Geology, mineralogy and mineral chemistry of the NYF-type pegmatites at the Gabal El Faliq area, South Eastern Desert, Egypt

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The Gabal El Faliq area is located in the Eastern Desert of Egypt and belongs to the Wadi Hafafit–Wadi Ghadir fold belt. Lenses and dykes of rare metal-bearing pegmatites occur along major weak zones in the area (Wadi El Gemal strike-slip and its Riedel shears), by which they are structurally controlled. These pegmatites are mineralogically complex. Moreover, they are zoned, and the intermediate zones of the pegmatite varieties contain high amounts of zircon, fergusonite-(Y), columbite-(Fe), allanite-(Ce), xenotime-(Y), monazite-(Ce and Nd), thorite and gummite. These minerals carry Zr, rare earth element (REE), Nb, Ta, Th, U > Ta, F, and they are classified into rare-metal pegmatites, specifically, the niobium, yttrium and fluorine (NYF) enrichment family. The geochemistry aspects of these pegmatites indicate Zr up to 28,984 ppm, Hf up to 1171 ppm, REE up to 19,206 ppm, Nb up to 9962 ppm, Ta up to 152 ppm, U up to 565.5 ppm, Th up to 850 ppm and Ga up to 90 ppm. They were formed under an intracontinental tectonic setting from a basaltic magma similar to the oceanic island basalts (OIBs) contaminated with the crust. Consequently, this study could be used as a guide for the exploration of rare metals in similar regions of the world.

Keywords. Gabal El Faliq; Wadi Hafafit–Wadi Ghadir fold belt; Pan-African orogeny; rare-metal NYF pegmatites.

1. Introduction

Pegmatite is a term used to describe very coarse (crystals that are at least 1 cm in diameter) intrusive igneous rocks, which are mafic–syenitic–granitic in composition.

Granitic pegmatites have been classified by Černý (1991) according to their composition into four classes or categories. This classification is a combination of depth of emplacement, metamorphic grade and minor elements. The classes are abyssal (high-grade metamorphism, high to

low pressure), muscovite (high pressure, lower temperature), rare-element (low temperature and pressure) and miarolitic (shallow level). Moreover, rare-element pegmatites are subdivided into lithium, caesium and tantalum (LCT) and niobium, yttrium and fluorine (NYF) families. The LCT family is characterised by Li, Rb, Cs, Be, Ga, Sn, Hf, Nb <> Ta, B, P, F, and the NYF family is characterised by Y, rare earth element (REE), Ti, U, Th, Zr, Nb > Ta, F (Černý 1991). Consequently, some pegmatite types are a source of REEs and gemstones.

Rare metal (Nb, Ta, U, Th, Zr and REE) deposits were discovered in felsic igneous rocks from Egypt, including alkaline/peralkaline igneous rocks and hydrothermally altered parts of them (e.g., Abu Elatta 2018), apogranites and pegmatites. The alkaline/peralkaline rocks in Egypt were formed during the period between the late Pan-African orogeny (~600 Ma) to post-Cretaceous (50 Ma, Vail 1989) along the continental breakup by the major shear zones (e.g., Garson and Kris 1976; Hashad and El Reedy 1979; Serencits *et al.* 1979; Abu Elatta *et al.* 2013; El Sayed *et al.* 2014). Apogranite is a specialised granitic type (albitised granite) having rare metals (Sabet and Tsogoev 1973; Beus 1982), and was studied by, e.g., Omar (1995), Attawiya *et al.* (2000), Raslan (2008) and Ali *et al.* (2011). Rare metal-bearing pegmatites were recorded in many localities in Egypt (figure 1), especially in the Eastern Desert, and they are of LCT or NYF families. For example, pegmatites of the LCT family occur at Abu Rusheid (Sabet and Tsogoev 1973), and pegmatites of the NYF family occur at Ras Baroud (Omar 1995) and Kadabora (El-Sayed 2003).

This paper presents the geology of the pegmatites that cross-cut the Gabal El Faliq leucogranites and their tectonic relation with the Wadi Hafafit–Wadi Ghadir rock units. The mineralogical, mineral chemistry and geochemical

studies of the intermediate zones of the pegmatites show mineralisation with Zr, Hf, REE, Nb, Ta, U, Th and Ga. This work was to understand if these pegmatites were formed by the *in-situ* partial melting or by the fractionation from mantle-derived magma.

2. Geological setting

2.1 Regional geology

The late Pan-African Gabal El Faliq leucogranites occur along the south-western side of the Wadi Hafafit–Wadi Ghadir area, in the South Eastern Desert of Egypt (figure 2a). The Wadi Hafafit–Wadi Ghadir area is part of the Arabian–Nubian shield of the Eastern Desert of Egypt and consists of two major groups, which are separated by the Nugrus thrust. The structurally lower group in the foot wall of this thrust is characterised by the Migif–Hafafit gneisses and associated rocks which have been affected by medium-grade metamorphism, whereas the structurally upper group or hanging wall is characterised by the Wadi Ghadir mélange which has experienced low-grade metamorphism (El-Ramly *et al.* 1993). These groups were formed as a result of the Pan-African convergent and marginal ocean basin processes. In addition to these, there are a group of Phanerozoic

Figure 1. Locations of some rare-metal pegmatites in the Eastern Desert of Egypt.

Figure 2. (a) Geologic map of the Hafafit area (El-Ramly *et al.* 1993; Abd El-Naby and Frisch 2006) and (b) detailed geologic map of Gabal El Faliq (modified from El-Ramly *et al.* 1993) showing sample locations in black numbers.

intrusive (unmappable) and alluvium (El-Ramly *et al.* 1984).

The Migif–Hafafit gneisses form the Hafafit domes. These comprise the orthogneiss of tonalitic, trondhjemitic and granodioritic composition, banded amphibolites overthrust by ultramafic rocks, alternating bands of biotite and hornblende and psammitic gneisses. In some parts, the amphibolite is associated with metagabbros, while the upper group, the Wadi Ghadir mélange forming the Ghadir ophiolite includes mica-schists and metavolcanics (Abd El-Naby and Frisch 2006).

The sequence of rock units in the Wadi Hafafit–Wadi Ghadir fold belt was formed as follows: Firstly, the ophiolites of the Arabian Nubian shield (including the Hafafit–Ghadir area) by the rifting of Rodinia (Abdelsalam and Stern 1996) between 900 and 740 Ma (Ries *et al.* 1983; Kroner *et al.* 1994; Loizenbauer *et al.* 2001). The event was followed by the development of sutures associated with the arc accretion between 750 and 650 Ma (Abdelsalam and Stern 1996; Blasband *et al.* 2000), when the tonalitic, trondhjemitic and granodioritic granites at the core of the Hafafit domes were formed as a result of the convergence and associated orogen-parallel extension (figure 3a). Secondly, the late- to post-tectonic magmatic activity (Stern and Hedge 1985; Hassan and Hashad 1990) and molasse basin formation (Grothaus *et al.* 1979; Willis *et al.* 1988; Rice *et al.* 1993) occurred between 620 and 580 Ma (Fritz *et al.* 2002) as a result of the arc–continent or arc–arc collision. This event is responsible for overthrusts taking place in the Hafafit–Ghadir area, where the Ghadir ophiolite mélange was thrust over the Hafafit units and led to a medium grade of amphibolite facies regional metamorphism that caused the orthogneisses (previously tonalitic, trondhjemitic and granodioritic granites) in the Hafafit units. In addition, this event increased the crustal thickening and caused a widespread partial-melting along the thrust zones, where leucogranites in the Hafafit area (including the Gabal El Faliq granites; figure 3b) were emplaced along the thrusts around 600 Ma (Abd El-Naby and Frisch 2006). This stage of crustal thickening and the emplacement of the leucogranites were accompanied by active strike-slip faults or shear zones, especially along the previous crustal weaknesses (Moghazi *et al.* 2004). These strike-slip faults or shear zones were deeper, so that, they were associated with magmas that have been represented by pegmatite dykes and lenses in the study area

within the fault zones and related Riedel shears (figure 3c). In addition, these faults or shear zones may have been reactivated during the Red Sea opening or before, as explained by the cracks in the pegmatites.

