.05 < CaO < 0.17$ %, $0.02 < FeO < 0.22$ %). The composition of nepheline suggests a crystallization temperature below 1068°C, and even below 700°C (empirical geothermometer of Hamilton and McKenzie, 1965). Sodalite and cancrinite, both colourless under the microscope, are secondary phases and alteration products of nepheline and alkali feldspar. In one specific area (west of Kiziba; Fig.1B), both minerals are easily recognized in the hand specimen because of their coarseness and typical colour, respectively intense blue for sodalite and pink-reddish or

yellow for cancrinite (Fig.2G). In this area, centimetric garnet, associated with minerals of the epidote-clinozoisite group, has also been observed in one of the outcrops.

4.3. Carbonatite

Midende (1984) and Midende *et al.* (2014) provided a detailed description of the Matongo carbonatite, including a comprehensive petrographical study. A brief summary of the data published by these authors is presented here.

Several facies have been recognized in the Matongo carbonatite. (1) The dominant facies is a fine-grained saccharoidal sövite, comprising mostly calcite (with $0.5 < \text{FeO} < 3$ wt.%; $0.1 < \text{MgO} < 1.6$ wt.%; $0.8 < \text{SrO} < 1.5$ wt.%) and discrete zones of vanadiferous aegirine (partly transformed into Mg-arfvedsonite) and small grains or prismatic crystals of apatite. Apatite and aegirine locally form cumulates in association with pyrochlore and K-feldspar. Richterite is another mineral phase that may be abundant and constitutes up to 30% of the sövite rock volume. (2) A coarse-grained sövite made of almost pure calcite (CaO content close to 55 wt%) is found in highly fractured zones of the carbonatite body. At the contact with the host rocks, it may be silicified and Fe-stained. "Late" hydrothermal calcite, slightly enriched in Fe, Mn and Mg (up to 1.2 wt.% FeO, 0.7 wt % MgO and 0.6 wt.% MnO), has also been identified in the coarse-grained sövite variety. (3) Ferrocarbonatites form decimeter-wide veins cross-cutting the sövites. They consist of fine-grained dolomite and ankerite (13.7-15.4 wt.% FeO, 11.3-13.7 wt.% MgO, 0.3-0.7 wt.% MnO), associated with abundant euhedral V-rich magnetite (V_2O_3 up to 2 wt.%) and ilmenite crystals.

In addition, two types of fenites occur at the contact with the host metasediments: biotite-bearing fenite and K-feldspar-bearing fenite. In the former biotite forms well-developed

crystals, and is partly converted into phlogopite. Other associated minerals consist of titanite, richterite, calcite and apatite. Both phyllosilicates have in turn been replaced by feldspars (mainly K-feldspar, with accessory albite grains), inducing eventually the formation of K-feldspar-bearing fenite.

The Matongo carbonatite is locally affected by ductile deformation, resulting in the development of a mylonitic foliation. Late fluid circulations have been documented as well (Decrée *et al.*, 2015).

5. Whole rock chemistry

5.1 Major elements

A total alkalis versus silica (TAS) diagram (Fig.4A) was used to discriminate the analysed URAPC rocks (Le Maitre *et al.*, 2002). All rocks attributed petrographically to the gabbro-diorite field group plot in the gabbro, diorite, syenodiorite and quartz diorite fields. The quartz-bearing syenites and granites fall in the quartz diorite and granite fields; the feldspathoidal syenites in the nepheline syenite, syenite, syenodiorite and ijolite fields.

Harker diagrams in which the weight percent of eight major and minor elements are plotted against wt.% silica in abscissa are illustrated in Figs.4B-I (for analytical data, see Table 2). The SiO₂ concentrations range from 43.9 to 77.0 wt.%, without any compositional gap. The rocks of the gabbro-diorite group and feldspathoidal syenites have rather low SiO₂ contents, 43.9-59.0 % and 47.5-62.1 %, respectively.

Two differentiation trends can be distinguished: 1) one for the silica-saturated rocks, compositionally varying from gabbro-diorite, through quartz-bearing syenite to granite; and 2) another for the silica-undersaturated rocks, including feldspathoidal syenites s.s., feldspathoidal monzonite and ijolite. However, the overall evolutions are similar for both

series: the MgO, Fe₂O₃, CaO, TiO₂ and P₂O₅ contents decrease more or less regularly with increasing SiO₂ (Figs. 4B, 4C, 4G, 4H and 4I), while the K₂O contents increase (Fig.4F). The Al₂O₃ and Na₂O contents first increase slightly (Figs.4D-E) and then decrease when the SiO₂ content reaches 55 to 60 wt.%. It is worth mentioning that the Na₂O content of the feldspathoidal syenite is significantly higher than those of the silica-saturated rocks, resulting in a very distinctive trend.

The Maranga granite (NW part of the outer unit) appears to be richer in silica (mean = 75.6 wt.%) than the Buraniro granite (SE part of the outer unit) (mean = 71.6 wt.%), and there is no overlap. In addition, the studied Maranga granite samples are characterized by higher Na₂O contents (> than 6 wt.%) than those belonging to the Buraniro granite do (< 5 wt.%).

5.2 Trace elements

Rocks of the gabbro-diorite group have moderate REE contents, with Σ REE ranging from 137 to 371 ppm (Table 3). Their chondrite-normalized REE patterns (Fig.5A) show La_N values ranging from 28 to 193. The LREE enrichment is moderate (La_N/Yb_N ratios ranging from 9.6 to 20), with generally positive Eu anomalies (Eu/Eu* ratios up to 2.4), except for a few samples that present slightly negative anomalies (Eu/Eu* down to 0.73). Sample LT95, collected at the contact with a feldspathoidal syenite dyke, displays a particular trend with Σ REE=67 and La_N/Yb_N=2.4, which probably results from alteration as suggested by the high Na₂O/K₂O value (> 8) of the rock. Quartz-bearing syenites are more enriched in REE than the rocks of the gabbro-diorite group (Fig.5B), with values ranging from 308 to 6978 ppm (Table 4). They have La_N/Yb_N ratios going from 9.8 to 48 and more pronounced negative Eu anomalies (0.26 < Eu/Eu* < 0.90; Table 4). Sample GPS36, which is strongly albitized, differs considerably from the others in having a less enriched REE pattern and a strong positive Eu

anomaly ($\text{Eu}/\text{Eu}^* = 2.09$). The granites are characterized by moderately fractionated patterns ($1.7 < \text{La}_N/\text{Yb}_N < 18$; Fig. 5C), lower REE contents ($102 < \Sigma\text{REE} < 671$) and similar to deeper negative Eu anomalies. The feldspathoidal syenites are also moderately enriched in REE, with ΣREE varying from 102 to 588 ppm (Table 3), but they appear to be more fractionated ($6.8 < \text{La}_N/\text{Yb}_N < 55$) than the saturated rocks (Fig. 5D), with La_N/Yb_N ratios ranging from 7 to 55 exhibiting variable negative Eu anomalies ($0.24 < \text{Eu}/\text{Eu}^* < 1$). The Eu anomaly is overall correlated to the SiO_2 content (Fig. 4J). Regarding the rocks of the gabbro-diorite group and the quartz-bearing syenites, the variation diagram shows an overall increase of the ΣREE content versus SiO_2 , and for the granites a decrease of the ΣREE versus SiO_2 (Fig. 4K).

Finally, there is no significant difference between the evolution displayed by samples from the NW and SE parts of the outer unit. Primitive mantle-normalized spidergram patterns (Fig. 5E-5H) show that most URAPC rocks share common characteristics: (1) an overall enrichment compared to the primitive mantle, both in large-ion lithophile (LILE), i.e. Rb_N values from 9 to 534, and in high-field-strength elements (HFSE), i.e. Nb_N values from 6 to 545; (2) common and strong negative Sr, Ba, P, and Ti anomalies. These anomalies are less pronounced than in rocks of the gabbro-diorite group. Abnormal patterns were recorded for the altered samples LT95 and GPS36.

6. Isotope geochemistry

The Sr, Nd and Hf isotopic compositions of the URAPC rocks are listed in Table 4. The initial isotopic ratios have been calculated on the basis of an emplacement age of ~ 705 Ma (Midende *et al.*, 2014). The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios vary considerably, - from 0.7040 to 0.7077 - for rocks of the gabbro-diorite group, from 0.7077 to 0.7096 for quartz-bearing syenites, from 0.7088 to 0.7214 for granites, and from 0.7031 to 0.7090 for feldspathoidal syenites. Three samples

(quartz-bearing syenite LT693, and the granites LT929 and LT 977) show abnormally low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (< 0.700). The two granites are significantly enriched in Rb (> 300 ppm). Their Rb/Sr isotopic system has obviously been disturbed (open-system behaviour, with possible addition of Rb, with other alkalis by late-stage fluids). These data are therefore not plotted on Figs.6 and 7 and will not further be discussed here.

