Neoproterozoic U-Th-REE-bearing Pegmatites in Wadi Ibib, South Eastern Desert, Egypt: Structural and Geochemical Measures for a Syn-Tectonic Anatectic Model of Formation

Waled S. IBRAHIM*
Nuclear Materials Authority, El Maadi, Cairo, Egypt

Abstract: The Wadi Ibib area is situated in the northern part of the Neoproterozoic Hamisana Shear Zone (HSZ), which is a high strain zone evolved during the late stages of the Pan-African orogeny, likely as a tectonic escape structure. Amphibolite facies pelitic metasedimentary windows crop out in the axial parts of the HSZ and are noticeably associated with numerous N-trending pegmatite dikes. Whole-rock geochemistry of the pegmatites reveals a peraluminous (S-type) affinity, with low K/Rb ratios and elevated concentrations of U, Th, REE, Rh, Li, Cs, Y, Nb and Ta. Structurally, the pegmatite sets intrude along the shear plane of the HSZ, corresponding to the regional N-trending tectonic fabrics, such as axial planar foliation and dextral-shearing in the metasedimentary host rock. Field relationships, including structural context, coupled with geochemical characteristics of the Wadi Ibib pegmatites, do not support their formation as a complementary part of evolved granitic magmas. Space-localized decompression-induced partial melting of peraluminous garnet-bearing metapelites was alternatively the underlying process for formation of these pegmatites. Such decomposition was associated with regional escape tectonics and stress axes permutations during the late deformation stage (D3) in the evolution of the south Eastern Desert terrane, due to end-orogeny system pressure-release.

Key words: U-Th-REE mineralization, anataxis, pegmatite origin, Hamisana Shear Zone, Wadi Ibib, Egypt


1 Introduction

Pegmatites may form significant reservoirs of rare metals (Linnen et al., 2012; Dill, 2015), rare earth elements (REE) and radioactive metals (U and Th). Nevertheless, the origin of pegmatites and the associated mineralization is still a matter of debate (Černý and Erceit, 2005). Mineralized pegmatites can be classified into two suites (end-members): i) pegmatites with concentrations of niobium-yttrium-fluorine (referred to as the NYF suite); and ii) pegmatites with concentrations of lithium-cesium-tantalum (referred to as the LCT suite) (Černý et al., 2012). Two hypotheses are posited for the generation of pegmatic melts: 1) continued fractionation of granitic magmas during plutonism and 2) limited partial melting of peraluminous (S-type) magmas during plutonism and 2) limited partial melting of peraluminous (S-type) magmas during plutonism and 2) limited partial melting of peraluminous (S-type) magmas during plutonism and 2) limited partial melting of peraluminous (S-type) magmas during plutonism. These hypotheses can practically represent two end-members of pegmatite formation. Generally, the host lithology, tectonic setting and geological setup of pegmatite swarms are important parameters that may control their formation and element budget.

In the Egyptian Eastern Desert, most of the abundant pegmatite sets are spatially associated with post-tectonic Pan-African granites (e.g., Abd El-Naby and Saleh, 2003; Saleh et al., 2008; Ghazaly et al., 2015). Alternatively, the pegmatite sets associated with amphibolite facies gneissic and schistose rocks across the southern Eastern Desert are rarely recorded in some localities, e.g., Hafafit and Nugrus areas (e.g., Raslan and Ali, 2011; Ibrahim et al., 2015, 2017). The structural pattern of Neoproterozoic rocks (e.g., propagation of thrusts and sutures) along the south Eastern Desert reveals a polyphase deformation history (Abd El-Naby et al., 2000; Ibrahim et al., 2016; Abd El-Wahed et al., 2019). It implies a complex tectonic configuration, due to successive collision events during the acceleration of the Pan-African orogeny (Abdelsalam and Stern, 1996; Abd El-Naby et al., 2000; Kusky and Ramadan, 2002; Abdeen and Abdelghaffar, 2011). The early collision phase between the south Eastern Desert and the Gabgaba terranes formed the E–W-trending Allaqi-Heiani–Sol Hamed suture (Fig. 1a; Kröner et al., 1987; Stern, 1994; Abdelsalam and Stern, 1996), which occurred between 750 Ma and 720 Ma (Abdelsalam and Stern, 1996; Abdelsalam et al., 2003). This collision event is superimposed by the N-trending HSZ (Fig. 1a), due to an E–W shortening related to the collision of East and West Gondwana, during the late stages of the Pan-African orogeny, in terms of escape tectonics (Burke and Sengor, 1986; Kröner et al., 1987; Stern, 1994; Abdelsalam and Stern, 1996; Kusky and Ramadan, 2002; Ibrahim et al., 2016; Hamimi et al., 2019; Abd El-Wahed and Hamimi, 2021). The geochronological data (Rb-Sr/U-Pb zircon) established by Stern et al. (1989) along the northern part
of the HSZ, indicate that the N-trending shearing event occurred between 660 Ma and 550 Ma, during the late- to post-stage terrane accretion. An amphibolite facies metamorphic phase may have been associated with this event, which occurred at ~600 Ma, according to K/Ar ages derived from hornblende grains of the amphibolite schist in the central part of the Allaqi–Hetani suture, farther west of the HSZ (Fig. 1a; Abd El-Naby et al., 2000).

The Wadi Ibib area lies within the uppermost northern part of the HSZ (Fig. 1a). The area is hosted by Neoproterozoic amphibolite facies of metavolcanic and metasedimentary rocks, that are structurally overlain by ophiolitic nappe and intruded by syn- and post-tectonic gabbroic and granitoid varieties (Fig. 1b, c). The Wadi Ibib pegmatites occur as steeply dipping, N-trending, dike-like bodies. They mainly intrude into the pelitic metasedimentary rocks, that have a parallel strike to the axial plane of the HSZ (Fig. 1b, c). Based on whole-rock geochemical data and classification by Černý and Ercit (2005), the Wadi Ibib pegmatites are akin to the NYF-suite, with significant concentrations of U, Th and REE.
This study uses field relationships, structural analysis, tectonic, mineralogical and geochemical aspects of the mineralized pegmatites in the Wadi Ibib area, to decode the role of tectonism and anatexis in the formation of pegmatites and concentration of U, Th, REE and some rare metals.

2 Geological Setting

2.1 Regional geological setting and tectonic evolution

The Wadi Ibib area is part of the HSZ (Hamisana Shear Zone; Fig. 1a, b). The latter is a high strain zone in the Neoproterozoic Arabian-Nubian Shield (ANS) and is proposed to be a transpressional zone, associated with late collisional events (Stern et al., 1989; Miller and Dixon, 1992; de Wall et al., 2001; Johnson and Woldehaimanot, 2003). Ibrahim et al. (2016) suggested that the HSZ represents a tectonic escape structure, according to a reconstructed paleostress configuration of the oblique shearing across a fold-thrust belt. Based on superimposed structures, the HSZ is suggested to have experienced a polyphase deformation history (Stern et al., 1989; Miller and Dixon, 1992; de Well, 2001; Johnson and Woldehaimanot, 2003; Ali-Bik et al., 2014; Ibrahim et al., 2014b, 2016).

