

Article

Petrogenesis and Mineralization Potential of the Early Silurian Syenogranites in the Yama Beryllium Deposit Within the Tataleng Granitic Batholith, Qilian Orogen, NW China: Insights from U-Pb Geochronology and Geochemistry

Yong-Liang Cao ¹, He Jiao ¹, Wei Ma ¹, Guo-Biao Huang ¹, Ming-Jun Sun ², Zhuang-Zhuang Yang ² and Qing-Feng Ding ^{2,*} 

¹ Qinghai Provincial Key Laboratory of Salt Lake Resources Exploration and Research in Qaidam Basin, Qaidam Integrated Geological Exploration Institute of Qinghai Province, Golmud 816099, China

² College of Earth Sciences, Jilin University, Changchun 130061, China

* Correspondence: dingqf@jlu.edu.cn

Abstract

Recently, significant beryllium mineralization was discovered along faults within syenogranites in the Yama area of the Qilian Orogen, NW China. This study presents a comprehensive analysis of the geochemical compositions of these syenogranites, along with zircon U–Pb geochronology, Hf isotopic data, and trace element analyses. Barren and beryllium-bearing syenogranites yielded weighted mean $^{206}\text{Pb}/^{238}\text{U}$ zircon ages of 437.7 ± 3.3 Ma and 432.0 ± 2.3 Ma, respectively, indicating a consistent emplacement age in the early Silurian. Their geochemical characteristics are slightly peraluminous, with high alkali contents; relatively low $(\text{K}_2\text{O} + \text{Na}_2\text{O})/\text{CaO}$ and $\text{Fe}_2\text{O}_3^{\text{T}}/\text{MgO}$ ratios, low concentrations of $\text{Zr} + \text{Nb} + \text{Ce} + \text{Y}$, strongly negative Eu anomalies, high Rb/Sr and Rb/Ba ratios, and negative correlations between Rb and Th, all indicative of fractionated S-type syenogranites. Furthermore, geochemical data combined with negative zircon $\varepsilon_{\text{Hf}}(\text{t})$ values suggest derivation from the partial melting of Paleoproterozoic pelitic-rich protoliths within a relatively shallow, reworked lower crust during an early Silurian syn-collisional setting. Negative ΔFMQ values and low $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratios, along with relatively high La contents and $\text{Sm}_{\text{N}}/\text{La}_{\text{N}}$ ratios in a few zircon grains, as well as relatively low whole-rock Nb/Ta and Zr/Hf ratios, imply a certain initial-stage magmatic–hydrothermal beryllium-mineralization potential for the studied syenogranites.

Keywords: zircon U–Pb geochronology; petrogenesis; redox state; mineralization potential; syenogranite; Yama beryllium deposit



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1. Introduction

Beryllium is an indispensable critical and strategic metal extensively used in national defense and emerging strategic industries [1–3]. Its primary sources are two minerals: bertrandite and beryl [4], which constitute the main ore minerals in beryllium deposits [1–3,5,6]. Most beryllium deposits are typically associated with geological settings influenced by magmatic–hydrothermal processes and are commonly classified as magmatic-related beryllium deposits [1]. These deposits predominantly include beryllium-bearing rare-metal pegmatite deposits [7–9], skarn- and greisen-hosted beryllium–fluorine deposits [3,10–13],

volcanogenic-hosted beryllium deposits [5,14,15], and carbonate-hosted beryllium deposits [16,17]. Furthermore, beryllium occurrences have been identified in various nonmagmatic hydrothermal, metamorphic, and sedimentary environments, collectively termed nonmagmatic-related beryllium deposits; however, most of these are not economically exploitable [1]. Given the increasing demand for beryllium resources, further investigation into the metallogenic processes governing the formation of diverse beryllium deposit types is imperative.

In recent years, notable geochemical anomalies involving fluorine, tin, beryllium, tungsten, and lithium have been identified within the Tataleng granitic batholith, located in the South Qilian Belt (SQB) of the Qilian Orogen (QLO), NW China (Figure 1, ref. [18]). These anomalies indicate significant potential for the mineralization of these elements [18–23]. In addition to the previously documented Yamatu–Yansenhada quartz vein deposit, Yama fluorite deposit, Gariligen gold occurrence, and several Au–Cu–Pb–Zn occurrences [18–20,24], recent investigations have revealed significant tin and beryllium mineralizations closely associated with specific granitic rocks in the Yama, East Elikendagou, and West Datouyang exploration zones within the Tataleng batholith (Figure 2, refs. [21,23,25–27]). Nevertheless, comprehensive geological and geochemical studies focusing on the tin and beryllium mineralizations and their related granitoids remain insufficient. To date, only the mineralization age (435–432 Ma) and petrogenetic characteristics of the syenogranites associated with the East Elikendagou tin deposits have been partially documented [23,25,27], whereas research on the beryllium deposits remains largely lacking.

The Yama fluorite deposit, located at the eastern boundary of the Tataleng granitic batholith, constitutes a significant fluorite resource within the SQB [18–20]. Recent studies have identified beryllium mineralization along faults within the syenogranite in the northwestern part of the Yama area [19,28]. Previous investigations have briefly assessed the fluorite mineralization potential and petrogenetic age of the Wulandawu granitic host rock adjacent to the fluorite veins in the northwest Yama region, concluding that this granitic unit contributes almost nothing to the fluorite mineralization [18]. Nevertheless, comprehensive and systematic research focusing on beryllium mineralization and the petrogenesis of the beryllium-associated syenogranite in the northwest Yama area remains lacking.

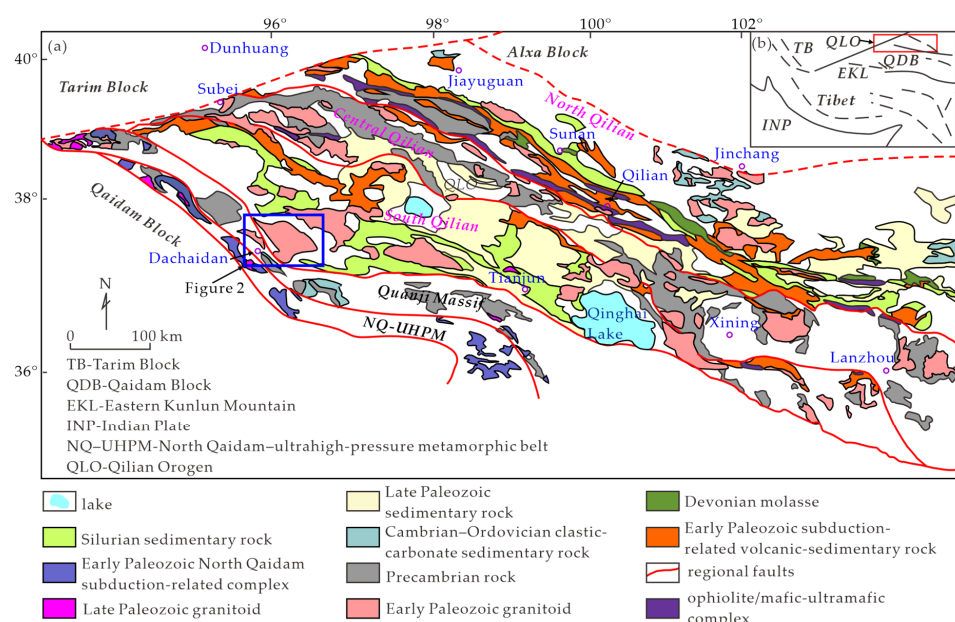


Figure 1. (a) Regional geological map of the Qilian Orogen and (b) schematic map showing major tectonic units in China (revised from ref. [29]).

This study presents in situ zircon U–Pb ages, trace element compositions, and Lu–Hf isotopic data, along with whole-rock major and trace element compositions of the syenogranites hosting beryllium mineralization in the northwest Yama area, located within the Tataleng granitic batholith of the SQB. The primary objectives are to elucidate the geochronology, petrogenetic classification, petrogenesis, redox state, and tectonic significance of the investigated syenogranite, as well as to evaluate its potential for beryllium mineralization.