2.2 Gabal El Faliq leucogranites

The Gabal El Faliq leucogranitic rocks occur as an elongated body along the NW–SE oriented Wadi El Gemal thrust and cover an area of about 5.65×1.67 km (figure 2b), forming a moderate relief terrain with the Gabal El Faliq peak of 664 m above sea level. The leucogranites intruded into the Wadi Hafafit rock units and are dissected by a major sinistral strike-slip fault (the Wadi El Gemal strike-slip fault) and associated Riedel shears. The main strike of this fault is the NW–SE trend, whereas the trends of synthetic Riedel shear strikes are E–W and N–S trends, respectively. The zones of these faults vary in width and are characterised by fault breccias, gouge, ferrugination (figure 4a) and pegmatite bodies that have abnormal concentrations of rare metals, such as Nb, Ta, Th, U, Zr, HF and REE in a few parts, while the other parts are covered by the Wadi sediments. The leucogranitic rocks contain many large xenoliths from the country rocks along their margins (figure 4b). Additionally, some basic dykes with a NW–SE trend also have been intruding into these leucogranites.

Saleh *et al.* (2014) divided the Gabal El Faliq leucogranites into two varieties according to their mineral contents: the first is represented by monzogranite and the second, by alkali feldspar (AF) granite.

2.3 Pegmatites

The granitic pegmatites at the Gabal El Faliq area occur as dykes and lenses. Granitic pegmatitic dykes cut all the rock units in the Gabal El Faliq area (figure 4c), where there is no relationship among them in the studied area. These pegmatites intrude into the Wadi El Gemal fault and the related Riedel shears, so that they are coincident with the NW–SE, E–W, and N–S strikes of these faults. Their widths range from 0.20 cm to 2 m and the lengths vary. They have sharp contacts with the host rocks (figure 4d). They are zoned (figure 4d

Figure 3. Plate-tectonic model for the evolution of the Wadi Hafafit–Wadi Ghadir rock units; where (a) and (b) are sketch diagrams (after Abd El-Naby and Frisch 2006) illustrating the development of the Gabal El Faliq leucogranites and their tectonic relation with the Wadi Hafafit–Wadi Ghadir rock units; (a) >650 Ma: subduction of the oceanic crust, ophiolite detachment and thrusting with arc volcanism, arc-related plutonism and back-arc basin development in the Hafafit-Ghadir region, (b) 600 Ma: collision and thrusting of the Nugrus unit over the Hafafit unit, leucogranite intrusion along thrust zones and regional metamorphism and (c) is sketch diagram showing trends of the Wadi El Gemal strike-slip fault and its Riedel shears where the pegmatites were injected.

and e) and are divided into complex-type granitic pegmatites. The pegmatites show complex mineralogy and textures, where their mineral composition and texture vary from the wall zones to the core zones. Additionally, they contain xenoliths from host rocks, and especially those found on the north-western side of Gabal El Faliq (figure 4e). The

intermediate zones of these pegmatites have high concentrations of rare metal-bearing pegmatites (figure 4f–h). Pegmatite lenses up to 30 m in length and 10 m in width occur along the NW–SE sinistral strike-slip fault and its related shears. Furthermore, they are of the zoned complex type with the same mineral composition of pegmatitic dykes.

Figure 4. (a–e) are field photographs showing, ferrugination and fault gouge, a large xenolith from the older mafic rocks in leucogranites, a sharp contact between a pegmatitic dyke and the host rock and zoned pegmatites, respectively, whereas (f–h) are photographs of samples from pegmatite intermediate zones that contain high concentrations of rare metal-bearing minerals.

3. Methodology

The locations of all samples collected from the pegmatites are shown in figure 2(b). A total number of 17 thin sections (5 from country rocks and 12 from pegmatite intermediate zones) were studied petrographically by using an optical microscope and a scanning electron microscope attached to a JEOL JXA-8900L electron microprobe equipped with a Si(Li) energy dispersive spectrometer (EDS) and five wavelength dispersive spectrometer (WDS, Department of Earth and Planetary Sciences, McGill University). Most of the minerals except main rock-forming minerals were identified from compositional data using EDS. Quantitative analyses were conducted in the WDS mode with a beam current of 50 nA, an acceleration voltage of 20 kV and a dwell time of 60 ms. An environmental scanning electron microscope (ESEM) coupled with an energy dispersive X-ray spectroscope (EDX) (Laboratory of Nuclear Materials Authority, Egypt) was also used to identify some minerals. Separated heavy minerals were hand-picked under the microscope and analysed by ESEM/EDX. Acme Labs., Canada using ICP-OES and ICP-MS methods to analyse eight representative samples from the pegmatite intermediate zones for whole-rock compositions (two samples of Nos. 2 and 10 from lenses and the other six samples from dykes) (table 1).

4. Results

4.1 Mineralogy

4.1.1 Leucogranites mineralogy

The two types of leucogranites are highly sheared by faults, and all crystals are deformed and cracked (figure 5a). The monzogranite is pink in colour and is composed essentially of K-feldspars, sodic plagioclase and biotite with a minor amount of muscovite and hornblende. The accessory minerals are titanite, zircon, garnet and opaque minerals (figure 5b). K-feldspars are orthoclase perthite dominant and also contain microcline perthite. Two types of quartz were found, particularly, the samples were taken from the fault zones: the first one occurs as subhedral to anhedral crystals showing undulatory extinction. It causes corrosion of feldspar minerals, and the second is smaller, relative to the first one. It is squeezed among the primary minerals (figure 5c) and fractured crystals. AF granite is brick-red in colour, medium- to coarse-grained and mainly

consists of K-feldspars, quartz and albite, with minor traces of alkali amphibole (arfvedsonite). Anhedral opaque crystals, zircon and monazite are the accessory minerals. K-feldspars are represented by predominantly string (figure 5d) and patchy types of microcline perthite more than orthoclase perthite.