The D1 event is an early N–S shortening event. Although this event is poorly preserved within the HSZ, some traces of its fabrics, especially fold-interference patterns, were recorded in the ophiolitic units (serpentinites, metagabbros, amphibolites and mélange; Table 1; Fig. 2a). Moreover, this event has been documented in the Allaqi–Heiani belt, farther west of the study area (e.g., Abdelsalam et al., 2003; Abdeen and Abdelghaffar, 2011, Fig. 1a). The D1 event is characterized by E–W thrust and E–W folding (DF1) (Fig. 2a). The D1 event represents an early episode of the collision between the Eastern Desert and Gabgaba terranes (Fig. 1a), during the Pan-African orogeny accretion (Kusky and Ramadan, 2002; Abdelsalam et al., 2003; Johnson and Woldehaimanot, 2003; Abdeen and Abdelghaffar, 2011).

The D2 event is an E–W to ENE–WSW shortening stage, that has a compressional regime with an ENE-trending σ1 paleostress axis (Ibrahim et al., 2014b). The structures evolved under this regime are the NW–thrust faults and related NW–folds, as well as NW-trending foliation (DF2) (Fig. 2a), which have been recorded only in the ophiolitic and island arc rocks (Table 1; Figs. 1b and 2a). The D2 event represents an early episode of the collision between the Eastern Desert and Gabgaba terranes (Fig. 1a), during the Pan-African orogeny accretion (Kusky and Ramadan, 2002; Abdelsalam et al., 2003; Johnson and Woldehaimanot, 2003). The D2 event is an E–W shortening deformation phase, expressed by N–S upright folds (DF3) (Fig. 2c) and conjugate NW-sinistral and NNE-dextral fault sets (Fig. 2d), deforming the ophiolitic and island arc rocks, as well as granodiorite intrusions (Table 1; Figs. 1b, c and 2). The D2 NW-thrusts were reworked and reactivated as strike-slip faults by the transpressional regime during D3 (Fig. 2d). The syn-shearing regime during D4 (Fig. 2d) shortening.

Table 1 Tectonic calendar of the Wadi Ibib area, based on the correlation between deformation events, tectonic regime, regional tectonic setting and related fabrics recorded in its rocks

<table>
<thead>
<tr>
<th>Deformation</th>
<th>Orogenic events (Neoproterozoic)</th>
<th>Tectonic Setting</th>
<th>Anogenic event</th>
</tr>
</thead>
<tbody>
<tr>
<td>D1 Tectonic</td>
<td>Compression N–S σ1</td>
<td>Collision of the Eastern Desert and Gabgaba terranes</td>
<td>Transpression E–W σ1/N–S σ3</td>
</tr>
<tr>
<td>D2</td>
<td>Compression ENE–WSW σ1</td>
<td>Collision of East and West Gondwana (early-stage collisional shortening)</td>
<td>Transtension E–W/N–S σ3</td>
</tr>
<tr>
<td>Fabrics</td>
<td>NW-minor fold (DF2)</td>
<td>Northwards escape tectonics (late-stage collisional shortening)</td>
<td>E–W dextral fault</td>
</tr>
<tr>
<td></td>
<td>NW–foliation (S2)</td>
<td>Red Sea rifting</td>
<td>NNW–SSE, and</td>
</tr>
<tr>
<td></td>
<td>F1/F2 interference</td>
<td></td>
<td>E–W normal faults</td>
</tr>
<tr>
<td></td>
<td>patterns</td>
<td></td>
<td>N–S and E–W</td>
</tr>
<tr>
<td></td>
<td>NW–thrust</td>
<td></td>
<td>joint/fracture</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>systems and</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>associated dike</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>swarms</td>
</tr>
<tr>
<td></td>
<td>NW–minor fold (DF1)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>NW–foliation (S1)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>NW–thrust</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>NW–minor fold (DF2)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>NW–foliation (S2)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>F1/F2 interference</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>patterns</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>NW–thrust</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>NW–minor fold (DF3)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>NW–foliation (S3)</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Conjugate crenulation cleavage</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N–S crenulation cleavage</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N–S mineral lineation</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>S/C foliation with dextral geometries</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Mylonic foliation</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Asymmetric δ- and σ- porphyroclasts of hornblende and deformed garnet rotate dextrally</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Asymmetric Z-shaped veins associated with N-dextral shearing</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>N–S dextral shearing</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Transpression conjugate set of NW-sinistral and NNE-dextral faults</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*Post-tectonic granite*
*Post-tectonic gabbro*
*Syn-tectonic granite*
*Metasediments*
*Metavolcanics*
*Métangé*
*Metagabbros*
*Serpentinites*
emplacement of the granodiorite related to this event is manifested by its deformation state, where the sheared slabs were intruded by less-deformed equivalents (Ibrahim et al., 2016; Fig. 2e). A localized amphibolite facies metamorphic stage may have been associated with this event along the HSZ (equivalent to D2 of Ali-Bik et al., 2014). Based on transpressional oblique-shearing geometry across a fold-thrust belt (Ibrahim et al., 2016), the D3 event refers to northward escape tectonics, in response to late-stage collisional shortening of Gondwana (e.g., Burke and Sengor, 1986; Kusky and Ramadan, 2002; Ibrahim et al., 2016; Hamimi et al., 2019).

The D4 event was a post-orogenic transtensional phase, characterized by E–W right-lateral and NNW–SSE, as well as E–W normal faults (Table 1; Figs. 1b, c, 2f and g). The D4 event likely began after a post-tectonic alkali granite emplacement, affecting all rock units in the study area (Table 1). The structural systems related to this event are concordant with a tectonic model (simple-shearing) of the rift-system of the Red Sea (e.g., Ghebreab, 1998).

2.2 Detailed geological setting

The Neoproterozoic basement of the study area comprises highly tectonized, allochthonous ophiolite blocks, island arc metavolcanic and metasedimentary sequences and different phases of syn- and post-tectonic gabbroic and granitoid intrusions (Fig. 1b).

Pegmatite sets in the Wadi Ibib area extend for more than 8 km and vary in width from 5 m to 15 m (Figs. 3–4). These pegmatites were emplaced parallel to the schistosity of the metasedimentary host rocks and do not extend beyond their boundaries (Figs. 1b–c, 3 and 4). The metasedimentary host rocks form a N–S belt of extensively foliated, folded and mylonitized rocks in the
central or axial part of the HSZ (Figs. 1b–c, 3 and 4a–b). The metasedimentary rocks are fine- to medium-grained texture, mainly represented by garnet–biotite and garnet-muscovite-biotite schists interleaved with quartzite bands (Figs. 3 and 4a–b). Biotite, feldspar, hornblende, quartz and muscovite are the main constituents, with disseminated garnet porphyroblasts, as well as a lesser amount of staurolite and monazite (Table 2). The quartzite bands are concordant with the pervasive foliation (schistosity) of the host metasedimentary rocks (Fig. 3). The quartzite is mainly composed of quartz, plagioclase and iron oxy-hydroxide with minor biotite, muscovite, kyanite and sillimanite, however zircon is also found as an accessory (Table 2).