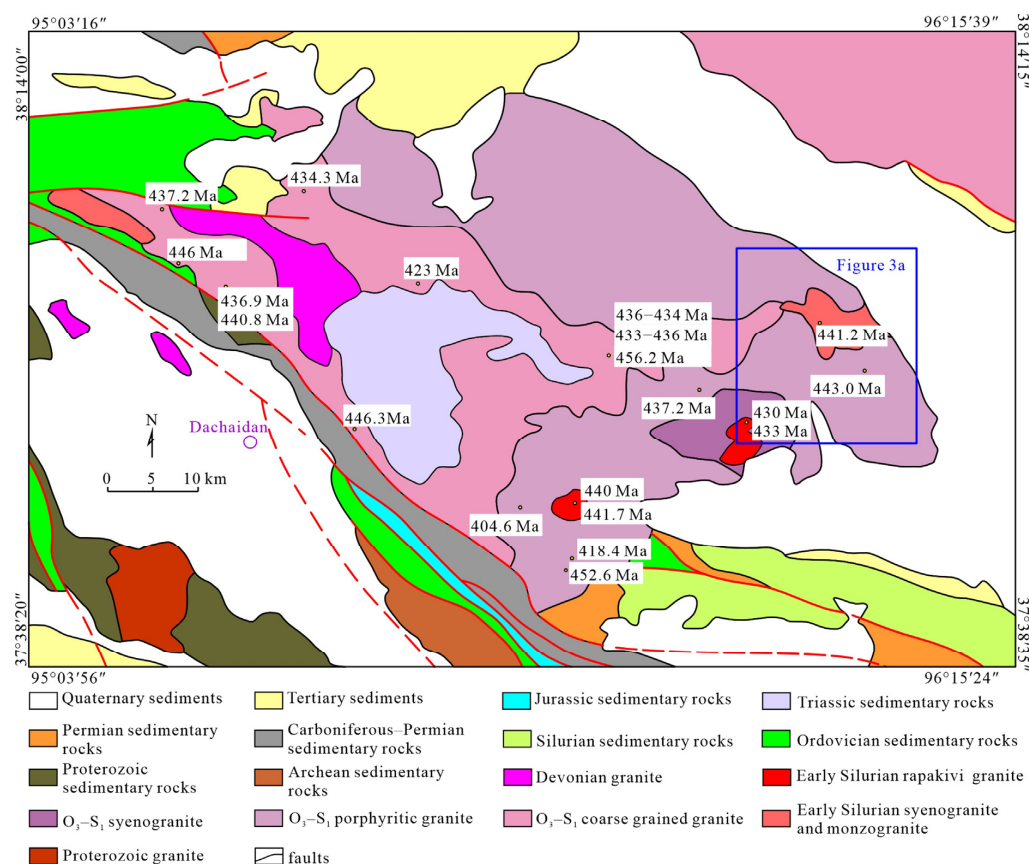


Figure 2. Tataleng granite batholith in the Southern Qilian Belt (revised by ref. [19]), showing the extent of the East Elikendagou area. O₃–S₁: the late Ordovician to early Silurian. Published ages are cited from ref. [18,22,23,25,27,30–38].

2. Geological Background and Geology of the Studied Syenogranites

2.1. Geological Background

The Qilian Orogen (QLO), extending approximately 350 km in width and 1200 km in length, represents the northeastern boundary of the Qinghai–Tibet Plateau. Tectonically, it is positioned between the Alxa Block to the north and the Qianji Massif (QM) to the south (Figure 1, refs. [29,39–42]. The QLO is extensively acknowledged for its well-preserved record of Early to Middle Paleozoic subduction, collision, and accretion events linked to the closure of the Proto-Tethys Ocean [41,43–46]. Conventionally, the QLO is divided from north to south into three principal tectonic units: the North Qilian Belt (NQB), the Central Qilian Block (CQB), and the South Qilian Belt (SQB). This subdivision is based on variations in basement rocks, lithostratigraphic sequences, and structural features (Figure 1a, refs. [29,39,43,44]).

The NQB constitutes the most thoroughly investigated portion of the QLO [43]. It is primarily composed of Early Paleozoic ophiolitic sequences, high-pressure metamorphic belts, and island-arc-related volcanic rocks and granitic plutons. These units are overlain

by Silurian flysch deposits and Devonian molasse formations, which are subsequently covered by Carboniferous to Triassic sedimentary sequences [29,30,44,47–49]. The CQB predominantly comprises Precambrian basement rocks overlain by volcanic sequences from the Cambrian to Ordovician periods, as well as Paleozoic sedimentary strata and Early Paleozoic granitic intrusions [29,48,50]. The SQB includes the Neoproterozoic Hualong Complex microcontinent, early Paleozoic volcanic–sedimentary sequences, and extensive emplacement of early Paleozoic arc-related granitic bodies [29,30,39,46,51].

A substantial body of research has focused on numerous early Paleozoic granitoids within the SQB, as these granitoids are frequently interpreted to be linked with magmatic events associated with subduction, collisional, or post-collisional processes related to the Proto-Tethys Ocean [18,22,23,30,43,52–54]. It is widely accepted that the South Qilian oceanic plate underwent northward subduction beneath the SQB [18,22,23,30,39,44,47,52,55,56]. The initiation of this early Paleozoic subduction is evidenced by the presence of the Tanjianshan arc basalts, dated to approximately 540 Ma, located along the southern margin of the QLO [57,58]. Subsequently, a series of early Paleozoic granitoids, including the extensive Tataleng granitic batholith (Figure 2), the Delingha pluton, the Huangyuan pluton, and several smaller granitic intrusions, were emplaced between approximately 450 Ma and 418 Ma. These granitoids are interpreted to have formed in a syn-collisional tectonic setting subsequent to the subduction of the South Qilian oceanic plate [18,22,23,47,53,55,58–61].

2.2. Geology of the Studied Syenogranites

The Yama study area is situated in the eastern sector of the Tataleng granitic batholith (Figure 2). Previous studies have provided concise descriptions of the geological characteristics of the Tataleng granitic batholith [18,22,23,30]. This batholith is bounded to the southwest by northwest-oriented ductile faults adjacent to Carboniferous to Permian sedimentary sequences and is centrally overlain by Triassic sedimentary strata (Figure 2) [18,22,23,30]. It is widely recognized that the extensive Tataleng granitic batholith underwent a prolonged crystallization history, predominantly from the Late Ordovician to the Early Silurian (O_3 – S_1) [18,22,23,25,27,30–35,41].

In the Yama area, faults oriented northeast and northwest have developed within extensive O_3 – S_1 granitic rocks. These faults are predominantly filled with numerous barren quartz veins and fluorite–quartz veins (Figure 3a). The fluorite–quartz veins are mainly concentrated in the central part of Yama, with minor occurrences in the northwest and southeast sectors. In total, 81 fluorite–quartz veins have been identified, collectively contributing to a substantial inferred fluorite reserve within the Yama fluorite deposit [19]. More recently, significant additional beryllium mineralization has been discovered along the northwest-oriented faults in the northwestern sector of the Yama area, designated as the Yama beryllium deposit (Figure 3b) [28]. The host rock for this beryllium mineralization is syenogranite (Figure 3b). Our previous investigations briefly examined the syenogranite located near the southeastern segment of the fluorite-bearing fault in the study area, known as the Wulandawu syenogranite (Figure 3b), and concluded that the Wulandawu syenogranite likely plays a minimal role in fluorite mineralization [18]. However, studies focusing on the syenogranite adjacent to the northern segment of the fluorite–beryllium-bearing fault and its associated potential for beryllium mineralization remain insufficient.

Within the medium- to coarse-grained syenogranite host rocks, hereafter referred to as barren syenogranite, the MBe 1 and MBe 2 beryllium ore bodies have been delineated along the northwest segment of the fluorite–beryllium-bearing fault (Figure 3b) using a BeO cutoff grade of 0.04%. The ores in these beryllium ore bodies primarily consist of limonitic fault breccia (Figure 4a) and relatively unoxidized fault breccia (Figure 4b) within fault zones. The MBe 3 beryllium ore body has been delineated within the altered syenogranite

near the northwest segment of the fluorite–beryllium-bearing fault, hereafter referred to as beryllium-bearing syenogranite (Figure 3b). Ores in this type of ore body primarily consist of bleached beryllium-bearing syenogranites (Figure 4c) near the surface and relatively unbleached beryllium-bearing syenogranites underground (Figure 4j,m). These three ore bodies range from 120 to 640 m in length and 0.90 to 9.21 m in width, exhibiting BeO grades between 0.059% and 1.153%. Bertrandite ($\text{Be}_4[\text{Si}_2\text{O}_7](\text{OH})_2$) was identified as the principal beryllium-bearing mineral [62]. We did not detect bertrandite in the Yama beryllium deposit under the microscope. It was only after sending the beryllium ore samples for mineral separation tests (which are beyond the scope of this study) and using an automated process mineralogy parameter testing system that we identified bertrandite as the primary beryllium-bearing mineral [62]. Despite preparing numerous thin sections, we were unable to observe any beryllium minerals within them, likely due to the low beryllium content, which may result in only trace amounts of bertrandite, making slicing and cutting challenging. However, since all beryllium-bearing syenogranite exhibits alterations such as silicification, sericitization, and albitization within these faulted zones or in the altered beryllium-bearing syenogranite [62], we infer that bertrandite is of hydrothermal origin.

