U–Pb zircon geochronology and petrology of granitoids from Mitwaba (Katanga, Congo): implications for the evolution of the Mesoproterozoic Kibaran belt

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Abstract

In Katanga (Congo), Mesoproterozoic Kibaran granitoids were previously divided into: (1) type A or G₁ granitoids (1370 ± 25 Ma); (2) type B or G₂ granitoids (1308 ± 25 Ma); (3) type C or G₃ granitoids (1094 ± 50 Ma); and (4) type D or G₄ tin granites (ca. 977 ± 18 Ma). Our field investigations indicate that igneous bodies previously described as G₁ (A), G₂ (B) and G₃ (C) granitoids in the Mitwaba region were emplaced during the same early Kibaran compressional orogenic event (D₁). The heterogeneous distribution of strain during D₁ produced orthogneissic rocks along high strain zones versus slightly to unfoliated granitoids in low strain zones. New SHRIMP U–Pb zircon analyses yield the following igneous crystallisation ages for Mitwaba G₁–G₃ granitoids: Kisele monzogranite gneiss 1386.3 ± 8.1 Ma, Kabonvia granodioritic augen gneiss 1385.5 ± 7.4 Ma, Nyangwa monzogranite 1383.3 ± 4.8 Ma, Kungwe-Kalumengongo monzogranite 1377 ± 10 Ma and Fwifwi foliated leucomonzogranite 1372 ± 9.6 Ma. All these granitoids were emplaced over a short period of time and could be coeval at 1381 ± 8 Ma. This age dates the early Kibaran orogenic event (D₁) in the Mitwaba region. Metamorphic zircons from the strongly deformed Kisele monzogranite gneiss yield an age of 1079 ± 14 Ma, which possibly dates the D₂ orogenic event in the Mitwaba area. The tin granites are younger and were emplaced at ~1.0 Ga.

The Mitwaba granitoids are two-micas and ilmenite-bearing strongly peraluminous S-type granitoids. The rare-earth elements (REE) patterns show a strong Eu negative anomaly, an enrichment and fractionation of light REE (LREE) and slightly fractionated to almost flat Heavy REE (HREE) patterns. Primordial-mantle normalized spidergrams are characterized by negative anomalies of Ti, Sr, Nb and subsidiary Ba. The emplacement of arc mafic magmas into the crust triggered the dehydration melting, at high temperatures, of a highly fertile and heterogeneous metasedimentary source material producing a felsic melt which was mixed with the arc mafic magmas in variable proportions to produce these granitoids. The petrogenesis of these rocks and similar coeval occurrences defining a ca. 3000 km granitoid belt stretching from SE Congo to NE Congo, Burundi, Rwanda and Uganda indicates similar processes throughout the Kibaran orogenic system in central and eastern Africa.

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1. Introduction

The Kibaran belt represents one of the few well-preserved Mesoproterozoic belts in the world (Kampunzu, 2001). The orogenic history of this belt is poorly constrained in the Democratic Republic of Congo (hereafter Congo), because of the lack of modern structural, petrological, geochemical and precise geochronological data. The type area of the Kibaran orogenic system of Africa, known as the Kibaran belt in Katanga, Congo (Cahen et al., 1984 and references therein), runs over 700 km with an average width of ~300 km. The total Kibaran orogenic system of central and southern Africa is more than 3000 km long and hosts large volume of granitoids previously suggested to be emplaced along a plate convergent margin (Kampunzu et al., 1986) or to mark an extensional rift setting (e.g. Klerkx et al., 1987). To evaluate these

conflicting interpretations, we conducted multidisciplinary studies of granitoids in the Mitwaba region of SE Congo. This paper provides: (1) the first precise U–Pb zircon crystallisation ages of a representative set of intrusive bodies allowing constraints to be placed on the timing of deformation and emplacement of granitoid batholiths in the Kibaran type-area; (2) petrological data constraining the genesis and tectonic setting during the emplacement of granitoids.

2. Regional geology and previous geochronology

The Kibaran orogenic system of Africa includes the following belts (Fig. 1): Kibaran (SE Congo), Burundian (E Congo, Rwanda, and Burundi), Karagwe-Ankolean (SE Uganda and NW Tanzania), Luhule Mo-bissio and Maniema (NE Congo), Irumides (Zambia),

Fig. 1. Sketch map of the Kibaranide belt in Central Katanga (Congo) showing the distribution of Mesoproterozoic granitoids, mafic igneous and metasedimentary rocks. The upper inset shows the Mesoproterozoic Kibaran belts in southern Africa and the pre-Kibaran basements. C, Congo, T, Tanzania, B, Bangweulu and K, Kalahari cratons.

Namaqua-Sinclair-Rehoboth (Namibia), Namaqua-Natal (South Africa) and Lurio (Mozambique-Malawi). The Kibaran belt defines a NE-trending belt exposed in SE Congo (Fig. 1). It is bounded by the Archaean/Paleoproterozoic Congo craton to the northwest. To the southeast, it is in contact with a mosaic of older crustal provinces amalgamated before the Kibaran orogeny, including the Palaeoproterozoic Bangweulu Block (Kapenda et al., 1998) and the Archaean Tanzania craton (Fig. 1). There are no geological similarities between the rock assemblages of the Bangweulu block/Tanzania craton and those of the Congo craton on the two margins of the Kibaran belt.

The Kibaran Supergroup is a thick supracrustal sedimentary succession (>10,000 m; Cahen et al., 1984), consisting mainly of metamorphosed pelites, siltstones, sandstones and conglomerates, with minor carbonate rocks. Large felsic and mafic igneous bodies intrude these metasedimentary rocks (e.g. Cahen, 1954; Kampunzu et al., 1986). Kibaran metasedimentary and igneous rocks were affected by two main compressional events (e.g. Kampunzu et al., 1986, 1998; Rumvegeri, 1991). The timing of these contractional events was inferred to be in the 1400–1000 Ma bracket (e.g. Cahen et al., 1984). Kampunzu et al. (1998) consider that the first event marks the development of an active continental margin, between 1.4 and 1.25 Ga. The second event is taken to mark a continental collision (1.25–1.0 Ga), with production of late- to post-kinematic “fertile” granites.

The paroxysmal Kibaran compressional deformation (D₂) generated NE-trending and NW-verging isoclinal folds and thrusts, defining the main physiographic feature of the belt over more than 3000 km, from Katanga in SE Congo to Uganda (Cahen et al., 1984; Kampunzu et al., 1986; Rumvegeri, 1991). Early Kibaran granitoids and related orthogneisses were emplaced during a protracted period between 1.4 and 1.25 Ga (e.g. Cahen et al., 1984; Kampunzu et al., 1998). This suggestion was mainly based on: (1) low-precision whole-rock Rb–Sr isochron and K–Ar dates (e.g. Cahen et al., 1984); (2) conventional U–Pb bulk zircon dates (e.g. Ledent, 1979).

In the Kibaran belt, four major granitoid groups were defined by previous workers (e.g. Boutakoff, 1939; Cahen et al., 1984): (1) type A or G₁ granitoids yielded a Rb–Sr whole-rock isochron of 1331 ± 25 Ma in the Mitwaba area. For the whole Kibaran orogenic

system, Cahen et al. (1984) suggested that an average age of 1370 ± 25 Ma could be considered for this group; (2) type B or G₂ granitoids for which an average age of 1308 ± 25 Ma was proposed; (3) type C or G₃ granitoids inferred to be emplaced at around 1094 ± 50 Ma; and (4) type D or G₄ tin granites emplaced at ca. 977 ± 18 Ma. This subdivision was followed, with minor modifications, by several authors (e.g. Klerkx et al., 1987; Fernandez-Alonso et al., 1986; Tack et al., 1994), despite the fact that it was poorly constrained by field data and based on low-precision geochronological data. Kampunzu et al. (1986), using a large amount of structural and petrofabric data, suggested that G₁, G₂ and G₃ Kibaran granitoids were part of the same syn-D₁ igneous event. According to these authors, the stronger versus weaker solid-state fabric recorded in G₁ versus G₃ (respectively) result from the heterogeneous distribution of strain within the belt instead of reflecting any absolute age difference. The new geochronological data presented in this paper allow us to constrain these conflicting interpretations.

3. Geology of the Mitwaba area

Two informal lithostratigraphic units separated by a disconformity were defined within the Kibaran supracrustal metasedimentary sequences in the Mitwaba area (Fig. 2): the older Mitwaba group and the younger Mulumbi group (Kokonyangi et al., 2001). The Mitwaba group includes metapelitic rocks, with minor intercalations of quartzite. This supracrustal metasedimentary assemblage is intruded by gabbros and granitoids. Two main mafic igneous bodies occur within the Mitwaba group (Fig. 2): (1) the Lwabwe metagabbro, enveloping the western margin of the Kifinga-Kisele granitic batholith, intrudes (cf. chilled margin) greenschist facies Mitwaba group metasedimentary rocks; and (2) the Kidilo amphibolite intruding the Mitwaba group supracrustal assemblage to the southeast of this batholith. The contact between the metagabbro and the granitoids is locally well exposed (e.g. near Komeshia), a sharp intrusive contact is observed and the granite contains metagabbro enclaves. The metagabbro is affected by contact metamorphism (static biotite-garnet-hornblende porphyroblasts) and hydrothermal silicification near the granite body (Kokonyangi et al., 2001).

Fig. 2. Geological map of the Mitwaba degree-sheet showing the studied granitoids and associated mafic igneous and sedimentary rocks.

Geochemical characteristics of the Mitwaba mafic igneous rocks indicate a continental arc signature and these mafic rocks yielded U–Pb single zircon igneous crystallisation ages between 1.4 and 1.38 Ga (Kokonyangi, unpublished data), supporting field studies indicating that they are closely related to the emplacement of the granitoids.

Two main deformation events (D_1 and D_2) affected the Mitwaba group supracrustal metasedimentary units and related igneous rocks. The oldest contractional deformation (D_1) is characterized by ENE-trending penetrative structures, asymmetric mesoscopic folds (F_1) with an axial planar foliation (S_1) steeply dipping (50 – 60°) to the NW or SE. This fabric is well developed in upper amphibolite facies rocks of the Mitwaba group to the southeast. To the northwest, the S_1 fabric occurs within the Mitwaba group and is totally absent within the Mulumbi group lithologies (Fig. 2).

The Mulumbi group is made of: (1) the Kataba conglomerate containing pebbles from the underlying Mitwaba group metasedimentary rocks and related orthogneisses; (2) quartzites with conglomerate intercalations; and (3) metapelites. The studied granitoids and the mafic igneous bodies do not intersect the Mulumbi group metasedimentary rocks. Therefore, the siliciclastic rocks of the Mulumbi group were deposited after D_1 and after the emplacement of the igneous rocks intruding the Mitwaba group. The Mulumbi group rocks are affected by the regional D_2 Kibaran compressional structures, e.g. NE-trending isoclinal folds (F_2) and related axial planar foliation (S_2) forming the main physiography of the Kibaran orogenic system in the type-area. Mineral parageneses defining this NE-trending foliation indicate an increase in metamorphic grade from greenschist facies rocks in the northwest, to amphibolite facies to the southeast, including staurolite, kyanite and sillimanite-bearing metasedimentary rocks (e.g. Kokonyangi et al., 2001).

Granitic rocks exposed in the Mitwaba area (Fig. 2) form the Kifinga-Kisele batholith and subordinate satellite bodies, including Nyangwa, Kungwe-Kalumengongo, Kisandji and Fubwe plutons. Among them, the Kifinga-Kisele batholith, Nyangwa and Kungwe-Kalumengongo plutons were investigated for this study. The Kifinga-Kisele batholith includes

lations; and (3) metapelites. The studied granitoids and the mafic igneous bodies do not intersect the Mulumbi group metasedimentary rocks. Therefore, the siliciclastic rocks of the Mulumbi group were deposited after D_1 and after the emplacement of the igneous rocks intruding the Mitwaba group. The Mulumbi group rocks are affected by the regional D_2 Kibaran compressional structures, e.g. NE-trending isoclinal folds (F_2) and related axial planar foliation (S_2) forming the main physiography of the Kibaran orogenic system in the type-area. Mineral parageneses defining this NE-trending foliation indicate an increase in metamorphic grade from greenschist facies rocks in the northwest, to amphibolite facies to the southeast, including staurolite, kyanite and sillimanite-bearing metasedimentary rocks (e.g. Kokonyangi et al., 2001).