Based on observed field relationships, the Wadi Ibib pegmatites have deformed with a well-developed set of conjugate shear fractures and a complete absence of internal zoning (Fig. 4f, h–j). They exhibit a relatively variable grain-size, however locally they are coarse-to very coarse-grained with inequigranular to equigranular textures, consisting of plagioclase, quartz and K-feldspar with variable amounts of muscovite (up to 2 cm; Table 2). Detached patches of garnet-muscovite (up to 30 cm; Fig. 4e) and variable amounts of disseminated coarse-grained garnet (up to 3 cm; Fig. 4f) are common and more concentrated near the pegmatite contacts. The accessory phases bearing mineralization (U, Th, REE and rare-metals) are commonly xenotime, uraninite, kasolite and zircon, as well as monazite and columbite (Table 2). These accessory phases are described in detail below (see section 4.3 below; Fig. 9 for description).

The pegmatites have very sharp planar contacts against their host schists, however the irregular and/or gradational contacts are completely absent (Fig. 4c–d, g–j). Pegmatite sets occur as steeply dipping (65°–80°W), N-trending, dike-like bodies, that are mostly concordant with the main pervasive fabrics in their host schists (e.g., schistosity; Fig. 4c–d, g–h). Generally, they occupy the F3-axial planar foliation (Fig. 4c–d). In some places, the pegmatites occur in the form of a boudinaged structure within the main N-trending dextral shear zone (Fig. 4i). Locally, they occur as discontinuous N-trending sheath-like bodies (nearly horizontal), filling the trough of F3-folds at the

---

Table 2: Summary of the typical petrographic characteristics of metasedimentary host rocks (metapelites and quartzites), pegmatites and granitoids in the Wadi Ibib area

<table>
<thead>
<tr>
<th>Rock type</th>
<th>Metasedimentary host rocks</th>
<th>Pegmatites</th>
<th>Granodiorites</th>
<th>Graniteoids</th>
<th>Alkali granites</th>
</tr>
</thead>
<tbody>
<tr>
<td>Essential minerals</td>
<td>Biotite</td>
<td>Quartz</td>
<td>Plagioclase</td>
<td>Plagioclase</td>
<td>K-feldspar</td>
</tr>
<tr>
<td></td>
<td>Plagioclase</td>
<td>K-feldspar</td>
<td>Quartz</td>
<td>quartz</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Muscovite</td>
<td>Plagioclase</td>
<td>K-feldspar</td>
<td>Muscovite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Hornblende</td>
<td>Muscovite</td>
<td>Muscovite</td>
<td>Muscovite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Quartz</td>
<td>Kaolinite</td>
<td>Quartz</td>
<td>Sericite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Staurolite</td>
<td>Sillimanite</td>
<td>Sericite</td>
<td>Sericite</td>
<td></td>
</tr>
<tr>
<td>Secondary minerals</td>
<td>Chlorite</td>
<td>Chlorite</td>
<td>Sericite</td>
<td>Chlorite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Epidote</td>
<td>Epidote</td>
<td>Kasolite</td>
<td>Sericite</td>
<td></td>
</tr>
<tr>
<td>Accessory minerals</td>
<td>Garnet</td>
<td>Garnet</td>
<td>Garnet</td>
<td>Garnet</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Xenotime</td>
<td>Xenotime</td>
<td>Xenotime</td>
<td>Xenotime</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Zircon</td>
<td>Zircon</td>
<td>Zircon</td>
<td>Zircon</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Opaques</td>
<td>Opaques</td>
<td>Uraninite</td>
<td>Uraninite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Apatite</td>
<td>Apatite</td>
<td>Zircon</td>
<td>Zircon</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Monazite</td>
<td>Monazite</td>
<td>Monazite</td>
<td>Monazite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sphene</td>
<td>Sphene</td>
<td>Columbite</td>
<td>Columbite</td>
<td></td>
</tr>
<tr>
<td>Common features</td>
<td>Garnet-rich metapelites</td>
<td>Garnet-rich</td>
<td>Garnet-rich</td>
<td>Garnet-free granitoids</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Opaque-rich quartzites</td>
<td>Disseminated accessory phases</td>
<td>Decreasing in the plagioclase, biotite and hornblende abundance from granodiorites to alkali granites</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
contact between metavolcanic and metasedimentary rocks (Fig. 4j).

3 Materials and Methods

For an updated geological map, data from ASTER false-color composite images and band ratios as well as ASTER-Digital Elevation Models (DEM) were fused and co-registered within an ArcGIS-environment. The detailed map of the mineralized pegmatites is based on a grid pattern of 25 m × 25 m (Fig. 3). Fieldwork included making structural measurements, recording cross-cutting relationships and sample collection. Petrographic investigation of both pegmatites and the host rocks was conducted on a bunch of thin and polished sections using a Nikon microscope along with Scanning Electron Microscope-Energy Dispersive Spectrometry (SEM-EDS) microanalyses, while SEM images were collected in backscattered electron mode (BSE) (Fig. 9). The analysis was done at the Center of Advanced Instrumental Analysis, Faculty of Engineering, Kyushu University (Japan). The operating conditions used for the SEM-EDS (SEM: Shimazdu Super Scan SS-550; EDS: Genesis 2000 EDS) were an acceleration potential of 15 kV with a beam...
current of about 60 nA and a beam diameter of 3 µm. All data were corrected with a ZAF matrix-correction program.

Seventeen representative pegmatites, 11 metasedimentary host rocks (metapelites and quartzites), and 10 granitoid samples were selected for a geochemical study. Pegmatite-rock samples of about 5 kg per sample were collected, in order to get a representative geochemical analysis of the coarse-grained rocks studied. Major and trace element composition was determined using a Rigaku RIX-3100, series VR 25006 X-Ray Fluorescence (XRF) spectrometer, at the Department of Earth Resources Engineering, Kyushu University (Japan). Samples were crushed and milled using a vibration mill, with loss of ignition (L.O.I) measured as the weight-difference after ignition. The operating conditions of the XRF analysis comprised an accelerating voltage of 50 kV, a current of 50 mA and carrier gas flow of 50 mL/min. The JA-3 (Japan-igneous rocks standard) was used as the standard sample. REEs and some trace-element determination was carried out on an Agilent 7500ce ICP-MS at the Central Laboratory, Department of Material Science and Engineering, Kyushu University (Japan). 0.5 gm sample was split and digested in Aqua Regia (a mixture of HF, HNO₃ and HClO₄ acids), with the CEM-MARS-6 system. Multi-element JA-3 (Japan - igneous rocks standard) was again used as a standard sample.