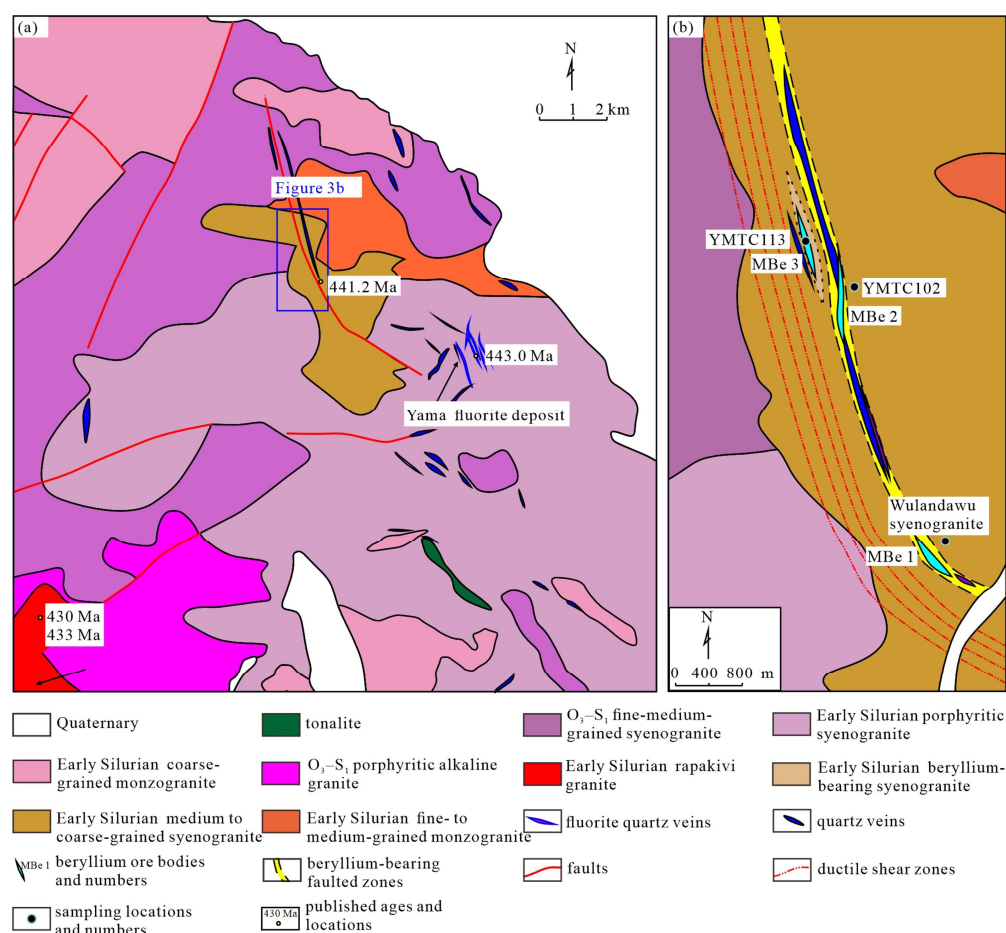


Figure 3. (a) Geological map of the eastern segment of the Tataleng granitic batholith; (b) geological map of the Yama beryllium deposit, showing sample locations. Simplified and revised after [62]. $\text{O}_3\text{--S}_1$: the late Ordovician to early Silurian. Published ages are cited from ref. [18,22].

The barren syenogranite is characterized by a typical medium- to coarse-grained granitic texture and massive structure (Figure 4d,g). Its mineralogical composition includes 45%–65% alkali feldspar, predominantly perthite with minor orthoclase, 5%–10% plagioclase, 20%–30% quartz, and approximately 5% muscovite, accompanied by minor accessory minerals such as zircon and epidote (Figure 4e,f,h,i). It is termed syenogranite

because its alkali feldspar content is significantly greater than that of plagioclase. The perthite grains are subhedral, measuring 1–4 mm in length, commonly exhibiting striped twins and localized sericitization. Plagioclase is subhedral to euhedral, with lengths of 0.5–2.0 mm, usually exhibiting polysynthetic twins and localized sericitization. Quartz grains are anhedral, with lengths of 0.5–2.0 mm. Additionally, the syenogranite shows localized sericitization and muscovitization of feldspar (Figure 4e,f,h,i).

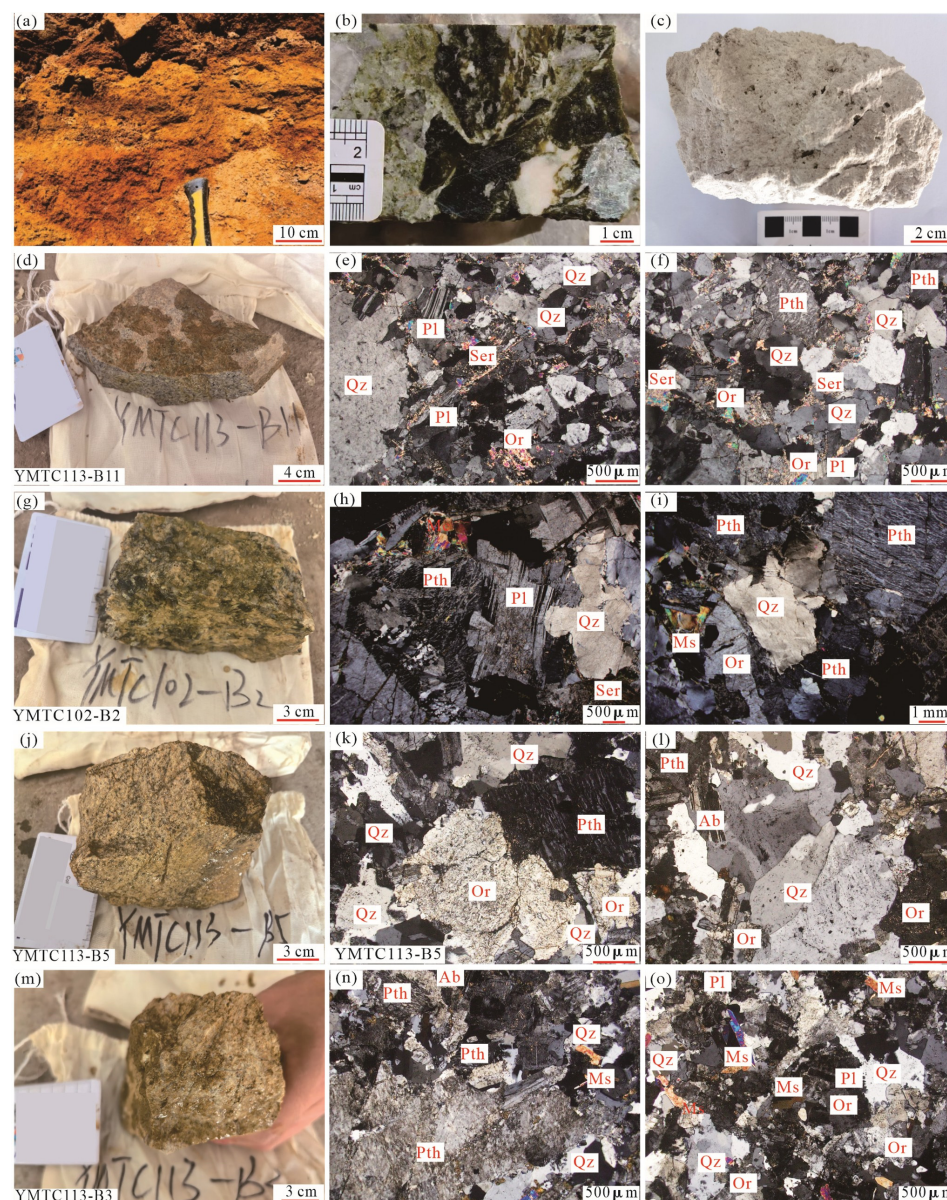


Figure 4. Typical specimen photographs and photomicrographs of the studied ore and syenogranite samples from the Yama beryllium deposit. Specimen photographs for ores from limonitic fault breccia (a), unoxidized fault breccia (b), and bleached beryllium-bearing syenogranites (c). Specimen photographs (d,g) correspond to samples YMTc113-B11 and YMTc102-B2 from the barren syenogranite, respectively. Photomicrographs under cross-polarized light (e,f) and (h,i) correspond to samples YMTc113-B11 and YMTc102-B2 from the barren syenogranite, respectively. Specimen photographs (j,m) correspond to samples YMTc113-B5 and YMTc113-B3 from the beryllium-bearing syenogranite, respectively. Photomicrographs under cross-polarized light (k,l), and (n,o) correspond to samples YMTc113-B5 and YMTc113-B3 from the beryllium-bearing syenogranite, respectively. Abbreviations: Qz—quartz, Or—orthoclase, Pth—perthite, Pl—plagioclase, Ab—albite, Ser—sericite, and Ms—muscovite.