Fig. 3. (a) Composition of biotites from the Mitwaba granitoids in the diagram $\text{Fe}/(\text{Fe} + \text{Mg})$ vs. Al . (b) Discrimination diagram for biotites from granitoids using FeO^* and Al_2O_3 (Abdel-Rahman, 1994). Most biotites from Mitwaba granitoids fall in the field for peraluminous S-type granitoids. See text for details.

three distinct plutons: (1) Kabonvia granodioritic augen gneiss; (2) Kisele monzogranite gneiss; and (3) Fwifwi foliated leucomonzogranite (Fig. 2). The main petrographic and mineralogical features of these intrusive bodies are summarized in Table 1. The biotites from Mitwaba granitoids display chemical characteristics of biotites from peraluminous granitoids (Fig. 3); their relatively high Al and low Fe contents are similar to those reported in peraluminous granites derived from partial melting of sedimentary rocks (Abdel-Rahman, 1994). Large euhedral muscovite crystals imbricated or in parallel growth with primary biotite and containing inclusion of accessory magmatic minerals contain high Ti, Fe and Si atomic values pointing towards their magmatic origin (e.g. Speer, 1984). Ilmenite occurs in all granitoids and rutile (99% TiO_2) is included in biotite and its alteration product chlorite.

All the plutons were affected by D_2 deformation event and related metamorphism M_2 . However, as stated above, the grade of M_2 is variable within the belt; the Fwifwi Nyangwa and Kungwe-Kalumengongo plutons were metamorphosed under greenschist facies (chlorite zone) whereas the Kabonvia granodioritic augen gneiss (staurolite zone) and Kisele monzogranite gneiss (sillimanite zone) were metamorphosed under amphibolite facies P-T con-

ditions. All the plutons exhibit sharp contact with Mitwaba group metasedimentary rocks and contain abundant foliated xenoliths of country rocks close to the contact.

Near the margins of the intrusive bodies, the S_1 foliations in the granitoids and in the metasedimentary country rocks are parallel. These granitoids contain foliated biotite-muscovite rich restitic clots (surmicaceous enclaves of Didier, 1973). The central parts of the plutons show weaker foliation whereas the margins of the granitoid bodies display a stronger solid-state fabric (e.g. Fig. 2). The contact aureole includes porphyroblasts of biotite and garnet in metapelitic rocks and hornblende and garnet in mafic igneous rocks (Kokonyangi et al., 2001). Hornblende and garnet are strongly poikiloblastic and display pressure shadows developed during the main fabric-forming event D_2 . Some garnet porphyroblasts exhibit an internal fabric defined by quartz and opaque minerals, oblique to the D_2 fabric. This internal fabric is related to an early event (D_1), which was contemporaneous to the emplacement of the batholith.

The Kabonvia granodioritic augen gneiss forms an ENE-oriented orthogneiss body elongated parallel to the trend of the D_1 fabric in the metasedimentary country rocks. A solid-state foliation (S_1) is pronounced throughout this body and locally the

Table 1
Summary of field and petrographic characteristics of Mitwaba granitoids

	Kifinga-Kisele batholith (ca. 75 km × 60 km)			Nyangwa	Kungwe-Kalumengongo
	Kabonvia	Kisele	Fwifwi		
Deformation features	D ₁ and F ₁ minor folds, D ₂	D ₁ , D ₂ and F ₁ minor folds	D ₁ , D ₂ and marginal mylonitic foliation	D ₁ , D ₂ , mesoscopic folds	D ₁ , D ₂ , minor shear zones subparallel to S ₂
Contact aureole	Grt-Bt hornfels	N.O.	Grt-Hb hornfels	N.O.	N.O.
Regional metamorph. zone of occurrence	Staurolite zone	Sillimanite zone	Grt-Bt-Chl zone	Chl zone	Grt-bt-Chl zone
Mineralogy (vol.%)					
Quartz	35–44	30–35	Up to 55	38–43	39–42
Alkali feldspar	10–15 (Or _{92–96} Ab _{7–3})	29–30 (Or _{95–96} Ab _{5–4})	10–19 (Or ₉₃ Ab ₆ An ₁)	12–22 (Or ₉₈ Ab ₂)	8–22 (Or _{74–89} Ab _{26–11})
Plagioclase	35–40 (An _{20–30} Ab _{79–20})	32–40 (c: An _{12–13} Ab ₈₆ ; r: An ₂ Ab ₉₈)	23–37 (An _{12–13} Ab _{81–85} Or _{6–1})	25–35 (An _{7–10} Ab _{91–84} Or _{1–6})	
Biotite	10–19 (X _{Mg} : 0.50–0.53)	3–6 (X _{Mg} : 0.52)	4–7 (X _{Mg} : 0.41)	~10 (X _{Mg} : 0.43–0.47)	8–18 (X _{Mg} : 0.42–0.43)
Primary muscovite	~ 5 (X _{Mg} : 0.62–0.66)	1–4 (X _{Mg} : 0.54)	1–4 (X _{Mg} : 0.52)	1–2 (X _{Mg} : 0.60)	2–3 (X _{Mg} : 0.66)
Accessories					
Apatite	xx	x	xx	xx	xx
Titanite			x		xx
Monazite	xx	xx	xx	x	xx
Tourmaline	x	xx	x	x	xx
Zircon	xx	x	xx	xx	xx
Chlorite	s	s	s	s	s
Musc/sericite	s	s	s	s	s
Ilmenite	xx	x	x	x	xx
Magnetite	x				
Rutile	x	x		x	
Metased. xenoliths	xx	x	x	x	xx
Classification (QAP)	Granodiorite	Monzogranite	Monzogranite	Monzogranite	Monzogranite
Inheritance	Zircon core	Zircon core	Zircon core	Zircon core	Zircon core
Enclaves	Surmicaceous (Didier, 1973)	Surmicaceous	Surmicaceous	Surmicaceous	Surmicaceous

Notes. c: core; r: rim; x: present; xx: abundant; s: secondary; Grt: garnet; Bt: Biotite; Chl: Chlorite; Hb: Hornblende; N.O.: not observed; metased.: metasedimentary rocks; XMg = Mg/Mg + Fe; calculated in molar ratios.

orthogneiss grades to banded migmatitic gneiss. The mineral paragenesis of sedimentary rocks metamorphosed along with this granitoid equilibrated in the staurolite zone (Kokonyangi et al., 2001). According to the classification of Cahen et al. (1984), this body belongs to the type A (G_1) granitoids supposedly emplaced at 1331 ± 25 Ma.

The Kisele monzogranite gneiss is pink to pinkish-grey and contains enclaves of the Kabonvia granodiorite augen gneiss close to their mutual contact. This observation and geometrical relationships between these intrusions (Fig. 2) suggest that the Kisele monzogranite gneiss postdate the emplacement of the Kabonvia granodioritic augen gneiss. The Kisele monzogranite gneiss contains also surmicaceous enclaves and enclaves of metasedimentary rocks that could not be identified within the country rock successions and are inferred to represent material brought from the source of the granite. The mineral paragenesis (Table 1) of sedimentary rocks metamorphosed along with this granitoid equilibrated in the sillimanite zone. This granitoid displays crosscutting relations between S_1 (70°N – 80°E) and S_2 (25°N – 40°E) and therefore, according to Cahen et al. (1984), it can be classified as type A or G_1 Kibaran granite.

Compared to the Kisele monzogranite gneiss, the Fwifwi foliated leucomonzogranite is quartz-rich, weakly foliated in the centre and exhibits higher strain at the margin of the pluton where it grades to a gneissic facies. The western margin of the pluton is in contact with greenschist facies metamorphosed mafic igneous rocks intruding the Mitwaba group. In contrast to other plutons, this body contains titanite close to its contact with the mafic rocks (Fig. 2). However, away from the contact, titanite is absent. The Fwifwi foliated leucomonzogranite contains angular enclaves of the Kisele monzogranite gneiss along the contact between the two intrusive bodies. Pioneering conventional Pb–Pb dating done almost half a century ago on this pluton (Eberhardt et al., 1956) yielded dates between 1050 ± 50 Ma and 1144 ± 10 Ma. According to the above dates and based on the classification criteria of Cahen et al. (1984), the Fwifwi foliated leucomonzogranite belongs to type C (G_3).

The Nyangwa pluton occurs to the northwest of the study area (Fig. 2). It is elongated $\text{N}30^\circ\text{E}$, parallel to the trend of the regional D_2 fabric. The pluton occurs within greenschist facies (chlorite-zone) metasedimentary rocks. The margins of the intrusion are buried

beneath a Neoproterozoic and Phanerozoic sedimentary cover. The rocks are foliated, medium- to coarse-grained, and pinkish to greyish. Biotite flakes and K-feldspar phenocrysts define the foliation. The pluton contains elongated elliptical enclaves of metasedimentary rocks distributed parallel to the foliation of the granite. Some quartz veins observed in the granite are isoclinally folded with the axial plane trending NE, parallel to the solid-state foliation in the granitoid and in the country rocks. No previous age determination was done on this pluton, but based on its regional NE elongation, it was inferred to be part of the ca. 1308 ± 25 Ma type B (G_2) granitoids (Cahen et al., 1984).

The Kungwe-Kalumengongo stock is a small granitic gneiss body intruding greenschist facies metasedimentary rocks. The granite is leucocratic, medium to fine grained, with sporadic K-feldspar phenocrysts. It is chilled against the Kalumengongo black shale formation (Mitwaba group) and contains abundant xenoliths of supracrustal metasedimentary country rocks. Metasedimentary enclaves observed in the central portion of this stock are unknown in the local lithostratigraphic sequence. The stock is particularly rich in tourmaline close to the contact with the country rocks. Two foliations are developed throughout this granitic body. The oldest trends 70°N – 90°E and the youngest 20°N – 30°E respectively, parallel to

Fig. 4. Classification of the Mitwaba granitoids using the modal composition plotted in the quartz-alkali feldspar-plagioclase (Q–A–P) diagram of Streckeisen (1976).

S₁ and S₂ in supracrustal country rocks. No previous age determination has been performed on this granitic stock, but according to Cahen et al. (1984), it would be classified as type A or G₁ because it carries both foliations (S₁ and S₂). The modal and mineralogical compositions of the Mitwaba granitoids are given in Table 1 and illustrated in Fig. 4. In this diagram, the Mitwaba granitoids fall in the fields of granodiorites for the Kabonvia pluton and monzogranites for the remaining plutons.

4. Sampling and analytical procedures

Ion Microprobe (SHRIMP) U–Pb zircon dating of representative samples from the Mitwaba granitoids (Table 2) was performed at the Research School of Earth Sciences (RSES), The Australian National University. Separated zircon grains were mounted in epoxy together with the RSES SHRIMP standard AS3 (Duluth Complex gabbroic anorthosite; Peaces and Miller, 1989) and SL13 (radiogenic $^{206}\text{Pb}/^{238}\text{U} = 0.0928$ and 238 ppm U, Roddick and Van Breemen, 1994). Data have been reduced in a manner similar to that described by Compston et al. (1992) and Williams and Claesson (1987), and using the software SQUID (Ludwig, 2000). U/Pb in zircon unknowns were normalized to a $^{206}\text{Pb}^*/^{238}\text{U}$ value of 0.1859 (equivalent to an age of 1099.1 Ma) for AS3. The U and Th concentrations were determined relative to those measured in the SL13 standard and the common Pb correction was determined from the measured ^{204}Pb and using the appropriate common Pb isotopic composition assuming the Cumming and Richards (1975) model. The age calculations and statistical assessment of the data have been done utilizing the geochronological statistical software package Isoplot/Ex (version 2.00) of Ludwig (2000). Uncertainties in the isotopic ratios and ages in the data table (and in the error bars in the plotted data) are reported at 1σ , but where ages are calculated from combined data or from regression analysis the error quoted are 95% confidence limits. Where appropriate, Concordia Ages for some data sets are calculated using Ludwig (1998), again with 95% confidence limit quoted for uncertainties.