4 Results

4.1 Structural analysis of the pegmatites and their host rocks

Pegmatites in the Wadi Ibib are concordant with the main structural fabrics of the metasedimentary host rocks. The metasediments have experienced a protracted history of collisional deformation, which has been detailed in several studies (e.g., Ibrahim et al., 2014a, b; 2016).

The ductile and brittle structural elements recorded and analyzed in the metasedimentary host rocks include foliation, lineation, folding, faulting and shear-related structures (Fig. 5), indicating that they mostly developed under amphibolite facies conditions. Based on trend analysis, these rocks have two foliation attitudes (S₁ and S₃), corresponding to different deformation events in the study area (Fig. 3). However, the S₁ fabrics related to the earlier deformation event in the Wadi Ibib area are rare or absent from these rocks (see Table 1). The S₂-foliation striking N40°–56°W with steep to moderate dip (45°–60°) towards the NW and SE, associated with the main NW-thrust faults in the study area (Figs. 1a–b). These foliation planes are parallel to the axial plane of the NW-directed σ2 minor folds, with a fold axis plunging 27° towards the NW (Fig. 5b). Near the thrust belt, the S₂ foliation is cut by a sub-vertical shear foliation (S3) (Fig. 5b). The S₃-foliation strikes N-S and is sub-vertical to steeply-dipping towards E and W directions along the axial part of the HSZ. The sub-vertical S₃ foliation is regarded as axial planar to F3 major folds in the study area (Fig. 5a–b). The F3 folds in the metasediments exist as open to tight, upright with syncline and anticline forms (Fig. 5a–b). On the stereonet, S₃-poles show sub-vertical to steeply E- and W-dipping planes, gently plunging N-wards (Fig. 5a–b). These folds have a crenulation cleavage along their axial planes, occurring as a conjugate set associated with an asymmetric Z-shaped crenulation (Fig. 5c). Well-developed N–S lineations (L₂) are commonly parallel to the F₃ fold axes, however within the shear zone these lineations occur at an acute angle (∼20°) to the measured F₃ fold axes (Fig. 5a). This indicates that the F₃ fold-event coincided with the dominant dextral oblique shearing (Fig. 5a–b).

According to the brittle deformation analysis, the oldest deformation fabrics in the metasedimentary rocks are NW-thrust faults, F₂ folds and the associated S₂-foliation. These fabrics were developed under an ENE–WSW compressional regime (Ibrahim et al., 2014b). The metasedimentary rocks represent the footwall of the major NW-thrust in the study area (Figs. 1b, 2a–b, 3 and 5a–b). Away from the axial plane of the HSZ, the NW-trending thrust planes are well-preserved, however within the shear zone these planes are gradually rotated N-wards and reactivated by a dextral oblique shearing (Fig. 5a). This regime is overprinted by a dominant dextral oblique-slip transpressional regime along the N-trending shear zone (HSZ; Figs. 1c, 2d and 5a–f). Based on fault-slip data, this regime is characterized by a paleostress field that has an E–W trending σ₁ (steeply plunging), N-trending σ₃ and σ₂ perpendicular to Earth’s surface (Fig. 5a, d and f). The regime-controlled a conjugate set of oblique-slip faults, including NW left-lateral and NNE–SSW to N-S trending right-lateral faults (Figs. 3 and 5a–d). In the late stage of the tectonic event, the regime has likely switched to pure extension, as a result of a permutation of the stress axes from an E–W directed σ₁ to an E–W directed σ₃, due to system pressure-release (Fig. 5d). A post-orogenic transtensional regime has been recorded in the metasedimentary host rocks, characterized by an E–W right-lateral and NNW–SSE, as well as E–W normal faults (Figs. 3, 5a and f–g).

Kinematic indicators are observed at different scales from outcrop to microscopic in both pegmatites and metasedimentary host rocks. In some places, the pegmatites occur in the form of asymmetric boudinaged structures within the main NS-dextral shear zone (Fig. 5d). Furthermore, the pegmatites contain sigmoidal aggregates of quartz and feldspar, as well as uncommon schist enclaves, parallel to the mylonitic foliation (S₃), indicating a dextral sense of shearing (Fig. 6a). The N–S–striking S–C shear-fabrics are common in the metasediments (Fig. 6b). The S-surfaces (schistosity) are recognized by a preferentially oriented hornblende, biotite, recrystallized quartz and garnet grains that are oblique to the C-surfaces (at ∼27°). The geometrical relationship between S- and C-surfaces (shear plane) reflects a dextral sense of shearing in the N-direction (Fig. 6b). Well-developed σ₅- and δ-porphyroclasts of hornblende and deformed garnet indicate that they have rotated dextrally during the shearing event (Fig. 6c–d).

4.2 Whole-rock geochemistry

4.2.1 Metasedimentary host rocks

The geochemical data of the metasedimentary...
Ibrahim / Syn-tectonic Anatectic Pegmatites in Southern Egyptian Nubian Shield

Samples of metapelites and quartzites are given in Supp. Table 1. The metapelite and quartzite samples show silica contents of 53.15–60.32 wt% and 83.80–85.81 wt%, respectively (Supp. Table 1). Al content in the metapelite samples are 13.43–15.17 wt% and 7.88–9.82 wt% in quartzite samples. The metapelite samples have high P2O5 concentrations (up to 0.96 wt%; Supp. Table 1), likely as a result of xenotime and monazite. The contents of Na2O and K2O (averages are 4.39 wt% and 2.16 wt%, respectively) and their ratios (Na2O/K2O > 1) in the metapelite samples may be explained by the relative abundance of plagioclases (whereas the quartzite samples have Na2O/K2O < 1), reflecting the abundant K-feldspar. The calculated alumina saturation index (ASI) of Shand (1943), [A/CNK; molar% A12O3/(Na2O + K2O + CaO)] for the metapelite designates their strongly peraluminous character. The trace element composition of the examined metapelite rocks is variable. The highest Th, Cr and V...
contents are 59 ppm, 380 ppm and 236 ppm, respectively (Supp. Table 1). The Ba values range from 132 ppm to 710 ppm, Sr from 286 ppm to 893 ppm and Rb from 33 ppm to 478 ppm, Y-contents are from 192 ppm to 518 ppm (Supp. Table 1). Also, REE, Zr, Nb, W and Co trace elements have average contents of 127.45 ppm, 182.41 ppm, 68.13 ppm and 41.75 ppm, respectively (Supp. Table 1).

