U-Pb geochronology, REE and trace element geochemistry of zircon from El Fereyid monzogranite, south Eastern Desert, Egypt

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The monzogranite of El Fereyid is one of the rare metal-rich granites, where zircon is one of the ore minerals. Thus, studying zircon here is of vital importance. Most of the studied zircon grains are metamict and thus the loss of radiogenic Pb is detected for them. Nevertheless, our study allowed us to obtain the U-Pb (SHRIMP-II) age of magmatic crystallization for the El Fereyid monzogranite: 626 ± 13 Ma. These data allowed the correction of earlier determined age (K-Ar system in biotite) for the El Fereyid massif.

Zircon grains of El Fereyid monzogranite demonstrate heterogeneous structure in CL images, they are rich in rare earth elements (REE, concentration 3000–22500 ppm), trace elements (U: 2000–14800 ppm, Th: 300–2500 ppm), and has low Th/U ratios (average 0.18). Most zircon grains exhibit multiple internal oscillatory zoning in CL images, indicating a typical magmatic origin. The core of zircon grain has a dark tone on CL imaging and magmatic-type REE spectra. Zircon cores are enriched in REEs, U, Th, and Y, thus recording that the magma was rich in incompatible elements. Light CL rims of zircon grains are depleted, relative to the cores in most of the trace elements, except for P, Ca, and Ti. REE spectra in rims show similar patterns that are more often demonstrated by hydrothermal zircon grains. This explains their crystallization in the late-magmatic stage when magma was rich in fluids but depleted in most of the trace elements.

Keywords: El Fereyid monzogranite, Zircon, U-Pb method, Magmatic zircon, Hydrothermal zircon, Trace elements, Rare-earth elements

INTRODUCTION

The Arabian Nubian Shield (ANS) forms the northern segment of the East African Orogen (EAO) and covers mainly Egypt, Sudan, Eritrea, Ethiopia, Somalia, Saudi Arabia, Yemen, and Oman, whereas the Mozambique belt occupies the southern segment of EAO. The ANS is made up of igneous, sedimentary, and metamorphic rocks, and its formation is considered as one of the greatest Neoproterozoic episodes of crustal growth in the Earth’s history (Johnson and Woldehaimanot, 2003). The Eastern Desert (ED) of Egypt is a part of the northern ANS (Figs. 1a and 1b). It consists of abundant granitic plutons containing rare metal and gold mineralizations. These late- to post-orogenic calc-alkaline/peralkaline granites were dated at 630–590 Ma (U-Pb, zircon) (Meert, 2003; Küster, 2009; Moghazi et al., 2011; Melecher et al., 2015). Rare metal mineralization occurs as small stocks or micro-veinlets commonly associated with younger granites and pegmatites (e.g., Helba et al., 1997; Sami et al., 2020).

In the Eastern Desert of Egypt, rare metal granites (RMG) fall into several categories: (1) metaluminous, Nb + Zr + Y-enriched alkali granites; (2) peraluminous, Ta > Nb + Sn ± W ± Be-enriched Li-albite granites; and (3) metasomatized, Nb >> Ta + Sn + Zr + Y + U ± Be ± W-enriched granites (Abdalla et al., 1998). Hassan and Hashad (1990) reported that the RMG intruded between
620 and 530 Ma during post-orogenic magmatism in Egypt. However, Nb, Ta, Y, U, and Th mineralizations are mainly limited in younger granites and associated pegmatite pockets that are widely distributed in the ED and southern Sinai, Egypt (El–Kammar et al., 1997; Dardier, 1999; Abdalla et al., 2008; Ali et al., 2011; Raslan et al., 2017; Saleh et al., 2018; Abu Elatta 2019; Fawzy et al., 2020).