The beryllium-bearing syenogranite similarly exhibits a typical medium- to coarse-grained granitic texture and massive structure (Figure 4j,m). Its mineral assemblage consists of 40%–55% alkali feldspar, primarily perthite with minor orthoclase, 5%–10% plagioclase, 20%–30% quartz, and 5% muscovite, along with minor accessory minerals such as zircon and epidote (Figure 4k,l,n,o). The perthite grains are subhedral, measuring 1–3 mm in length, and commonly exhibit striped twins as well as localized sericitization and kaolinization. Plagioclase is subhedral to euhedral, with lengths of 0.5–1.0 mm, usually exhibiting polysynthetic twins and localized sericitization. Quartz grains are anhedral, with lengths of 0.5–1.5 mm. The altered beryllium-bearing syenogranite shows localized to pervasive sericitization and muscovitization of feldspar, with minor albitization of perthite (Figure 4k,l,n,o).

3. Sampling and Analytical Methods

3.1. Sampling

For this study, four barren syenogranite samples and six beryllium-bearing syenogranite samples were collected from east to west along near east–west trenches at sampling sites YMTC113 and YMTC102, located within trenches in the Yama beryllium deposit (Figure 3b). Initially, the specimens were sectioned and polished into thin sections for petrographic examination. The remaining portions of these samples were then crushed, and conventional magnetic and heavy liquid separation methods were used to extract zircon minerals. Zircon grains were carefully handpicked under a binocular microscope, embedded in epoxy resin, and polished in preparation for subsequent analytical procedures.

3.2. CL Imaging of Zircons

Zircon grains were analyzed using transmitted light microscopy and cathodoluminescence (CL) imaging to investigate their internal features. The CL images were acquired at Langfang Chengxin Geology Survey Co., Ltd., Langfang, Hebei, China, employing a TESCAN Panchromatic CL detector integrated with a TESCAN MIRA 3 LMH scanning electron microscope (TESCAN Group, Brno, Czech Republic).

3.3. LA-ICP-MS Zircon U–Pb Dating and Trace Element Analysis

U–Pb zircon dating analyses were performed at Beijing Createch Testing Technology Co., Ltd., Daxing, Beijing, China, employing Laser Ablation–Inductively Coupled Plasma–Mass Spectrometry (LA-ICP-MS) with an Analytik Jena PQMS Elite ICP-MS (Inductively Coupled Plasma–Mass Spectrometry) system (Analytik Jena, Jena, Germany) coupled to a RESOLUTION 193 nm ArF excimer laser ablation system (Applied Spectra, Inc., West Sacramento, CA, USA). Detailed operational parameters for both the laser ablation system and the ICP-MS equipment, along with data reduction procedures, are documented by Jiao et al. [23]. Each analytical run consisted of approximately 15–20 s of background (gas blank) acquisition, followed by 45 s of sample signal collection. Post-acquisition data processing, including raw data selection, integration of background and analytical signals, correction for instrumental time drift, and quantitative calibration for U–Pb dating, was conducted using LADR_1.1.07 software [63]. Zircon GJ-1 was utilized as the primary reference standard, while zircon Plesovice served as the secondary standard for U–Pb age calibration. The Plesovice zircon yielded a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 337.0 ± 2.0 Ma ($n = 17$, 2σ), consistent with the recommended $^{206}\text{Pb}/^{238}\text{U}$ age of 337.13 ± 0.37 Ma (2σ) reported by Sláma et al. [64]. For trace element analyses, NIST SRM 610 was used as the primary external standard, with zirconium (Zr) employed as the internal standard. Instrumental drift in U–Th–Pb isotope ratios was corrected every 10 analyses by linear interpolation based on variations observed in zircon GJ-1. Propagated uncertainties associated with the

recommended values of the primary reference material GJ-1 were incorporated into the final sample age determinations [65]. Correction for common lead was deemed unnecessary due to negligible common ^{204}Pb and the high $^{206}\text{Pb}/^{204}\text{Pb}$ ratio observed in the analyzed zircon grains. Concordia diagrams and weighted mean age calculations were performed using IsoplotR software (version 7.0) [66].

3.4. Lu–Hf Isotopes in Zircons

Hf isotope analyses of zircon, previously dated via LA–ICP–MS U–Pb methods, were conducted in situ at Beijing Createch Testing Technology Co., Ltd. using Multi-Collector Inductively Coupled Plasma Mass Spectrometry (MC–ICP–MS) equipped with a Neptune Plus coupled to a RESOLUTION SE 193 nm ArF excimer laser ablation system (Thermo Fisher Scientific, Waltham, MA, USA). Detailed instrumental parameters and data acquisition protocols are provided by Jiao et al. [23]. For this study, a stationary laser spot with a diameter of 30 μm , an energy density of 6 J/cm², and a frequency of 8 Hz was employed. Helium was used as the carrier gas to transport the ablated sample aerosol, which was subsequently mixed with argon, from the laser-ablation cell to the ICP–MS torch. Corrections for isobaric interferences from Lu and Yb were applied following the methodology outlined by Hou et al. [67]. Zircon standard GJ-1 (with a $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.282003 ± 0.000018 (2σ) [68]) served as the primary reference material, yielding a weighted mean $^{176}\text{Hf}/^{177}\text{Hf}$ ratio of 0.282008 ± 0.000027 (2σ , $n = 10$). The procedures for calculating $\varepsilon_{\text{Hf}}(t)$ values, depleted mantle model age (T_{DM1}), and the two-stage mantle extraction model age (T_{DM2}) are described by Jiao et al. [23].

3.5. Major and Trace Elements in Whole Rock Samples

The major elemental compositions of the samples were determined through X-ray fluorescence spectroscopy (XRF) at ALS Chemex Co., Ltd. (Guangzhou, China), utilizing the PANalytical PW5400 XRF instrument from Malvern Panalytical, Almelo, the Netherlands, the Netherlands. This analysis was complemented by a loss-on-ignition analysis conducted at 1000 °C. Trace element concentrations, excluding rare earth elements (REEs), were measured using inductively coupled plasma–atomic emission spectroscopy (ICP–AES, Agilent 5110, Agilent Technologies, Santa Clara, CA, USA) and ICP–MS (Agilent 7900 ICP–MS, Agilent Technologies, Santa Clara, CA, USA) at the same facility. The detection limit for beryllium ranges from 0.05 to 990 ppm. Appropriate corrections for spectral interelement interferences were applied to ensure data accuracy. REE compositions were specifically analyzed via ICP–MS (Agilent 7900 ICP–MS, Agilent Technologies, Santa Clara, CA, USA) at ALS Chemex Co., Ltd. (Guangzhou, China). Detailed methodological protocols for XRF, ICP–AES, and ICP–MS analyses are documented by Gou et al. [69]. The analytical precision for major element determinations is reported to be better than 5%, while the precision for trace elements and REEs is within 10% [23].

3.6. Zircon Thermometry and Oxybarometry

Zircon thermometer temperatures, including zircon saturation temperatures (T_{Zr} , °C) and Ti-in-zircon thermometer temperatures (T_{Ti} , °C), were determined following the methodologies of Watson and Harrison [70] and Ferry and Watson [71], respectively, as described by Jiao et al. [23]. The fayalite–magnetite–quartz (FMQ) mineral oxidation buffer is widely used in empirical formulations to estimate the oxidation state of magmatic systems [72]. Values of ΔFMQ (defined as $\log f\text{O}_{2(\text{sample})} - \log f\text{O}_{2(\text{FMQ})}$), as well as $\log f\text{O}_{2(\text{sample})}$ and $\log f\text{O}_{2(\text{FMQ})}$ values for zircon samples, were calculated according to the approach outlined by Loucks et al. [73]. The $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratios in zircon were computed using the method described by Ballard et al. [74], implemented in the Geo-fO₂ software [75]. Comprehensive descriptions of all zircon thermometer and oxybarometry techniques employed are provided by Jiao et al. [23].