The Nd isotopic compositions of the URAPC rocks also vary a lot. The initial ϵNd values span a fairly large range of values, from positive to negative. The gabbro-diorite group, the quartz-bearing syenites and the granites display largely overlapping ranges: from +3.2 to -5.7, from +2.2 to -4.6 and from +2.2 to -4.4, respectively. The feldspathoidal syenites yielded less scattered values, most values ranging from +1.5 to -2.2.

The new data are consistent with the isotopic data of Tack *et al.* (1984), which are also mentioned in Table 4 and on Fig.6A. In the Nd-Sr isotopic diagram (Fig. 6A), most samples - including the Matongo carbonatites ($0.7028 < ^{87}\text{Sr}/^{86}\text{Sr} < 0.7040$, $+5.2 < \epsilon\text{Nd}_{(i)} < +0.2$; Demaiffe, 2008) - plot more or less along a mixing hyperbola (the dashed orange line on Fig. 6A). The carbonatites, with the lowest Sr initial composition and the most positive ϵNd values, represent the end-member. The Sr isotopic compositions of some saturated and undersaturated samples are shifted to the right of the hyperbola (the dashed blue lines on Fig.6A).

The initial ϵHf of the URAPC samples also vary substantially. The dispersion of the data and the lack of correlation can be seen in a $\epsilon\text{Hf}_{(i)}$ vs. $\epsilon\text{Nd}_{(i)}$ diagram (Fig.6B). The gabbro-diorite group shows $\epsilon\text{Hf}_{(i)}$ values ranging from 9.2 to -5.8. The quartz-bearing syenites and the granites yielded $\epsilon\text{Hf}_{(i)}$ values varying from 3.1 to -12.2 and from 0.7 to -10.2, respectively. The feldspathoidal syenites finally yielded $\epsilon\text{Hf}_{(i)}$ values going from -10.7 to 6.1.

7. Discussion

7.1. Fractional crystallization

It is evident that the saturated and undersaturated series are distinct in most of the variation diagrams (Fig. 4), and that both series fractionate independently of each other. However, the shapes of the geochemical evolution trends revealed by the major, minor and trace element data of these series (Figs.4, B-I) are overall comparable. The decrease of the MgO, Fe₂O_{3tot}, CaO and TiO₂ contents with increasing SiO₂ contents (Figs.4B,C,G and H) recorded in rocks of the gabbro-diorite group is likely to result from fractionation of mafic phases, olivine first, followed by clinopyroxene, amphibole, biotite and oxides. The decrease of the P₂O₅ content (Fig.4I) with increasing silica is to be linked with apatite fractionation.

The bell-shaped evolution of the Na₂O content (Fig.4E) and to a lesser extent of the K₂O content (Fig.4F) with increasing silica content suggests early fractionation of calcic plagioclase followed, at an intermediate stage of evolution, by fractionation of alkali feldspar (and biotite). Regarding the undersaturated series, nepheline crystallized together with alkali feldspar during the final stages, still with a notably higher Na₂O content compared to the saturated rocks for a given SiO₂ content.

The marked negative anomalies of Sr, Ba, Ti and P in the quartz-bearing syenites, the granites and the feldspathoidal syenites are consistent with fractionation of plagioclase, alkali feldspar, Fe-Ti oxides and apatite (Figs.5F-H). The deepening of the Eu anomaly (from ~0.8 to ~0.1) in the quartz-bearing syenites and the granites with increasing SiO₂ content (Fig.4J) points to the continuous fractionation of plagioclase in the saturated series.

As to the saturated series, the bell-shaped evolution of the Σ REE content with increasing SiO₂ content (Fig.4K) suggests that the REEs behave as incompatible elements during the early stages of the magmatic evolution and as compatible elements at an intermediate stage (at

about 65% SiO₂). Regarding the quartz-bearing syenites - the most REE-enriched rocks - positive correlations between the La_N/Yb_N ratios and the total REE content (Fig.4L), and between ΣREE and P₂O₅ contents (Fig.4M) are indicative of fractionation of LREE-enriched phosphates such as apatite and/or monazite. However, the fractionation of other REE-minerals, allanite e.g. (also strongly enriched in LREE; Gieré and Sorensen, 2004), cannot be ruled out.

7.2. Assimilation/contamination processes and coexistence of saturated and unsaturated series

In the εNd_(i) vs. ⁸⁷Sr/⁸⁶Sr_(i) diagram, the URAPC samples roughly plot along a hyperbolic trend (the dashed orange line on Fig.6A). The two end-members correspond, on one hand, to the Matongo carbonatite (with the most positive εNd_(i) values and the least radiogenic Sr isotopic ratios, Demaiffe, 2008), which are largely overlapping with the feldspathoidal syenites and, on the other hand, to a crustal component (negative εNd_(i) values and radiogenic Sr ratios). The ⁸⁷Sr/⁸⁶Sr initial ratios increase generally with increasing SiO₂ content (Fig.7A); moreover, these ratios are roughly linearly correlated with 1/Sr (Fig.7B), pointing to a progressive contamination in the course of the fractionation process (AFC process). This contamination trend is not clearly seen in the εNd_(i) vs. εHf_(i) diagram (Fig.6B), where the Hf isotopic composition shows a large vertical variation compared to the Nd isotopic composition. Large variations in radiogenic εHf is common in cratonic domains, as the Lu-Hf system is very sensitive to long-term complex depletion of the lithospheric root (e.g. Pivin *et al.*, 2013). On the other hand, as Lu and Hf are not hosted in the same mineral phases (contrary to the two rare earths Sm and Nd that behave more coherently), it cannot be excluded that the lack of correlation in the εNd_(i) vs. εHf_(i) diagram actually reflects secondary processes, as discussed in paragraph 7.3.

Besides the general hyperbolic trend in the $\epsilon\text{Nd}_{(i)}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ diagram, some samples (whatever their mineralogical composition) plot distinctly to the right of the trend toward high $^{87}\text{Sr}/^{86}\text{Sr}$ values (the dashed horizontal blue lines on Fig.6A). This may result from late stage/hydrothermal processes that affected the URAPC (see section 7.3 below).

The country rocks of the URAPC are potential contaminants in case of upper crustal contamination/assimilation. Deeper crustal rocks of the Archaean Tanzanian Craton can also be invoked if, during the magma ascent, contamination started in the deep crust. The country rocks of the Karagwe-Ankole Belt, compositionally comparable to the rocks of the Kibara Belt (Fernandez-Alonso *et al.*, 2012), define a very large domain with typical crustal signature (recalculated at 705 Ma: $^{143}\text{Nd}/^{144}\text{Nd} < 0.5113$, $^{87}\text{Sr}/^{86}\text{Sr} > 0.710$; Fig.8A).

There is no systematic and/or regular increase of the contamination rates, neither towards the outer part of the URAPC, nor in the vicinity of the large shear zones that played a crucial role in the emplacement of the URAPC (Boulvais *et al.*, 2015; Decrée *et al.*, 2015). Therefore, it seems unlikely that contamination by/or assimilation of country rocks has taken place during emplacement at high crustal level. Most probably it occurred during magma ascent or in (a) intermediate magma chamber(s).

Several alkaline plutonic complexes occurring along the Western Rift contain both saturated and undersaturated rocks (Tack *et al.*, 1984, 1996), as it is observed worldwide (Platt, 1996). Combination of assimilation and fractional crystallization (AFC process) is invoked by some authors (e.g., Chen *et al.*, 1994; Harris *et al.*, 1999; Landoll *et al.*, 1994; Xu *et al.*, 2016; Zhu *et al.*, 2016) to explain the coexistence of both series. Other authors (e.g., Beccaluva *et al.*, 2017; Eby, 1985), however, take for granted the existence of two different sources (parental magmas).

In the URAPC, the rocks of the saturated series display more radiogenic Sr isotope compositions and more negative $\epsilon\text{Nd}(i)$ values than the undersaturated series. This argues for a larger degree of contamination for the saturated rocks and is consistent with AFC processes.

There is field evidence (Tack *et al.*, 1984) that the saturated rocks of the outer unit were emplaced before the undersaturated rocks of the inner unit (e.g., xenoliths deriving from the outer unit occur in the inner unit, whereas numerous dykes of feldspathoidal syenite intrude the outer unit), the latter followed by the Matongo carbonatite. Assuming that assimilation/contamination has taken place during magma ascent or in intermediate magma chambers (see above), crystallization of the contaminated magma close to the wall of the plumbing system could have limited the interaction between the primary undersaturated magma (tapped from the underlying parental magma chamber) and the country rocks. This would then lead to emplacement of successive magma pulses with decreasing crustal contamination, in line with the saturated series significantly more contaminated than the undersaturated series. Such armouring of the plumbing system, inducing a decrease of crustal contamination with time, is in line with the subvertical concentric magmatic layering observed in the feldspathoidal syenites of the N-S trending internal unit (Tack *et al.*, 1984). Such a mechanism has also been proposed for the Messum igneous complex in Namibia (Harris *et al.*, 1999) and the Kangerlussuaq complex in East Greenland (Riishuus *et al.*, 2006).