El Fereyid area, which is the subject of this study, is located in the southern part of the ED of Egypt. Geochemically, El Fereyid monzogranite belongs to the second type of peraluminous granite as mentioned above, according to Abdalla et al. (1998). El Fereyid area is well studied from structural, gamma-ray spectrometry, geochemical and mineralogical points of view (Soliman et al., 1985a, 1985b; El Amawy, 1991; El-Baraga, 1992; Abdel Karim and Sos, 2000; Abd El-Naby and Saleh, 2003; Dawoud et al., 2018; Saleh et al., 2018). However, the age and geochemistry of zircon grains from El Fereyid remain insufficiently studied. It is well known that zircon contains trace lithophile and high field strength elements, especially Y, Ti, Hf, Th, U, Nb, as well as REE, as shown by many authors (Hinton and Upton, 1991; Omar et al., 2011; El–Bialy and Ali, 2013; Levashova et al., 2016; Skublov and Li, 2016; Wu et al., 2020). According to Geisler et al. (2003), zircon from 619 ± 17 Ma Gabel Hamradom granite massif located in the vicinity of El Fereyid massif (Fig. 1b) revealed primary oscillatory growth zoning with no further internal complexity on CL imaging. They are anomalously rich in U and Th, reaching concentrations up to 30000 and 8000 ppm (SIMS), respectively, metamictized and had lost significant amounts of radiogenic Pb. The chemical enrichment of Ca, Sr, Ba, Al, Fe, and Mn, also REE (averages LREE content is 1500 ppm, HREE – 2500 ppm) in the dark–CL bands of Hamradom zircon was resulting in an unusually flat REE pattern without any Ce-anomaly typical for natural zircon (Geisler et al., 2003).

The main objective of the present study is to a) define the crystallization age for El Fereyid monzogranites by U–Pb isotopic system in zircons, b) specify the crystallization parameters for the zircons, as one of the ore-bearing minerals in El Fereyid.

Figure 1. (a) Map of North East Africa showing the Arabian Nubian Shield (ANS), the Saharan Metacraton, and Archaean and Palaeoproterozoic crust that was remobilized during the Neoproterozoic. (b) Geological map showing the distribution of the Neoproterozoic basement rocks from the Eastern Desert (ED) of Egypt (Ali et al., 2015; Sami et al., 2017). (c) Geological map of El Fereyid, South ED, Egypt (Saleh et al., 2018). Color version is available online from https://doi.org/10.2465/jmps.210320.
GEOLOGICAL SETTING

The basement rocks in the ED of Egypt are outcropping along flanks of the Red Sea and the border with Sudan. Most of the ED basement rocks are Neoproterozoic, 900–550 Ma, in age. The Nd isotope data indicate that the basement rocks are composed of juvenile Neoproterozoic crust sandwiched between reworked older crust (Stern, 2002). The ED is subdivided into three lithologically and structurally distinct terranes (Figs. 1a and 1b): the North, the Central, and the South ED (SED) (Stern and Hedge, 1985). The SED terrane, a part of the ANS, consists of Neoproterozoic fragmented ophiolites, island arc metavolcanics with less abundant gneisses, and metasedimentary schists. All these units are intruded by syn– to late–orogenic granites. The next stage of the SED terrain formation is marked by the mafic and intermediate volcanic eruptions (Dokhan volcanic suites). Volcanic complexes are partly overlain by molasses-type sediments (Abd El-Wahed and Kamh, 2010; Bezenjania et al., 2014). After the craton stabilization numerous bodies of granites, enriched in Nb, Sn, W, Zr, Y, Mo, Ta, Cu, Au, and REE were intruded into the ANS rocks in the ED (Abdalla et al., 2008; Moghazi et al., 2011; Raslan et al., 2017; Zoheir et al., 2019; Sami et al., 2020).

The tectonic terrain of the SED was mainly shaped during the final stage of East– and West-Gondwana formations (Abdelsalam and Stern, 1996; Abdelsalam et al., 2003; Johnson et al., 2011). Five structural episodes affected the SED (Abdelsalam and Stern, 1996; Abdelsalam et al., 2003; Zoheir and Klemm, 2007; Abdeen and Abdelghaffar, 2011; Johnson et al., 2011; Ali, 2013; Ali et al., 2018; Hamimi et al., 2019; Abd El-Wahed and Hamimi, 2020). (1) Folding–thrusting episode in which syntectonic (tonalite–granodiorite–quartz diorite) granites were intruded. It was associated with the cratonization of the arc/inter–arc rock association. Allaqi–Heiani and Sol Hamed–Onib sutures have NW–WNW trends and were formed due to the collision event between Gabgaba–Gebeit and Gerf terrains at >715 Ma (Stern et al., 1989). (2) Upright folding episode in which late–tectonic granites were intruded. It was associated with the compression and shortening of the terrain to the NE–SW direction. The Hamisana shear zone has N–S trending upright folds that were dated between 580 and 660 Ma (Stern et al., 1989). (3) Post–tectonic granitic intrusion episode, which produced syenogranite, monzogranite, alkali feldspar granite, and microgranite dikes. (4) Syncline folding along ENE–WSW axis (Early Cretaceous to Post Pleistocene episode) which is detected in Fileita Nubian sandstone. (5) Fracturing, faulting, and post–granitic dike extrusion episodes that are expressed by multi–trends and types of fault populations that have been taking place since cratonization and till present.