4. Results

The results of LA–ICP–MS zircon U–Pb dating conducted on syenogranite samples from the Yama beryllium deposit are summarized in Table 1. Additionally, in situ trace element analyses of zircon are provided in Table S1. The in situ Hf isotope data for zircon, obtained concurrently with LA–ICP–MS U–Pb dating, are presented in Table 2. Comprehensive whole-rock geochemical data are compiled in Table 3.

Table 1. LA–ICP–MS U–Pb data for zircon from the syenogranite in the Yama beryllium deposit.

Spot	²³² Th	²³⁸ U	Th/U	Isotopic Ratios						Isotopic Ages (Ma)					
	(ppm)	(ppm)		²⁰⁷ Pb/ ²⁰⁶ Pb	2σ	²⁰⁷ Pb/ ²³⁵ U	2σ	²⁰⁶ Pb/ ²³⁸ U	2σ	²⁰⁷ Pb/ ²⁰⁶ Pb	2σ	²⁰⁷ Pb/ ²³⁵ U	2σ	²⁰⁶ Pb/ ²³⁸ U	2σ
YMTC113-B5 barren syenogranite															
YMTC113-B5-1	423	356	1.19	0.0576	0.0018	0.4031	0.0125	0.0507	0.0011	493.1	69.3	343.1	9.0	319.0	6.6
YMTC113-B5-2	161	331	0.49	0.0678	0.0022	1.1242	0.0369	0.1205	0.0030	844.7	66.4	763.3	17.5	733.5	17.0
YMTC113-B5-3	71	144	0.49	0.1157	0.0024	5.4458	0.1174	0.3410	0.0071	1883.7	37.1	1889.7	18.4	1891.1	33.9
YMTC113-B5-4	657	723	0.91	0.0580	0.0015	0.5290	0.0138	0.0661	0.0013	515.2	59.7	430.5	9.1	412.7	8.0
YMTC113-B5-5	56	298	0.19	0.0634	0.0076	0.3384	0.0449	0.0387	0.0012	618.9	227.8	292.9	32.6	244.5	7.6
YMTC113-B5-6	174	202	0.86	0.0599	0.0021	0.5895	0.0210	0.0711	0.0016	572.1	75.6	468.7	13.4	442.9	9.4
YMTC113-B5-7	575	477	1.20	0.0660	0.0014	1.2043	0.0315	0.1318	0.0031	794.2	46.2	800.7	14.5	797.9	17.6
YMTC113-B5-8	154	324	0.47	0.0530	0.0022	0.3024	0.0132	0.0413	0.0009	278.3	98.8	267.1	10.2	260.7	5.8
YMTC113-B5-9	66	145	0.46	0.0584	0.0029	0.5653	0.0269	0.0708	0.0017	478.2	104.0	451.7	16.8	441.1	10.2
YMTC113-B5-10	219	663	0.33	0.0555	0.0016	0.5323	0.0174	0.0695	0.0019	420.6	63.4	432.7	11.4	432.8	11.4
YMTC113-B5-11	107	107	1.00	0.0560	0.0026	0.5500	0.0257	0.0712	0.0018	362.8	117.2	441.8	16.9	443.5	10.7
YMTC113-B5-12	146	416	0.35	0.0569	0.0020	0.5429	0.0193	0.0690	0.0015	445.7	76.6	438.7	12.5	430.3	9.3
YMTC113-B5-13	122	390	0.31	0.0613	0.0021	0.5878	0.0226	0.0690	0.0016	631.5	80.7	467.9	14.4	430.0	9.8
YMTC113-B5-14	119	203	0.58	0.0646	0.0021	0.9560	0.0473	0.1064	0.0042	748.9	67.2	673.9	25.6	650.9	24.7
YMTC113-B5-15	92	130	0.71	0.1032	0.0063	1.0553	0.0593	0.0750	0.0022	1577.5	114.5	721.3	29.5	466.2	13.4
YMTC113-B5-16	169	513	0.33	0.0556	0.0013	0.5483	0.0134	0.0717	0.0015	434.1	54.4	444.3	9.2	446.6	9.1
YMTC113-B5-17	104	197	0.53	0.0563	0.0023	0.5268	0.0210	0.0685	0.0015	399.9	95.8	427.6	13.9	426.8	9.3
YMTC113-B5-18	109	200	0.54	0.0566	0.0025	0.6060	0.0256	0.0783	0.0018	430.4	97.2	478.1	16.3	485.8	11.0
YMTC113-B5-19	139	197	0.71	0.0608	0.0028	0.6205	0.0335	0.0734	0.0016	589.5	104.5	488.2	20.7	456.5	9.8
YMTC113-B5-20	108	224	0.48	0.0573	0.0019	0.5614	0.0196	0.0713	0.0016	481.7	77.4	452.4	13.3	444.2	9.6
YMTC102-B2 beryllium-bearing syenogranite															
YMTC102-B2-1	88	223	0.39	0.0551	0.0022	0.5263	0.0215	0.0694	0.0016	368.3	92.4	431.2	14.2	432.7	9.8
YMTC102-B2-2	98	222	0.44	0.0555	0.0021	0.5472	0.0198	0.0715	0.0016	384.1	82.7	441.4	13.0	445.4	9.5
YMTC102-B2-3	120	259	0.46	0.0604	0.0018	0.6359	0.0182	0.0764	0.0017	590.0	66.0	498.5	11.3	474.7	9.9
YMTC102-B2-4	178	410	0.43	0.0565	0.0015	0.5710	0.0180	0.0731	0.0017	448.3	60.7	457.3	11.6	454.8	10.2
YMTC102-B2-5	139	403	0.35	0.0607	0.0022	0.6605	0.0286	0.0786	0.0019	586.7	79.2	511.5	17.4	487.9	11.2
YMTC102-B2-6	225	789	0.29	0.0569	0.0014	0.5305	0.0126	0.0679	0.0014	473.0	52.8	431.6	8.4	423.5	8.3
YMTC102-B2-7	88	606	0.14	0.0618	0.0014	0.8499	0.0460	0.0990	0.0044	653.5	50.0	617.6	24.3	607.3	25.6
YMTC102-B2-8	61	123	0.49	0.0549	0.0028	0.4984	0.0229	0.0669	0.0016	350.9	106.8	410.0	16.2	417.7	9.7
YMTC102-B2-9	159	643	0.25	0.0572	0.0015	0.5083	0.0123	0.0650	0.0014	481.2	56.8	417.8	8.6	405.9	8.2
YMTC102-B2-10	122	386	0.32	0.0564	0.0016	0.5386	0.0151	0.0698	0.0014	442.2	62.2	436.6	9.8	434.8	8.6
YMTC102-B2-11	136	244	0.56	0.0563	0.0021	0.5212	0.0207	0.0674	0.0017	432.7	91.0	423.9	13.8	420.6	10.2
YMTC102-B2-12	150	276	0.54	0.0567	0.0056	0.5079	0.0473	0.0659	0.0018	403.6	245.5	414.8	32.3	411.2	11.1
YMTC102-B2-13	64	374	0.17	0.0884	0.0018	1.8997	0.0534	0.1566	0.0044	1383.8	37.5	1080.2	19.4	937.5	24.4
YMTC102-B2-14	114	247	0.46	0.0555	0.0021	0.5179	0.0196	0.0681	0.0015	394.7	84.3	423.7	13.6	424.9	9.0
YMTC102-B2-15	119	381	0.31	0.0553	0.0016	0.5490	0.0165	0.0719	0.0016	395.9	65.2	443.2	10.7	447.8	9.5
YMTC102-B2-16	154	429	0.36	0.0543	0.0016	0.5020	0.0139	0.0669	0.0014	355.3	65.5	412.2	9.4	417.5	8.4
YMTC102-B2-17	103	153	0.67	0.0548	0.0025	0.5468	0.0251	0.0721	0.0016	324.5	109.9	439.9	16.4	448.9	9.8
YMTC102-B2-18	142	381	0.37	0.0568	0.0018	0.5670	0.0183	0.0723	0.0016	450.5	71.1	456.2	12.2	450.0	9.5
YMTC102-B2-19	78	166	0.47	0.0547	0.0023	0.5630	0.0255	0.0743	0.0018	334.7	98.2	450.5	16.2	462.0	10.9
YMTC102-B2-20	124	410	0.30	0.0552	0.0017	0.5279	0.0153	0.0693	0.0014	399.0	65.8	430.8	9.8	432.1	8.6

Table 2. In situ Lu–Hf isotopic compositions of zircon from the syenogranite in the Yama beryllium deposit.