In conclusion, the new data strongly suggest that the saturated and the undersaturated series are cogenetic: (i) they are spatially and temporally (respectively 748 and 739 Ma) associated. Long-lived alkaline complexes (~15 Ma) have been documented in East Greenland (Kangerlussuaq complex; Riishuus *et al.*, 2006) and southern Brazil (Lages complex; Morbidelli *et al.*, 2000); and (ii) they originate from the same source (see section 7.5 below), even if the alkalinity and the silica content of the parental magma have been modified during

the magmatic evolution of the complex by armouring of the plumbing system and crustal assimilation/contamination.

7.3. Late contamination/alteration of the URAPC rocks

The three units of the URAPC were emplaced syntectonically in shear-zones recognized to be the main plumbing system for late-magmatic/metasomatic/hydrothermal fluids (Decrée *et al.*, 2015). The circulation of fluids at the scale of the whole URAPC induced the mobilization and redistribution of many elements, including HFSEs and REEs (Decrée *et al.*, 2015) and resulted at the end in abnormal concentrations (either enrichment or depletion) of certain elements. This may explain the particular outlook of the REE patterns and spidergrams of the strongly albitized quartz-bearing syenite GPS36 (Figs.5, B and F) and gabbro LT95 (Figs.5, A and E), which were altered at the contact with a syenite dyke. This could also be related to the lack of correlation in the diagram $\epsilon\text{Nd}(i)$ vs. $\epsilon\text{Hf}(i)$ (Fig.6B) as phosphate will be more prone to disturbance in Lu-Hf than in Sm-Nd (Debaille *et al.*, 2017). Late fluids may also have induced significant changes in the initial Sr isotope composition of the rocks of the URAPC due to interaction with altered country rocks, at different times and for the different rocks of the complex.

In the Nd-Sr isotope diagrams (Figs.6A and 8A), some samples plot to the right of the hyperbolic mixing curve, along roughly horizontal lines (the blue dashed on the figures), e.g. the strongly albitized quartz-bearing syenite GPS36 (which contains calcite grains and veinlets) and the granite LT1011. A shift of the primary Sr isotopic signature (with minor or no influence on the Nd system) can be attributed to syn- or late-fluid interaction with the rocks during emplacement (by addition of radiogenic Sr), as it is also known to occur in other complexes (Chen *et al.*, 1994; Landoll *et al.*, 1994).

7.4. The Matongo carbonatite: the final pulse of the URAPC

The Sr-Nd-Hf isotope data strongly suggest that the rocks of the undersaturated inner unit, the Matongo carbonatite and the rocks composing the saturated outer unit are cogenetic. Similar conclusions have been proposed for other alkaline provinces, e.g. the Grønnedal-Ika complex in Greenland (Halama *et al.*, 2005) and several complexes of the Kola Peninsula (Downes *et al.*, 2005) (see also Bell, 1998). However, the origin and the coexistence of such magmas is still a matter of debate. It is commonly assumed that carbonatite could result either from the separation by immiscibility of the carbonatitic melt from a carbonated silicate melt (at crustal pressures below 1 Gpa) (e.g., Freestone and Hamilton, 1980; Kjarsgaard, 1998), or by intensive crystal fractionation of a similar parental magma (e.g., Doroshkevich *et al.*, 2017; Kjarsgaard, 1998; Lee and Wyllie, 1994; Verhulst *et al.*, 2000). Alternatively, silicate and carbonated magmas could likely be produced by distinct melting events (e.g., Harmer and Gittins, 1998).

However, the hypothesis of a carbonatitic magma evolving from an undersaturated silicate magma is largely favoured (e.g., Beccaluva *et al.*, 1992; Bodeving *et al.*, 2017; Guzmics *et al.*, 2015) and is also adopted for the URAPC. The relatively high Nb and Zr contents of the feldspathoidal syenites of the inner unit (52-368 ppm Nb and 111-744 ppm Zr) as compared to those of the Matongo carbonatite (1-168 ppm Nb and 11-339 ppm Zr; Midende *et al.*, 2014; Fig.5H) argue for an origin of the carbonatite by immiscibility. These elements are indeed expected to strongly partition into the silicate melts as shown by experimental studies between immiscible carbonate and silicate liquids (Martin *et al.*, 2012; Veksler *et al.*, 2012).

In the URAPC, the carbonatite has been emplaced after the two associated silicate magmas (the saturated and the undersaturated rocks) but spatially and genetically belongs to the

URAPC. This is easily explained because (i) in many alkaline complexes silicate and carbonated magmas use the same plumbing system (e.g., Bell *et al.*, 1998; Harmer, 1999; Woolley and Kjarsgaard, 2008), and (ii) a carbonated melt has a higher melt-solid interfacial energy than the coexisting silicate melt, which restricts its mobility independently of the silicate melt. As a result, melt separation can be achieved only when the silicate melt has solidified: this however remains questionable considering the large time gap, 30 Ma between the emplacement of the feldspathoidal syenite and the carbonatite (note however that state-to-the-art dating would help specifying this time gap), or within liquid-dominated reservoirs, such as sills, dykes or subsurface chambers (Minarik, 1998).

7.5. Sources of the URAPC magma(s)

To determine the nature of the source of the URAPC primary magma (or magmas), only data provided by uncontaminated (or slightly contaminated) samples have been used. The latter include the Matongo carbonatite and, to a lesser extent, the undersaturated feldspathoidal syenites. The ranges of the initial Sr, Nd and Hf isotope compositions of these rocks are respectively $0.7029 < {}^{87}\text{Sr}/{}^{86}\text{Sr}_{(i)} < 0.7036$; $+4.6 < \epsilon\text{Nd}_{(i)} < +0.2$ (Demaiffe, 2008) and $0.7031 < {}^{87}\text{Sr}/{}^{86}\text{Sr}_{(i)} < 0.7056$; $+1.5 < \epsilon\text{Nd}_{(i)} < -2.2$; $+0.1 < \epsilon\text{Hf}_{(i)} < -5.2$. In agreement with Demaiffe (2008) and Midende *et al.* (2014), these isotopic data attest for the role of an asthenospheric mantle source less depleted than the depleted mantle (DM) and possibly contaminated by the subcontinental lithospheric mantle (SCLM).

In the $\epsilon\text{Hf}_{(i)}$ - $\epsilon\text{Nd}_{(i)}$ diagram (Fig.8B), almost all the samples plot in the field of the Mantle Array, with most of them falling in the field of the modern OIB, thus emphasizing involvement of the asthenospheric mantle. There is however a dispersion of the data points towards the field of the crust on one hand and towards less radiogenic $\epsilon\text{Hf}_{(i)}$ compositions (-

12.2 <math>\epsilon\text{Hf}_{(i)} < -5.4</math>) on the other hand. The URAPC data thus falls in the large field of the Kaapvaal craton xenoliths (Simon *et al.*, 2007). This large field reflects the geochemical (and thus the isotopic) heterogeneity of this mantle domain which results from its long and complex history with several melting and metasomatic enrichment episodes (Simon *et al.*, 2007). The contribution of an old SCLM component could explain the rather wide distribution of the URAPC rocks in the Nd-Hf diagram (Figure 8B), and the lack of correlation between $\epsilon\text{Nd}_{(i)}$ and $\epsilon\text{Hf}_{(i)}$ compositions. An SCLM source is consistent with the observed trace elements (LREE, LILE and HFSE) enrichment of the URAPC rocks. A contamination of the mantle-derived magmas with crustal material is also attested.

Enriched SCLM is a common source for alkaline complexes (e.g., Harmer and Gittins, 1998; Miyazaki *et al.*, 2003), in combination or not with plume involvement (e.g., Tappe *et al.*, 2007; Roulleau and Stevenson, 2013; Saha *et al.*, 2017; Chandra *et al.*, 2018). The involvement of a plume has in particular been mentioned to explain the emplacement of temporally and spatially associated alkaline complexes and flood basalts in the western branch of the East African Rift (Bell, 2001; Ernst and Buchen, 2003), as a superplume could be present under Rodinia (Li *et al.*, 2008). However, its role in the genesis of the alkaline URAPC remains uncertain (Midende *et al.*, 2014).

7.6. Differentiation and late-stage circulations of fluids with respect to geometry of the URAPC

In the outer unit of the URAPC, the NW and SE parts can be spatially distinguished by several features: (i) the size of the area of exposure of the gabbro-diorite field group (see 4.1), (ii) the more evolved nature of the Maranga granite, in the NW, compared to the Buraniro granite, in the SE (Fig.3E), in combination with a significantly higher Na₂O content (> 6 wt % and < 5 wt %, respectively); (iii) the habit of the quartz crystals included of both granites:

idiomorphic (Maranga) vs. xenomorphic (Buraniro) (see 4.1.). The distinct magmatic evolution of the two parts of the outer unit is even more evidenced by the occurrence - in-between these two parts - of the inner unit of feldspathoidal syenite itself bordered to the East by several large shear zones (Fig.1B). In the NW part of the outer unit, the gabbro-diorite group is located to the North and the Maranga granite to the South with the quartz-bearing syenites in-between. This suggests that the most evolved rocks are encountered towards the south and, at the same time, it points to the fact that the rocks of the gabbro-diorite group corresponds to a deeper and the Maranga granite to a more apical portion of the intrusion.