4.1.2 Pegmatites mineralogy

The intermediate zones of the pegmatite dykes and lenses were studied in detail because they contain a large amount of rare metal-bearing minerals. The intermediate zones are composed essentially of quartz (53–80% by volume), magnetite and titanomagnetite (5–25% by volume), patch perthite (3–17.5% by volume), with a minor amount of arfvedsonite (about 0.5% by volume) and haematite (3% by volume). Chlorite and biotite are occasionally found in the pegmatite bodies that were intruded into biotite gneisses. The rare metal-bearing minerals are zircon (0.1–5% by volume), fergusonite-(Y) (0.3–3% by volume), allanite-(Ce) (1–4% by volume), monazite-(Ce and Nd), (0.5% by volume), with a small amount of columbite-(Fe), xenotime-(Y), thorite and gummite that form about 1% by volume.

Quartz occurs as colourless crystals that have been deformed and show undulatory extinction. These crystals are accompanied by oriented chlorite (figure 6a) and/or feldspar (patch perthite). These textures suggest that the Gabal El Faliq pegmatites were emplaced into fault zones, where these faults reactivated prior to the complete crystallisation of the pegmatites. Magnetite and titanomagnetite occur as dark anhedral mega crystals (up to 3 cm in length and 1.5 cm in width). These crystals are fractured with two or more sets of fractures and contain zircon and fergusonite inclusions (figure 6b). Additionally, fractures are filled with microvioletlets of fergusonite (figure 6c) and/or quartz. Zircon occurs as euhedral to subhedral crystals with pyramidal terminations. These crystals vary in size from medium to mega-crystals (up to 2.0 cm in length and 1.2 cm in width; figure 6d and e) and colours, with colour zoning (figure 5d and e). The zircons contain many inclusions of fergusonite (figure 6f), columbite, xenotime and haematite (figure 6g). Allanite-(Ce) shows a shade of brown colour with a dark core in the thin sections (figure 6h) and is pleochroic in brown, red-brown and yellow-brown colours. Most of the allanite crystals were altered and occur in

Table 1. Whole-rock geochemical data in wt% (major oxides) and ppm (trace and REEs) of the Gabal El Faliq pegmatite intermediate zones.

Sample/oxide	2	4	6	7	8	9	10	12
SiO ₂	62.75	78.41	73.73	83.99	69.83	71.41	83.01	62.16
TiO ₂	0.28	0.31	0.53	0.35	0.7	0.59	0.2	0.67
Al ₂ O ₃	6.89	4.25	0.86	2.73	0.80	0.92	3.76	1.12
Fe ₂ O _{3t}	22.68	12.7	21.43	9.22	27.4	24.67	4.7	28.97
MgO	0.07	0.06	0.03	0.1	0.03	0.08	0.14	0.06
CaO	0.76	0.51	0.13	0.61	0.05	0.08	0.39	0.23
Na ₂ O	1.16	1.63	0.06	0.88	0.02	0.1	0.12	0.7
K ₂ O	3.56	0.54	0.03	0.45	0.02	0.07	0.38	0.1
P ₂ O ₅	0.04	0.02	<0.01	0.05	<0.01	<0.01	0.16	<0.01
MnO	0.06	0.12	0.07	0.07	0.08	0.13	0.02	0.14
L.O.I	0.5	0	0.5	0.8	0.5	0.7	2.1	1.2
Total	98.75	98.55	97.37	99.25	99.43	98.75	94.98	95.35
<i>Trace elements</i>								
Ba	352	74	9	118	5	27	108	13
Cs	0.2	0.1	<0.1	0.2	<0.1	<0.1	0.3	0.1
Ga	31.7	20.4	68.6	16	89.8	59.8	10	53.3
Hf	65.3	25.4	211.8	18	46.6	36.3	1170.9	384.8
Rb	59.2	8.6	2	6.8	0.8	1.7	8.7	4.6
Sr	114	40	10.6	60	3.3	9.9	52	21.6
Nb	2911.4	4285	4795.6	2108	915	3066.4	1914.8	9661.8
Zn	124	166	430	132	428	185	153	588
Ta	43.2	140.8	152.4	25.5	85	65.3	67	146.5
Zr	1695.2	579	4151.3	443.6	656.4	877.3	28,983.8	8964
U	123.4	184.9	332.3	87.1	90.3	192	240.4	565.5
Th	220.1	301.3	850	174.5	597	257.7	195.9	881.5
<i>REE</i>								
La	51.5	21.8	343.5	11.2	7.5	90.4	291	990.2
Ce	173.6	92.3	1121.8	34.7	31.8	316.4	745.3	3113.3
Pr	32.11	22.55	157.78	8.12	6.7	45.18	107	450.93
Nd	189.6	166.6	732.3	69.9	44.7	236.6	487.9	2012.2
Sm	123.99	144.89	289.48	69.13	36.79	125.91	172.8	688.2
Eu	6.43	8.18	5.84	3.77	0.33	5.52	6.46	13.56
Gd	227.11	297.94	472.87	164.94	68.73	245.4	233.75	1028
Tb	55.34	75.48	109.2	40.75	16.56	60.11	54	229.7
Dy	381.8	551.48	748.25	290.17	110.5	433.5	395.27	1609.15
Ho	80.5	118.4	166.35	61.69	24.17	99	90.32	350.52
Er	228	347.17	454.79	171.17	71.5	292.18	288.5	973.52
Tm	30.37	50.16	63.68	22.23	11	41	47.96	126.52
Yb	156.79	286.32	341.84	114.19	70.84	228.77	301.39	664.83
Lu	18.72	37.19	41.21	13.24	9.5	28.21	43.45	76.95
Y	1703.8	2914.7	3427	1378.2	565.9	199.4	2182	6878.2
∑LREE	577.23	456.32	2650.7	196.82	127.82	820.01	1810.46	7268.39
∑HREE	2882.43	4678.84	5825.19	2256.58	948.7	1627.25	3636.64	11,937.39
∑REE	3459.66	5135.16	8475.89	2453.4	10,767.52	2447.58	5447.1	19,205.78
Eu/Eu*	0.12	0.12	0.05	0.11	0.02	0.1	0.1	0.05
Th/U	1.78	1.63	2.58	2.00	6.61	1.34	0.81	1.59
Rb/Cs	296	86	u.d.	34	u.d.	u.d.	27.7	46
Nb/Ta	67.39	30.4	31.46	82.67	10.78	46.95	28.78	65.95
Zr/Hf	25.96	22.6	19.6	24.64	14.1	24.17	24.75	23.3
Y/Nb	0.59	0.68	0.71	0.65	0.62	0.07	1.14	0.71

Fe₂O_{3t} = total iron, HREEs include Y, u.d. = under detection limit.

Figure 5. Optical microphotographs (CN) showing (a) deformed polysynthetic twinning of plagioclase, (b) opaque minerals, (c) squeezed quartz, and (d) deformed string perthite (CN = Crossed Nicols).

different sizes from medium to mega-crystals (up to 1.6 cm in length and 0.5 cm in width; figure 6h–j). These crystals contain inclusions of monazite-(Ce) (figure 6i and j). Fergusonite-(Y) occurs as reddish-brown euhedra to subhedral prismatic crystals and the aggregates vary in size from medium to mega-crystals (up to 2.4 cm in length and 0.6 cm in width; figure 6k and l). Columbite crystals are locally found as large black prismatic crystals but they commonly exist as inclusions in zircon as mentioned above. Xenotime occurs as euhedral to subhedral prismatic grey colour crystals, commonly fractured (figure 6m and n). These crystals are also found as inclusions in thorite and zircon crystals (figure 6d and g). Monazite occurs as minor reddish-brown monazite-(Ce) crystals (figure 6i) and monazite-(Nd) which is more abundant either in the form of large or small pale yellow prismatic crystals (figure 6o). Thorite occurs as anhedral black or brown prismatic crystals. Gummite occurs as yellow-coloured grains (figure 6p) and may be formed as a result of uraninite alteration.