Field studies have shown that the basement rocks outcropped in El Fereyid are older granites (tonalite–granodiorite), younger granites (monzogranite), and pegmatites cut by mafic and felsite dikes in decreasing mode of occurrences, respectively (Fig. 1c). The tonalite–granodiorite rocks occur as some scattered small hills surrounding the peripheries of monzogranite (Fig. 2). They form low to moderate relief, are grey to whitish–grey, coarse–grained, highly exfoliated, fractured, and exhibit cavernous weathering. These rocks are composed mainly of plagioclase, quartz, K–feldspar, hornblende, and biotite. They were intruded by the monzogranite with sharp intrusive contacts. The monzogranite forms the most abundant granitic plutons in the mapped area. They crop out in high mountains, which are usually elongated along the
NW–SE trend, as shown on the geological map (Fig. 1c). The monzogranite rocks are hosting late magmatic pegmatites occurring as pockets and veins (Abd El-Naby and Saleh, 2003; Raslan et al., 2010a, 2010b; Saleh et al., 2018; Dawoud et al., 2018). Pegmatites are often very coarse-grained and are reddish to buff-colored. They occur in the northern, northwestern, and southern parts of the mapped area (Fig. 1c). The pegmatite veins occur as irregular bodies varying in width from 0.5 to 1 m to over 5 m. They are mainly composed of quartz, K–feldspar, muscovite, and biotite megacrysts. Pegmatites contain higher radioactive element concentrations, especially in U and Th, as compared to other varieties of rocks in the study area due to the presence of radioactive minerals such as thorite, uranothorite, samarskite, ishikawaite and are enriched by accessories such as zircon and monazite (Saleh et al., 2018; Dawoud et al., 2018). The rare earth elements concentration in pegmatite is less than that of monzogranite due to magmatic fractionation and absence of monazite (Dawoud et al., 2018).

The study area is affected by heterogeneous ductile and brittle deformation (Saleh et al., 2018). The WNW–ESE strike-slip fault, affecting the northern part of the El Fereyid area (Fig. 1c). El Amawy (1991) and Saleh et al. (2018) suggested ductile deformation that affected the study area by a prominent set of folds with their axes striking WNW to NW and others extending ENE and NNW. At the intersection of NW–SE and NNW–SSE trends, gold mineralization is observed. Gold is mostly concentrated in hydrothermally altered rocks (chlorite-epidote alteration) in the northern and northwestern parts of El Fereyid monzogranite (Zoheir et al., 2019).

The studied monzogranite rocks are medium– to coarse-grained, reddish grey to pink and are highly fractured with predominant NW–SE trends of joints. They were cut by basic and felsite dikes, as well as quartz veins with NW–SE, and NE–SW trends. Microscopically, they are composed essentially of K–feldspar, plagioclase, quartz, biotite, and hornblende. Myrmekitic and perthitic textures are well noticeable. The monzogranite of El Fereyid was affected by hydrothermal processes including chloritization, sericitization, kaolinization, and hematitization. Hydrothermal alterations led to the crystallization of rock-forming minerals like biotite, hornblende, and feldspar as well as formation of uranopane, pyrite, epidote, and iron oxides (Abd El-Naby and Saleh, 2003; Saleh et al., 2018). Zircon, apatite, titanite, allanite, and monazite are accessories, while chlorite, epidote, and sericite are secondary minerals. Zircon occurs as colorless prismatic crystals with high relief; some crystals are surrounded by dark pleochroic haloes within biotite.