Samples were powdered using an automatic agate mill at OCU. Two sets of new whole-rock chemical data are used in this study. The first set includes

21 whole-rock analyses determined at OCU using a Rigaku RIX-2000 X-ray fluorescence spectrometer on glass beads. Precision is $\sim 1\%$ for major elements and $\sim 5\%$ for trace elements. These data corresponds to analyses without rare-earth-element composition in Table 3. The second set consists of 17 remaining whole-rock analyses (Table 3) performed at Chemex Laboratories (Canada). Major element compositions were determined using an ICP-AES and trace elements were analysed using an ICP-MS. Pb, Ni, Cr and Li were analysed by AAS. The precisions are ± 1 and $\pm 5\%$ for major and trace elements, respectively. All analyses were recalculated to 100% anhydrous for inter-sample comparisons in the paper. The data were normalized to chondrite and primordial mantle using the reference values of Sun and McDonough (1989).

5. Geochronology

Most of the granite samples processed for this study yielded good quality zircons suitable for U–Pb dating, with one sample (K2) producing the most complicated population. Inheritance, in the form of cores, is a common feature in the zircon grains from these granitoids. However, the use of cathodoluminescence imaging enables these xenocrystic components to be identified and generally avoided for this study mainly focussed on the crystallisation age of the granitoids.

5.1. Kisele monzogranite gneiss

The zircons from this sample are complicated by structure, alteration (through metamictisation), possible metamorphic rounding and the presence of metamorphic overgrowths. The general population comprises subhedral grains of various sizes that are yellow to brown, and translucent. Cathodoluminescence imaging shows strong concentric compositional zoning interpreted to represent the magmatic zircon, but with some grains showing large altered areas. A few grains have small, distinct overgrowths or zones that appear uniform (unzoned) in cathodoluminescence.

Twelve U–Th–Pb analyses were done on different zircons with the data reported in Table 2 and plotted on a conventional Wetherill U–Pb concordia shown in Fig. 5a. As anticipated, most analyses are highly

Table 2

U–Pb SHRIMP isotopic data on zircons from Kibaran granitoids Mitwaba Katanga (Congo)

Grain spot	$^{206}\text{Pb}/^{238}\text{U}$ (%)	U (ppm)	Th (ppm)	$^{232}\text{Th}/^{238}\text{U}$	$^{206}\text{Pb}^*$ (ppm)	(1) $^{207}\text{Pb}^*/^{206}\text{Pb}^*$	±%	(1) $^{207}\text{Pb}^*/^{235}\text{U}$	±%	(1) $^{206}\text{Pb}^*/^{238}\text{U}$	±%	Err corr	(1) $^{206}\text{Pb}/^{238}\text{U}$ age (Ma)	(1) $^{207}\text{Pb}/^{206}\text{Pb}$ age (Ma)	Dis- cordant (%)
Kabonvia granodioritic augen gneiss (K12)															
1.1	0.55	228	33	0.15	46.0	0.08654	0.93	2.785	1.2	0.2334	0.82	0.660	1352.20 ±10.0	1350 ±18	0
2.1	0.94	256	56	0.23	52.1	0.0890	3.5	2.88	3.6	0.2347	0.75	0.208	1359.10 ±9.2	1404 ±68	3
3.1	0.07	635	46	0.07	135	0.08829	0.33	2.998	0.73	0.2463	0.65	0.895	1419.10 ±8.3	1388.8 ±6.3	-2
4.1	3.91	376	42	0.12	73.6	0.0892	1.7	2.695	1.9	0.2191	0.70	0.377	1277.00 ±8.1	1409 ±33	9
5.1	0.28	214	42	0.20	43.9	0.08665	0.72	2.851	1.0	0.2386	0.75	0.725	1379.30 ±9.3	1353 ±14	-2
6.1	0.46	251	63	0.26	53.3	0.08909	1.0	3.021	1.3	0.2459	0.83	0.632	1418.00 ±11	1406 ±19	-1
7.1	0.44	179	20	0.12	37.6	0.08729	0.97	2.922	1.2	0.2428	0.78	0.625	1401.00 ±9.8	1367 ±19	-2
8.1	1.00	88	148	1.74	26.3	0.1182	1.2	5.634	1.6	0.3457	1.0	0.659	1914.00 ±17	1929 ±21	1
8.2	0.13	286	28	0.10	60.7	0.08873	0.69	3.019	0.98	0.2467	0.70	0.714	1421.60 ±8.9	1398 ±13	-2
9.1	0.20	420	41	0.10	85.6	0.08873	0.58	2.898	0.89	0.2369	0.68	0.756	1370.40 ±8.3	1398 ±11	2
10.1	0.33	236	36	0.16	47.7	0.08795	0.79	2.839	1.1	0.2341	0.72	0.677	1356.10 ±8.8	1381 ±15	2
11.1	0.22	369	48	0.13	76.0	0.08774	0.57	2.892	0.89	0.2391	0.68	0.770	1381.80 ±8.5	1377 ±11	0
12.1	0.43	283	46	0.17	57.9	0.08720	0.95	2.850	1.2	0.2370	0.71	0.600	1371.30 ±8.8	1365 ±18	0
13.1	2.54	161	45	0.29	30.7	0.0927	2.2	2.765	2.3	0.2162	0.77	0.330	1261.80 ±8.9	1483 ±42	15
14.1	0.42	146	64	0.45	30.2	0.08885	1.0	2.943	1.4	0.2402	1.0	0.718	1388.00 ±13	1401 ±19	1
15.1	0.59	213	30	0.14	43.9	0.08717	0.94	2.872	1.3	0.2390	0.87	0.679	1381.00 ±11	1364 ±18	-1
Kisele monzogranitic gneiss (K2)															
1.1	0.9	536	107	0.21	75.8	0.08529	0.76	1.92	1	0.1632	0.7	0.679	974.70 ±6.4	1322.00 ±15	26
2.1	0.32	2436	318	0.13	227	0.07736	0.44	1.151	0.94	0.10794	0.83	0.882	660.80 ±5.2	1130.50 ±8.9	42
3.1	0.36	1562	23	0.02	224	0.07536	0.37	1.727	0.73	0.1662	0.63	0.864	991.00 ±5.8	1078.00 ±7.4	8
4.1	0.14	430	72	0.17	89.2	0.08774	0.61	2.921	1	0.2415	0.81	0.798	1394.00 ±10	1377.00 ±12	-1
5.1	0.14	350	146	0.43	72.6	0.08792	0.61	2.92	0.91	0.2409	0.68	0.749	1391.30 ±8.6	1381.00 ±12	-1
6.1	5.29	2504	144	0.06	184	0.075	4	0.837	4.1	0.08087	0.72	0.177	501.30 ±3.5	1070.00 ±81	53
7.1	0.13	388	43	0.11	79.7	0.08834	0.46	2.911	0.82	0.239	0.68	0.826	1381.50 ±8.4	1390.00 ±8.9	1
8.1	1.13	6647	1269	0.2	162	0.05916	0.99	0.2294	1.2	0.02813	0.64	0.542	178.80 ±1.1	573.00 ±22	69
9.1	0.23	861	23	0.03	134	0.07542	0.43	1.875	0.79	0.1804	0.66	0.841	1068.90 ±6.5	1079.70 ±8.6	1
10.1	2.51	11334	3042	0.28	233	0.05958	1.3	0.1917	1.4	0.02334	0.68	0.473	148.70 ±1.0	588.00 ±27	75
11.1	0.7	5555	1447	0.27	123	0.06658	0.86	0.2349	1.1	0.02559	0.63	0.593	162.90 ±1.0	825.00 ±18	80
12.1	1.39	9865	3610	0.38	210	0.05912	1.1	0.1991	1.5	0.02443	1	0.683	155.60 ±1.6	572.00 ±24	73
Fwifwi foliated leucomonzogranite (K23)															
1.1	0.64	158	52	0.34	32.6	0.08687	1.1	2.852	1.3	0.2381	0.77	0.576	1377.10 ±9.6	1358.00 ±21	-1
2.1	0.66	127	33	0.27	26.7	0.0874	1.2	2.931	1.4	0.2433	0.8	0.559	1404.00 ±10	1369.00 ±23	-3
3.1	0.41	159	51	0.33	32.9	0.08887	0.97	2.937	1.2	0.2397	0.76	0.619	1385.10 ±9.5	1401.00 ±19	1
4.1	0.14	144	50	0.36	30.1	0.09089	0.88	3.041	1.2	0.2426	0.78	0.663	1400.30 ±9.8	1444.00 ±17	3
5.1	0.49	149	47	0.33	31	0.08683	1.1	2.876	1.5	0.2402	1.1	0.699	1388.00 ±13	1357.00 ±21	-2
6.1	0.29	205	60	0.3	42.3	0.08862	0.77	2.92	1.1	0.239	0.75	0.7	1381.60 ±9.4	1396.00 ±15	1
7.1	0.36	226	60	0.28	47.1	0.0877	0.8	2.922	1.1	0.2416	0.72	0.669	1395.20 ±9.1	1376.00 ±15	-1
8.1	0.43	155	49	0.32	31.8	0.08767	1.1	2.868	1.5	0.2373	1	0.68	1372.00 ±12	1375.00 ±21	0
9.1	0.47	155	53	0.35	31.9	0.08676	1.1	2.846	1.3	0.2379	0.77	0.588	1375.60 ±9.5	1355.00 ±20	-2
10.1	0.45	171	56	0.34	35.4	0.0875	0.97	2.898	1.3	0.2402	0.79	0.63	1387.60 ±9.8	1372.00 ±19	-1
11.1	0.55	165	57	0.36	34	0.08648	1.1	2.841	1.5	0.2383	0.97	0.657	1378.00 ±12	1349.00 ±21	-2
12.1	0.28	234	56	0.25	48.6	0.08747	0.7	2.909	1	0.2412	0.72	0.717	1393.10 ±9.0	1371.00 ±13	-2
13.1	0.47	181	58	0.33	38.1	0.08735	0.97	2.94	1.2	0.2442	0.75	0.613	1408.20 ±9.5	1368.00 ±19	-3
14.1	0.65	140	49	0.36	29.2	0.087	1.2	2.903	1.6	0.242	1.1	0.661	1397.00 ±13	1360.00 ±23	-3
15.1	0.78	114	30	0.27	23.4	0.0876	1.4	2.857	1.6	0.2365	0.85	0.52	1369.00 ±11	1374.00 ±27	0
16.1	0.67	143	44	0.32	30.4	0.08715	1.1	2.953	1.5	0.2458	0.93	0.631	1417.00 ±12	1364.00 ±22	-4

Table 2 (Continued)