4.2 Pegmatites mineral chemistry

Quantitative electron microprobe analyses were carried out on fergusonite [(Y,REE)NbO₄],

allanite [(Ce, Ca, Y, La)₂(Al, Fe³⁺)₃(SiO₄)₃(OH)] and zircon (ZrSiO₄), and semi-quantitative analyses by ESEM/EDX for titanomagnetite [Fe²⁺(Fe³⁺, Ti)₂O₄], xenotime (YPO₄), monazite (Nd, Ce, La, Th)(P, Si)O₄, thorite (ThSiO₄) and gummite (U₃O₈).

The electron microprobe analyses of fergusonite crystals (15 crystals; table 2) reveal that they contain Nb₂O₅ (46.72–50.03 wt%), followed by Y₂O₃ (27.29–28.84 wt%), with total heavy rare-earth oxides (HREO) including Y₂O₃ (40.38–44.86 wt%) and total light rare-earth oxides (LREO: 1.73–3.71 wt%) as well. Furthermore, they contain UO₂ (1.62–2.19 wt%) with a considerable amount of ThO₂ (0.81–2.01 wt%), and they represent fergusonite-(Y). The chondrite-normalised REE patterns of fergusonite (figure 7) show high heavy rare-earth elements (HREEs) with pronounced negative Eu anomalies. In addition, some crystals show positive Ce anomalies and negative Pr anomalies at the same time.

The electron microprobe analyses of zircon crystals (7 spots; table 3) contain (63.54–64.34 wt%) ZrO₂ and (32.49–33 wt%) SiO₂ with a significant amount of (2.22–2.29 wt%) HfO₂. These crystals have significant concentrations of rare earth oxides (REO) (0.429–1.245 wt%), mostly HREO. In addition, they are very rich in U relative to Th.

Figure 6. (a–c) Optical microphotographs (normal light (NL)) showing chlorite and fergusonite, magnetite/titanomagnetite contains zircon and fergusonite as inclusions, and a fracture in magnetite/titanomagnetite filled by fergusonite, respectively, (d and e) are optical microphotographs (CN) showing mega-crystals of zircon containing xenotime as inclusions, (f and g) are backscattered electron (BSE) images showing zircon crystals containing inclusion of fergusonite, xenotime and haematite, (h and i) are optical microphotographs showing mega-crystals of allanite with a dark core in h and contains monazite-(Ce) in (i), and (j) is BSE for allanite containing monazite-(Ce), (k) and (l) are optical microphotographs (NL) showing aggregates of fergusonite mega-crystals, and (m–p) are BSE images for thorite grain, rock fragment contains cracked xenotime, zircon and fergusonite, minute crystals of monazite-(Nd) and grains of gummite, respectively.

Furthermore, Fe_2O_3 (total iron as Fe_2O_3) and TiO_2 are very limited. The chondrite-normalised REE patterns of zircon crystals (figure 8) show an enrichment in the HREE relative to the light rare-earth element (LREE). Additionally, Eu is depleted in most crystals. These results indicate that the zircon crystals were formed in

an oxygenated environment from hydrothermal solutions which came from highly differentiated magma.

The electron microprobe analyses of allanite crystals (eight crystals; table 4) showed that they contain SiO_2 (31.5–34.09 wt%), followed by REEO (22.51–24.00 wt%; LREO = 21.05–22.38 wt% and

Figure 6. (Continued.)

HREO = 1.46–1.97 wt%), Al_2O_3 (15.27–18.06 wt%), Fe_2O_3 (10.22–16.05 wt%) and CaO (7.51–9.46 wt%). MnO concentration is greater than 1 wt% and many other oxides are less than 1 wt%. ThO_2 is present whereas UO_2 is not present or exists in a very limited amount, and ThO_2/UO_2 ratios are 0–199. Accordingly, they represent allanite-(Ce). The chondrite-normalised REE patterns of these crystals (figure 9) indicate that $(\text{LREE})_N$ values are very high relative to $(\text{HREE})_N$ values with a depletion of Eu. Additionally, the crystals show slight positive Ce anomalies. These indicate that the crystals originated from an oxygenated environment.

The semi-quantitative analyses by ESEM/EDX for magnetite (figure 10a) indicate that it contains Fe_2O_3 (98 wt%) and TiO_2 (2 wt%). Titanomagnetite contains Fe_2O_3 (85.52 wt%) and TiO_2 (14.48 wt%; figure 10b). Xenotime contains Y_2O_3 (46.4 wt%), P_2O_5 (35.35 wt%), Gd_2O_3 (2.61 wt%), Dy_2O_3 (5.78 wt%), Er_2O_3 (3.69 wt%) and Yb_2O_3 (2.81 wt%), with some impurities of Al_2O_3 and Fe_2O_3 (figure 10c). Monazite-(Nd) contains P_2O_5 (17.97 wt%), Nd_2O_3 (24.28 wt%), Ce_2O_3 (13.96 wt%), La_2O_3 (8.71 wt%), Pr_2O_3 (3.43 wt%) and SiO_2 (5 wt%) as well as a high proportion of ThO_2 (22.57 wt%) and UO_2 (2.48 wt%; figure 10d). This mineral also has Al_2O_3 (0.45 wt%) with high

Figure 6. (Continued.)

concentrations of Fe_2O_3 (12.49 wt%) and SiO_2 (11.67 wt%), and these oxides may be a result of the groundmass. The thorite contains UO_2 (7.65 wt%), ThO_2 (38.36 wt%), SiO_2 (14.85 wt%) and Fe_2O_3 (11.0 wt%; figure 10e), and additionally, P_2O_5 (15.27 wt%) and Y_2O_4 (4.86 wt%), reflecting xenotime inclusions in the mineral, with some impurities of CaO , Al_2O_3 and TiO_2 . Finally, gummite contains UO_2 (74 wt%) and other oxides such as SiO_2 (15.7 wt%), Al_2O_3 (4.4 wt%) and CaO (4 wt%; figure 10f).