In the southern portion of the inner unit, enrichment of the feldspathoidal syenites in sodalite and cancrinite (both with pronounced colours) as well as in garnet and associated clinzoisite-pyroxene (Fig. 1B), is indicative of a local increase of fluid circulation during the magmatic evolution. The southern portion thus likely also corresponds to a more apical part of the intrusion. This is in agreement with the previously proposed setting for the western part of the outer unit.

In the URAPC, most of the late-magmatic/metasomatic/hydrothermal activity is thus recorded along submeridian shear zones affecting the eastern border of the inner unit (see 7.3.; Fig.1B) and, more to the ENE, in dykes of feldspathoidal syenite with blue sodalite crystals forming part of the outer unit. This large corridor of shear zones has been the locus of fluid circulations during 1) the late-stage evolution of the emplacement of the undersaturated magma, and 2) the diversified late-stage activity postdating the emplacement of the carbonatite (see 7.3.). Crustal-scale shear zones thus played a prominent role during the emplacement of the URAPC, controlling the pattern of magmatic differentiation of the complex and - even more important - channeling successive late-stage circulations of hydrothermal fluids.

7.7. The URAPC in the context of the other alkaline complexes along the Western Rift

The magmatic evolution proposed for the URAPC is similar to what happened in many other alkaline complexes (e.g., Beccaluva *et al.*, 1992; Chen *et al.*, 1994; Landoll *et al.*, 1994). Nevertheless, the URAPC differs from numerous alkaline-carbonatite complexes by its syntectonic way of emplacement, elongate shape and large size (~ 24 km long and ~ 10 km wide).

Other elongate Neoproterozoic complexes exposed along the present-day western branch of the EAR occur south-east of the URAPC, in Tanzania, e.g. : 1) Sangu-Ikola (18.3 x 1.6 km); 2) Mbozi (32.2 x 1.6 to 12.9 km) and 3) Nachendezwaya (2.4 x 0.8 km) (see location and references in Tack *et al.*, 1984; Fig.11; Table 5). Although their structural setting versus magmatism has not been studied recently, it was previously recognized, e.g. for the Sangu-Ikola complex, that “the tectonic structure predates the carbonatite and controlled the emplacement”. Alternatively, to the N of the URAPC, the complexes display the more “normal”, subcircular geometry and have a smaller size, e.g. 1) Bingo-Mont Home (6 x 4.5 km); 2) Lueshe (3.1 x 2.1 km) and 3) Kahuzi (9 x 7 km). As to the south-eastern group of complexes, with “structural control” during emplacement, a link can be suggested with the peculiar geologic setting of the region corresponding to the NW-SE trending Ubende Shear belt (2.1 Ga), which experienced repeatedly reactivation, amongst others, during the Mesoproterozoic, Neoproterozoic, Karoo and Cenozoic (Klerkx *et al.*, 1998; Delvaux *et al.*, 2012). However, the discussion of the differences between both groups of complexes is outside the scope of this paper.

Conclusion

The Neoproterozoic Upper Ruvubu Alkaline Plutonic Complex (URAPC) in Burundi consists of three main units: an outer unit with silica-saturated plutonic rocks, an inner unit with silica-undersaturated plutonic rocks, and a carbonatite, each with a specific emplacement age and petrological composition.

This petrological, geochemical and radiogenic (Sr, Nd and Hf) isotope study focuses on the processes that lead to the formation of the URAPC: (1) Fractional crystallization is evidenced by the chemical evolution trends emanating from the major, minor and trace element data, and from clear P, Ti and Ba anomalies appearing in the trace element patterns; (2) crustal assimilation/contamination is suggested by the large range of Nd isotope compositions characterizing the analysed rocks, and the general increase of the Sr isotope ratio with increasing SiO₂ contents, and finally (3) late-magmatic/hydrothermal alteration is consistent with the increase of the Sr isotope composition of the samples without changing significantly their Nd isotope composition.

The two silicate magmatic series of the URAPC (an undersaturated one with minor carbonatite and a saturated one) appear to be cogenetic. The source of the magma(s) could be an old SCLM. The series constituting the outer unit is more contaminated than the inner, undersaturated series. The carbonatite could have evolved by liquid immiscibility.

Unlike many alkaline-carbonatite complexes, the URAPC is rather large-sized. It was emplaced syntectonically under extensional regime. Several structural levels have been identified in its western part: the north-western part of the outer unit, where gabbros predominate, would correspond to a deep zone whilst the granites outcropping in the south-eastern part would correspond to apical zones. The new data provide keys of critical importance for a better understanding of the magmatic differentiation history and late-stage circulations of fluids of alkaline massifs in general.

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Figure and table captions

Figure 1. (A) Geological sketch map of the study area (after Tack *et al.*, 1984, modified and redrawn, and Fernandez-Alonso *et al.*, 2012). The insert shows the location of alkaline complexes occurring along the Western Rift, including the URAPC (after Tack *et al.*, 1984, redrawn). (B) Location of all rock samples from the URAPC reported in this work.

Figure 2. Selected field pictures and macrophotographs of URAPC rocks. (A) An undeformed block of granite in the Buraniro area. (B) A cataclastic granite from the Buraniro area. (C) A gabbroic xenolith in the Buraniro granite. (D) A strongly deformed quartz-bearing syenite (mylonitic facies). (E) A feldspathoidal syenite presenting magmatic layering. (F) A deformed and foliated syenite. (G) Blue sodalite and yellow to pinkish cancrinite in a rock sample collected west of Kiziba (see Fig. 1B).

Figure 3. Photomicrographs of URAPC rocks using plane-polarized (A, G and H-insert) and crossed-polarized (B, C, D, E, F and H) light. Abbreviations are the following: Amp: amphibole, Bt: biotite, Ccn: cancrinite, Cpx: clinopyroxene, Grt: garnet, Kfs: K-feldspar, Nph: nepheline, Ol: olivine, Pl: plagioclase, Pcl: pyrochlore, Qz: quartz, Sdl: sodalite, and Ttn: titanite. (A and B): Hypidiomorphic-granular texture in a gabbro (LT26); the Bowen discontinuous reaction series is illustrated by rims of amphibole and biotite around olivine; C: A mafic cluster in a quartz-bearing syenite (LT566); D: Cataclastic texture in a granite (LT956); E: Hypidiomorphic-granular texture and typical mineral assemblage of a feldspathoidal syenite (LT36); F: Replacement of nepheline by cancrinite in a foid-bearing syenite (LT259); G: Interstitial garnet in a feldspathoidal syenite (LT36); H: Feldspathoidal syenite with layering evidenced by grain size variation and orientation of the constituent minerals. Notice a large perthitic alkali feldspar to the left, and a fine-grained mix of alkali feldspar and nepheline to the right. The insert shows euhedral crystals of titanite and pyrochlore, two accessory minerals frequently appearing in the layered syenites (LT1076).

Figure 4. (A) The analysed rocks in the $\text{Na}_2\text{O}+\text{K}_2\text{O}$ vs. silica classification diagram introduced by Le Maitre *et al.*, 2002; (B-I, K) Binary plots showing major, minor and ΣREE contents vs. SiO_2 , and (J, L and M) plots of Eu/Eu^* ratios vs. SiO_2 , $\text{La}_\text{N}/\text{Yb}_\text{N}$ ratios vs. ΣREE , and ΣREE vs. P_2O_5 , respectively. Data from this work, and from Vergauwen (1990) and Buyschaert (1993).

Figure 5. REE patterns (A-D) and spidergrams (E-H) of URAPC rocks (data from this study, and from Vergauwen, 1990 and Buyschaert, 1993). A and E: rocks of the gabbro-diorite group; B and F: quartz-bearing syenites; C and G: granites; D and H: feldspathoidal syenites. The domain where the Matongo carbonatite (see H) plots is given for comparison purposes

(Midende *et al.*, 2014). Normalization to the chondrites (Sun, 1982, and McDonough, 1990) for the REE, and to the Primitive Mantle (McDonough and Sun, 1995).

Figure 6. (A) Initial ϵNd vs. $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ and (B) Initial ϵHf vs. $\epsilon\text{Nd}_{(i)}$ isotopic diagrams for the URAPC samples. Data related to the Matongo carbonatite are from Demaiffe (2008).

Figure 7. Correlation between (A) $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ and SiO_2 content, and (B) between $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ and $1/\text{Sr}$. Data from this study and from Tack *et al.* (1984) for the Sr isotopic data, and from this study, Vergauwen (1990) and Buyschaert (1993) for the SiO_2 contents (cf. Tables 2, 3 and 4).

Figure 8. (A) Initial $^{143}\text{Nd}/^{144}\text{Nd}_{(i)}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$ diagram for the URAPC samples. Data with regard to the Matongo carbonatite are from Demaiffe (2008). The country-rocks field is defined after data from Tack *et al.* (1994) and Debruyne *et al.* (2015). (B) Initial ϵHf vs. ϵNd for the URAPC samples. Fields reported for comparison: Depleted Mantle (DM), Modern OIB, HIMU Reservoir, and a Neoproterozoic (582 Ma) carbonatite suite associated with ultramafic lamprophyres (UML) exposed along the Labrador Sea (Tappe *et al.*, 2007); SCLM xenolith from Mingxi peridotite (Liu *et al.*, 2017), Somerset Island kimberlites (Schmidberger *et al.*, 2002), and Kaapvaal Craton peridotites (Simon *et al.*, 2007).