Grain spot	$^{206}\text{Pb}/^{238}\text{U}$ (%)	U (ppm)	Th (ppm)	$^{232}\text{Th}/^{238}\text{U}$	$^{206}\text{Pb}^*$ (ppm)	(1) $^{207}\text{Pb}^*/^{206}\text{Pb}^*$	$\pm\%$	(1) $^{207}\text{Pb}^*/^{235}\text{U}$	$\pm\%$	(1) $^{206}\text{Pb}^*/^{238}\text{U}$	$\pm\%$	Err corr	(1) $^{206}\text{Pb}/^{238}\text{U}$ age (Ma)	(1) $^{207}\text{Pb}/^{206}\text{Pb}$ age (Ma)	Dis- cordant (%)
Kungwe-Kalumengongo monzogranite (K15)															
1.1	0.35	133	41	0.32	27.2	0.08778	1	2.877	1.3	0.2377	0.82	0.616	1375.00 ± 10	1378.00 ± 20	0
2.1	0.78	219	76	0.36	39.9	0.08807	1.1	2.557	1.3	0.2106	0.73	0.552	1232.00 ± 8.2	1384.00 ± 21	11
3.1	0.45	180	64	0.37	37.7	0.08716	0.99	2.921	1.6	0.2431	1.2	0.769	1403.00 ± 15	1364.00 ± 19	−3
4.1	0.21	315	81	0.26	64.9	0.08766	0.64	2.891	0.95	0.2392	0.7	0.736	1382.30 ± 8.7	1375.00 ± 12	−1
5.1	0.42	199	62	0.32	41.6	0.08681	0.92	2.902	1.2	0.2424	0.74	0.627	1399.30 ± 9.3	1356.00 ± 18	−3
6.1	0.35	129	65	0.52	26.9	0.0892	1.2	2.966	1.4	0.2412	0.81	0.562	1393.00 ± 10	1408.00 ± 23	1
7.1	0.3	172	62	0.37	35.6	0.08871	0.89	2.935	1.2	0.24	0.76	0.649	1386.70 ± 9.4	1398.00 ± 17	1
8.1	0.5	174	57	0.34	36	0.08604	1.1	2.846	1.3	0.2399	0.75	0.578	1386.20 ± 9.4	1339.00 ± 21	−4
9.1	0.14	127	41	0.33	26.3	0.0883	1.3	2.941	1.5	0.2416	0.8	0.538	1395.00 ± 10	1389.00 ± 24	0
10.1	0.56	139	48	0.35	29.6	0.0879	1.1	2.977	1.5	0.2457	0.89	0.613	1416.00 ± 11	1379.00 ± 22	−3
11.1	0.37	178	58	0.34	37.6	0.08634	1.1	2.922	1.3	0.2455	0.75	0.578	1415.10 ± 9.6	1346.00 ± 21	−5
12.1	0.54	118	44	0.39	24.9	0.0874	1.6	2.94	1.8	0.244	0.82	0.465	1407.00 ± 10	1369.00 ± 30	−3
13.1	0.14	173	57	0.34	36.2	0.08845	0.96	2.97	1.2	0.2435	0.76	0.618	1404.90 ± 9.6	1392.00 ± 18	−1
14.1	0.31	159	51	0.33	33.3	0.089	1.1	2.984	1.4	0.2431	0.77	0.558	1403.00 ± 9.7	1405.00 ± 22	0
15.1	0.33	198	54	0.28	41.6	0.08805	0.85	2.963	1.1	0.244	0.74	0.658	1407.60 ± 9.4	1384.00 ± 16	−2
Nyangwa monzogranite (K30)															
1.1	0.13	429	108	0.26	80.6	0.08885	0.62	2.673	0.92	0.2182	0.69	0.745	1272.20 ± 7.9	1401.00 ± 12	9
2.1	0.32	343	41	0.12	70.4	0.08778	0.72	2.884	1.6	0.2383	1.4	0.891	1378.00 ± 18	1378.00 ± 14	0
3.1	0.32	215	67	0.32	45	0.08788	0.79	2.946	1.2	0.2432	0.95	0.77	1403.00 ± 12	1380.00 ± 15	−2
4.1	0.13	263	104	0.41	53.8	0.08861	0.6	2.904	1.1	0.2377	0.91	0.836	1375.00 ± 11	1396.00 ± 11	2
5.1	0.15	384	54	0.14	79.1	0.08859	0.52	2.924	0.85	0.2394	0.68	0.797	1383.50 ± 8.5	1395.20 ± 9.9	1
6.1	0.22	450	52	0.12	91.6	0.08786	0.51	2.864	1.1	0.2364	0.92	0.875	1368.00 ± 11	1379.50 ± 9.8	1
7.1	0.33	272	95	0.36	56.7	0.08859	0.73	2.957	1.1	0.2421	0.88	0.768	1398.00 ± 11	1395.00 ± 14	0
8.1	0.18	389	43	0.11	80.3	0.0871	0.49	2.882	0.91	0.24	0.77	0.842	1386.70 ± 9.6	1362.60 ± 9.5	−2
9.1	0.31	476	66	0.14	98	0.08814	0.5	2.905	0.83	0.239	0.67	0.798	1381.50 ± 8.3	1385.60 ± 9.6	0
10.1	2.43	305	36	0.12	60.7	0.0884	3.3	2.753	3.4	0.2259	0.88	0.26	1313.00 ± 11	1391.00 ± 63	6
11.1	0.23	128	44	0.36	26.2	0.08953	0.97	2.933	1.3	0.2376	0.83	0.648	1374.00 ± 10	1416.00 ± 19	3
12.1	0.36	329	122	0.38	64	0.08696	0.64	2.705	0.94	0.2256	0.69	0.731	1311.40 ± 8.2	1360.00 ± 12	4
13.1	0.14	329	40	0.13	68.9	0.08747	0.56	2.938	1.1	0.2436	0.99	0.871	1406.00 ± 13	1371.00 ± 11	−3
14.1	0.18	295	56	0.2	60.8	0.08756	0.57	2.886	1	0.239	0.84	0.829	1382.00 ± 10	1373.00 ± 11	−1
15.1	0.5	316	99	0.32	61.1	0.08659	0.74	2.676	1.1	0.2242	0.8	0.736	1303.80 ± 9.4	1351.00 ± 14	4
16.1	0.05	325	30	0.1	66.8	0.08828	0.5	2.912	0.87	0.2392	0.71	0.817	1382.80 ± 8.8	1388.60 ± 9.6	0

Errors are 1 σ ; Pbc and Pb* indicate the common and radiogenic portions, respectively. Error in standard calibration was 0.19% (not included in above errors but required when comparing data from different mounts). (1) Common Pb corrected using measured ^{204}Pb .

Table 3
Representative major and trace element compositions of Mitwaba granitoids (Katanga, Congo)

	Plutons													
	Kabonvia						Kisele						Fwifwi	
	K12.1	K12.2	K12.3	K12.4	K12.5	K12.6	K2.1	K2.2	K2.3	K2.4	K2.5	K2.6	K23.1	K23.2
SiO ₂	69.39	68.32	67.69	69.78	67.35	68.68	74.16	75.46	74.68	75.45	73.35	74.88	70.84	74.21
TiO ₂	0.51	0.53	0.46	0.51	0.52	0.47	0.17	0.16	0.17	0.18	0.19	0.18	0.48	0.17
Al ₂ O ₃	13.79	14.25	13.06	14.57	14.49	13.68	13.7	12.66	12.84	14.36	14.97	13.27	13.71	12.35
Fe ₂ O ₃	3.47	3.82	3.71	3.46	3.74	3.77	1.44	1.22	1.23	1.48	1.19	1.22	2.95	1.46
MnO	0.07	0.07	0.04	0.07	0.07	0.04	0.03	0.02	0.02	0.03	0.02	0.02	0.03	0.01
MgO	1.66	1.72	1.67	1.74	1.82	1.85	0.57	0.49	0.51	0.63	0.51	0.51	1.01	0.41
CaO	1.36	1.32	1.95	1.34	1.31	1.95	0.52	0.58	0.68	0.57	0.58	0.58	2.12	0.72
Na ₂ O	2.68	2.66	1.72	2.72	2.66	1.76	4.55	3.49	3.87	4.80	3.48	3.58	2.25	1.78
K ₂ O	4.71	5.03	4.4	4.64	4.85	4.34	2.93	4.43	3.82	2.95	4.22	4.35	5.05	6.46
P ₂ O ₅	0.12	0.13	0.1	0.12	0.14	0.11	0.08	0.09	0.08	0.07	0.08	0.08	0.18	0.06
LOI	1.87	1.74	3.34	1.49	1.094	1.55	1.4	1.04	1.5	1.02	0.98	1.01	1.37	2.27
Total	99.76	99.71	98.26	100.44	98.04	98.20	99.59	99.68	99.43	101.54	99.56	99.69	100.1	99.95
Cs	18.2	19.6	6.8				1.3	1.3	1.2				4.2	11.8
Rb	309	329	225	255	268	186	132	185.5	168	119	151	158	254	249
Ba	999	1085	836	909	958	771	242	384	332	229	338	342	947	375
Th	14	16	17				16	20	25				17	27
U	3	3.5	4				13	16	19				3	7.5
Ta	3	4	3				6	6	6				6	9
Nb	13	14	13	12	12	11	9	9	10	7	6	6	15	16
Sr	160.5	159	226	141	140	196	65.8	72.6	74.5	61	61	63	284	183
Hf	5	6	6				5	4	6				6	3
Zr	279	281	296	177	185	207	267	241	278	172	126	131	300	192
Y	26	34	28.5	27	33	30	20	43.5	52	22	43	42	35.5	46.5
V	65	70	60	60.2	63.2	59.8	10	20	25	19.9	17.8	20.1	50	15
Cr	64	64	56	70	70	56	4	6	4	8	7	8	46	2
Ni	30	30	25	21	22	18	15	15	10	5	7	7	10	10
Cu	25	25	25	20	16	17	15	10	5	1	2	3	5	10
Pb	20	20	7	31	32	12	7	9	11	18	24	24	7	9
Zn	75	85	50	62	66	28	30	20	30	12	10	10	25	40
Co	104	71	109.5				96	103	67				66.5	150.5
La	50	52	56.5				41.5	44	57.5				50.5	59.5
Ce	96	105	112				83	87.5	112.5				104.5	129
Pr	11	12.5	13				9.4	9.8	12.5				12.6	15.7
Nd	41	46	44.5				31.5	34.5	44.5				46.5	58.5
Sm	7.5	8.2	8				6.2	7.3	8.8				8.9	12
Eu	1.4	1.4	1.3				0.9	1	1.1				1.6	1
Gd	6.7	7.1	6				4.9	6.9	8.4				7.2	9.9
Tb	1	1.1	0.9				0.8	1.2	1.5				1.1	1.4

Table 3 (Continued)