The bulk rock sample of El Fereyid monzogranite is enriched in large ion lithophile elements (Ba, Rb, and Sr), high field strength elements (Y, Zr, Nb, and REE), and depleted in Ca, Mg, K, P, and Ti. Monzogranite can be attributed as calc-alkaline series, has peraluminous characteristics and represents I-type granite (Dawoud et al., 2018). Monzogranite in El Fereid is plotted in the granite compositional field based on the TAS diagram (Cox et al., 1979 and Middlemost, 1994) (Fig. 3). The REE concentration in monzogranite is 150–240 ppm. The concentration of LREE in monzogranite range from 130 to 220 ppm, while HREE range from 15 to 27 ppm. REE spectra demonstrate enrichment of LREE and low content of HREE with moderate negative Eu-anomaly (Eu/Eu* = 0.38–0.67) due to the presence of some accessory minerals like monazite, apatite, zircon (Dawoud et al., 2018).

### ANALYTICAL METHODS

Around 300 zircon grains were extracted from monzogranite in the El Fereyid area. Standard methods, such as the crushing of samples, heavy liquid separation, and magnetic separation were carried out in the Institute of Precambrian Geology and Geochronology Russian Academy of Sciences, St. Petersburg. A binocular microscope was used for picking up zircon grains. The zircon grains were analyzed to determine U-Pb age and REE contents.

The U-Pb age dating of separated zircon grains was conducted using a SHRIMP-II ion microprobe at the Centre of Isotopic Research of the Karpinsky Geological Research Institute, St. Petersburg. Zircon grains were implanted manually into epoxy resin with TEMORA and standard 91500 zircon grains. The grains were polished to approximately half of their thickness. To select surface spots for age dating, we used optical (transmitted and re-

**Figure 3.** The total alkali versus silica diagram (Middlemost, 1994) with plotted data for El Fereyid granites (Dawoud et al., 2018).
Reflected light) and cathodoluminescence (CL) images showing the internal structure and zoning of zircon grains, and the presence of inclusions. CL images were carried out on a Camscan MX2500S SEM equipped with a QLI/QUA2 CL. U-Pb age dating was done using the method of Williams (1998). The intensity beam was 4 nA of the primary O2, and the spot or crater was 20 µm in diameter. The data obtained were processed using the SQUID program according to Ludwig (2001). Differential fractionation between U and Pb was monitored by reference value of 0.0668 for interspersed analyses of the Temora zircon standard (416.8 ± 0.3 Ma; Black et al., 2003), based on the logarithmic law relationship $^{206}$Pb/$^{238}$U versus $^{207}$U/$^{235}$U. The $\sigma$ level was used to show errors in single analyses of U/Pb ratios and ages, whereas the 2$\sigma$ level was used to calculate errors in concordant ages, as well as intercepts with the concordia. The ISOPLOT/EX program was used for concordia plots according to Ludwig (2003). The 1$\sigma$ level was used to show errors in single analyses of U/Pb ratios and ages, whereas the 2$\sigma$ level was used to find errors in calculated concordant ages and weighted mean age calculation, as well as intercepts with the concordia.

Analysis of rare earth and trace elements in the zircon grains was carried out by secondary-ion mass spectrometry (SIMS) on a Cameca IMS-4F ion microprobe (Cameca, Gennevilliers, France) at the Yaroslavl Branch of the Institute of Physics and Technology, Russian Academy of Sciences. Analytical procedures are mainly described in Hinton and Upton (1991) and Fedotova et al. (2008). The primary O$^{2-}$ ion beam was accelerated to about 14.5 keV and focused to a spot size of ~ 20 µm. The ion current was 5–8 nA. Each analysis was carried from 3 cycles of measurements with a discrete transition between mass peaks within the given set of elements. The total measurement time of one spot is about 30 min and varies depending on signal intensity and was determined automatically by a statistical control. The absolute concentrations of each element were calculated from the measured intensities of positive atomic secondary ions, which were normalized to the intensity of secondary $^{30}$Si$^+$ ions, using calibration curves based on a set of reference glasses (Jochum et al., 2000, 2006). NIST-610 reference glass (Rocholl et al., 1997) was used as a daily monitor for trace element analyses. Accuracy of the trace element measurements for most elements with a concentration >1 ppm is 10–15 and 15–20% – for the concentration range 0.1–1 ppm; minimum detection limit is 5–10 ppb. To construct REE distribution spectra, the composition of zircon was normalized for that of chondrite CI (McDonough and Sun, 1995). Zircon crystallization temperature was estimated with a Ti-in-Zr thermometer (Watson et al., 2006).