Table 1. Succinct description of all examined URAPC samples.

Table 2. Major element composition (in wt. %) of the studied URAPC rocks. Analyses carried out at the ULB “G-Time” Laboratories. Reference numbering is as follows: (*) This study, the others are from Buyschaert (1993) and Vergauwen (1990).

Table 3. Trace element content (in ppm), La_N/Yb_N ratio, Ce^*/Ce ratio, Eu^*/Eu ratio and ΣREE contents of the examined URAPC rocks. Analyses carried out at the ULB “G-Time” Laboratories. Reference numbering is as follows: (*) This study and from Buysschaert (1993) and Vergauwen (1990).

Table 4. Measured and initial Sr, Nd and Hf isotopic composition of the studied URAPC rocks. Initial ratios are recalculated at 705 Ma (Midende *et al.*, 2014). Analyses were carried out at the ULB “G-Time” Laboratories. Reference numbering is as follows: (*) This study, the others are from Tack *et al.* (1984).

Sample no.	RMCA code	Coordinates (Long.; Lat.)	Rock type
Outer Unit			
LT 1	RG134240	29°33'52.0"E; 2°58'46.3"S	Monzodiorite
LT 26	RG134262	29°34'12.7"E; 2°57'06.8"S	Gabbro
LT 95	RG134324	29°33'29.6"E; 3°00'21.4"S	Gabbro in contact with quartz-bearing syenite
LT 210	RG134425	29°38'03.1"E; 2°58'58.8"S	Gabbro
LT 317	RG134525	29°32'55.3"E; 2°58'06.2"S	Gabbro
LT 340	RG134546	29°34'09.1"E; 2°57'09.5"S	Quartz-bearing monzogabbro
LT 342	RG134548	29°32'26.0"E; 2°57'15.0"S	Diorite
LT 347	RG134552	29°32'22.0"E; 2°57'14.6"S	Diorite
LT 350	RG134554	29°32'17.3"E; 2°57'36.0"S	Diorite
LT1006	RG144616	29°39'14.8"E; 3°05'49.8"S	Gabbro
LT 1017	RG144626	29°40'10.7"E; 3°05'33.9"S	Quartz monzodiorite
LT 1061	RG144670	29°37'47.9"E; 3°05'36.2"S	Quartz monzodiorite
GPS 9+		29°38'48.5"E; 3°02'2.3"S	Gabbro
GPS 11		29°38'57.5"E; 3°02'18.0"S	Gabbro
GPS 38		29°32'37.3"E; 2°57'09.4"S	Gabbro
GPS 80		29°37'55.1"E; 3°0'34.1"S	Gabbro
LT 60	RG134291	29°33'34.5"E; 2°58'52.6"S	Quartz-bearing syenite
LT 288	RG134496	29°33'45.3"E; 2°57'16.4"S	Quartz-bearing syenite
LT 318	RG134526	29°32'59.0"E; 2°57'57.4"S	Quartz-bearing syenite
LT 693	RG134685	29°33'38.2"E; 2°59'53.3"S	Quartz-bearing syenite
LT 993	RG144605	29°34'22.0"E; 3°01'21.0"S	Quartz-bearing syenite
LT 1133	RG144738	29°32'58.6"E; 2°58'06.4"S	Quartz-bearing syenite
LT 485	RG134681	29°39'47.6"E; 3°06'37.3"S	Quartz-bearing syenite
LT 486	RG134682	29°39'03.1"E; 3°02'29.2"S	Quartz-bearing syenite
LT 566	RG134760	29°38'45.6"E; 3°01'35.8"S	Quartz-bearing syenite
LT 1041	RG144650	29°39'04.9"E; 3°02'28.4"S	Quartz-bearing syenite
GPS 6		29°38'47.3"E; 3°01'24.1"S	Quartz-bearing syenite
GPS 9		29°38'47.0"E; 3°01'37.2"S	Quartz-bearing syenite
GPS 36		29°32'34.4"E;	Strongly albitized quartz-bearing syenite

2°57'08.4"S

LT 96	RG134325	29°33'29.2"E; 3°00'22.3"S	Granite
LT 321	RG134528	29°32'17.5"E; 2°57'55.2"S	Granite
LT 956	RG144569	29°33'29.9"E; 3°00'22.0"S	Granite
LT 1011	RG144621	29°39'47.6"E; 3°06'37.3"S	Granite
LT 1111	RG144716	29°37'54.3"E; 3°02'38.2"S	Granite
LT 1112	RG144717	29°37'54.7"E; 3°02'38.4"S	Granite
LT 929	RG144542	29°33'38.2"E; 2°59'53.3"S	Granite
LT 977	RG144590	29°34'03.3"E; 3°01'05.2"S	Granite
LT 1034	RG144643	29°39'38.9"E; 3°03'15.4"S	Granite
Inner Unit			
LT 36	RG134268	29°36'07.4"E; 2°58'50.8"S	Feldspathoidal syenite - hypidiomorphic-granular
LT 46	RG134278	29°36'53.9"E; 2°58'21.5"S	Feldspathoidal syenite - trachytic
LT 105	RG134333	29°37'51.7"E; 2°56'54.4"S	Feldspathoidal syenite - hypidiomorphic-granular
LT111	RG134338	29°37'25.8"E; 2°58'10.6"S	Feldspathoidal syenite - hypidiomorphic-granular
LT134	RG134360	29°35'39.3"E; 2°57'42.6"S	Feldspathoidal syenite - hypidiomorphic-granular
LT629	RG134820	29°35'35.6"E; 3°01'43.1"S	Feldspathoidal syenite - hypidiomorphic-granular
LT253	RG134465	29°36'46.9"E; 3°01'24.4"S	Feldspathoidal syenite - hypidiomorphic-granular
LT 259	RG134471	29°36'35.6"E; 3°01'55.8"S	Feldspathoidal syenite - hypidiomorphic-granular
LT479	RG134675	29°36'53.2"E; 3°02'27.8"S	Feldspathoidal syenite - hypidiomorphic-granular
LT512	RG134708	29°35'00.9"E; 2°57'32.0"S	Feldspathoidal syenite - porphyritic
LT590	RG134782	29°36'02.3"E; 3°01'48.5"S	Feldspathoidal syenite - hypidiomorphic-granular
LT595	RG134787	29°36'00.9"E; 3°01'35.5"S	Feldspathoidal syenite - porphyritic
LT596	RG134788	29°36'01.0"E; 3°01'34.9"S	Feldspathoidal syenite - porphyritic
LT623	RG134814	29°35'12.7"E; 3°01'30.8"S	Feldspathoidal syenite - porphyritic
LT1076	RG144684	29°36'38.0"E; 3°00'09.2"S	Feldspathoidal syenite - layered
GPS 49		29°33'19.9"E; 2°57'14.1"S	Feldspathoidal syenite

Sample	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	Fe ₂ O ₃
Outer Unit					
LT1	54.74	0.95	20.80	4.53	1.83
LT26	46.67	2.15	17.91	8.55	2.56
LT 95*	52.15	1.13	13.23	15.45	
LT210	53.91	1.80	16.72	8.95	1.61
LT317	56.46	1.03	16.06	5.39	1.65
LT340	46.42	2.62	17.95	8.01	2.65
LT342	43.83	4.00	14.31	11.85	3.24
LT347	57.96	0.80	19.28	4.44	1.75
LT350	56.21	1.27	20.97	3.93	1.64
LT1006	49.76	1.50	15.50	7.41	1.43
LT1017	54.61	1.62	16.24	6.11	2.38
LT1061	58.70	1.70	15.06	6.90	2.43
GPS 9+*	58.99	1.37	14.79	9.12	
GPS 11*	51.65	2.39	14.78	11.96	
GPS 38*	49.93	2.55	14.10	13.09	
GPS 80*	46.57	3.21	13.94	14.99	
gabbro-diorite group					
LT60	60.76	1.14	16.63	5.16	1.18
LT288	67.43	0.39	14.59	0.40	4.17
LT318	64.95	0.31	16.44	2.01	2.88
LT693	69.87	0.25	15.05	1.13	1.39
LT993	66.59	0.32	16.12	2.01	1.14
LT1133	68.59	0.33	15.63	2.21	0.75
LT485	64.66	0.52	16.46	0.51	5.41
LT486	62.31	0.57	17.22	4.31	1.75
LT566	61.51	0.90	16.37	0.94	6.33
LT1041	64.86	0.65	15.42	3.86	1.67
GPS 6*	67.36	0.27	17.17	2.63	
GPS 9*	64.11	0.43	16.27	4.66	
GPS 36*	60.49	0.50	19.50	5.80	
quartz-bearing syenite					
LT96	77.01	0.13	12.66	0.03	1.20
LT321	74.83	0.13	11.64	0.48	1.03
LT956	75.91	0.11	12.46	0.46	0.93
LT1011	73.29	0.26	13.51	1.38	0.84
LT1111	70.92	0.31	13.86	2.27	0.94
LT1112	69.53	0.16	14.05	0.79	4.31
LT929*	75.69	0.10	13.63	1.77	
LT 977*	74.49	0.11	12.50	1.06	
LT 1034*	72.68	0.33	12.91	3.25	
granite					