4.2.2 Pegmatites and granitoids

The results of the geochemical analysis of the selected pegmatite and granite samples are given in Tables 4 and 5, respectively. SiO$_2$ concentrations in the pegmatite samples range between 71.02 wt% and 78.83 wt% and show a granitic composition (Fig. 7a; Gillespie and Styles, 1999). However, the granitoid rocks show two distinctive ranges of silica content of 64.15–68.09 wt% and 75.78–78.29 wt%, reflecting a compositional variation in these rocks, from granodiorite to granite, respectively (Fig. 7a).

Major oxides of the pegmatites TiO$_2$, MnO and MgO are commonly low (<0.59 wt%), however FeO and CaO are more variable, having ranges between 0.26–1.31 wt% and 0.73–1.16 wt% respectively (Supp. Table 2). Broadly, Na$_2$O is the prevailing alkali in all pegmatites, except for two samples (PE-05 and PE-07) that have K$_2$O > Na$_2$O, which might be attributable to the presence of muscovite patches (e.g., Fig. 4e). The pegmatite samples mostly contain high P$_2$O$_5$ concentrations (up to 1.11 wt%), likely as a result of the accessory phases (e.g., xenotime and monazite). On the other hand, the granodiorites have higher Al$_2$O$_3$, MgO, CaO, FeO, TiO$_2$ oxides, with lower total alkalis-SiO$_2$ content, compared to those of the alkali granites (Supp. Table 3). Both, however, are far from those of the pegmatites (major oxides).

The calculated ASI (Shand index) for the pegmatite samples (A/CNK) range between 1.11 and 1.26 (Supp.
Table 2), designating their peraluminous character (Fig. 7b). Furthermore, this index (A/ CNK ≥ 1.1) indicates S-type affinity for the studied pegmatites (Fig. 7b), according to Vetter and Tessensohn (1987). However, the Shand index for the granodiorites ranges from 0.91 to 0.97, and the alkali granites from 0.82 to 0.98 (Supp. Table 3), indicating that the granitoid rocks have a metaluminous character (Fig. 7b). Also, in Fig. 7c (Frost et al., 2001), pegmatites are scattered in the S-type affinity field, however the granodiorites and alkali granites show I- and A-type affinities, respectively (Fig. 7c). The pegmatites and granodiorites show a separate differentiation trend, according to the K2O/Rb versus Rb binary relationship (Fig. 7d). The pegmatites are commonly scattered in the mineralized granitic field, however the granitoid rocks plot in the barren granitic field of Stavrov et al. (1969) (Fig. 7d).

The trace-element (mantle-normalized; McDonough and Sun, 1995) signature of the pegmatite samples is distinguished by a positive anomaly of Rb, U and Th, but strongly depleted in Ti, Sr and Ba, also showing a pronounced enrichment in Cs, Li, Y, Nb and Ta (Fig. 8a). These signatures reflect an abundant garnet, xenotime, zircon and columbite accessory phase, also the Cs-enrichment signatures may be caused by the muscovite patches in the pegmatites. Notably, the pegmatite samples are characterized by low K/Rb ratios (mostly < 62; Supp. Table 2). The granodiorites are characterized by Rb, Sr, Y and Ti enrichments, with Cs, Ba, Nb and Ta depletion (Fig. 8b), however the alkali granites show more Rb, Cs and Y enrichment, with Ba, Nb, and Ti depletion, compared to those of granodiorite (Fig. 8b). K/Rb ratios in granodiorites (145.83–182.04) are lower than those from alkali granites (232.23–292.67; Supp. Table 3), however both are below the mantle ratio (K/Rb = 1000) of Shaw (1968), indicating that both are crust-derived rather than mantle-derived varieties.

The total REE contents of the pegmatite samples are commonly high (ΣREE up to 1354.96 ppm), with an average of around 600 ppm (Supp. Table 2). The REE patterns (chondrite-normalized; Sun and McDonough, 1989) of these pegmatites (Fig. 8c) are characterized by extreme enrichment of HREE over LREE and strong negative Eu anomalies (mean Eu/Eu* = 0.13). LREE depletion and HREE enrichment (Supp. Table 2) likely reflect the abundance of garnet and xenotime in the bulk pegmatites. Furthermore, the REE signatures (chondrite-normalized; Sun and McDonough, 1989) of the granitoids (Fig. 8d) show that the granodiorites are enriched in LREE and exhibit distinct less-fractionated flat HREE patterns and moderate negative Eu anomalies (0.66–0.68). However, the alkali granites are more enriched in LREE and exhibit a more distinctive Eu negative anomaly (0.31–0.40) than the granodiorites (Supp. Table 3).

McKeough et al. (2013) have described the correlation between the hybridization index (CaO + MgO + FeO) of pegmatites and the abundance of U-Th-REEs-bearing minerals, which is considered to be evidence of a host
rock assimilation (i.e., pegmatites/host rock interaction). However, there is no clear correlation between the enrichment of U, Th and REE in the Wadi Ibib pegmatites and their hybridization index (Fig. 8e–f).

4.3 Mineralization of the pegmatites

Field relationships and geochemistry of the Wadi Ibib pegmatites show an elevated concentration of U, Th, REE, Nb and Ta. The mineralization-bearing phases are mostly represented by xenotime, uraninite, kasolite and zircon with variable amounts of monazite and columbite-tantalite. They are concentrated and visibly disseminated within the investigated pegmatites (Fig. 4). The presence of these accessory phases was confirmed and described using SEM-EDS examination (Fig. 9).

Garnet mostly occurs as a coarse- to locally medium-grained, fractured, unzoned crystal (Fig. 9a, c–d). The well-developed inclusion-rich garnets prevail, containing anhedral crystals of uraninite, kasolite (Fig. 9c–d) and/or xenotime (Fig. 9a). Based on EDS analysis, garnets consist essentially of Fe₂O₃ (46.84 wt%), SiO₂ (27.12 wt%), Al₂O₃ (15.53 wt%), with minor MnO (5.51 wt%), CaO (3.80 wt%) and MgO (1.19 wt%), which in turn reflects an almandine-rich garnet composition (Fig. 9a). Xenotime occurs as anhedral medium- to fine-grained disseminated crystals, commonly associated with garnet-rich patches and/or occurring as inclusions in garnet and feldspars (Fig. 9a–b). According to EDS data (Fig. 9b), it consists of Y₂O₃ (48.61 wt%), P₂O₅ (34.35 wt%), Dy₂O₃ (10.13 wt%), Gd₂O₃ (3.44 wt%) and Er₂O₃ (2.34 wt%) with minor SiO₂ (1.13 wt%). Uraninite is the most abundant U-mineral in the Wadi Ibib pegmatites.