**RESULTS**

**Zircon morphology**

Studying zircon grain’s morphology and the inner structure is an essential step before their local isotope-geochemical analyses. Integrated investigation using the SEM-BSE (Back-scattered electron image) and CL brings out useful information about the inner structure of each grain, allowing to distinguish several generations inside it. Such studies are necessary for the primary characterization of grains of one population and further genetic interpretations of geochemical data. Zircon is known for the broad formation of rims around the primary core grains, which can be crystallized both in magmatic and metamorphic stages, as well as during hydrothermal metasomatism (Corfu et al., 2003). When recrystallization of zircon takes place in fluid-enriched conditions, very unusual inner structures are formed, which are affecting primary growth zoning. These are complex, curved, and spongy structures with inclusions of surrounding rock-forming minerals (Kaulina, 2010; Skublov et al., 2010; Alfimova et al., 2011). These secondary zones in zircons are usually different from the primary ones by darker BSE or lighter in CL images and greater changes in chemical compositional enrichment with a broad range of non-formula elements.

Figure 4. Cathodoluminescence (SEM-CL) images of zircon grains (sample F36) from El Fereyid monzogranite with the analytical points numbered for U-Pb age ($^{206}$Pb/$^{238}$Pb) at the same spots. Ion microprobe crater is approximately 20 µm in size. The analytical points numbers are the same as the points numbers in Tables 1 and 2.
Zircon crystals from El Fereyid monzogranite are usually enclosed in quartz, biotite, and hornblende. They are characterized by a large size range between 120 to 600 µm. They commonly occur as prismatic and bipyramidal grains. Grains show multiple oscillatory growth zones in CL imaging (Fig. 4), which are consistent with typical magmatic zircon grains (Corfu et al., 2003). A few zircon grains show curved structures (e.g., rim of zircon F36_16), which are believed to be a result of the hydrothermal alteration. Zircon grains consist of the center of grain (cores) and marginal zones (rims). They are heterogeneous on CL imaging: core yields a darker CL response area, the rim yields a lighter CL response, and vice versa (Fig. 4). BSE imaging does not show significant differences between rims and cores. The cores of zircon grains are white, light grey as compared to their light grey rims. The cores of zircon contain an abundance of very fine mineral microinclusions.

U–Pb age of zircon

The uranium content of the zircon grains analyzed ranges between 378 and 13959 ppm, and the average concentration of U is 7500 ppm in cores and 600 ppm in rims. The thorium content varies between 54 and 2474 ppm, the average concentration of Th is 1650 ppm in cores and 120 ppm in rims, and the Th/U ratio is between 0.09 and 0.32. Ten measurements (5 analytical points in cores and 5 analytical points in rims) were carried out on zircon grains using the U–Pb geochronology method. The results obtained are summarized in Table 1. Nine analyses of zircon grains (the concordant cluster for point of cores F36_5 was not considered) yield the upper intercept of Concordia and Discordia age of 626 ± 13 Ma (MSWD = 0.44) (Fig. 5a). Two data points (cores F36_3 and F36_9), plotted away from the Concordia have younger 206Pb/238U–ages, which is very likely to be due to radiogenic

| Spot | 206Pb*, % | U, ppm | Th, ppm | 238U/206Pb* | % | 207Pb/206Pb Age, Ma | % |
|------|-----------|--------|---------|-------------|---|---------------------|---|
| F36_1 | 0.03 | 1507 | 323 | 0.22 | 133 | 628.5 | 2.6 |
| F36_2 | 0.12 | 616 | 137 | 0.23 | 54.0 | 625.3 | 3.7 |
| F36_3 | 0.54 | 5806 | 1180 | 0.21 | 368 | 456.4 | 1.6 |
| F36_4 | 0.12 | 378 | 116 | 0.32 | 32.8 | 620.0 | 4.5 |
| F36_5 | 0.01 | 13959 | 2474 | 0.18 | 1150 | 591.2 | 1.4 |
| F36_6 | 0.15 | 599 | 53.8 | 0.09 | 52.8 | 628.9 | 3.8 |
| F36_7 | 0.02 | 8689 | 2474 | 0.29 | 892 | 727.8 | 2.1 |
| F36_8 | 0.14 | 751 | 170 | 0.23 | 66.3 | 629.8 | 3.5 |
| F36_9 | 0.24 | 7513 | 1820 | 0.25 | 481 | 462.3 | 1.5 |
| F36_10 | 0.21 | 631 | 141 | 0.23 | 54.0 | 611.6 | 3.7 |