	Plutons													
	Kabonvia						Kisele						Fwifwi	
	K12.1	K12.2	K12.3	K12.4	K12.5	K12.6	K2.1	K2.2	K2.3	K2.4	K2.5	K2.6	K23.1	K23.2
Dy	5.2	6.5	4.8				4.7	7.9	9.5				6	8.9
Ho	0.9	1.2	1.1				0.8	1.6	1.9				1.2	1.6
Er	3	3.7	3				2.2	4.9	5.7				3.6	4.7
Tm	0.4	0.6	0.4				0.3	0.7	0.9				0.5	0.6
Yb	2.6	3.8	3.3				2	5.2	6.2				3.2	4.1
Lu	0.4	0.5	0.4				0.3	0.8	1				0.5	0.6
K ₂ O/Na ₂ O	1.8	1.9	2.6	1.7	1.8	2.5	0.6	1.3	1.0	0.6	1.2	1.2	2.2	3.6
ASI	1.2	1.2	1.2	1.2	1.2	1.2	1.2	1.1	1.1	1.2	1.3	1.1	1.1	1.1
Cor.norm	1.8	2.1	2.0	3.0	2.6	2.8	2.3	1.2	1.2	2.4	1.8	1.8	1.0	1.2
Mg#	49	47	47	50	49	49	44	44	45	46	46	45	40	36
Rb/Sr	1.9	2.1	1.0	1.8	1.9	0.9	2.0	2.6	2.3	2.0	2.5	2.5	0.9	1.4
Rb/Ba	0.3	0.3	0.3	0.3	0.3	0.2	0.5	0.5	0.5	0.5	0.4	0.5	0.3	0.7
Rb/Th	22	21	13				8	9	7				15	9
Rb/Zr	1.1	1.2	0.8	1.4	1.4	0.9	0.5	0.8	0.6	0.7	1.2	1.2	0.8	1.3
Rb/Hf	62	55	38				26	46	28				42	83
Rb/Y	12	10	8	9	8	6	7	4	3	5	4	4	7	5
K/Rb	127	127	152	151	150	194	184	198	189	206	232	229	165	215
Sr/Ba	0.2	0.1	0.3	0.2	0.1	0.3	0.3	0.2	0.2	0.3	0.2	0.2	0.3	0.5
Th/U	5	5	4				1	1	1				6	4
Ta/Nb	0.2	0.3	0.2				0.7	0.7	0.6				0.4	0.6
REEt _{total}	253	284	284				209	257	324				283	354
(La/Sm) _N	4	4	4				4	4	4				4	3
(La/Yb) _N	13	9	11				14	6	6				11	10
(Gd/Yb) _N	2	2	1				2	1	1				2	2
Eu/Eu*	0.6	0.6	0.6				0.5	0.4	0.4				0.6	0.3
Fwifwi						Kungwe-Kalumengongo			Nyangwa					
	K23.3	K23.4	K23.5	K23.6	K23.7	K15.1	K15.2	K30	K35.1	K35.2	K35.3	K36.1	K36.2	K36.4
SiO ₂	71.38	70.98	71.81	69.34	75.74	69.2	70.26	70.88	74.05	76.73	76.53	68.58	68.01	67.31
TiO ₂	0.43	0.48	0.42	0.45	0.19	0.59	0.60	0.53	0.43	0.44	0.45	0.56	0.56	0.55
Al ₂ O ₃	13.35	13.75	13.22	13.73	13.22	13.19	13.78	12.87	13.1	13.53	13.52	13.36	13.16	13.69
Fe ₂ O ₃	2.38	2.88	2.07	2.9	1.54	3.66	3.67	3.82	2.13	2.15	2.15	4.83	4.79	4.83
MnO	0.04	0.03	0.03	0.04	0.02	0.06	0.07	0.06	0.01	0.02	0.02	0.05	0.05	0.06
MgO	0.99	1.01	0.86	1	0.44	1.35	1.47	1.37	0.72	0.74	0.74	2.12	2.18	2.21
CaO	1.77	2.13	1.83	2.01	0.71	2.55	2.57	0.24	0.63	0.65	0.65	2.29	2.29	2.24
Na ₂ O	2.44	2.27	3.12	1.9	1.81	2.12	2.19	2.54	1.85	1.89	1.93	1.82	1.81	1.84
K ₂ O	5.2	5.13	3.75	5.9	6.50	3.82	3.75	4.89	5.02	5.03	5.02	4.09	4.02	3.94

P ₂ O ₅	0.21	0.16	0.16	0.19	0.06	0.2	0.21	0.16	0.09	0.09	0.09	0.16	0.13	0.15
LOI	1.6	1.27	1.54	1.68	1.23	3.29	2.62	1.89	1.52	0.48	0.38	1.55	2.18	1.45
Total	99.96	100.2	98.97	99.33	101.45	100.1	101.18	99.42	99.66	101.74	101.48	99.53	99.36	98.25
Cs	6.9	3.8	6.9	5.2		13.3		2.4	23.1			6.6	6.4	
Rb	220	255	184.5	257	211	243	194	169	301	253	252	201	195	163
Ba	1300	936	1065	1235	371	805	755	1345	587	594	588	896	887	870
Th	12	16	13	14		17		14	13			14	14	
U	2.5	2.5	3.5	3		5.5		6	3			4	3.5	
Ta	4	6	5	12		4		5	8			7	12	
Nb	12	15	13	16	12	16	14	15	15	12	12	15	16	13
Sr	272	290	238	313	166	281	239	105	91	80	81	288	276	249
Hf	5	6	5	5		7		6	5			6	5	
Zr	272	301	286	288	72	323	220	313	290	213	209	312	300	219
Y	29.5	35.5	32	32	48	40	39	37.5	35	38	35	33.5	29	30
V	50	55	55	55	10	55	50.5	70	70	41.1	41.9	105	100	71.5
Cr	46	44	44	42	4	64	61	42	30	37	30	82	84	84
Ni	15	10	15	15	7	25	10	15	5	7	9	30	30	22
Cu	5	5	5	20	2	5	0	10	10	1	3	25	30	19
Pb	10	8	9	14	20	7	14	9	9	19	18	12	14	18
Zn	30	20	35	90	26	50	31	45	25	17	16	70	65	48
Co	86	69.5	84.5	137		86		49	80.5			114	126.5	
La	41	50	49.5	43		49.5		52.5	44.5			55.5	54.5	
Ce	78.5	101.5	89	82.5		104		118.5	86.5			108	106.5	
Pr	10	12.5	11.3	10.2		12.4		12.9	10.1			12.5	12.4	
Nd	37	46.5	41	38.5		47		47	35.5			45	45	
Sm	7.6	8.7	7.7	7.2		9.3		9.2	6.7			8.1	8.2	
Eu	1.3	1.3	1.4	1.2		1.3		1.4	0.9			1.2	1.3	
Gd	5.9	7.8	6.6	6.4		7.9		7.4	7.3			6.9	6.7	
Tb	0.9	1.1	1	0.9		1.1		1.1	0.9			1	0.9	
Dy	4.9	6.2	5.4	5.3		6.7		6.8	5			5.3	5.3	
Ho	1	1.2	1.1	1		1.3		1.3	1.1			1.2	1	
Er	2.9	3.5	3	2.7		3.8		3.6	3.5			3.4	3	
Tm	0.4	0.5	0.5	0.4		0.6		0.5	0.4			0.5	0.4	
Yb	2.7	3.1	2.7	3.2		3.7		3.5	3.2			3.3	2.7	
Lu	0.4	0.5	0.4	0.4		0.6		0.5	0.5			0.4	0.3	
K ₂ O/Na ₂ O	2.1	2.3	1.2	3.1	3.6	1.8	1.7	1.9	2.7	2.7	2.6	2.3	2.2	2.1
ASI	1.0	1.1	1.1	1.0	1.2	1.1	1.1	1.3	1.4	1.4	1.4	1.5	1.2	1.2
Cor.norm	0.8	0.8	1.0	0.9	2.1	1.3	2.0	3.2	3.5	4.0	3.9	2.0	1.8	2.7
Mg#	45	41	45	41	36	42	44	42	40	40	41	47	47	48
Rb/Sr	0.8	0.9	0.8	0.8	1.3	0.9	0.8	1.6	3.3	3.2	3.1	0.7	0.7	0.7
Rb/Ba	0.2	0.3	0.2	0.2	0.6	0.3	0.3	0.1	0.5	0.4	0.4	0.2	0.2	0.2
Rb/Th	18	16	14	18		14		12	23			14	14	
Rb/Zr	0.8	0.8	0.6	0.9	2.9	0.8	0.9	0.5	1.0	1.2	1.2	0.6	0.7	0.7

Table 3 (Continued)

	Plutons													
	Fwifwi					Kungwe-Kalumengongo			Nyangwa					
	K23.3	K23.4	K23.5	K23.6	K23.7	K15.1	K15.2	K30	K35.1	K35.2	K35.3	K36.1	K36.2	K36.4
Rb/Hf	44	43	37	51		35		28	60			34	39	
Rb/Y	7	7	6	8	4	6	5	5	9	7	7	6	7	5
K/Rb	196	167	169	191	255	130	161	240	138	165	165	169	171	200
Sr/Ba	0.2	0.3	0.2	0.3	0.4	0.3	0.3	0.1	0.2	0.1	0.1	0.3	0.3	0.3
Th/U	5	6	4	5		3		2	4			4	4	
Ta/Nb	0.3	0.4	0.4	0.8		0.3		0.3	0.5			0.5	0.7	
REEt _{total}	224	280	253	235		289		304	241			286	277	
(La/Sm) _N	3	4	4	2		3		4	4			4	4	
(La/Yb) _N	10	11	12	9		9		10	9			11	14	
(Gd/Yb) _N	2	2	2	2		2		2	2			2	2	
Eu/Eu*	0.6	0.5	0.6	0.4		0.5		0.5	0.4			0.5	0.5	

Samples with REE contents provided were analysed using ICP at Chemex Laboratory (Canada). The other samples were analysed using X-ray fluorescence (XRF) at Osaka City University (Japan).

Fig. 5. Wetherill U–Pb concordia diagrams for: (a) Kisele monzogranitic gneiss (K2). The dashed line represents the regression of the data for the three low Th/U (metamorphic) analyses; (b) Kabonvia granodioritic augen gneiss (K12); (c) Fwifwi foliated leucomonzogranite (K23). The age shown is calculated from the dotted error ellipses; (d) Kungwe-Kalumengongo stock (K15). The black error ellipse represents the data for analysis 2.1, which was not included in the age calculation for this sample; (e) Nyangwa monzogranite (K30). The discordant data represented by the dotted error ellipses were not included in the age calculations. For all these diagrams, the data-point error ellipses are plotted with 1σ errors.

discordant, having clearly suffered severe Pb-loss as a consequence of high and sometimes extreme U content (e.g. analysis 10.1 has an U concentration of over 1.1 wt.%). For these analyses, cores were avoided, and most analyses were sited in the zoned magmatic area of the grains. Three such analyses (4.1, 5.1 and 7.1) yielded data within error of concordia and these can be combined to give a Concordia age of 1386.3 ± 8.1 Ma (probability of concordance = 0.582). This date is interpreted to represent the best estimate of the age of emplacement of the protolith of this granite.

Three analyses (3.1, 6.1 and 9.1) have unusually low Th/U composition compared to the remainder of the population. Two of these analyses were sited on the rims of uniform zircon and the third (9.1) in a uniform zone, i.e. without the concentric compositional zoning seen in the magmatic population. These three data points range from 1 to 53% discordant and lie on a discordia for which an upper intercept age of 1079 ± 14 Ma can be calculated. The low Th/U values in these zircon overgrowths are commonly observed in metamorphic zircons and thus these three analyses are interpreted to date a metamorphic event that affected this granite, and the zircons within it.

5.2. Kabonvia granodioritic augen gneiss

The zircons from this sample are euhedral to subhedral, with clear elongate grains mixed with larger, oblate and somewhat chunky, brown crystals. Cathodoluminescence revealed numerous cores of variable dimensions, with the general population comprising strongly zoned (concentric compositional zoning) zircons.

Sixteen U–Th–Pb analyses were done on 15 different zircons from this ortho gneiss. The data are listed in Table 2 and are plotted on a conventional Wetherill U–Pb concordia plot (Fig. 5b). Most of the analyses plot as a group near the concordia, for which a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ date of 1385.5 ± 7.4 Ma (MSWD = 1.09; $P = 0.36$) can be calculated. Two analyses (4.1 and 13.1), which were excluded from the above age calculation, are significantly discordant and have relatively high common Pb content. A single large, sector-zoned core (8.2) was analysed and yielded a concordant $^{207}\text{Pb}/^{206}\text{Pb}$ crystallisation age of 1929 ± 21 Ma. The zoned, magmatic overgrowth on this xenocrystic zircon formed during the crystallisa-

tion of the gneiss protolith, giving a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1398 ± 13 Ma (1σ), an age indistinguishable from the magmatic age calculated above.

5.3. Fwifwi foliated leucomonzogranite

Zircons from this sample are euhedral to subhedral, and light pink to brown. The general population comprises strongly zoned zircon, consistent with crystallisation from felsic magma. Many grains have cores of inherited zircon. Sixteen U–Th–Pb analyses were done on different grains from this granite, targeting the zoned magmatic area of the grains. The results are shown in Table 2 and plotted in Fig. 5c. All analyses are very uniform with respect to Th/U with an average value of 0.32, and apart from one analysis (3.2), the data form a group around concordia. Fifteen analyses can be combined to give a weight mean $^{207}\text{Pb}/^{206}\text{Pb}$ date of 1372 ± 9.6 Ma (MSWD = 0.61; $P = 0.86$).