4.3 Whole-rock chemistry of pegmatites

4.3.1 Geochemical aspects of the pegmatite intermediate zones

The rock-forming minerals of the pegmatite intermediate zones are quartz, magnetite and feldspars with a small amount of rare arfvedsonite. Zircon, fergusonite-(Y), allanite-(Ce) monazite, xenotime, columbite, thorite and gummite occur as trace/accessory minerals in some of the samples. The alteration product is represented by haematite and chlorite. The pegmatite intermediate zones contain a moderately high concentration of SiO_2

(62.2–84.0 wt%) and Fe_2O_3 (total iron as Fe_2O_3 , 4.7–29.0 wt%), where the iron concentration of the samples increased with the decreasing concentration of SiO_2 (table 1), and low concentrations of Al_2O_3 , CaO , Na_2O and K_2O and the presence of P_2O_5 with limited concentrations or below detection limits were noticed. The concentration of Ga (up to 90 ppm) may be because Ga^{3+} was substituted for Fe^{3+} in the tetrahedral cation sites in magnetite and titaniferous magnetite (Vincent and Nightingale 1974). The pegmatites have high concentrations of Zr up to 28,984 ppm (about 3 wt%) with the lowest value of 579 ppm, Nb up to approximately 1 wt% with the lowest value of 915 ppm, Th up to 882 ppm with the lowest value of 258 ppm, U up to 566 ppm with the lowest value of 87 ppm, REE up to 19,206 ppm with the lowest value of 2453 ppm and the HREE are more predominantly over the LREE in most samples, while the Ta ranges from 26 to 152 ppm. These concentrations of all studied samples indicate that they are classified as rare-element pegmatites, NYF family (Černý 1991), where Zr, REE, Nb, Ti, Th, U > Ta, F.

There is a strong correlation between U and Th in the studied pegmatites (figure 11a). These pegmatites have Th/U ratios of 0.81–2.58, which

Table 2. Microprobe analyses of fergusonite crystals (wt%).

Sample	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
SiO ₂	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.1	0.75	0.0
TiO ₂	0.333	0.376	0.322	0.364	0.338	0.408	0.415	0.401	0.445	0.489	0.424	0.431	0.412	0.354	0.396
Fe ₂ O ₃ t	0.0	0.0	0.64	0.0	0.03	0.033	0.0	0.445	1.346	0.0	0.0	0.468	0.571	0.573	0.0
CaO	0.311	0.307	0.304	0.318	0.328	0.290	0.353	0.302	0.354	0.305	0.381	0.385	0.319	0.428	0.340
PbO	0.105	0.123	0.074	0.069	0.07	0.095	0.104	0.057	0.076	0.086	0.106	0.139	0.146	0.106	0.087
Nb ₂ O ₃	49.79	49.54	49.06	49.16	50.03	48.98	49.30	48.6	47.13	48.79	49.81	48.47	48.73	46.72	49.19
Ta ₂ O ₃	0.494	0.517	0.565	0.715	0.593	0.708	0.632	0.620	0.638	0.504	0.558	0.652	0.396	0.311	0.589
ThO ₂	0.861	0.893	0.992	0.878	0.812	0.956	1.336	1.236	2.014	1.252	1.422	0.806	0.871	0.890	0.836
UO ₂	2.089	2.012	1.936	1.962	2.023	1.926	1.953	1.617	2.172	2.057	2.189	2.222	2.174	2.035	1.773
La ₂ O ₃	0.025	0.018	0.007	0.001	0.019	0.021	0.021	0.0	0.048	0.0	0.0	0.028	0.025	0.089	0.034
Ce ₂ O ₃	0.063	0.090	0.071	0.089	0.094	0.12	0.143	0.113	0.36	0.096	0.082	0.116	0.108	0.413	0.143
Pr ₂ O ₃	0.025	0.01	0.0	0.041	0.052	0.072	0.026	0.060	0.152	0.004	0.054	0.047	0.076	0.108	0.007
Nd ₂ O ₃	0.684	0.720	0.811	0.800	0.760	0.708	0.890	0.777	1.567	0.684	0.747	0.699	0.82	0.947	0.789
Sm ₂ O ₃	1.033	1.046	1.13	1.003	1.091	1.047	1.12	1.169	1.578	0.946	1.108	0.934	1.067	1.078	1.028
Eu ₂ O ₃	0.0	0.09	0	0	0	0.057	0.0	0.061	0.0	0.0	0.0	0.0	0.046	0.0	0.0
Gd ₂ O ₃	2.765	2.783	3.013	2.753	2.795	3.063	2.063	2.839	2.839	3.04	2.789	2.71	2.666	2.836	2.765
Tb ₂ O ₃	0.599	0.676	0.646	0.578	0.666	0.615	0.616	0.590	0.553	0.531	0.762	0.625	0.609	0.663	0.656
Dy ₂ O ₃	5.766	5.595	5.692	5.697	5.688	5.612	5.314	5.228	5.248	5.40	5.56	5.552	5.386	5.434	5.68
Ho ₂ O ₃	1.173	1.141	1.056	1.139	1.217	1.101	1.071	1.131	0.945	1.136	1.173	1.172	1.065	1.101	1.131
Er ₂ O ₃	3.595	3.501	3.628	3.553	3.602	3.508	3.284	3.055	3.434	3.584	3.205	3.506	3.377	3.302	3.187
Yb ₂ O ₃	1.967	1.936	1.96	2.03	2.061	1.974	2.097	2.171	2.034	2.067	1.998	2.077	2.033	1.93	2.073
Y ₂ O ₃	28.74	28.81	28.39	28.22	28.84	28.67	28.58	28.19	25.33	28.09	27.33	28.29	27.64	27.29	28.73
Total	100.418	100.148	100.297	99.37	101.109	99.318	98.662	98.263	99.061	99.061	99.329	99.698	98.637	97.357	99.434
ΣLREO	1.83	1.974	2.019	1.934	2.016	2.025	2.2	2.18	3.705	1.73	1.991	1.824	2.142	2.635	2.001
ΣHREO	44.605	44.442	44.385	43.025	44.864	44.543	43.025	43.204	40.384	43.848	42.817	43.932	42.776	42.556	44.222
ΣREO	46.435	46.416	46.404	44.959	46.88	46.568	45.225	45.384	44.089	45.578	44.808	45.756	44.918	45.191	46.223
ΣLREO/ΣHREO	0.041	0.044	0.045	0.045	0.045	0.045	0.051	0.05	0.092	0.04	0.046	0.042	0.05	0.062	0.045
ThO ₂ /UO ₂	0.412	0.444	0.512	0.448	0.401	0.496	0.684	0.763	0.927	0.609	0.651	0.363	0.401	0.437	0.472

Fe₂O₃t = total iron, HREOs include Y₂O₃.

Figure 7. Chondrite-normalised REE profiles of the fergusonite crystals (Boynton 1984).

Table 3. Microprobe analyses of zircon crystals (wt%).