Fig. 8. Geochemical data of pegmatites and selected granitoids in the Wadi Ibib area. (a) Primitive mantle-normalized trace-element spectrum of pegmatites; (b) Primitive mantle-normalized trace-element spectrum of selected granitoids; (c) Chondrite-normalized REE diagram of pegmatites; (d) Chondrite-normalized REE diagram of selected granitoids; (e and f) A hybridization index of CaO + MgO + FeO contents in the Wadi Ibib pegmatites (n = 17) show weak to no relationship between the hybridization effects and enrichment of the U, Th and REE mineralizing phases. Normalizing factors for trace elements are from McDonough and Sun (1995) and normalizing factors for REEs are from Sun and McDonough (1989).
occurring either as anhedral disseminated crystals and/or as inclusions in garnet, occasionally some of the anhedral crystals of columbite being associated with it (Fig. 9c). It may be slightly altered to a secondary U-bearing constituent (kasolite, Fig. 9d). The EDS results (Fig. 9c) indicate that the examined uraninites typically consist of UO$_2$ (78.37 wt%), PbO (12.18 wt%), ThO$_2$ (2.73 wt%), with minor Fe$_2$O$_3$ (1.92 wt%), CaO (1.07 wt%), SiO$_2$ (0.88 wt%) and K$_2$O (0.31 wt%). Additionally, a minor content of LREE was detected (Fig. 9c), this content including La$_2$O$_3$ (0.74 wt%), Ce$_2$O$_3$ (0.48 wt%) and Nd$_2$O$_3$ (1.32 wt%). Kasolite rarely occurs as a secondary U-bearing constituent in the Wadi Ibib pegmatites, as a secondary alteration product of uraninites. It occurs either as anhedral disseminated crystals and/or as filling voids in corroded garnet with small traces of columbite (Fig. 9c). The EDS analysis (Fig. 9d) shows that kasolite essentially consists of UO$_2$ (47.79 wt%), PbO (36.41 wt%), SiO$_2$ (0.88 wt%), with minor K$_2$O (3.56 wt%) and ThO$_2$ (0.92 wt%). Zircon mostly occurs as disseminated euhedral prismatic to subhedral grains associated with garnet-rich segregations, occasionally occurring as inclusions in garnet (Fig. 9e–f). The EDS results (Fig. 9e) show that the zircons typically consist of ZrO$_2$ (48.68 wt%), SiO$_2$ (20.22 wt%), ThO$_2$ (18.77 wt%), UO$_2$ (6.51 wt%), CaO (3.12 wt%) and HfO$_2$ (2.70 wt%). This result shows high Th- and U-contents, reflecting a Th-rich metamict zircon. Monazite is rare in the Wadi Ibib pegmatites, occurring either as minute anhedral discrete crystals or as inclusions within garnet (Fig. 9f). On the basis of EDS data (Fig. 9f), monazite consists of P$_2$O$_5$ (21.92 wt%), Ce$_2$O$_3$ (30.79 wt%), La$_2$O$_3$ (16.36 wt%), Nd$_2$O$_3$ (9.58 wt%) and Pr$_2$O$_3$ (2.98 wt%), with minor SiO$_2$ (1.60 wt%) and CaO (0.72 wt%). However, it shows high contents of ThO$_2$ (13.01 wt%) and UO$_2$ (3.04 wt%), which in turn reflects metamict monazite. Columbite-tantalite occurs either as anhedral small disseminated crystals and/or as inclusions in garnet (Fig. 9g–h).

The geochemical characteristics of the pegmatite samples show high U-concentrations (up to 573 ppm) and Th-concentrations (up to 232 ppm) (Supp. Table 1). However, the pegmatites show a very high concentration of REEs (up to 1354.96 ppm) compared to their U- and Th-contents (Supp. Table 2). Pegmatite samples show a moderate correlation between U and Th ($R = 0.63$; Fig. 10a) and have low Th/U ratio values, mostly less than unity (0.38–0.75; Supp. Table 2). The Th/U ratio is weakly correlated with silica contents ($R = -0.14$), however this ratio tends to decrease with increases in the SiO$_2$ content (Fig. 10b), which likely reflects the high concentrations of U and Th phases in the pegmatitic melt (e.g., McKeough et al., 2013). The total REE-content in the pegmatites shows a strong correlation with the U-content ($R = 0.83$; Fig. 10c), which likely reflects the high concentrations of U and Th phases in the pegmatitic melt (e.g., xenotime). These strong and weak correlations may be due to the level of abundance of U- and Th-bearing phases (e.g., uraninite and zircon) associated with the REE-bearing phase (e.g., xenotime).

5 Discussion

5.1 Tectonic emplacement controls

Although the Wadi Ibib pegmatite was emplaced into syn-tectonic structures (D3), this does not necessarily
imply that their formation age is congruent with those host structures. However, the Wadi Ibib pegmatites equally show clear evidence supporting their syn- to late-tectonic emplacement (i.e., shear zone-supported melt transfer) rather than dike-opening in the post-tectonic conditions. In the Wadi Ibib area, structural controls in the pegmatite occurrences have been observed at various scales, from map to outcrop (Figs. 1b–c, 3–4). Pegmatites were emplaced within a N to NNE-trending regional-scale shear zone and are strike-parallel to the shearing axis (i.e., HSZ), likely acting as a basic control on pegmatite distribution in the Wadi Ibib area.

Structural analysis of the mineralized site and the surroundings indicates a predominantly transpressional deformation event (D3), with a paleostress field of an E–W-trending σ1 (plunging moderately to the W), N–S-trending σ3 and σ2 perpendicular to Earth’s surface (Fig. 5a–d). During the late stage of the D3 event, a permutation of their stress axes has occurred, the emplacement of the Wadi Ibib pegmatite having taken place along N-trending dextral shearing. The regime has switched to pure extension, where its stress axes have permuted from an E–W directed σ1 to an E–W directed σ3, with σ2 in a N–S trend (Fig. 5e). In addition to structural analysis, this hypothesis (stress axes permutation) can be supported by: 1) an increase in the confining pressure caused by a major crustal thickening related to D3-tectonics, which is followed by a decrease in collision intensity (system pressure-release), thus the field itself tends to be inverted and transformed into an extensional regime (e.g., Saintot and Angelier, 2002), towards the end of a D3-tectonic phase; 2) a local magmatic stress field occurred due to magma-squeezing during the emplacement of granodiorite and terrain uplifting, this field likely supporting the regime reflection, thus the extensional structures have developed.

The pegmatite modes of occurrence are typically parallel to those of the host rock D3-tectonic fabrics e.g., F3-axial plane, S3-mylonitic foliation, L3-lineation and NS-dextral shearing (Figs. 4–5). This shows that their emplacement was controlled by D3-tectonics, rather than dike-opening in post-tectonic extension conditions. Additionally, the pegmatites show no evidence of cross-cut field relationships with D4-structures (post-tectonic fabrics). In contrast, they show evidence indicating that they themselves have been transected by D4-tectonic structures (Table 1; Figs. 3–4).