5.4. Nyangwa monzogranite

The zircons from this sample are very variable in size and shape, and are generally brown in colour. Some of the grains are fragments of presumably once larger grains. The zoning patterns vary between areas of weak, broad zonation and areas of rapidly changing concentric compositional zoning. Core of inherited fragments of various sizes are preserved.

Most of the 16 analyses performed on the zircon from this sample (Table 2) plot as a cluster on concordia (Fig. 5d), and gives a weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1382 ± 8.1 Ma (MSWD = 1.3; $P = 0.23$). The age from the same data set can be further constrained by the calculation of the Concordia age of 1383.3 ± 4.8 Ma (probability of concordance = 0.713). Four discordant analyses (1.1, 10.1, 12.1 and 15.1) were excluded from all age calculations. The concordia date of 1383.3 ± 4.8 Ma is the best estimate of the age of crystallisation of this pluton.

5.5. Kungwe-Kalumengongo monzogranite

This sample produced a good crop of excellent quality euhedral and subhedral zircons. The zircons are clear, transparent and light pink. Only a few, small apparent core (not analysed) are observed through the use of cathodoluminescence imaging, with most

zircons comprising simple concentric compositional zoning.

Fifteen different zircons were analysed from this sample. The results are plotted in Fig. 5e and are listed in Table 2. The zircons appear to be simple magmatic grains with very uniform U, Th and Th/U contents. Apart from a single discordant analysis (2.1), the data plot as a group on or near concordia. A weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ date of 1377 ± 10 Ma (MSWD = 1.05; $P = 0.40$) calculated from this group of analyses is interpreted to be the magmatic age of this granite.

6. Whole-rock geochemistry

Representative whole-rock analyses are listed in Table 3. The composition of the Mitwaba granitoids falls between 67.24 and 76.73 wt.% SiO_2 . All the rocks are potassic, with $\text{K}_2\text{O}/\text{Na}_2\text{O} > 1$, except for sample K2.1 in which K_2O loss is suspected. In the Qz–Ab–Or normative diagram (Fig. 6a), the Mitwaba granitoids show a near eutectic minimum-temperature melt composition and they overlap with the field of experimental felsic melts originating from dehydration melting of metasedimentary rocks (e.g. Stevens et al., 1997). All Mitwaba granitoids have normative corundum in the range 0.8–4.8 and its content is not correlated to loss-on-ignition values. The alumina saturation index (ASI: molar $\text{Al}_2\text{O}_3/\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}$) is in the range 1.1–1.5, indicating strongly peraluminous compositions (Fig. 6b). These granitoids have low total Fe_2O_3 , Na_2O , MgO and CaO (Table 3) and are characterized by: (1) $\text{CaO}/\text{Na}_2\text{O}$ in the range 0.3–1.2, except in the Kisele pluton where lower values occur; and (2) $\text{Al}_2\text{O}_3/\text{Na}_2\text{O}$ between 22 and 31, except in the Kisele pluton marked by higher ratios (71:81). The composition and inter-oxide ratios of major elements show that Mitwaba strongly peraluminous (SP) granitoids are similar to Hercynian and Lachlan Fold belts SP granitoids (e.g. Fig. 6c and d) and have characteristics of synorogenic anatectic SP granitoids originating from partial melting of metasedimentary rocks (Fig. 6d and e). Mg# versus CaO diagram (Fig. 6f) indicates distinct evolution trends between and within plutons from the Mitwaba SP granitoids.

Rb and Sr contents range between 119–329 and 61–313 ppm, respectively. K/Rb ratios are in the

range 150–260, except in two samples marked by lower ratios close to 130. $\text{K}/\text{Rb} > 150$ usually mark granitoids not affected by post-magmatic activity of aqueous phase. Ba concentrations are between 229 and 1000 ppm, except in three samples containing 1100–1400 ppm Ba. Rb/Ba ratios are in the range 0.2–0.7, excluding the three samples with high Ba content leading to lower $\text{Rb}/\text{Ba} < 0.2$. Rb/Sr is in the range 0.7–3.3 whereas Sr/Ba ratios are between 0.1 and 0.3 in Mitwaba granitoids. With a few exceptions, the above trace element contents compare with values reported for S-type granitoids. Indeed, granitoids resulting from dehydration melting of a metapelitic precursor are characterized by $\text{Rb} > 100$ ppm, $\text{Sr} < 300$ –400 ppm, $\text{Ba} < 600$ –1000 ppm and Rb/Ba ratios > 0.25 (Miller, 1985).

Zr shows wide scatter and is not correlated with SiO_2 . Its concentration is medium to high, frequently between 131 and 323 ppm (except in sample K23.4 containing 72 ppm). S-type granites are commonly characterized by low Zr content (< 100 ppm; e.g. Scaillet et al., 1990) because of the low solubility of zircon in peraluminous melts formed at relatively low ($< 800^\circ\text{C}$) temperature (Watson and Harrison, 1983). However, large volume SP granitoids generated at higher temperatures ($> 875^\circ\text{C}$) as documented in the Lachlan and Hercynian belts display higher Zr contents, similar to the high Zr values recorded in the Mitwaba SP granitoids. Rb/Zr ratios in Mitwaba granitoids are between 0.5 and 1.4 and this ratio is usually > 1 in Himalaya-type S-type leucogranites (e.g. Harris and Inger, 1992) and < 1 in Lachlan-type SP granitoids. In the diagrams using Nb, Y and Rb, the Mitwaba granitoids cluster at the boundary between within-plate, collisional and volcanic-arc granitoids (Fig. 7) and specifically overlap with the field for post-collisional granitoids of Pearce (1996). We are aware that the position of the data in these diagrams precludes their use for tectonic setting interpretation but we are more interested by showing that again the Mitwaba granitoids overlap the average composition of the Lachlan SP granitoids. These granitoids are characterized by Nb content between 6 and 16 ppm and Ta concentrations between ~ 3 and 12 ppm, yielding Ta/Nb ratios in the range 0.2–0.8. For the above range of Nb, granitoids containing Ta contents higher than 2 ppm are commonly of S-type affinity (Harris et al., 1986).

Symbols:

- ◇ Kabonvia granodioritic augen gneiss
- △ Kisele monzogranitic gneiss
- ▤ Fwifwi foliated leucomonzogranite
- Nyangwa monzogranite
- LFB: Field for Lachlan Fold Belt SP
- HL: Field for Himalayan belt SP

- ◆ Kungwe-Kalumengongo monzogranite
- Average Himalayan leucogranites
- Average I-type granites from LFB
- Average S-type granites from LFB
- HR: Field for Hercynian belt SP
- SP: Strongly Peraluminous Granites

Fig. 7. (a) Nb vs. Y; and (b) Rb vs. (Y + Nb) tectonic setting discrimination diagrams of Pearce et al. (1984) and Pearce (1996). See the text for more details.

Th contents vary between 13 and 25 ppm, except in one sample containing 27 ppm Th. Uranium ranges between 2.5 and 7.5 ppm, except in three samples containing between 13 and 19 ppm. The Th/U ratio is variable and range from 1 to 6, with an average of 3.7, which is close to the crustal average of 3.8 (Taylor and McLennan, 1985). S-type felsic magmas derived from the melting of sedimentary rocks are usually characterized by Th < 25 ppm and U < 10 ppm (e.g. Taylor and McLennan, 1985). Rb/Th ratios in Mitwaba granitoids are higher than the chondritic value (~8). They range generally from 13 to 23, with five samples showing lower values (~8–12).

The rare earth elements (REE) patterns are illustrated in Fig. 8a and b. The Mitwaba granitoids are enriched in REE and display moderately to strongly fractionated patterns ($La_N = 59–84$, $Yb_N = 4–12$, $La_N/Yb_N \sim 6–14$). The REE fractionation is greater for LREE ($La_N/Sm_N = 2–4$) than for HREE ($Gd_N/Yb_N = 1–2$). The Kisele monzogranite is mainly characterized by sub-flat HREE patterns (Fig. 8b). LREE and HREE are not correlated with silica content and a negative europium anomaly occurs in all the samples ($Eu/Eu^* = 0.3–0.6$). For comparison, S-type leucogranites are commonly characterized by $La_N < 100$, $Yb_N < 10$ and are marked by strong

Fig. 6. (a) Normative Qz–Ab–Or diagram (part of the Qz–Ab–Or–An–H₂O system) showing the low-temperature melt area near the cotectic region (E5–P) at 5 kb (Winkler et al., 1975). Small numbers give anorthite contents at the intersection of the isotherms (dashed lines) with the cotectic planes. Most of the Mitwaba granitoid samples (black diamonds) plot in the minimum melt low-temperature region of this system. Also shown in this diagram are fields of felsic melts originating from experimental dehydration melting of metasedimentary rocks at 850 °C and 0.5 GPa (open field) and at 900 °C and 1 GPa (grey-shaded field) (Stevens et al., 1997). See text for details. (b) Plot of SiO₂ against ASI [molar Al₂O₃/(CaO + Na₂O + K₂O)] for Mitwaba granitoids. (c) CaO/Na₂O vs. Al₂O₃/TiO₂ ratios for experimental vapour-absent melts from natural rocks (data from Sylvester, 1998). Temperatures and percentage of melt produced are shown. The mixing curve (dashed-dotted line), with percentage of mixing between end members of Mitwaba mafic and felsic igneous rocks is shown in comparison to the field for strongly peraluminous granites. Fields for Himalayan (HL), Hercynian (HR) and Lachlan Fold belt (LFB) SP granitoids are drawn. Vectors indicate that Al₂O₃/TiO₂ ratios decrease with the increase of melt temperature, whereas CaO/Na₂O ratios decrease with the increase of clay-rich source-rocks. See the text for discussion. (d) R1 and R2 chemical parameters (De La Roche et al., 1980). $R1 = 4Si - 11(Na + K) - 2(Fe + Ti)$; $R2 = 6Ca + 2Mg + Al$. Also shown in this diagram is the field of syn-orogenic granitoids originating from partial melting of crustal rocks (Batchelor and Bowden, 1985). (e) Molar Al₂O₃/(MgO + FeO_T) vs. CaO/(MgO + FeO_T) for Mitwaba granitoids. The fields shown are from Altherr et al. (2000). See the text for further details. (f) Mg# vs. CaO diagram for Mitwaba granitoids (see the text for details).

Fig. 8. (a and b) Chondrite-normalized REE patterns; and (c and d) primordial mantle-normalized spider diagrams for Mitwaba granitoids. Normalizing values of Sun and McDonough (1989).

negative Eu anomaly shown by $\text{Eu}/\text{Eu}^* < 0.5$ (e.g. Williamson et al., 1996). Zr content shows a good correlation with HREE as illustrated in Zr versus Yb diagram (Fig. 9), whereas there is no obvious correlation with LREE such as La.

Primordial mantle-normalized diagrams (Fig. 8c and d) show marked negative anomalies in Ti, Sr, Nb and subsidiary Ba. The Kisele monzogranite gneiss exhibits a positive U anomaly. This whole-rock uranium anomaly correlates with high U content in zircons from this pluton (see Section 6).