Spot	Age (Ma)	¹⁷⁶ Yb/ ¹⁷⁷ Hf	¹⁷⁶ Lu/ ¹⁷⁷ Hf	¹⁷⁶ Hf/ ¹⁷⁷ Hf	2σ	ε _{Hf} (0) ⁽¹⁾	ε _{Hf} (t) ⁽¹⁾	2σ	t _{DM1} ⁽²⁾ (Ga)	t _{DM2} ⁽³⁾ (Ga)	f _{Lu/Hf}
YMTC113-B5 barren syenogranite											
YMTC113-B5-4	412.7	0.035061	0.001052	0.282566	0.000016	−7.8	1.2	0.6	972	1303	−0.97
YMTC113-B5-6	442.9	0.026112	0.000676	0.282298	0.000015	−17.2	−7.6	0.5	1335	1874	−0.98
YMTC113-B5-9	441.1	0.033242	0.000885	0.282313	0.000016	−16.7	−7.1	0.6	1322	1846	−0.97
YMTC113-B5-10	432.8	0.071688	0.001838	0.282361	0.000015	−15.0	−5.9	0.5	1286	1760	−0.95
YMTC113-B5-11	443.5	0.054985	0.001411	0.282278	0.000017	−17.9	−8.5	0.6	1389	1930	−0.96
YMTC113-B5-12	430.3	0.052913	0.001373	0.282365	0.000016	−14.8	−5.7	0.6	1265	1745	−0.96
YMTC113-B5-13	430.0	0.048060	0.001273	0.282370	0.000015	−14.7	−5.5	0.5	1254	1732	−0.96
YMTC113-B5-16	446.6	0.042974	0.001123	0.282337	0.000013	−15.8	−6.2	0.5	1296	1793	−0.97
YMTC113-B5-17	426.8	0.070444	0.001848	0.282626	0.000020	−5.6	3.4	0.7	907	1174	−0.94
YMTC113-B5-20	444.2	0.036315	0.000960	0.282319	0.000016	−16.5	−6.9	0.6	1316	1832	−0.97

Table 2. Cont.

Spot	Age (Ma)	$^{176}\text{Yb}/^{177}\text{Hf}$	$^{176}\text{Lu}/^{177}\text{Hf}$	$^{176}\text{Hf}/^{177}\text{Hf}$	2σ	$\varepsilon_{\text{Hf}}(0)$ ⁽¹⁾	$\varepsilon_{\text{Hf}}(t)$ ⁽¹⁾	2σ	t_{DM1} ⁽²⁾ (Ga)	t_{DM2} ⁽³⁾ (Ga)	$f_{\text{Lu/Hf}}$
YMTC102-B2 beryllium-bearing syenogranite											
YMTC102-B2-1	432.7	0.044944	0.001179	0.282349	0.000015	−15.4	−6.1	0.5	1281	1776	−0.96
YMTC102-B2-2	445.4	0.042038	0.001107	0.282352	0.000014	−15.3	−5.7	0.5	1274	1761	−0.97
YMTC102-B2-4	454.8	0.053581	0.001398	0.282322	0.000016	−16.4	−6.7	0.6	1327	1827	−0.96
YMTC102-B2-6	423.5	0.060328	0.001583	0.282319	0.000016	−16.5	−7.5	0.6	1338	1855	−0.95
YMTC102-B2-8	417.7	0.045477	0.001200	0.282272	0.000015	−18.1	−9.2	0.5	1390	1955	−0.96
YMTC102-B2-9	405.9	0.026651	0.000709	0.282311	0.000017	−16.8	−7.9	0.6	1318	1868	−0.98
YMTC102-B2-10	434.8	0.048554	0.001266	0.282326	0.000015	−16.2	−6.9	0.5	1316	1827	−0.96
YMTC102-B2-11	420.6	0.044115	0.001131	0.282210	0.000016	−20.3	−11.3	0.6	1474	2089	−0.97
YMTC102-B2-12	411.2	0.045611	0.001200	0.282350	0.000015	−15.4	−6.6	0.5	1280	1787	−0.96
YMTC102-B2-14	424.9	0.034835	0.000929	0.282325	0.000016	−16.3	−7.1	0.6	1306	1829	−0.97
YMTC102-B2-15	447.8	0.050263	0.001330	0.282346	0.000015	−15.5	−5.9	0.5	1291	1777	−0.96
YMTC102-B2-16	417.5	0.060568	0.001597	0.282336	0.000014	−15.9	−7.0	0.5	1314	1821	−0.95
YMTC102-B2-17	448.9	0.045851	0.001207	0.282397	0.000016	−13.7	−4.1	0.6	1215	1661	−0.96
YMTC102-B2-18	450.0	0.049027	0.001291	0.282359	0.000017	−15.1	−5.4	0.6	1271	1746	−0.96
YMTC102-B2-19	462.0	0.056574	0.001483	0.282326	0.000014	−16.2	−6.4	0.5	1324	1816	−0.96
YMTC102-B2-20	432.1	0.047390	0.001261	0.282288	0.000014	−17.6	−8.3	0.5	1370	1913	−0.96

(1) $\varepsilon_{\text{Hf}}(0)$ and $\varepsilon_{\text{Hf}}(t)$ values were calculated using the present-day ($^{176}\text{Hf}/^{177}\text{Hf}$)_{CHUR} = 0.282785 and ($^{176}\text{Lu}/^{177}\text{Hf}$)_{CHUR} = 0.0336 [76]. (2) T_{DM1} values were calculated using present-day ($^{176}\text{Hf}/^{177}\text{Hf}$)_{DM} = 0.28325 and ($^{176}\text{Lu}/^{177}\text{Hf}$)_{DM} = 0.0384 [77] (3) T_{DM2} values were calculated using present-day ($^{176}\text{Hf}/^{177}\text{Hf}$)_{DM} = 0.283250, ($^{176}\text{Lu}/^{177}\text{Hf}$)_{DM} = 0.0384 [77] and ($^{176}\text{Lu}/^{177}\text{Hf}$)_{cc} = 0.015 [78].

Table 3. Major (wt.%) and trace element (ppm) compositions of the syenogranite in the Yama beryllium deposit.