Errors are 1–σ; Pb* and Pb* indicate the common and radiogenic portions, respectively. Error in Standard calibration was 0.35% (not included in above errors but required when comparing data from different mounts). Common Pb corrected using measured 204Pb.
Pb loss from the metamict zircon domain. All these data (except point F36_5) yield a weighted average age of 626 ± 10 Ma (MSWD = 2.0) (Fig. 5b).

Zircon geochemistry

In zircon grains from El Fereyid monzogranite, the rims of zircon grains are depleted relative to the cores in almost all the trace elements analyzed, except for P, Ca, Ti with average values of 3625, 5880, 413 ppm, respectively (Table 2). The chemical composition of zircon grains from cores shows a higher concentration of Li, Y, Nb, Ba, Hf, Th, and U with averages of 75, 7449, 94, 17, 1593, 1351, and 7969 ppm, respectively (Table 2).

Both Ti and Nb can substitute for Zr, whereas Nb in the zircon rims is similar to those of unaltered magmatic zircon (Nb ≤ 62 ppm) (Hoskin and Schaltegger, 2003; Van Lichtervelde et al., 2011). Ti concentration exceeds the empirical limit of 50 to 70 ppm in zircon grains according to Page et al. (2007), which could be due to the addition of Ti by fluids, and does not indicate an elevated zircon crystallization temperature. According to the Ti–zircon thermometer (Watson et al., 2006), the estimated temperature of zircon with a low Ti concentration (cores F36_5, F36_11, F36_12, F36_14, and rim F36_6 in Table 2) showed averages of approximately 743 °C. This temperature can be considered as the crystallization temperature of monzogranite. The Ti–Ca binary diagram shows a distinctive positive correlation in the cores and the rims of zircon with higher concentrations in the latter one (Fig. 6a). The cores of zircon grains are enriched in U and Th, as compared to the rims (Fig. 6b). The cores and rims of zircon grains contain more U than Th, and thus a low Th/U ratio with average values of 0.17 and 0.19, respectively. The binary diagram of U–Th concentrations in both the zircon cores and the rims shows a linear and positive correlation trend with a marked enrichment in the cores of zircon, while the Th versus Th/U ratio shows fluctuation in the zircon grains analyzed (Fig. 6c). The binary diagram of the Hf versus Th/U ratio shows fluctuation in the zircon cores and rims with higher Th/U ratios in the cores (Fig. 6d).

The zircon grains from El Fereyid monzogranite exhibit high REE concentrations (average 7465 ppm for cores and 5375 ppm for rims). The cores of zircon grains have higher HREE concentrations, as compared with the zircon rims (average 5533 and 2783 ppm). On the other hand, the rims contain similar LREE concentrations (Table 2). The zircon rims show a strong distinctive positive correlation in the distribution of HREE and LREE, while the zircon cores display a positive correlation (Fig. 7a). The LuN/LaN ratio shows the average values of 899 and 77.4 for the cores and rims of zircon. The LuN/GdN and SmN/LaN ratios have higher contents in the cores with average values of 25.6 and 3.74, respectively than in rims of the zircon grains analyzed with the average values of 17.2 for LuN/GdN ratios and 1.75 for SmN/LaN ratios (Fig. 7b).

The majority of studied zircon grains display different patterns in LREE to HREE contents with an increase in the amount of HREE in the cores compared to the rims of the same zircon grains (Fig. 8). The zircon grains exhibit a negative Eu-anomaly (Eu/Eu*) with the average values of 0.50 for the rims and 0.37 for the cores (Table 2). The analyzed zircon grains also display a positive Ce-anomaly (Ce/Ce*) with the average values of 1.79 for the rims and 2.36 for the cores (Table 2). The REE distribution patterns in the zircon cores (F36_1 and F36_1D) respectively.
have a higher content of HREE than in the rims (F36_2), as opposed to a high LREE content in the zircon rims (Fig. 8a).