Kinematic indicators identified in both pegmatites and metasedimentary host rocks support syn-to late-shearing characteristics of the Wadi Ibib pegmatites. As is evidenced by the field relationships and structural analysis, pegmatites have deformed with a well-developed set of conjugate shear fractures coeval with the D3-shearing regime (Figs. 4d–j and 5d). The occurrence of asymmetric boudinaged pegmatites within the sub-vertical S3-shearing foliation points to a dextral sense of shearing (Figs. 4i and 5d). These pegmatites also contain stretched (sigmoid) aggregates of feldspar and quartz, as well as uncommon schist enclaves, being deformed and serrated in a top-to-N, parallel to the host shear zone boundaries and its S3-foliation, that further reflects a dextral sense of...
shear (Fig. 6a).

5.2 Mineralogical and geochemical constraints

According to the microscopic studies, the common mineralogy of the pegmatites is distinguished by the predominance of albitic plagioclase, microcline, quartz, muscovite and abundant accessories with high amounts of garnet. There is a mineralogical similarity in the accessory phases between the pegmatites and their metapelite host rocks which have a high amount of garnet (Figs. 4 and 7). In contrast, there is a distinct mineralogical difference found in accessory phases between these pegmatites and those in the surrounding granitoids, which are typically garnet-free. This may suggest that the mineralogical composition of the pegmatites, especially their accessory phases, are controlled by a partially-melted material (e.g., metapelites), rather than by evolved granitic magmas (e.g., Bea and Montero, 1999; Baba et al., 2012).

The metapelite host rocks are peraluminous, have high major oxide contents such as Al₂O₃, TiO₂, MnO, FeO, MgO, CaO and P₂O₅, with low silica content, as well as notable enrichment in Cr, V, Sr, Rb, Zr, Y and Nb. These amphibolite facies rocks contain hornblende, biotite, muscovite, garnet, zircon and monazite (Ali-Bik et al., 2014), which are considered to be significant reservoirs of incompatible elements (REE and HFSE) in the metamorphic and magmatic systems (Bea, 1996). The abundance of these mineralogical phases in the metapelites, coupled with their geochemical characteristics, points to them likely being fertile enough to be a possible source for partial melting at depth, producing the pegmatite in the Wadi Ibib area (e.g., Bea and Montero, 1999).

Traditionally, pegmatites are regarded as a continued fractionation of parent granitic magma on a plutonic-scale (Jahns and Burnham, 1969; Černý, 1989), a model that has widely been applied for most studied pegmatites in the Egyptian Eastern Desert (e.g., Abd El-Naby and Saleh, 2003; Ghazaly et al., 2015; and others). However, a growing number of worldwide studies evaluate pegmatites as a partial melting product, rather than as a final fractionation product of a parent granitic pluton (e.g., Lentz, 1996; McKeough et al., 2013). McKeough et al. (2013) pointed out that the abundance of rare elements in pegmatites should be statistically correlated with higher hybridization index (CaO + MgO + FeO) values, which reflect the degree of the host rock assimilation. However, these correlations have not been recorded in the Wadi Ibib pegmatites (Fig. 8e–f), by contrast they show low index values, reflecting an absence of host rock assimilation. This may be further supported by field evidence, such as the absence of gradational and/ or irregular contacts of the pegmatites against their host rock (Fig. 4).

5.3 Source controls on the pegmatite mineralization

The U, Th, REE and Nb-Ta mineralization in the Wadi Ibib pegmatites are controlled by abundant xenotime, uraniumite, kasolite and zircon with monazite and columbite-tantalite accessory phases. This reflects the composition of the source materials, from which the pegmatites have evolved (e.g., Novák et al., 2012).

Field relationships, mineralogy and bulk-chemistry of the Wadi Ibib pegmatites point to their likely formation by a localized partial melting of a garnet-bearing metapelitic source.

During the D3-tectonic event, the Wadi Ibib-HSZ area underwent an amphibolite facies metamorphism in response to intense collision (crustal thickening) and heat provided by terrain uplifting and shearing. The Wadi Ibib-HSZ rocks attained a peak temperature and pressure of about 600 ± 50°C and 5.0–6.5 kbar, during the retrogressive (cooling stage) reaching about 530°C (Ali-Bik et al., 2014). This syn-deformation thermal event supplied enough heat to create a local small-scale partial melting zone that had perhaps produced the pegmatites by the end of this event (e.g., Hibbard, 1987; Jamie et al., 2013). Accordingly, this process introduced and localized fluids into the pegmatitic-system at depth, where the retrograde metapelitic rocks had an abundance of hydrous mineral constituents (Ali-Bik et al., 2014). This decompression-melting process led to a local breakdown of phases e.g., hornblende, biotite, muscovite and garnet
(e.g., Vance and Mahar, 1998; Bea and Montero, 1999), thus the partial-melting process generated a melt enhanced by rare-metals, HFSE and HREE. Xenotimes may have grown early at the expense of partial garnet breakdown, however the melt may still have had reactants (biotite + muscovite + plagioclase + quartz) to produce more garnets (e.g., Vance and Mahar, 1998; Foster et al., 2000), which were crystallized with well-developed inclusion-rich xenotime, zircon, monazite and columbite (Fig. 9).

Garnets occur as anhedral unzoned grains (Fig. 9), which may reflect their rapid growth (Dahlquist et al., 2007). This points to a limited in-situ fractional crystallization of the pegmatites (e.g., Dahlquist et al., 2007), which further indicates that the Wadi Ibib pegmatite has rapidly ascended from its source level (anataxis situ) to its current erosional position. This may be attributed to an accelerated rate of escape tectonics (stress axes permutation) and/or the abundant P, F and H$_2$O in the pegmatitic melt, which in turn decreased both its viscosity and the crystallization temperature (Manning, 1981; Nabalek et al., 2010).

Pegmatites also have elevated concentrations of U and Th, in addition to REE and some rare metals (Supp. Table 2; Figs. 7–10). As identified by Robb (2005), to attain around 300 ppm of U in a leucogranite partial melt, the partial melting degree should be very low (<5%) or the source material (protolith) should be enriched in U-Th-content (>10 ppm). The Wadi Ibib pegmatites have high concentrations of U and Th, up to 573 and 232 ppm, respectively (Supp. Table 2). Furthermore, the metasedimentary host rock (metapelites and quartzites) itself contains a slight concentration of U and Th, especially the quartzites (Supp. Table 1).

Hence, the pegmatites are likely derived from a fertile enough garnet-bearing pelitic source through low partial melting degrees (e.g., Nex et al., 2001) of a local limited-scale anatexitic event in the Wadi Ibib area. Conversely, the absence of host rock assimilation and/or secondary hydrothermal-bearing phases in the Wadi Ibib pegmatites, except for some traces of kasolite, point to the mineralization being controlled by internal fractionation processes (syngenetic), likely without an external addition to the pegmatitic system.