Sample	YMTC113-B1	YMTC113-B2	YMTC113-B3	YMTC113-B4	YMTC113-B5	YMTC113-B6	YMTC113-B9	YMTC113-B11	YMTC102-B2	YMTC102-B4
Rock type	Beryllium-bearing syenogranite						Barren syenogranite			
SiO ₂	71.61	70.61	70.99	76.73	77.61	77.24	74.93	74.05	76.72	75.70
TiO ₂	0.01	0.03	0.03	0.04	0.04	0.03	0.03	0.03	0.08	0.07
Al ₂ O ₃	13.32	13.12	13.62	12.30	12.16	12.25	12.88	12.78	11.89	12.82
Fe ₂ O ₃	0.95	1.12	1.16	1.06	0.94	0.81	1.30	1.43	1.35	1.28
MnO	b.d.l.	0.01	0.01	0.01	0.01	0.01	0.04	0.04	0.03	0.02
MgO	0.19	0.22	0.22	0.14	0.12	0.13	0.08	0.10	0.25	0.22
CaO	0.30	0.42	0.34	0.22	0.20	0.29	0.58	1.22	0.56	0.30
Na ₂ O	4.83	4.81	4.92	3.98	3.88	4.10	3.01	2.94	2.26	2.56
K ₂ O	3.93	3.89	3.93	4.01	4.21	4.02	5.15	4.90	5.24	5.96
P ₂ O ₅	0.07	0.08	0.08	0.10	0.09	0.10	0.08	0.08	0.06	0.06
LOI	1.34	1.49	1.22	0.72	0.69	0.81	1.08	1.78	1.17	1.00
Total	96.55	95.80	96.52	99.31	99.95	99.79	99.16	99.35	99.61	99.99
ASI	1.05	1.02	1.05	1.09	1.08	1.05	1.11	1.03	1.14	1.14
Mg#	0.28	0.28	0.27	0.21	0.20	0.24	0.11	0.12	0.27	0.25
Be	>1000	>1000	>1000	14.00	10.95	4.98	6.63	6.73	3.02	5.58
Sc	1.4	1.9	2.1	1.0	1.1	0.9	3.1	2.9	2.4	2.5
V	2	1	2	3	2	2	1	1	2	2
Cr	15	12	13	17	13	14	22	17	19	25
Mn	86	102	111	99	95	80	295	324	217	144
Co	0.5	0.7	0.7	0.6	0.5	0.4	0.4	0.4	0.7	0.6
Ni	1.1	1.5	1.4	1.3	1.0	1.0	1.0	0.7	1.1	1.1
Cu	3.2	3.7	4.3	7.8	5.0	3.9	4.4	3.7	1.5	1.7
Zn	16	20	20	18	18	13	37	41	25	24
Ga	14.7	14.2	15.1	14.4	15.1	14.3	23.0	22.6	18.0	18.8
Rb	298	289	289	384	419	382	780	751	454	476
Sr	33.9	43.2	37.9	23.1	20.4	24.5	17.3	49.6	34.6	36.3
Y	71.8	66.9	64.6	58.2	57.8	54.6	46.0	46.1	57.0	59.3
Zr	72	64	67	56	57	52	53	52	76	74
Nb	11.4	16.8	17.6	19.0	20.0	16.0	23.6	20.7	15.2	14.6
Mo	1.72	1.68	1.79	4.45	2.30	3.15	1.96	1.38	1.82	2.17
Cd	b. d. l.	b. d. l.	0.02	0.02	0.03	0.03	b. d. l.	b. d. l.	b. d. l.	b. d. l.
Sn	19	18	20	11	8	11	35	33	14	14
Cs	17.10	17.90	17.40	28.6	30.8	26.5	63.5	62.8	19.15	18.30
Ba	75.6	109.5	104.0	71.9	62.8	69.8	20.3	30.6	84.3	103.5
La	14.1	12.7	13.4	12.6	12.1	11.6	11.6	11.1	19.3	20.0
Ce	39.9	36.0	36.9	34.6	33.3	31.9	32.1	29.8	45.5	48.2
Pr	4.85	4.49	4.47	4.09	4.06	3.85	3.80	3.60	5.19	5.44
Nd	17.2	15.6	15.9	14.2	13.8	13.4	13.2	13.0	18.4	19.4
Sm	6.64	6.27	6.42	5.25	5.31	4.95	4.91	4.94	5.58	5.98
Eu	0.12	0.09	0.10	0.05	0.04	0.05	0.02	0.03	0.19	0.22
Gd	8.03	7.53	7.39	6.08	5.96	5.66	4.98	5.06	6.32	6.73
Tb	1.81	1.70	1.61	1.41	1.37	1.33	1.13	1.16	1.36	1.43
Dy	11.90	11.00	10.90	9.52	9.24	8.89	7.58	7.76	9.00	9.50

Table 3. Cont.

Sample	YMTC113-B1	YMTC113-B2	YMTC113-B3	YMTC113-B4	YMTC113-B5	YMTC113-B6	YMTC113-B9	YMTC113-B11	YMTC102-B2	YMTC102-B4
Ho	2.35	2.14	2.08	1.87	1.83	1.76	1.46	1.47	1.87	1.99
Er	7.27	6.44	6.45	5.64	5.49	5.22	4.30	4.30	5.32	5.73
Tm	1.25	1.14	1.11	0.94	0.91	0.88	0.74	0.72	0.83	0.88
Yb	8.57	7.88	7.55	6.07	6.02	5.79	5.00	4.70	4.98	5.49
Lu	1.20	1.07	1.07	0.83	0.85	0.79	0.69	0.66	0.68	0.75
Hf	3.4	2.8	3.1	2.8	2.8	2.6	2.7	2.4	2.9	2.7
Ta	2.8	3.9	4.1	4.2	4.1	3.6	4.9	3.7	1.8	1.8
W	1.3	1.4	2.0	3.1	2.5	2.0	3.9	2.9	2.6	2.4
Pb	9.4	8.9	9.0	17.5	13.8	13.4	11.2	14.2	21.2	24.1
Bi	0.42	0.34	0.56	2.16	2.18	1.61	2.18	1.86	0.14	0.11
Th	25.0	23.2	24.4	24.9	25.0	22.5	24.4	22.8	22.7	23.4
U	2.12	2.15	2.18	4.59	3.94	3.99	1.73	2.60	2.04	1.97
ΣREE	125.19	114.05	115.35	103.15	100.28	96.07	91.51	88.30	124.52	131.74
LREE/HREE	1.95	1.93	2.02	2.19	2.17	2.17	2.54	2.42	3.10	3.05
(La/Yb) _N	1.11	1.09	1.20	1.40	1.36	1.35	1.56	1.59	2.61	2.46
(La/Sm) _N	1.34	1.27	1.31	1.51	1.43	1.47	1.49	1.41	2.18	2.10
Eu/Eu*	0.05	0.04	0.04	0.03	0.02	0.03	0.01	0.02	0.10	0.11
Tzr	722	711	716	711	711	703	707	699	739	735

b. d. l. = below the detection limit. Eu/Eu^* (Eu anomalies) = $\text{Eu}_N / \sqrt{\text{Sm}_N \times \text{Gd}_N}$, N represents chondrite-normalized values from ref. [79].

4.1. Zircon U–Pb Dating of the Studied Syenogranites

Zircon crystals extracted from the barren syenogranite within the Yama beryllium deposit, observed under a polarizing microscope, are commonly gray or colorless. They exhibit euhedral to subhedral crystal forms, varying in shape from elongated to stubby (40–160 μm in length, mostly with length-to-width ratios of 1:1 to 2:1). These zircon grains display concentric zoning, a typical characteristic of magmatic oscillatory zonation (Figure 5a) [80,81]. The Th/U ratios of these zircon grains vary from 0.19 to 1.20, clustering around 0.30–0.60, which is also consistent with a magmatic origin [80,81]. Consequently, the U–Pb ages obtained from these magmatic zircons are interpreted to represent the emplacement ages of the barren syenogranites.

The U–Pb concordia diagrams for the barren syenogranite samples are presented in Figure 5a,b. Analysis of 10 zircon spots from sample YMTC113-B5 reveals a generally coherent cluster on the concordia diagram (Figure 5b), yielding a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 437.7 ± 3.3 Ma (95% conf. (confidence interval)). Notably, two spots, YMTC113-B5-1 and YMTC113-B5-5, yielded significantly younger $^{206}\text{Pb}/^{238}\text{U}$ ages of 319.0 ± 6.6 Ma and 244.5 ± 7.6 Ma, respectively. In contrast, several other spots exhibit significantly older $^{206}\text{Pb}/^{238}\text{U}$ ages ranging from 490.1 to 1745.4 Ma.

Zircon grains obtained from the beryllium-bearing syenogranite are predominantly gray or colorless and exhibit euhedral to subhedral crystal forms, ranging in shape from elongated to stubby (40–200 μm in length, mostly with length-to-width ratios of 1:1 to 2:1). These zircon grains also display concentric zoning, a typical characteristic of magmatic oscillatory zonation (Figure 5c) [80,81]. The Th/U ratios of the zircons vary from 0.14 to 0.67, clustering around 0.30–0.50, which is also consistent with a magmatic origin [80,81].

The U–Pb concordia diagrams for the beryllium-bearing syenogranite are presented in Figure 5c,d. The 16 spots analyzed on zircons from sample YMTC102-B2 cluster as a generally coherent group on the concordia diagram (Figure 5d), yielding a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 432.0 ± 2.3 Ma (95% conf.). Notably, additional analytical spots exhibit significantly older $^{206}\text{Pb}/^{238}\text{U}$ ages ranging from 474.7 to 937.5 Ma.

4.2. Zircon Trace Element Compositions of the Studied Syenogranites

The trace element compositions of zircon grains extracted from the barren syenogranite are presented in Table S1. This study specifically focuses on primary zircon grains that yield reliable ages; consequently, data from inherited, contaminated, or altered zircon grains were excluded from the REE patterns. Notably, three samples (YMTC113-B5-13,

YMTC113-B5-16, and YMTC113-B5-20) exhibited La concentrations below the detection limit. Following the criteria proposed by Yang et al. [82], zircon trace element data with La concentrations exceeding 1 ppm were considered indicative of ablated mineral and melt inclusions; therefore, these data points (YMTC113-B5-4, YMTC113-B5-6, and YMTC113-B5-12) were also omitted from the REE patterns. Overall, the primary zircon grains displayed comparable chondrite-normalized REE patterns, characterized by pronounced negative Eu anomalies ($\text{Eu}_N/\text{Eu}_N^*$ ratios of 0.03–0.51, mean of 0.16, where N denotes chondrite-normalized values) and distinctly curved, convex upward middle REE (MREE) to heavy REE (HREE) patterns, as illustrated in Figure 6a.