Sample	1 rim	1 core	2 rim	2 core	3rim	3 core	4 core
SiO ₂	32.97	32.93	32.96	33.00	32.81	32.49	32.84
TiO ₂	0.011	0.023	0.003	0.0	0.031	0.0	0.0
Al ₂ O ₃	0.002	0.0	0.0	0.055	0.0	0.0	0.0
Fe ₂ O _{3t}	0.0	0.0	0.005	0.003	0.0	0.0	0.0
CaO	0.003	0.0	0.0	0.0	0.003	0.012	0.0
P ₂ O ₅	0.312	0.395	0.337	0.740	0.293	0.892	0.377
ZrO ₂	64.07	64.19	63.54	64.17	64.34	63.80	63.58
HfO ₂	2.216	2.282	2.357	2.294	2.379	2.316	2.359
Nb ₂ O ₃	0.111	0.131	0.182	0.123	0.109	0.092	0.106
ThO ₂	0.0	0.032	0.017	0.049	0.036	0.008	0.084
UO ₂	0.131	0.178	0.179	0.216	0.150	0.227	0.263
Ce ₂ O ₃	0.026	0.021	0.0	0.0	0.037	0.022	0.01
Nd ₂ O ₃	0.0	0.005	0.0	0.026	0.069	0.008	0.026
Sm ₂ O ₃	0.0	0.02	0.029	0.0	0.0	0.052	0.0
Eu ₂ O ₃	0.0	0.0	0.0	0.0	0.0	0.011	0.0
Gd ₂ O ₃	0.0	0.0	0.0	0.135	0.089	0.086	0.093
Tb ₂ O ₃	0.047	0.139	0.149	0.149	0.102	0.083	0.194
Dy ₂ O ₃	0.101	0.046	0.060	0.093	0.024	0.044	0.029
Ho ₂ O ₃	0.037	0.014	0.015	0.0	0.0	0.027	0.0
Er ₂ O ₃	0.062	0.221	0.171	0.183	0.122	0.011	0.131
Yb ₂ O ₃	0.040	0.085	0.122	0.101	0.097	0.150	0.182
Y ₂ O ₃	0.116	0.184	0.196	0.149	0.113	0.222	0.580
Total	100.255	100.896	100.332	101.486	100.804	100.553	100.854
ΣLREO	0.026	0.046	0.029	0.026	0.106	0.093	0.036
ΣHREO	0.403	0.689	0.713	0.81	0.547	0.623	1.209
ΣREO	0.429	0.735	0.742	0.836	0.653	0.716	1.245
ΣLREO/ΣHREO	0.065	0.067	0.041	0.032	0.194	0.194	0.03
ThO ₂ /UO ₂	0	0.18	0.09	0.23	0.24	0.03	0.32

Fe₂O_{3t} = total iron, HREOs include Y₂O₃.

is lower than that of the upper part of the Earth's crust (3.8: Taylor and McClellan 1985), except sample 8 (Th/U ratio = 6.61). These ratios indicate that these rocks are enriched in U relative to Th, except sample 8, and they might be enriched

in the thorium minerals than uranium minerals or were subjected to the uranium migration as a result of the reactivation of the faults. The decrease in the Th/U ratios with the increase in SiO₂ (figure 11b) may be due to the uranium enrichment relative to

Figure 8. Chondrite-normalised REE profiles of the zircon crystals (Boynton 1984).

Table 4. Microprobe analyses of allanite crystals (wt%).

Sample	1	2	3	4	5	6	7	8
SiO ₂	33.27	31.54	31.64	34.09	32.22	31.5	32.08	31.86
TiO ₂	0.425	0.380	0.398	0.484	0.367	0.341	0.419	0.422
Al ₂ O ₃	18.06	16.57	15.27	17.61	17.17	16.23	16.87	16.64
Fe ₂ O ₃ t	10.22	12.21	16.05	11.36	11.99	13.22	12.01	12.01
CaO	7.517	8.265	9.458	7.882	7.865	7.695	8.061	8.249
MgO	0.283	0.570	0.966	0.347	0.688	0.702	0.590	0.603
Na ₂ O	0.204	0.334	0.046	0.179	0.378	0.257	0.410	0.358
K ₂ O	0.024	0.014	0.014	0.035	0.02	0.022	0.021	0.019
P ₂ O ₅	0.025	0.013	0.031	0.029	0.020	0.009	0.020	0.015
MnO	1.281	1.103	1.172	1.291	1.002	1.289	1.119	1.079
SrO	0.096	0.103	0.003	0.108	0.144	0.027	0.144	0.111
PbO	0.075	0.045	0.013	0.083	0.008	0.023	0.035	0.049
ThO ₂	0.656	0.762	0.530	0.597	0.697	0.737	0.769	0.711
UO ₂	0.0	0.0	0.0	0.003	0.0	0.018	0.0	0.02
F	0.149	0.173	0.155	0.131	0.170	0.073	0.080	0.150
Cl	0.326	0.027	0.023	0.156	0.024	0.047	0.042	0.024
La ₂ O ₃	3.404	3.278	3.760	3.472	3.425	3.489	3.432	3.129
Ce ₂ O ₃	10.74	10.32	10.90	10.28	10.52	10.78	10.70	10.28
Pr ₂ O ₃	1.317	1.263	1.412	1.314	1.308	1.444	1.347	1.347
Nd ₂ O ₃	5.359	5.601	5.40	5.201	5.689	5.066	5.552	5.594
Sm ₂ O ₃	0.887	1.051	0.909	0.780	1.029	1.222	1.056	1.156
Gd ₂ O ₃	0.515	0.635	0.531	0.487	0.613	0.728	0.622	0.692
Dy ₂ O ₃	0.099	0.159	0.199	0.122	0.146	0.169	0.199	0.229
Y ₂ O ₃	0.911	1.038	0.891	0.850	0.932	0.921	0.977	1.048
Total	95.843	95.454	99.771	96.891	96.425	96.009	96.555	95.795
∑LREO	21.707	21.513	22.381	21.047	21.971	22.001	22.087	21.506
∑HREO	1.525	1.832	1.621	1.459	1.691	1.818	1.798	1.969
∑REO	23.232	23.345	24.002	22.506	23.662	23.819	23.885	23.475
∑LREO/∑HREO	14.234	11.743	13.807	14.426	12.993	12.102	12.284	10.922
ThO ₂ /UO ₂	0	0	0	199	0	40.94	0	35.55

Fe₂O₃t = total iron, HREOs include Y₂O₃.

Figure 9. Chondrite-normalized REE profiles of the allanite crystals (Boynton 1984).

thorium because U entered into an aqueous phase as uranyl ions during the final stages of crystallisation of granitic magmas (pegmatite). Furthermore, the strong correlation between ΣREE and both Th and U (figure 11c and d) occurs as a result of the strong link between Th and U in monazite-(Ce and Nd), xenotime and fergusonite minerals which have considerable concentrations of Th and U. The correlation and variations of Nb/Ta and Zr/Hf indicate a moderate to high degree of fractionation in the NYF-type pegmatite melts of Gabal El Faliq (figure 12a–c). The Nb/Ta ratios in the studied pegmatites range from 10.76 to 81.67 with an average of 45.55. Nb *vs.* the Nb/Ta ratios show a moderate positive relation (figure 12b). The Zr/Hf ratios range from 14.9 to 25.96 with an average of 22.39; there is variability in Hf, with relatively constant Zr/Hf ratios (figure 12c).

The chondrite-normalised REE patterns of the pegmatite intermediate zones (figure 13) show relatively similar flat patterns with negative europium anomalies ($\text{Eu}/\text{Eu}^* = 0.05\text{--}0.12$). Samples with a lower total REE show a lower content of LREE. Overall, REE patterns are characterised by HREE enrichment relative to LREE.