Inner Unit

LT36		56.74	0.17	20.88	2.64	1.94
LT46*		58.25	0.08	20.84	4.16	
LT105		57.53	0.06	20.13	1.08	3.75
LT111		45.96	2.88	14.40	9.85	3.74
LT134		46.42	1.72	14.74	7.33	2.85
LT629		50.26	1.11	20.72	4.75	2.58
LT253		47.49	1.27	20.97	4.77	2.99
LT479*	feldspathoidal syenite	54.68	0.40	13.60	12.13	
LT512*		62.11	0.34	18.66	4.57	
LT590*		61.17	0.25	18.14	4.87	
LT595		57.14	0.06	21.95	0.66	0.73
LT596*		58.16	0.15	20.32	2.68	
LT623		56.87	0.24	21.70	1.49	1.12
GPS 49*		59.31	0.33	19.36	5.76	
BHVO-2		50.44	2.74	13.55		12.80
AGV-2		59.33	0.97	16.89		6.55

* Analysed at the Laboratoire G-Time-ULB. Reproducibility is better than 2%.

Other analyses are from Vergauwen (1990) for the Inner Unit, and from Buyschaert (1993) for the Outer Unit

Sample	V	Rb	Sr	Y	Zr	Ni
Outer Unit						
LT1	34	74	878	23	188	60
LT26	121	22	950	24	120	51
LT 95*		5.5	103	25	22.3	4.1
LT340	156	34	903	28	124	42
LT342	132	60	525	40	176	74
LT347	21	87	611	30	330	101
LT350	49	79	931	23	130	49
LT1006	172	41		19	133	29
LT1017	160	102		31	209	52
LT1061	116	134		47	283	81
GPS 9+*		134	311	42	310	68
GPS 11*		92	393	37	113	91
GPS 38		70	416	45	46	48
GPS 80*		66	348	39	113	65
gabbro-diorite group						
LT60	75	163	416	41	271	73
LT288	0.79	167	29	438	699	131
LT318	11	183	142	46	437	82
LT693	10	139	102	62	489	101
LT993	1.2	170		121	748	101
LT1133	5.5	189	72	58	483	111
LT485	2.4	116	56	45	996	101
LT486	8.6	128	132	64	1104	111
quartz-bearing syenite						

LT566		9.8	145	243	81	711	12
LT1041		19	175		215	402	83
GPS 6*			83	202	39	393	58
GPS 9*			134	219	51	330	10
GPS 36*			124	38	7.9	62	30
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LT96		1.2	207	31	56	213	90
LT321		2.2	176	19	46	205	68
LT956		1.5	199	32	66	183	81
LT1011		9.4	216	88	91	238	50
LT1111	granite	3.3	166		63	389	10
LT1112		21	108		43	534	13
LT929*			311	32	52	176	13
LT 977*			321	17	50	199	9
LT 1034*			228	103	58	246	5
<hr/>							
Inner Unit							
LT36		5.2	133	63	9.0	136	9
LT46*			151	51.88	19	111	5
LT105		0.8	225	17	25	549	18
LT253			169	1476	33	221	20
LT479*				203	548.17	44	35
LT512	feldspathoidal syenite	2.2	192	75	31	412	19
LT590		4.4	164	305	19	162	13
LT595		4.2	165	133	13	544	6
LT596		7.6	125	262	8.2	744	35
LT623		4.7	132	112	10	116	10
GPS 49*				131	59.63	20	14
<hr/>							
BHVO-2			8.97	393	25.75935694	169	17
AGV-2			64	662	20	231	14

Chondrite normalization values (Sun, 1982; McDonough, 1987)

Primary mantle normalization values (McDonough and Sun, 1995)

0.6 19.9 4.3 10.5 0.

* Analysed at the Laboratoire G-Time-ULB

Other analyses are from Vergauwen (1990) for the Inner Unit, and from Buyschaert (1993) for the Outer Unit

	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$ (measured)	2σ	$^{87}\text{Sr}/^{86}\text{Sr}_{(t)}$ (at 705 Ma)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$ (measured)	
Outer Unit							
GPS 9+*	1.028	0.719789	0.00001	0.707204385	0.104	0.511	
GPS 11*	0.557	0.713750	0.00001	0.706913311	0.116	0.512	
GPS 38*	0.382	0.709645	0.00001	0.704761304	0.114	0.512	
GPS 80*	gabbro-diorite group	0.458	0.711239	0.00001	0.705735144	0.116	0.512
LT 26		0.074	0.704600		0.703855465		
LT 95*		0.068	0.709259	0.00001	0.707689422	0.163	0.512
LT347		0.398	0.708010	0.00007	0.704007114		0.512
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LT693*		0.012	0.704130	0.00001	0.664467283	0.167	0.512
GPS 6*		0.978	0.720260	0.00001	0.708244255	0.086	0.511
GPS 9*	quartz-bearing syenite	1.458	0.725694	0.00001	0.70781169	0.113	0.512
GPS 36*		8.446	0.804757	0.00002	0.7096342	0.112	0.512
LT 566		1.754	0.725300	0.00003	0.707656446		0.512
LT485		6.000	0.759400		0.699032334		
<hr/>							
LT 96		27.300	0.981500		0.706827119		
LT929*		22.744	0.953237	0.00002	0.666032607	0.105	0.512
LT 956		0.126	0.868180		0.866912279		
LT 977*	granite	45.161	1.180739	0.00003	0.589423074	0.136	0.512
LT 996		4.026	0.750300	0.00009	0.709796453		0.512
LT1011*		6.961	0.793455	0.00002	0.721386129	0.108	0.512
LT 1034*		5.285	0.773699	0.00001	0.708834933	0.100	0.511
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Inner Unit							
LT46*		7.637	0.793742	0.00002	0.708058258	0.083	0.512
LT479*		0.882	0.713914	0.00001	0.703136499	0.076	0.512
LT512*		5.764	0.783991	0.00002	0.708954084	0.114	0.512
LT590*		1.269	0.720293	0.00002	0.704649186	0.089	0.511
LT596*		1.105	0.719544	0.00001	0.705617648	0.081	0.512
LT404	feldspathoidal syenite	1.310	0.718600	0.00060	0.705417496	0.083	0.512
LT36				0.767300		0.704417014	
LT105			1.063900		0.661448892		
LT253			0.706240		0.703593884		
LT595			0.721800		0.683969596		
LT623			0.738100		0.704193494		
GPS 49*		5.122	0.769123	0.00002	0.704994619	0.095	0.512

* Analysed at the Laboratoire G-Time-ULB

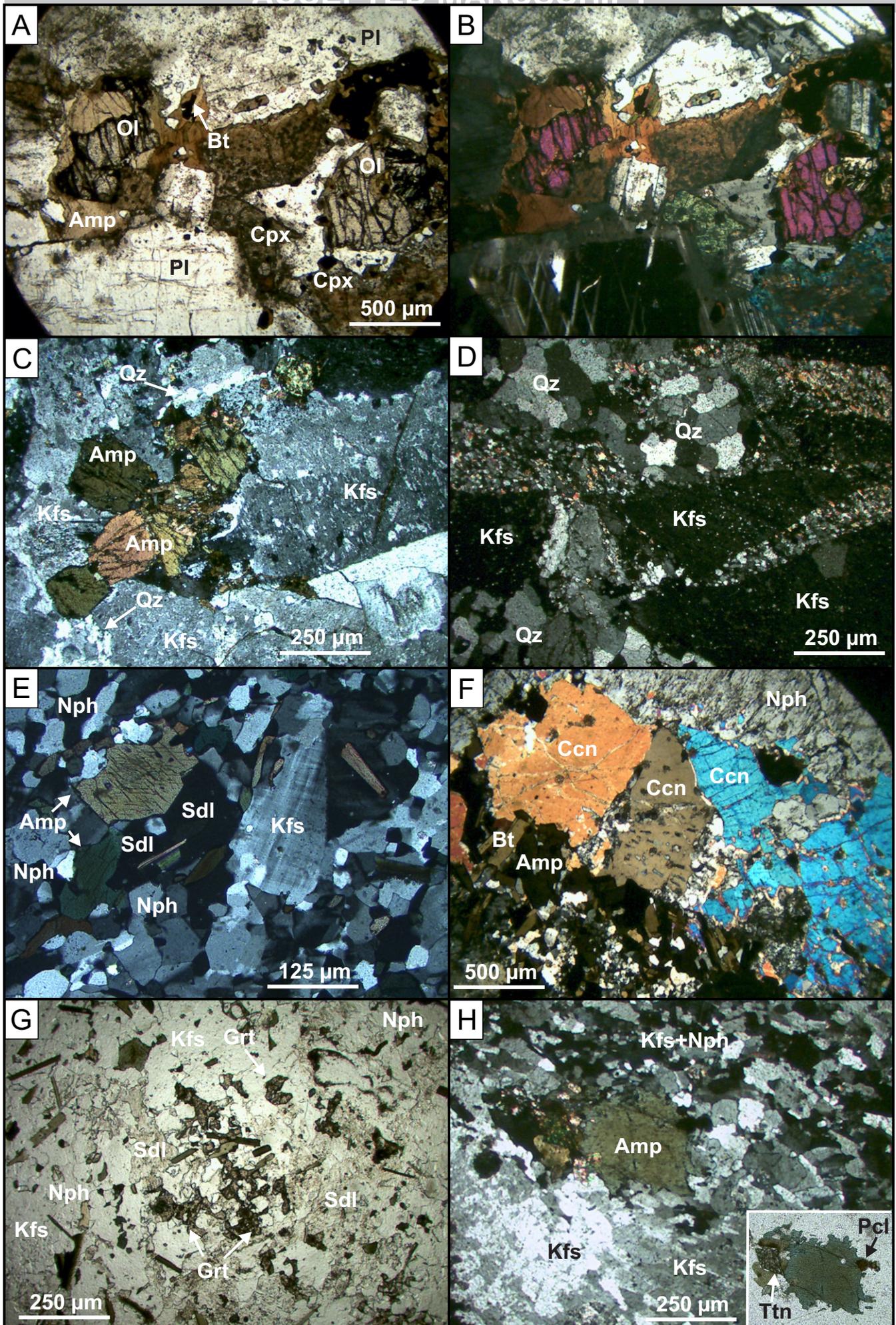
Other analyses are from Tack *et al.* (1984)