7. Discussion

7.1. Geochronological results

The following are the igneous crystallisation ages for the Mitwaba granitoids: 1386.3 ± 8.1 Ma (Kisele monzogranite gneiss), 1385.5 ± 7.4 Ma (Kabonvia granodioritic augen gneiss), 1383.3 ± 4.8 Ma (Nyangwa monzogranite), 1377 ± 10 Ma (Kungwe-Kalumengongo monzogranite) and 1372 ± 9.6 Ma (Fwifwi foliated leucomonzogranite). The dated gran-

Fig. 9. Zr vs. Yb plot for Mitwaba granitoids. Symbols: see Fig. 7.

itoids cover the spectrum from G_1 (type A) Kibaran granitoids previously inferred to be emplaced at 1331 ± 25 Ma up to G_2 (type B) granitoids supposedly intruded at 1308 ± 25 Ma and G_3 (type C) granitoids assumed to be younger and emplaced at around 1094 ± 50 Ma (e.g. Cahen et al., 1984; Klerkx et al., 1987; Fernandez-Alonso et al., 1986). It emerges from this SHRIMP U–Pb zircon geochronological study that all these granitoids were emplaced over a short period of time, between 1386.3 ± 8.1 Ma and 1372 ± 9.6 Ma, and considering the uncertainty on the individual dates, all these plutons could be coeval at 1381 ± 8 Ma. These geochronological data support the conclusion of Kampunzu et al. (1986) who indicated, based on structural data, that the so-called types A–B–C (G_1 – G_2 – G_3) Kibaran granitoids were early syn-kinematic D_1 Kibaran intrusive bodies. The age of these granitoids date the first contractional deformation in the Kibaran, at least in the Mitwaba area.

The age of the second Kibaran deformation event is younger than 1381 ± 8 Ma, taken above to be the age of D_1 , and is known to be older than the tin granite emplaced between ca. 1000 and 950 Ma (Cahen et al., 1984; Ikingura et al., 1992; Romer and Lehmann, 1995). The Kisele monzogranite gneiss contains metamorphic zircon overgrowths with low Th/U ratios and dated at 1079 ± 14 Ma. Because the Kisele foliated monzogranite is deformed, developing gneissosity with a strong NE-trending solid-state S_2 fabric marking the Kibaran paroxysmal compressional deformation (D_2) superimposed on an earlier fabric reported to D_1 , we tentatively conclude that the metamorphic age of 1079 ± 14 Ma dates the second Kibaran oro-

genic event (D_2). Additional geochronological data are required to constrain this interpretation. Metamorphic zircon overgrowth was recorded only in the Kisele pluton and is lacking in all other granitoids investigated. This difference probably reflects the fact that the Kisele pluton is the only one among the studied granitoid bodies that was reworked under sillimanite zone metamorphism (Table 1), i.e. under temperatures high enough for metamorphic zircon overgrowth. All other plutons were reworked under chlorite zone (Nyangwa, Kungwe-Kalumengongo) or staurolite zone (Kabonvia) metamorphism, i.e. below the temperature required for metamorphic zircon overgrowth (e.g. Mezger and Krogstad, 1997). The large number of low-precision Rb–Sr ages of 1094 ± 50 Ma reported by previous workers as the emplacement age for G_3 granitoids is, within the margin of errors, similar to the metamorphic zircon age reported above. In our interpretation, these old Rb/Sr G_3 Kibaran granitoid dates from the Kibaran belt in central and eastern Africa record the resetting of the Rb–Sr systematic during the D_2 deformation event.

This study also indicates that all the studied granitoids contain inherited zircons, one of them yielding a Paleoproterozoic age of 1929 ± 21 Ma. There is no field evidence to suggest that the Mitwaba granitoids were emplaced through an older basement. Thus, we infer that the “old” zircons in these granitoids are inherited from the metasedimentary source of the granitic magma.

7.2. Characterization of the Mitwaba granitoids

The Mitwaba granitoids have the following characteristics:

- (1) In the field, they define very large batholiths (e.g. Kifinga-Kisele batholith: $70 \text{ km} \times 65 \text{ km}$) surrounded by small satellite plutons. Apart from the Kisele foliated monzogranite gneiss, most of these granitoids intrude into low-to-medium grade metasedimentary rocks. Large batholiths such as Kifinga-Kisele are rimmed by contact metamorphic aureoles (garnet-hornblende hornfels). The granitoids contain restitic micas-rich clots as well as enclaves of metasedimentary rocks. These observations are similar to those described for the Lachlan and Hercynian belts

SP granitoids originating from partial melting of a metasedimentary precursor (e.g. Chappell and White, 1992; Sylvester, 1998).

- (2) The Mitwaba granitoids are devoid of hornblende and pyroxene and contain biotite and primary muscovite. Biotites are Al-rich and have Fe–Mg–Al relationship suggesting that they grew in peraluminous melts. The mineralogical composition of the Mitwaba granitoids shows that they belong to muscovite-bearing peraluminous group of Barbarin (1999). Monazite is a common accessory mineral, tourmaline occurs in several samples and these rocks are similar to the “ilmenite series” granitoids of Ishihara (1977).
- (3) Whole rock chemical compositions of these granitoids are characterized by high silica content (67.24–76.73 wt.%), ASI values between 1.04 and 1.6 (Fig. 6b) and up to 4.8% corundum in the normative composition. All the granitoids are potassic. Rb, Sr, Ba concentrations and ratios of trace elements such as Rb/Th and Rb/Ba suggest an S-type nature for the Mitwaba granitoids. For example, Rb/Th values higher than 20, low Th/U ratios (average 3.6), high Ta contents and high Rb/Hf (26–83), Rb/Y (3.2–12) and Ta/Nb (0.19–1.15) in these granitoids suggest that they are S-type felsic melts derived from a sedimentary protolith (e.g. Harris and Inger, 1992).
- (4) Primordial mantle normalized patterns (Fig. 8) exhibits Ba, Nb, Sr and Ti negative anomalies. The high field strength element (HFSE) anomalies (mainly Nb and Ti) are known to be characteristic of subduction zones (e.g. Saunders et al., 1991) and igneous rocks originating from the continental crust (e.g. Tarney and Jones, 1994). For S-type granitoids, it has been suggested that the Sr negative anomaly coupled to low Na₂O, CaO and high K₂O contents reflect the low versus high concentration of these elements in the source, most probably weathered metasedimentary rocks (e.g. Chappell and White, 1992). The LREE enrichment (e.g. La_N = 59–84 times chondrite) relative to HREE (e.g. Yb_N = 4–12 times chondrite) results most probably from restitic zircon (Harris and Inger, 1992) and this is supported by the correlation between Zr and HREE (e.g. Fig. 9). The composition of Mitwaba granitoids (e.g. Fig. 6) overlaps the field of fel-

sic melts originating from dehydration melting of metapelites/metagreywackes precursors (e.g. Stevens et al., 1997). The only potential metasedimentary source rocks for these granitoids are the Mitwaba Group supracrustal metasedimentary rocks. The Mulumbi group sedimentary rocks post-date the emplacement of these granitoids.

- (5) Cathodoluminescence and SHRIMP data indicate that igneous zircons in these granitoids contain nearly rounded inherited zircon cores, although in a few samples, cores of zircon show irregular shapes. As stressed above, there is no field evidence suggesting that these granitoids were emplaced through an older basement and thus we infer that the xenocrystic zircons were inherited from the metasedimentary source of the magma. The rounding of most of the “old” zircon cores could be of metamorphic origin (e.g. Mezger and Krogstad, 1997), although it could as well result partly from pre-metamorphic sedimentary rounding (e.g. Williams, 2001).

These petrographic and geochemical characteristics of the Mitwaba SP granitoids are consistent with an S-type classification. Investigations of S-type granites and rhyolites have flourished during the last two decades (e.g. Le Fort et al., 1987; Pichavant et al., 1988; Chappell and White, 1992; Harris and Inger, 1992; Williamson et al., 1996; Sylvester, 1998). The common point of agreement among experts of S-type felsic rocks is that they originate from the partial melting of a metasedimentary source, with or without mixing with mafic magmas.

7.3. Petrogenesis

The presence of two or three distinct correlation trends in some binary diagrams (e.g. Mg# versus CaO, Fig. 6) along with the parallelism between the REE and spidergram patterns (Fig. 8) suggest that the Mitwaba granitoids originated from different sources but having almost similar compositions and the petrogenetic processes involved were similar.

Several investigations (e.g. Shearer et al., 1987) have shown that fractional melting could produce geochemical trends observed in S-type leucogranites. The first liquid produced during fractional melting has the composition of minimum melt. In-

creased melting will produce magmas characterized by higher $\text{MgO} + \text{FeO} + \text{TiO}_2$ from dehydration melting of biotite at higher temperatures. Similarly, the CaO content of the melt will increase. The diagram Mg# versus CaO (Fig. 6) shows distinct variation trends within both different plutons and single pluton, each of the trends originating potentially from fractional melting of a single metasedimentary source or from different source materials but with the same composition. The three distinct evolution curves in the diagram Mg# versus CaO (Fig. 6) require at least three different metasedimentary source materials.

A critical melt fraction (>30–40 vol.%) is required in order to separate a viscous granitic magma from its source (Wickham, 1987). Such amount of melt cannot be supplied by dehydration melting of muscovite alone because both field and experimental works showed that this reaction produces small amounts of magma (e.g. Clemens and Vielzeuf, 1987). Dehydration and breakdown of biotite is required in order to generate a large amount (up to 40%) of melt from metasedimentary rocks (e.g. Stevens et al., 1997) to form large batholiths of S-type granites such as the Kifinga-Kisele batholith. Harris and Inger (1992) suggested that vapour-present melting of muscovite in a metapelitic source will result in peraluminous felsic melts characterized by low Rb/Sr (0.7–1.6) versus high Sr/Ba (0.5–1.6) ratios. Such melts should have in addition positive Eu anomalies. Vapour-absent melts from metapelites should have higher Rb/Sr (3–6) versus lower Sr/Ba (0.2–0.7) ratios and negative Eu anomalies resulting from residual plagioclase at the source. The Mitwaba granitoids have low Rb/Sr ratios similar to those recorded in vapour-present melting of muscovite. However, their low Sr/Ba ratios and the absence of positive Eu anomalies preclude the generation of these granitoids by vapour-present breakdown of muscovite. In addition, geochemical data (e.g. CaO, Sr contents and Eu/Eu^* values) suggest that these granitoids contain relict plagioclases, supporting that this mineral was among the residues during partial melting. Therefore, the data indicate vapour-absent partial melting during the genesis of the Mitwaba peraluminous granitoids. The low Rb/Sr values recorded in these granitoids probably indicate their derivation from a source with higher feldspar content than the Himalayan metapelites used by Harris and

Inger (1992) in their geochemical modelling of the collisional, Himalaya-type S-type granitoids.

Underplating and reaction of mafic magmas with metapelites (e.g. Patino Douce, 1995) is known to produce large volumes (up to 40%) of SP leucogranitic melts that resemble S-type granitoids of the Lachlan Fold belt in Australia. The discrimination of S-type granitoids generated by the two distinct processes is difficult but the average chemical composition of Himalayan and Lachlan/Hercynian SP granitoids display substantial differences (e.g. Fig. 6). The Hercynian and especially the Lachlan SP granitoids show strong similarities with the Mitwaba SP granitoids (e.g. Fig. 6). Furthermore, Himalayan S-type leucogranites occur as intrusive sheets closely linked to metamorphic packages and they usually have a crustal-like isotopic signature. In contrast, the Lachlan and the Mitwaba SP granitoids occur as large plutons associated with mafic bodies and they have isotopic signatures indicating mixing between crust and mantle sources. The Sr isotopic ratios of Kibaran sedimentary rocks (e.g. Klerkx et al., 1987) give values close to 0.720 when recalculated back at 1381 Ma. The Sr isotopic compositions of granitoids from the Kibaran belt in Katanga (Cahen et al., 1984 and references therein) yield initial $^{87}\text{Sr}/^{86}\text{Sr}_{(1381\text{ Ma})}$ between 0.7016 and 0.7198, suggesting mixing between crust- and mantle-originating materials. The diagram $\text{CaO}/\text{Na}_2\text{O}$ versus $\text{Al}_2\text{O}_3/\text{TiO}_2$ (Fig. 6c) suggests that Mitwaba SP granitoids could result from mixing of coeval mafic and felsic magmas exposed in the Mitwaba area (Fig. 2). To test this interpretation, we have conducted mixing calculations using: (1) the presumed mafic end-member represented by rocks with the highest Mg# and lowest silica and alkalis from the two mafic intrusions exposed in the Mitwaba region (Lwabwe LM9 and Kidilo KM5); (2) the presumed felsic end-member corresponding to the samples with the highest silica and alkalis and the lowest CaO abundance (e.g. samples K2, Table 3) and chemically overlapping with the field of pelite-derived melts of Patino Douce and Johnston (1991). The quantification results in Table 4 show that, mixing 10–20% Mitwaba mafic magmas with felsic magmas represented by the samples K2 reproduce the major elements and immobile trace elements compositions of the various petrographic groups of the Mitwaba SP granitoids ($\Sigma r^2 < 1$).