4.3.2 Tectonic setting and petrogenesis

The high Fe/Mg, Zr, Nb, Ga, REE, Y and F contents are features of A-type granite geochemistry (e.g., Collins *et al.* 1982; Whalen *et al.* 1987; Eby 1990). The NYF pegmatites are characterised by Nb > Ta (Černý and Ercit 2005). The ternary diagrams of Y–Nb–3Ga and the Y–Nb–Ce (Eby 1992) that discriminate between different types of

granites (figure 14a and b) indicate that the NYF pegmatites fall completely within the field A₁ of mantle-derived magma. The magma was emplaced along the Wadi El Gemal strike-slip fault within a continental plate (figure 14c) and was subjected to prolonged fractionation to form these pegmatite dykes (figure 15).

The petrogenesis of the Gabal El Faliq pegmatites may be reflected in the geochemistry of their trace elements. Melts derived from the mantle with little crustal contamination are reflected by low Cs (<5 ppm) and high Rb/Cs (>40; Scarrow *et al.* 2009). The studied Gabal El Faliq pegmatites have low Cs (<0.1–0.3 ppm) and high Rb/Cs. These ratios suggest a mantle-derived origin. Furthermore, the enrichment in REE with predominant HREE and the absence of Nb depletion in these rocks reflect an intracontinental tectonic setting similar to the Kola Peninsula, Russia (Kogarko *et al.* 2002). The depletion in Ba, Sr, P and Ti (figure 16) indicates a strong mantle-derived source for these rocks similar to that of the alkaline volcanic rocks derived from the alkali basalt magmas of Lake Chad (Mbowou *et al.* 2012). These results are also consistent with the opinion of Martin and De Vito (2005), who suggested that the NYF-type deposits occur in anorogenic tectonic settings and are ultimately mantle-derived (London 2008). Low Y/Nb ratios (<1.2) provide evidence for their origin from mantle-derived magma similar to the oceanic island basalts (OIBs) as suggested by Eby (1990; figure 15). This diagram shows the studied pegmatite sample plots within the OIB field and the differentiation trend of AF emanating from this field.

Figure 10. ESEM/EDX pattern of (a) magnetite, (b) titanomagnetite, (c) xenotime, (d) monazite-(Nd), (e) thorite and (f) gummite.

5. Discussion and interpretations

The Gabal El Faliq leucogranites host dykes and lenses of rare metal-bearing pegmatites, located in the Southeastern Desert of Egypt in the southwestern segment of the Wadi Hafafit–Wadi Ghadir fold belt. The Wadi Hafafit–Wadi Ghadir fold belt, formed during the Pan-African orogeny, consists of two main groups that were separated by a major thrust called Nugrus (figure 2a). In the

late Pan-African orogeny, many leucogranites were formed in the Wadi Hafafit–Wadi Ghadir fold belt along the Nugrus thrust zone and its branches (e.g., Wadi El Gemal thrust) as the Gabal El Faliq leucogranites, which were formed by the partial melting of the thickened crust (Abd El-Naby and Frisch 2006). This event was accompanied by the reactivation of the old thrust zones as strike-slip faults or shear zones (Moghazi *et al.* 2004). Accordingly, the major Wadi El Gemal strike-slip

Figure 11. (a) U vs. Th binary variation diagram, (b) SiO₂% vs. Th/U binary variation diagram, (c) U vs. total REEs binary variation diagram and (d) Th vs. total REEs binary variation diagram for the studied pegmatite intermediate zones.

motion would produce a mantle-derived magma. This magma is easily contaminated by the country rocks due to its relatively *prolonged* crystallisation time and/or shear heating. As a result, the rare metal-bearing pegmatites were formed from the residual melts of the prolonged fractionation of a previously contaminated magma. These rare metal-bearing pegmatites exist as dykes that were injected into the fractures of the Gabal El Faliq area.

The mineral chemistry indicated that fergusonite crystals (table 2) are composed of Nb₂O₅ (46.72–50.03 wt%), Y₂O₃ (27.29–28.84 wt%), with total HREO, including Y₂O₃ (40.38–44.86 wt%), and LREO (1.73–3.71 wt%), as well as a significant concentration of UO₂ (1.62–2.19 wt%) and ThO₂ (0.81–2.01 wt%). The chondrite-normalised REE patterns of fergusonite crystals show that the HREE values are very high relative to the LREE values with depleted or pronounced negative Eu-anomalies. On the other hand, some samples show positive Ce anomalies and negative Pr anomalies

at the same time which is due to the crystals that were subjected to the hydrothermal metasomatism which led to variable Ce anomalies (Bagiński *et al.* 2016). Accordingly, a positive Ce anomaly causes an apparent negative Pr anomaly. Variable Ce anomalies also can be attributed to an oxygenated environment, where redox conditions changed during the formation of pegmatite. This fact is due to the presence of arfvedsonite which indicates reducing conditions and high water pressure, but the presence of haematite indicates oxidising conditions. Accordingly, the mineral assemblage of pegmatites took place in two stages: the first stage was reducing conditions and it was appropriated to form arfvedsonite, but the second stage was oxidising conditions where the replacement of magnetite and mafic minerals by haematite took place and extensive quartz was formed. Zircon crystals are composed of ZrO₂ (63.54–64.34 wt%) and SiO₂ (32.49–33 wt%) with a significant amount of HfO₂ (2.22–2.29 wt%), an average of 2.31 wt%. This average is closer to the lower value of HfO₂ (2.4 wt%)

Figure 12. (a) Zr/Hf vs. Nb/ Ta binary variation diagram, (b) Nb vs. Nb/Ta binary variation diagram and (c) Hf vs. Hf/Zr binary variation diagram.

that was recorded in zircon from the Beauvoir pegmatite (France) by Wang *et al.* (1992). Allanite crystals are composed of SiO₂ (31.50–34.09 wt%), REO (22.51–24.00 wt%), mostly LREO (21.047–22.38 wt%), Al₂O₃ (15.27–18.06 wt%), Fe₂O₃t (10.22–16.05 wt%) and CaO (7.51–9.46 wt%). Additionally, ThO₂ exists in these crystals whereas UO₂ either exists or does not exist (table 4).

The fractional crystallisation processes and volatiles were responsible for the high concentrations of Zr, Hf, REE, U and Th in the pegmatite intermediate zones, where Zr was up to 3 wt%, Hf up to 0.12 wt%, Nb up to 1 wt%, Ta up to 0.015 wt%, REEs up to 2 wt%, U up to 0.06 wt%, Th up to 0.09 wt% and Ga was up to 0.009 wt%, so they were classified as the NYF pegmatites. The REE patterns of fergusonite (figure 7), allanite (figure 8), zircon (figure 9) and bulk samples of pegmatite intermediate zones (figure 13) show depleted, weak negative Eu anomalies and enrichment of REE. This indicated that the nature of residual pegmatite magma is characterised by (i) extreme enrichment in volatiles with very low viscosity (Baker and Vaillancourt 1995), despite its low temperature, (ii) strong depletion in feldspar-compatible elements and (iii) higher abundances of HFSE cations. Redistribution of elements took place by the hydrothermal solutions (Bau and Dulski 1999).

On the other hand, the internally derived mineralising fluids during the fractionation crystallisation processes have modified ratios, such as Nb/Ta and Zr/Hf (figure 12a). These ratios have been used to show fractionation in other rare metal granitic pegmatites (e.g., Linnen 1998).