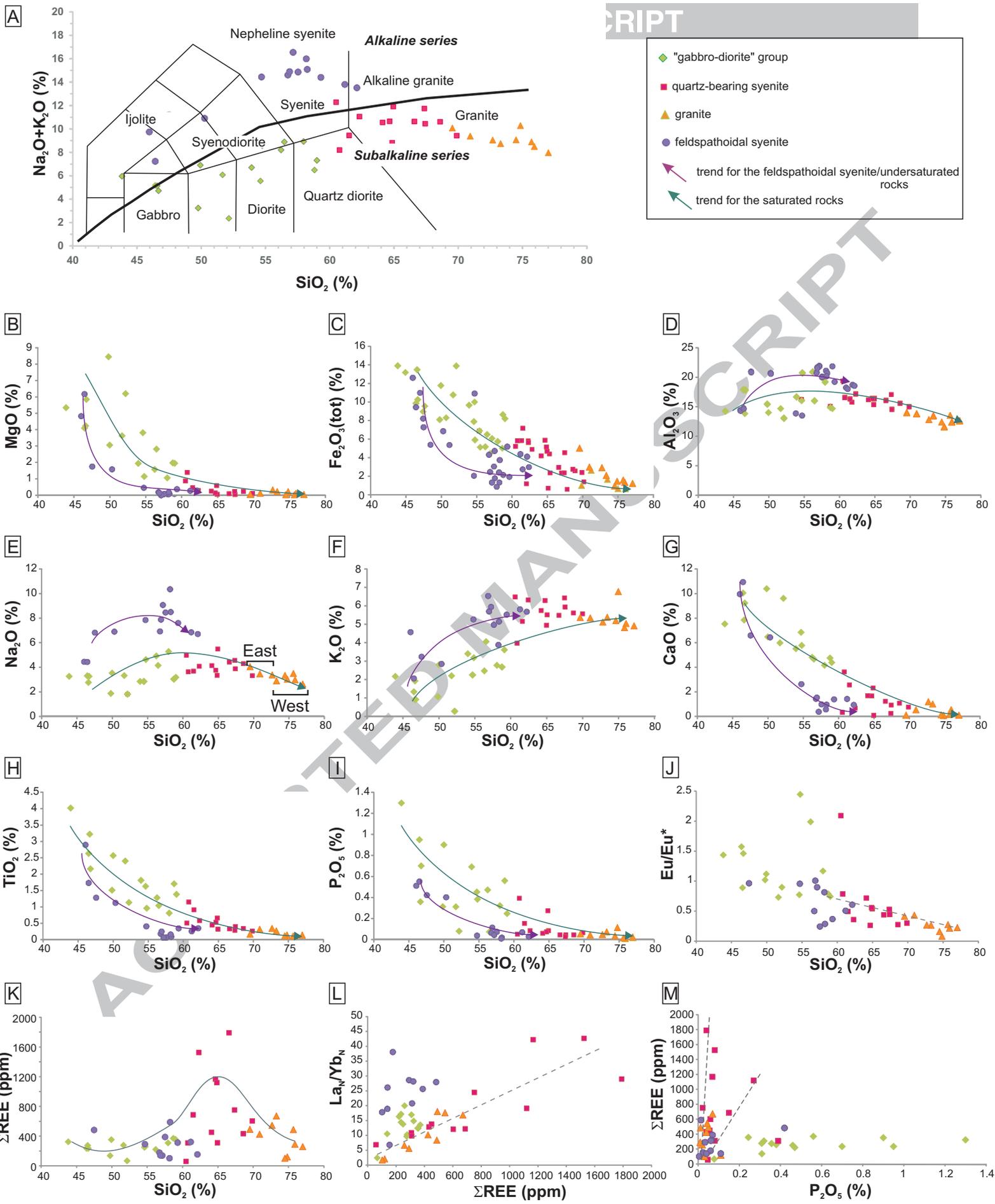
Ratios and decay constants used: $(^{147}\text{Sm}/^{144}\text{Nd})_{\text{CHUR}}^{\text{present day}} = 0.1967$; $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}}^{\text{present day}} = 0.512638$; $(^{176}\text{Lu}/^{177}\text{Yb})_{\text{CHUR}}^{\text{present day}} = 0.0243$