The REE content is higher in the cores of zircon grain (F36_3 and F36_3D) than in the rim (F36_4). The LREE in the cores of zircon grain is two orders of magnitude higher than the rims (Fig. 8b). The core and the rims are alike other grains from the population in the negative Eu/Eu* and positive Ce/Ce* anomalies. The rim of zircon (F36_4) has a well-defined positive Ce-anomaly (Ce/Ce* = 4.06), with the values not quite common for the Ce-anomaly in El Fereyid monzogranite zircons (Fig. 8c).

The core of grain F36_7 has differentiated REE spectra with typical low values for the Eu/Eu* and Ce/Ce* anomalies (Fig. 8d). The rim of the grain has a smooth REE pattern, with LREE concentration exceeding HREE content, which is typical for the zircon grains of hydrothermal-metasomatic origin (e.g. Hoskin and Schaltegger, 2003; Hoskin, 2005).

The core of zircon (F36_5) shows markedly differentiated REE patterns, in contrast to the rim (F36_6) with a more distinct enrichment in HREE in both the core and the rim of zircon grains than LREE. The core of zircon (F36_5) has a negative Eu-anomaly (Eu/Eu* = 0.14), as compared in the rim (Eu/Eu* = 0.44) with distinct positive Ce-anomalies in both of them. The values of Eu/Eu* and Ce/Ce* anomalies in this grain are not typical for El Fereyid monzogranite zircons (Fig. 8e).

The core of grain F36_7 has differentiated REE spectra with typical low values for the Eu/Eu* and Ce/Ce* anomalies (Fig. 8d). The rim of the grain has a smooth REE pattern, with LREE concentration exceeding HREE content, which is typical for the zircon grains of hydrothermal-metasomatic origin (e.g. Hoskin and Schaltegger, 2003; Hoskin, 2005).

The REE patterns in the core of zircon (F36_9 and F36_9D) are similar to the REE pattern in the rim of that grain (F36_10), but with a higher REE concentration of
the zircon cores. This grain has moderate negative Eu-anomaly and positive Ce-anomaly typical of the El Fereyid monzogranite zircons (Fig. 8e).

The REE patterns for the cores of various zircon grains (F36_11, F36_12, F36_14, and F36_16) show markedly differentiated REE patterns from LREE to HREE with a distinct enrichment in the latter one. The zircon cores F36_11 and F36_12 have a low LREE concentration and differ from the other studied grains by a distinct positive Ce-anomaly (Ce/Ce* = 4.69 and 6.72, respectively) and negative Eu-anomaly (Eu/Eu* = 0.07 and 0.06, respectively) (Fig. 8f).

The rims of zircon grains (F36_13, F36_15, and F36_17) have high REE content, with a gentle slope of the pattern where LREE is almost equal to HREE. This type of REE spectra is typical for the El Fereyid monzogranite zircon grains (Fig. 8g).
DISCUSSION

El Fereyid monzogranite belongs to granites with calc-alkaline and peraluminous composition and corresponds to the I-type granite (Dawoud et al., 2018). It is enriched in Ba, Rb, Sr, Zr, Nb, and REE, but depleted in K, P, and Ti. The magmatic emplacement of El Fereyid monzogranite was dated in this research by U–Pb zircon (SHRIMP II) method as 626 ± 13 Ma. Discordant U–Pb data for two analytical points (of cores F36_3 and F36_9 grains) are likely to be a result of partial Pb loss in metamict domains of zircons during modern-day weathering. Also, one cannot exclude the same effect from the onset of erosion and uplift associated with the major continental rifting of the Red Sea (Geisler et al., 2003). Also, Abdel Karim and Sos (2000) reported the K–Ar age of biotite from El Fereyid monzogranite as 610 ± 23 Ma, which is a significantly older age than that of syenogranites dated at 582 ± 22 and 570 ± 21 Ma. The presently studied El Fereyid massif is geochemically close to the Hamradon granite massif and their age is equal within the uncertainties of measurements (619 ± 17 Ma, U–Pb method; Geisler et al., 2003). Furthermore, the estimated age is consistent with the K–Ar age of El Fereyid monzogranite dated at 614 ± 23 Ma (Abdel Karim and Sos, 2000); the Sm–Nd age ~ 620 Ma of A-type granite from Humr Akarim and Humrat Mukbidof (Ali et al., 2012); the U–Pb age 622 ± 6 Ma of zircon from alkali-feldspar granite from Wadi Zaghra (Andresen et al., 2014) and the U–Pb age ~ 620 Ma of zircon from albite granite from Nuweibi (Emam et al., 2018). The ages obtained are much older than the previously reported whole-rock Rb–Sr age of monzogranite (587 ± 11 Ma; Stern and Hedge, 1985) from the study area.