Figure 13. Chondrite-normalised REE profiles of the pegmatite intermediate zones (after Boynton 1984).

Figure 14. (a) Ternary plot Y–Nb–3Ga for the pegmatite intermediate zones (Eby 1992). (b) Ternary plot Y–Nb–Ce for the pegmatite intermediate zones (Eby 1992). (c) Ternary plot HF–Rb/30–3Ta for the pegmatite intermediate zones (Harris *et al.* 1986). Symbols are as in figure 11.

The intermediate zones of Gabal El Faliq pegmatites show a distinct fractionation trend when comparing Nb/Ta *vs.* Nb (figure 12b). This diagram shows that the Nb/Ta ratio increases with increasing Nb. This result can be attributed to the low concentration of hydrofluoric acid (HF) at a low pH, where Nb is more soluble than Ta in acidic hydrothermal fluids that have low concentrations of HF (Timofeev *et al.* 2017). This is consistent with the present study, where fluorine is found only in allanite, and may also be due to the occurrence of phosphorus, which only increases columbite solubility (Wolf and London 1993) in the residual granitic melt that formed pegmatite. The retention of Nb in acidic hydrothermal fluids may be responsible for the increase of Nb/Ta

Figure 15. Y/Nb *vs.* Ba/La binary variation diagram (Eby 1990). Symbols are as in figure 11; OIB, IAB = island arc basalts, CPX and AF. Fractionation trends calculated using partition coefficients of Henderson (1982).

ratios in residual pegmatite melts. Accordingly, the Nb/Ta ratios for the studied intermediate zones of pegmatites (table 1) are more than the chondritic values (17 ± 1 , Weyer *et al.* 2002). The Zr/Hf ratios are relatively constant for the studied pegmatites (figure 12c) with an increasing Hf, and the fact could be explained by the lower crystallisation temperature, which increases the diffusion coefficient of hydrogen in hafnium (DHf) between zircon and pegmatite magma (Wang *et al.* 2010). Furthermore, their Zr/Hf ratios range from 14.1 to 25.96 with an average of 22.39. This average (about 25) is mostly close to the average Zr/Hf in 107 whole-rock analyses of pegmatites (Erlank *et al.* 1978), but it is less than the chondritic values (36.6 ± 2.9 ; Bau 1996) for unfractionated igneous rocks.

As for the origin, this study mentioned many evidences that supported the origin of the Gabal El Faliq NYF-type granitic pegmatite from prolonged fractionation basaltic magma (similar to the OIB magmas), coming from the mantle as a result of the reactivation of the Wadi El Gemal thrust as a strike-slip fault that was accompanied by the regional strike-slip faults and shear zones that were formed in the Hafafit–Ghadir fold belt after a late Pan-African orogeny (Moghazi *et al.* 2004). These suggest that pegmatites are emplaced into the non-orogenic setting (figure 14c), enriched in REE, Nb, Zr, Hf, Th, U and Ga and depletion of Ba, Sr, P and Ti (figure 16). The low Y/Nb ratios (<1.2 ; figure 15) that indicate the most samples of parental melt of Gabal El Faliq

Figure 16. Spider diagrams of the studied pegmatite intermediate zones (McDonough and Sun 1995).

rare metal-bearing pegmatites are located under the OIBs field. This phenomenon shows that the pegmatite melt underwent high AF fractionation in the presence of volatiles (Eby 1990, 1992). Additionally, the Y/Nb ratio is virtually unchanged over the wide range of silica values for any particular suite (Eby 1990), where clinopyroxenes (CPXs) and amphiboles of the common silicate minerals would be significant phases during fractional crystallisation. CPXs or/and amphiboles fractionation will significantly change the Y/Nb ratio but co-crystallisation of a Ti-rich phase would lower Nb relative to Y (Green and Pearson 1986), thus, the Y/Nb ratio remains relatively constant during the fractionation processes.

The enrichment of rare elements is usually associated with higher concentrations of fluxing components such as F, P and H₂O (Linnen 1998) such as the studied rocks. These fluxing elements played an important role in pegmatitic magma evolution by causing significant changes in the residual melt structure that affects melt properties such as viscosity and diffusivity (London 2008). The confirmation of fluxing element effects is derived from the recent experimental studies by Migdisov and Williams-Jones (2002, 2007) and Migdisov *et al.* (2009), which indicated that Zr and REE form complexes with fluoride and the fluids with HF concentration at 0.1 m that can dissolve 10 ppm of zircon and higher concentrations of REE at 200°C. Alternatively, only phosphorus increases columbite solubility (Wolf and London 1993). Accordingly, the high concentration mechanism for the saturation of the Zr-, Nb-, Th-, U- and REE-bearing

pegmatites at Gabal El Faliq is because phosphorus and fluorine reduce solidus temperatures, decrease viscosity and increase the solubility of H₂O in the melt (Dingwell *et al.* 1996; London 1997, 2009). Consequently, the fluorine could have led to rare elements transport in hydrothermal fluids mainly as fluoride complexes, and the deposition of these element-bearing minerals at Gabal El Faliq from the fluoride complexes could have taken place as a result of the interaction of such Ca-poor, F-fluids with Ca-bearing rocks (Salvi and Williams-Jones 2004).

In general, these rocks can be compared with the alkaline rocks in Egypt, in their alkaline parent magma, where El-Ramly and Hussein (1982) suggested that these rocks were formed under certain favourable geotectonic conditions (e.g., intracontinental hot spots) and a partial fusion of deeper levels in the upper mantle could produce enough heat and volatiles (Lameyre *et al.* 1984; Black *et al.* 1985).

6. Conclusions

- The Gabal El Faliq rare metal-bearing pegmatites were formed as a result of mantle-derived magma contaminated with the lower crust.
- The contaminated magma underwent prolonged crystallisation during its ascent in the Wadi El Gemal strike-slip fault and its Riedel shears and in so doing achieved a felsic composition. Early pulses of magma are thought

not to have contained any significant amounts of HFSEs (Zr, Hf, REE, Nb, Ta, U, Th and Ga) which led to the increase in their concentrations in the residual melt.

- (c) The pegmatite bodies are mineralogically complex types and consist mainly of quartz and magnetite, feldspars, haematite and rare arfvedsonite. They are zoned with a high abundance of rare metal-bearing minerals in their intermediate zones. These include fergusonite-(Y), zircon and allanite-(Ce), in addition to monazite-(Ce and Nd), xenotime-(Y) and columbite-(Fe), thorite and gummite.
- (d) The high concentrations of Zr-, Nb-, Th-, U- and REE-bearing pegmatites at the Gabal El Faliq are likely to be due to the presence of phosphorus and fluorine, which reduce solidus temperatures, decrease viscosity and increase the solubility of H₂O in the melt.
- (e) The binary diagrams of the geochemical data of the studied pegmatites support a fractionation model of the Gabal El Faliq NYF-type pegmatites. This fractionation was enhanced by the presence of phosphorus and fluorine. Moreover, this study documented the importance of the prolonged fractionation magma that was produced as a result of the motion of the regional strike-slip fault in the origin of the NYF-type pegmatites in the within-plate setting.

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