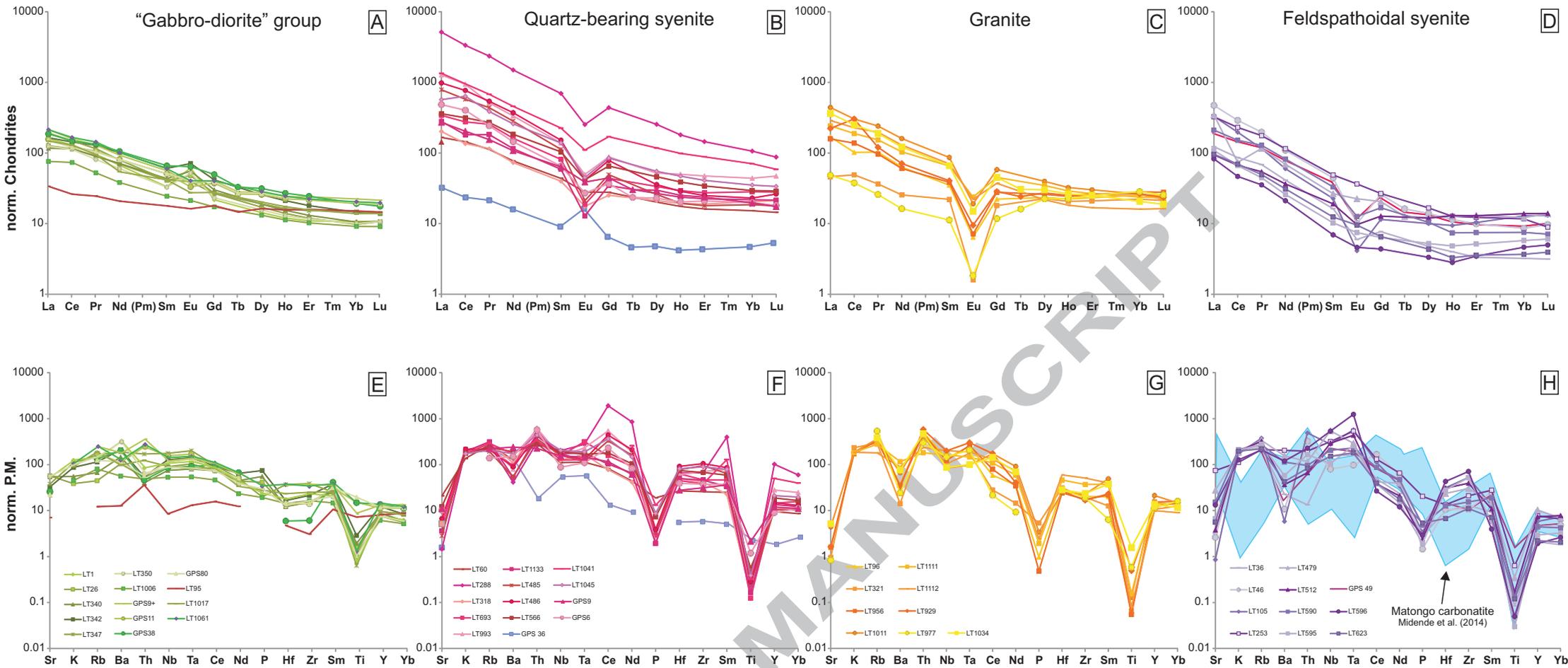
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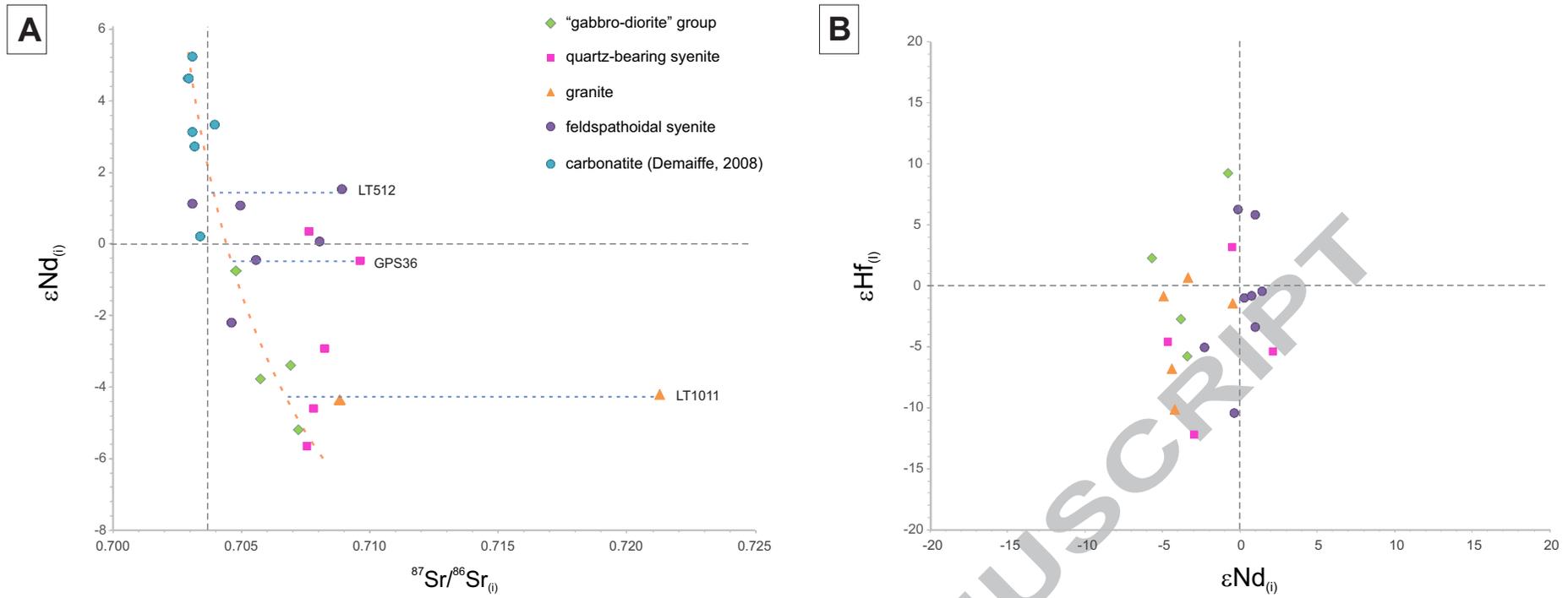
- This study focuses on the syntectonic Neoproterozoic Upper Ruvubu Alkaline Plutonic Complex (URAPC, Burundi)
- The petrological, geochemical and isotope study emphasizes the processes leading to its formation
- Fractional crystallization, crustal assimilation/contamination, and late-magmatic/hydrothermal alteration are evidenced
- The two silicate magmatic series and the carbonatite appear cogenetic

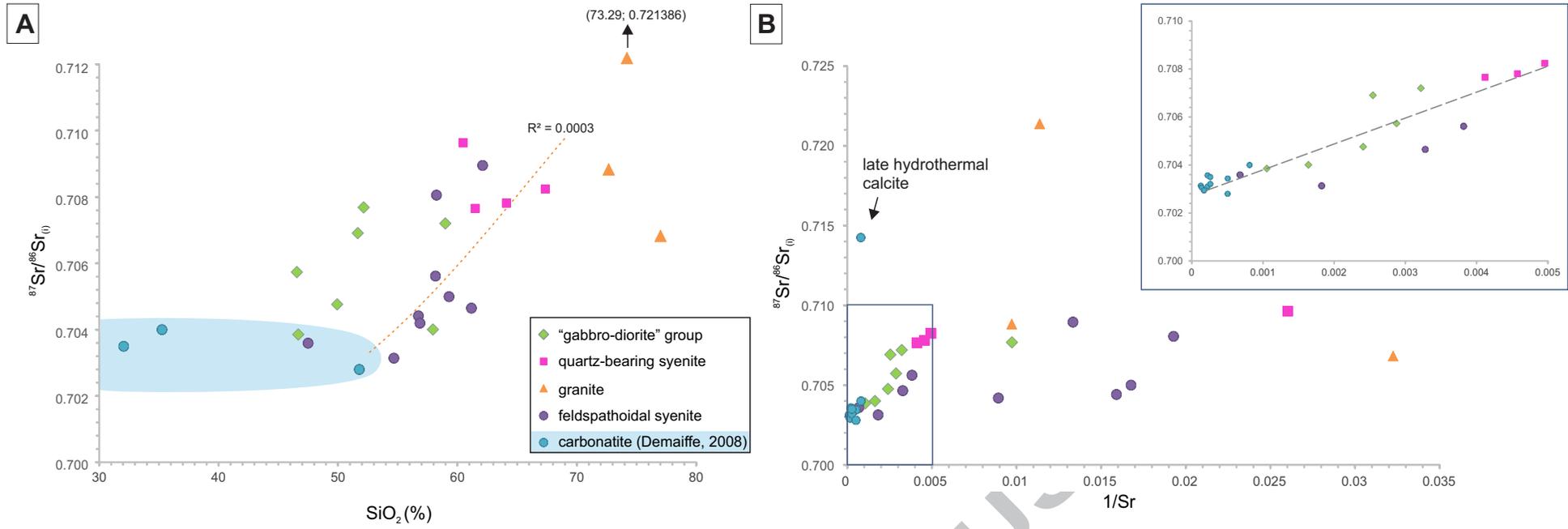




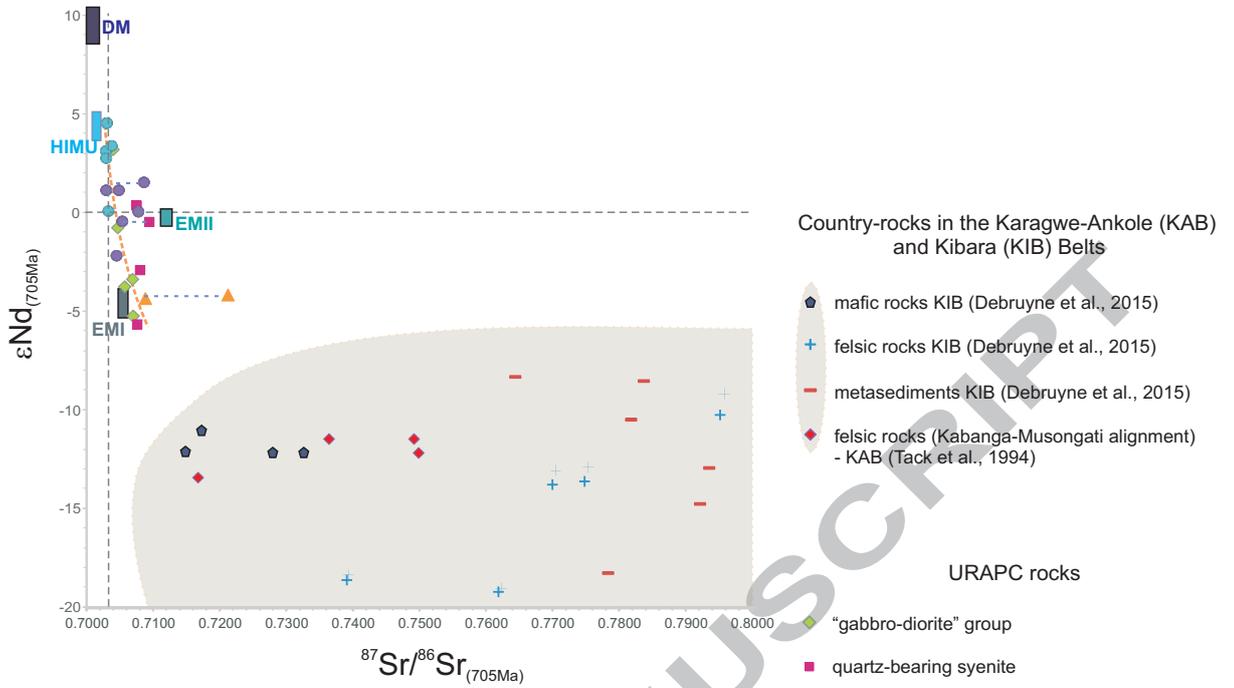








A



B