5.4 Pegmatitic origin and mechanism

As discussed earlier, although there are many granitic plutons in the Wadi Ibib area, they are unlikely to be genetically related to the investigated pegmatites. However, if it is reasonable to not ignore the possibility of the existence of an undetected granitic pluton at depth, linked to the pegmatites, then it is equally reasonable to suppose that they have been produced by a limited-scale anatexitic. This is dependent on the pegmatite setting, which is obviously concordant with the anatectic origin in the study area.

To enrich and achieve a better understanding of the Wadi Ibib pegmatitic origin and their mechanism of formation, a petrogenetic model has been constructed (Fig. 11), according to the preceding results and discussions. The model characterizes key controls on the pegmatites, e.g., field relationships, tectonic setting, mineralogical-geochemical features, as well as the absence of any relationship with the granitoid outcrops in the Wadi Ibib area.

Fig. 11 is a proposed multiphasic model for the Wadi Ibib pegmatites, illustrating four phases as a mechanism of formation. Phase (1) is interpreted as the evaporation of heat and fluids from depth, as a result of a crustal thickening and extensional amphibolite facies metamorphism associated with the early D3 (?) deep-seated tectonic event. The early D3 is characterized by an E–W σ1 and N–S σ3 transpressional stress-field (Fig. 11). This field likely leads to the evaporation and release of heat and fluids from the higher-grade metamorphic rocks at great depths (system-diaphoresis) into less-strongly metamorphosed rocks (e.g., Dill, 2016) synchronous with shearing and faulting along the HSZ. Phase (2) represents a short-lived localized anatectic event between the metamorphic peak of 600 ± 50°C and 5.0–6.5 kbar (Ali-Bik et al., 2014) and the emplacement of the pegmatites. This short-lived anatectic event (i.e., rapid start-stop system) may be attributed to an accelerated rate of escape tectonic (D3),
which in turn localized and accelerated the pegmatite emplacement. During the D3-tectonic event, the partial melting of a garnet-bearing metapelitic source has occurred as a result of decompression, associated with escape tectonics-stress axes permutation (system pressure-release). All these factors led to its production of a melt, introducing volatiles and the breakdown of minerals acting as reservoirs for incompatible elements (HFSE and REE), such as garnet, hornblende, muscovite and biotite. The absence of host rock assimilation indicates that melting has taken place at depth in the Wadi Ibib area, likely subjected to further fractionation during its ascent until it attained its emplacement position. Phase (3) is interpreted as a fractionation phase. At this phase, a cooling retrogressive stage (post-peak), which reached about 530°C in the Wadi Ibib area (Ali-Bik et al., 2014), the pegmatite fluid pressures allowing the partial melt to ascend upwards in the crust. The tectonic regime turned into pure extension as a result of the stress axes permutation during this stage. When the stress-field switched from an E–W directed σ1 to an E–W directed σ3 and σ2 perpendicular to Earth's surface (Fig. 11), the melts were transported over pathways (S3-foliation and faults), attaining the shallow crustal level before their solidification. During the melt ascent, the mineralization accessory phases (U-, Th- and REE-bearing) were evolved without external additions to the system. It is possible that the melt kept its characteristics, i.e., without external additions to the system, due to the melt being intruded into only one host rock type (no lithology contrast in the host) and reached their emplacement position rapidly. Phase (4) represents the emplacement of the Wadi Ibib pegmatites at their present-day erosional level, as a result of attaining their crystallization point.

6 Conclusions

The mineralized pegmatites in the Wadi Ibib area only intrude the garnet-bearing metapelite rocks along the shear plane of the HSZ, corresponding to the regional N-S-trending D3-tectonic fabrics of their host. These pegmatites are distinguished by the dominance of albite-plagioclase, muscovite and garnet, with accessory phases including xenotime, uraninite, kasolite, zircon with monazite and columbite. They are peraluminous, showing S-type characteristics, having low K/Rb ratios and significant concentrations of U, Th, REE, Rb, Li, Cs, Y, Nb and Ta, likely indicating a NYF-pegmatites signature. There is an absence of host rocks assimilation and a rare hydrothermal secondary mineralization-bearing phase (epigenetic). Thus, the mineralization seems to be controlled by an internal fractionation process in the Wadi Ibib pegmatites and is considered to be syngenetic in origin.

Field relationships and structural analysis, combined with the geochemical signatures of the Wadi Ibib pegmatites, do not support their evolution by an extensive fractionation of the surrounding granitoids, but indicate that they have an anatectic origin. This anatectic event was likely a short-lived limited-scale partial melting of a peraluminous garnet-bearing metapelitic source at a deep-seated level. It occurred as a result of decompression, associated with escape tectonics-stress axes permutation (regime switching to extension) during the syn- to late D3-tectonics.

This study to a significant extent proved the relationships between tectonics and anatexis, in the context of pegmatite origin and their mineralization potential, which emphasizes the syn-tectonic controls on the investigated anatectic pegmatite. However, the present results suggest an additional factor controlling pegmatite formation, namely the source lithologies from which the parent melt was derived. In this way, the priority in the exploration targets for rare-metal pegmatites should be given to the adjacent high-grade tectono-metamorphic setting, such as pelitic metasedimentary belts, governed by extensive amphibolite facies metamorphism.

Acknowledgements

This study was supported by a research cooperation (Grant No. 3TE1107T) between Kyushu University, Japan and the Egyptian Nuclear Materials Authority (NMA). Prof. K. Watanabe and Prof. K. Yonezu (Kyushu University, Japan) are highly thanked for allowing access to the different laboratory facilities and their valuable discussions. The author is grateful to Prof. G. Saleh and all members of the Abu Rushied project, NMA, for their assistance in the fieldwork. Many thanks to Assoc. Prof. Thomas Tindell, Kyushu University, Japan, for the improvement of the English phrasing.

Supplementary data to this article can be found online at http://doi.org/10.1111/1755-6724.14713.

References

in felsic pegmatites ultimately depend on tectonic setting.


About the first and corresponding author

Waleed Saad IBRAHIM; born in Ismailia City, Egypt. He is an associate professor of tectonics and economic geology in the Research Sector, Nuclear Materials Authority (NMA), Cairo, Egypt. He graduated in 2002 with a honors degree, obtaining his M.Sc. degree in 2009 from the Geology Department, Faculty of Science, Suez Canal University. He received his Ph.D. degree in 2014 from the Economic and Geology Laboratory, Department of Earth Resources Engineering, Graduate School of Engineering, Kyushu University, Japan. He joined the research group of the Economic Geology Laboratory and has joined many projects through this group and participated in many fieldwork activities (related to tectonics and ore geology) in Egypt, Japan, the Philippines, Mongolia and Indonesia. He has worked in different fields of Earth Sciences, including applied structural geology, ore geology, tectonic-paleostress analysis, geodynamics and remote-sensing. E-mail: dr.waleedelghazali@yahoo.com; dr.wellgeo@kyudai.jp; phone: +20-1027155133.