Table 4
Quantification of the mixing between mafic and felsic magmas to generate various Mitwaba granitoids

Major and trace elements	Fwifwi foliated leucomonzogranite ^a	Mixed magmas ^a		Calculated mixture ^a	Square difference ^a	Kabonvia Granodioritic augen gneiss ^b	Mixed magmas ^b		Calculated mixture ^b	Square difference ^b
	Sample K23.5 A	Felsic (%) 81.0	Mafic (%) 19.0	B	(A – B) ²	Sample K12.1 A	Felsic (%) 81.0	Mafic (%) 19.0	B	(A – B) ²
SiO ₂ (%)	71.81	60.49	10.52	71.01	0.64	69.39	60.49	9.14	69.63	0.06
TiO ₂	0.42	0.14	0.15	0.28	0.02	0.51	0.14	0.13	0.26	0.06
Al ₂ O ₃	13.22	10.40	2.69	13.09	0.02	13.79	10.40	3.14	13.54	0.06
FeO _T	2.07	1.00	1.39	2.38	0.10	3.47	1.00	2.38	3.37	0.01
MnO	0.03	0.02	0.04	0.06	0.00	0.07	0.02	0.02	0.04	0.00
MgO	0.86	0.41	0.82	1.24	0.14	1.66	0.41	1.37	1.78	0.02
CaO	1.83	0.55	1.10	1.65	0.03	1.36	0.55	1.01	1.56	0.04
Na ₂ O	3.12	3.13	0.14	3.28	0.03	2.68	3.13	0.02	3.15	0.22
K ₂ O	3.75	3.09	0.51	3.61	0.02	4.71	3.09	0.91	4.00	0.50
					$\sum r^2 = 0.99$					$\sum r^2 = 0.97$
La (ppm)	49.50	46.58	2.49	49.06	0.19	50	46.58	3.82	50.39	0.15
Hf	5.00	4.86	0.19	5.05	0.00	5	4.86	0.19	5.05	0.00
Sm	7.70	7.13	0.57	7.70	0.00	7.5	7.13	0.57	7.70	0.04
Nd	41.00	36.05	5.19	41.23	0.05	41	36.05	5.00	41.05	0.00
Pr	11.30	10.13	0.72	10.85	0.21	11	10.13	0.57	10.70	0.09
Lu	0.40	0.81	0.04	0.85	0.20	0.4	0.81	0.04	0.85	0.20
Eu	1.40	0.89	0.15	1.04	0.13	1.4	0.89	0.15	1.04	0.13
Tm	0.50	0.73	0.06	0.79	0.08	0.4	0.73	0.06	0.79	0.15
Tb	1.00	1.22	0.08	1.29	0.08	1	1.22	0.11	1.33	0.11
					$\sum r^2 = 0.95$					$\sum r^2 = 0.88$

^a Parent magmas used to generate the Fwifwi leucomonzogranite are: Kisele monzogranite (sample K2.3, Table 3), and the Kidilo mafic intrusive (KM) having the following chemical composition (%): SiO₂: 55.38; TiO₂: 0.77; Al₂O₃: 14.17; Fe₂O₃: 7.3; MnO: 0.22; MgO: 4.3; CaO: 5.81; Na₂O: 0.76; and K₂O: 2.7. Traces (ppm): La: 13.1; Hf: 1; Sm: 3; Nd: 27.3; Pr: 3.8; Lu: 0.2; Eu: 0.8; Tm: 0.3; and Tb: 0.4.

^b Parent magmas used to generate the Kabonvia granodiorite are: the Kisele monzogranite (sample K2.3, Table 3) and Lwabwe mafic intrusive (LM) having the following chemical composition (%): SiO₂: 48.10; TiO₂: 0.66; Al₂O₃: 16.50; Fe₂O₃: 12.50; MnO: 0.10; MgO: 7.21; CaO: 5.32; Na₂O: 0.10; and K₂O: 4.79. Traces (ppm): La: 20.09; Hf: 1; Sm: 3; Nd: 26.3; Pr: 3; Lu: 0.2; Eu: 0.8; Tm: 0.3; and Tb: 0.6. (Mafic rocks data from Kokonyangi et al., unpublished.)

8. Geotectonic implications

Three geotectonic settings will be evaluated for the Kibaran granitoids from the Mitwaba area and for coeval and petrologically similar Kibaran granitoids exposed in Burundi, Rwanda and NE Tanzania.

The first model is an extensional (plate divergent) setting, i.e. a continental rift setting proposed by Klerkx et al. (1987). In modern tectonic environments, nowhere very large S-type granitic batholiths unrelated to felsic A-type intrusions have been reported in plate divergent setting. We have shown that the Mitwaba granitoids were emplaced during a compressional (D_1) tectonic event, which preclude a pristine extensional environment. Plate divergent settings are marked by extensional tectonic structures and alkaline-peralkaline felsic rocks with a very specific high-field-strength geochemical signature unknown in the Kibaran granitoids.

The second model is a continental collision setting similar to the Himalaya where S-type granitoids are widespread (e.g. Le Fort et al., 1987). In such setting, S-type leucogranites mark an advanced stage of crustal thickening by stacking of thrust sheets. In the Mitwaba area, the S-type batholiths were emplaced ~ 300 Ma before D_2 , the climax of collisional deformation tentatively dated at ~ 1080 Ma in this paper. D_1 deformation event does not seem to represent an advanced stage of crustal thickening in the Kibaran belt. Himalayan-type S-type granitoids usually plot well inside the collisional granites field in Pearce (1996) discrimination diagram. The syn- D_1 Mitwaba granitoids originate also from partial melting of sedimentary rocks but in contrast to the Himalayan S-type granitoids, they plot at the boundary between volcanic-arc, collisional and within-plate granite fields (Fig. 7). We infer that this peculiar feature result from the mixing between mafic and felsic magmas during the genesis of the Mitwaba SP granitoids. Huppert and Sparks (1988) stressed the important role of mafic magmas in the generation of granitoids in the crust and experimental work (e.g. Patino Douce, 1995) showed that a large amount (up to 40 vol.%) of felsic melts ($>70\%$ SiO_2) could be produced by the interaction of hot mafic magma with metapelitic material. The degree of crustal melting required for generating, during the early stage of an orogen, the large S-type granitic batholiths known in the Kibaran belt could not be achieved without input of ex-

ternal heat. Although part of heat required to generate the large S-type granitic batholiths is probably from radioactive decay as suggested by high-U contents in the restitic zircons (see Table 2), we are suggesting that the ~ 1.38 Ga mafic rocks closely associated to the Kibaran SP granitoids (e.g. Fig. 2) were probably the major source of heat required for melting the Kibaran metasedimentary rocks of the Mitwaba group. The Mulumbi group metasedimentary rocks post-date the emplacement of the syn- D_1 Kibaran SP granitoids and thus cannot be considered as their potential source.

A mechanism that can supply heat for crustal melting is underplating of mafic magma at the Moho and/or within the crustal metasedimentary package. This process is important both in volcanic-arc (e.g. Hyndman and Foster, 1988) and in extensional setting (e.g. Bergantz, 1989). Structural investigations indicated that the Mitwaba area was undergoing compressional tectonics (D_1) during the emplacement of the S-type granitoids and thus cannot mark an extensional regime. Therefore, the Mitwaba S-type granitoids formed in a continental-arc setting as previously suggested by Kampunzu et al. (1986). Geochemical data related to mafic magmas emplaced at the same time as the SP granitoids also record a continental arc setting (Kokonyangi, unpublished data). The two cratons involved during this convergence are the Congo craton to the northwest and the Bangweulu block/Tanzania craton to the southeast. The absence of any possibility of correlation of geological units on the two sides of the Kibaran belt indicates that the two-cratonic blocks were exotic to each other. The formation of S-type felsic melts in subduction zones is a well-known fact in modern orogens (e.g. Pichavant et al., 1988). The arc signature documented in the mafic magmas emplaced at the same time as the SP granitoid melts in the Mitwaba batholiths supports this interpretation. The petrogenesis and geodynamic setting during the emplacement of Kibaran peraluminous granitoids present several analogies with SP granitoids in the Lachlan Fold belt and in the Variscan belt in Europe (e.g. Bea et al., 1999). The Variscan peraluminous granitoids were produced in a belt thickened by underthrusting of the lower crust (Azor et al., 1994), i.e. a crust thickened by subduction rather than by continental collision. Bea et al. (1999) emphasized the production of Variscan SP granitoids during an orogeny-related extensional event separat-

ing two compressional events. In this study, we do not see evidence of orogeny-related extensional tectonics during the emplacement of the Kibaran granitoids. Arc mafic magmas are almost coeval with SP crustal felsic melts and therefore we infer that this is an important parameter for the genesis of peraluminous granitoids in these belts. However, *Bea et al. (1999)* suggested that the amount of heat advected to the middle crust by mafic magmas was insufficient to cause significant melting of crustal rocks and preferred intra-crustal radioactive decay as the main heat source. An important observation that can be made from the data of *Bea et al. (1999)* is that the “mafic precursor” rocks in the Variscan belt show abnormal compositions indicating that they are hybrid rocks from various degree of mixing of mantle- and crust-originating melts. Such feature imply that a large proportion of this mixture could be trapped below the currently exposed crustal level because, in such a scenario, the most viscous felsic melt will impend the ascent to shallow level of the less viscous mafic melts. Whatever the model adopted, what is critical is that the Mitwaba S-type granitoids were formed and emplaced within a plate convergent setting, most probably during a subduction event.

9. Conclusion

New, high precision ages presented in this paper show that the emplacement of the three generations (G_1 , G_2 and G_3) of granitoids previously reported to span the period 1370 ± 25 Ma to 1094 ± 50 Ma in the Kibaran belt occurred within a short period of time at 1381 ± 8 Ma and date the first deformation event D_1 in the Kibaran orogenic system in central Africa. The Kisele monzogranite gneiss was strongly deformed and metamorphosed during D_2 and hosts metamorphic zircons yielding an age of 1079 ± 14 Ma, which sets the age of the D_2 deformation event and related metamorphism in the Mitwaba area.

The major- and trace-element compositions (e.g. high SiO_2 , Al_2O_3 , ASI ratios, normative corundum, low CaO, Th and U contents and high Rb/Zr and Rb/Th) of the syn-kinematic D_1 Kibaran granitoids from the Kibaran belt are consistent with their genesis by partial melting of metasedimentary rocks of the Mitwaba group. The emplacement of mafic and felsic

igneous rocks in the Mitwaba occurred within the same period of time were coeval. The source of heat for the production of the large S-type granitic batholiths in the Kibaran is inferred to be underplating of arc mafic magmas at the base and within the crust. However, additional heat could have been supplied by radioactive decay from high U, Th and K metasedimentary rocks as shown by high U-content of relict zircon and high K-content of these granitoids. Available data do not support a Himalaya-type continental collision setting during the genesis of these S-type granitoids. A model involving subduction, genesis of arc mafic magmas, substantial intra-crustal melting of metasedimentary rocks triggered by the emplacement of arc mafic magmas and mixing in variable proportions of these mafic and felsic magmas is taken to be the dominant factor during the genesis of these peraluminous granitoids.

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