Zircon grains from El Fereyid monzogranite are heterogeneous on CL images. They consist of cores and rims. Most cores yield a darker CL response, and the rim yields a lighter CL response. The zircon grains from El Fereyid monzogranite in general exhibit high REE and trace elements concentrations, especially for U, Th, HREE, and Y. The darker CL areas have higher REE, Th, U, Sr concentrations, than the rims, which are depleted in almost all trace elements analyzed, except P, Ca, and Ti. These characteristics are similar to those of the mineralized Wagone pluton (Li et al., 2018), suggesting slow crystallization from Th-depleted and U-enriched residual magmatic fluids under reducing conditions. As a rule, magmatic zircon grains display highly...
differentiated REE patterns with an increase in the amount of HREE in the cores compared to in the rims of the same zircon grains. Such grains exhibit a negative Eu-anomaly and positive Ce-anomaly. This pattern is referred to as characteristic of the magmatic zircon (Hoskin, Schaltegger, 2003; Hoskin, 2005). Another type of REE pattern in zircon, called ‘flattened’, with LREE almost equal to HREE, is characteristic for the hydrothermal and metasomatic zircon (Hoskin, Schaltegger, 2003; Hoskin, 2005). This type of REE spectra is observed in the rims of studied El Fereyid zircon grains and rarely, in their core (F36_1) (Fig. 8).

Binary diagrams of (SmN/LaN) versus La, (SmN/LaN) versus (Ce/Ce*) and Ce/Ce* versus Eu/Eu* could be used to distinguish magmatic and hydrothermal zircon grains as earlier proposed (Hoskin, 2005; Kirkland et al., 2009). The zircon grains analyzed tend to be located near the hydrothermal field (Hoskin, 2005; Kirkland et al., 2009) (Fig. 9). However, some of the cores and rims of zircon grains fall between the hydrothermal and magmatic fields. That position is possible for magmatic zircon, partially affected by hydrothermal fluids, or for the zircon crystallized at the final magmatic stage, when the melt is saturated with fluids (Levashova et al., 2016). Observed high REE and trace elements content, together with flattened spectra are not directly pointing to the hydrothermal genesis of all zircon grains from El Fereyid monzogranite. Enriched cores of zircon grains formed from REE and trace elements-rich magmatic melt. The rims of the zircon grains analyzed are depleted relative to the cores by almost all elements, except for some elements mobile in fluid (Ca, LREE, P, and Ti). Thus, rims were formed at the final stages of magmatic crystallization of the melt, which was rich in fluid but poor in rare and trace elements. Obtained U–Pb age of the rims supports this suggestion, being equal to the magmatic stage age.

CONCLUSIONS

1. The age calculated for the studied zircon grains (U–Pb) from El Fereyid monzogranite is 626 ± 13 Ma and is interpreted as an age of monzogranite crystallization. This conclusion supports earlier data for the K–Ar age of the massif (610 ± 23 Ma) and makes it more precise.

2. Zircon grains of El Fereyid monzogranite are heterogeneous in internal structure as revealed in CL images, they have elevated REE and trace elements content, especially REE (reaches the value of 22500 ppm), U (reaches the value of 14800 ppm), Th (reaches the value of 2500 ppm), and low Th/U ratios (average 0.18). The cores of zircon grains are enriched in almost all the trace elements analyzed, first of all in U, Th, HREE, and Y, compared to the rims. Rims are enriched only with Ca, P, LREE, Ti, and Sr, which are considered as mobile in the fluid.

3. The rims of zircon grains which demonstrate REE spectra typical for the hydrothermal zircons were nevertheless, crystallized on the final magmatic stage when the melt was rich in fluid and depleted in trace elements.

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SUPPLEMENTARY MATERIALS

Color version of Figures 1 and 2 is available online from https://doi.org/10.2465/jmps.210320.

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