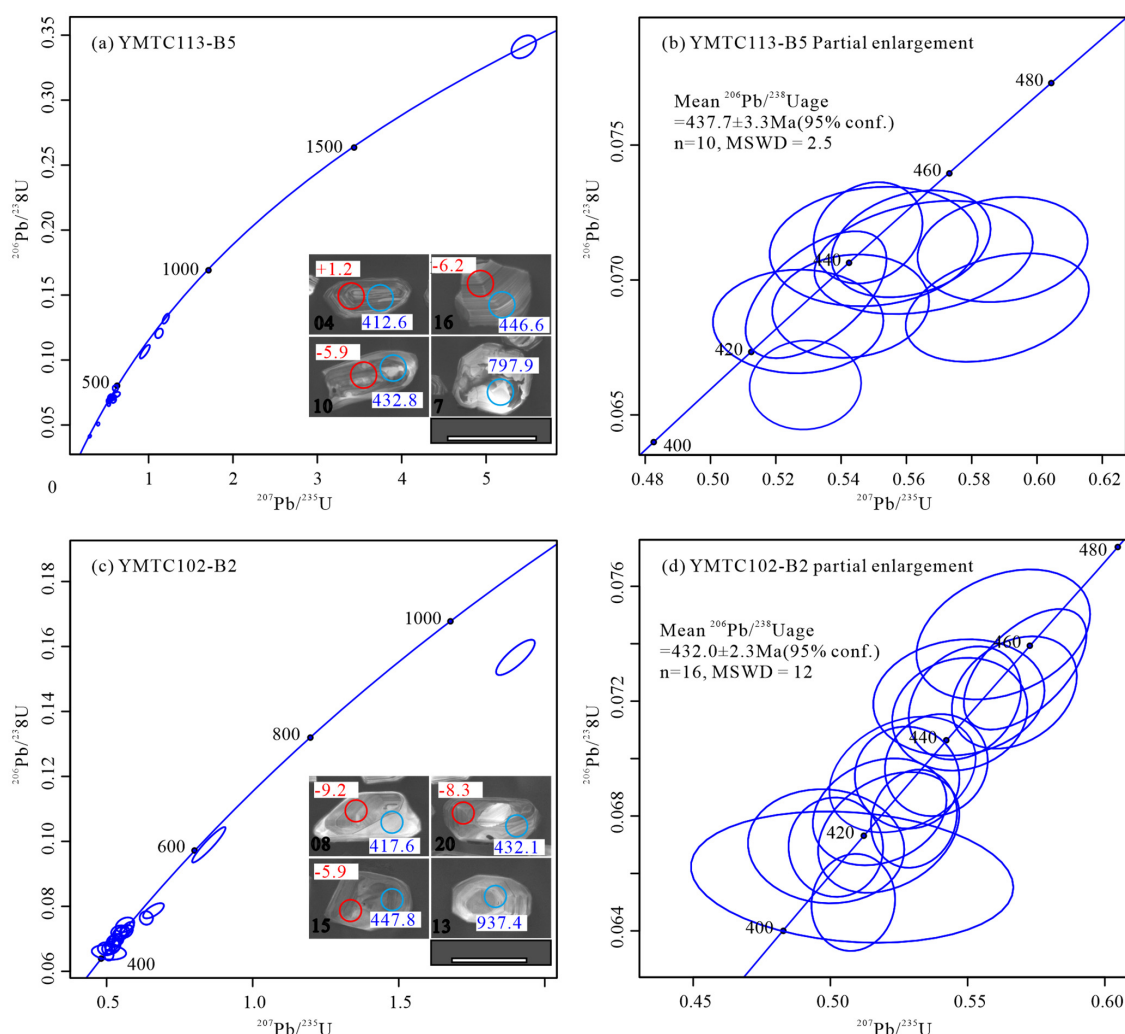


Figure 5. U-Pb concordia diagrams for zircon samples from the studied syenogranites in the Yama beryllium ore district within the Tataleng granitic batholith. (a) Sample YMTC113-B5 and (b) sample YMTC113-B5 (partially enlarged diagram) from the barren syenogranite; (c) sample YMTC102-B2 and (d) sample YMTC102-B2 (partially enlarged diagram) from the beryllium-bearing syenogranite. The insets display typical CL images of zircon, along with their corresponding age/ $\epsilon_{\text{Hf}}(t)$ values. The solid blue circles in the insets indicate the locations of the U-Pb age analysis spots, while the solid red circles mark the locations of the Hf isotope analysis spots. The white numbers in the insets represent grain numbers.

The trace element compositions of zircon grains extracted from the beryllium-bearing syenogranite are summarized in Table S1. Among these, six samples (YMTC102-B2-10, YMTC102-B2-13 to YMTC102-B2-16, and YMTC102-B2-20) exhibited La concentrations below the detection limit. Additionally, data from inherited zircon grains were also excluded

from the REE patterns. Overall, the primary zircon grains display similar chondrite-normalized REE patterns, characterized by pronounced negative Eu anomalies ($\text{Eu}_\text{N}/\text{Eu}_\text{N}^*$ ratios of 0.02–0.17 with a mean of 0.05), along with distinctly curved and convex upward MREE to HREE patterns (Figure 6a).

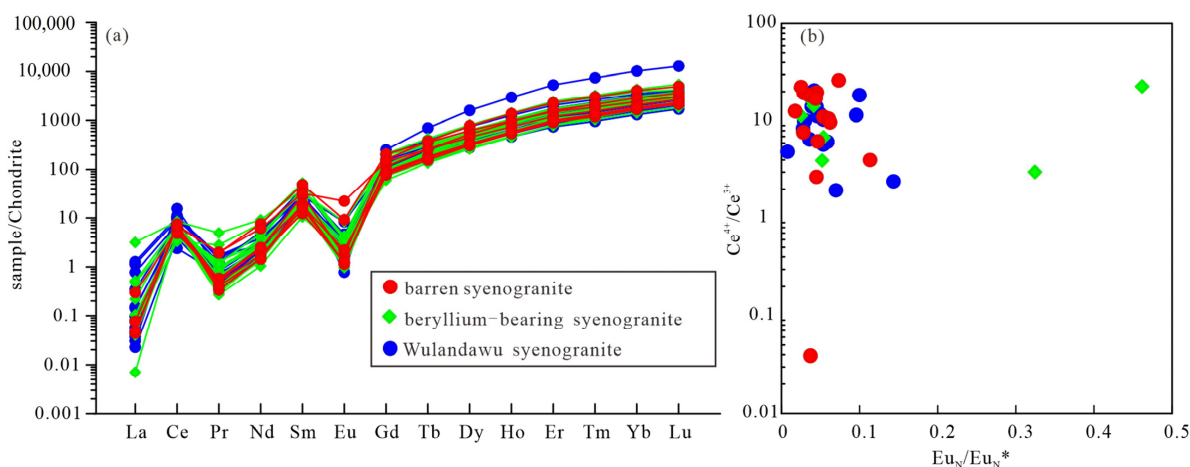


Figure 6. (a) Chondrite-normalized REE patterns and (b) $\text{Eu}_\text{N}/\text{Eu}_\text{N}^*$ vs. $\text{Ce}^{4+}/\text{Ce}^{3+}$ in zircons from the studied syenogranites from the Yama beryllium ore district. Chondrite normalization values are from [79]. $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratios were calculated using the method described in [74].

Moreover, the REE patterns of zircons from both barren and beryllium-bearing syenogranites closely resemble those of zircons from the Wulandawu syenogranite (Figure 6a), as reported by Jiao et al. [18]. Notably, neither the barren nor the beryllium-bearing syenogranite samples demonstrate a correlation between $\text{Eu}_\text{N}/\text{Eu}_\text{N}^*$ ratios and $\text{Ce}^{4+}/\text{Ce}^{3+}$ values in zircon grains (Figure 6b).

4.3. Zircon Hf Isotopes of the Studied Syenogranites

In situ Hf isotope analyses were performed on zircon grains previously dated by LA-ICP-MS U–Pb methods. Ten zircon grains from the barren syenogranite yielded reliable Hf isotope data. The results, summarized in Table 2, reveal that the magmatic zircons from sample YMTC113-B5 exhibit closely clustered initial $\varepsilon_{\text{Hf}}(t)$ values (corrected for age using the U–Pb ages of individual grains) ranging from -8.5 to $+3.4$, with an arithmetic mean of -5.7 (Figure 7a). These values correspond to t_{DM2} ages ranging 1174 Ma to 1930 Ma, with an arithmetic mean of 1748 Ma (Figure 7b).

Regarding the beryllium-bearing syenogranite, sixteen primary magmatic zircon spots yielded consistent Hf isotope data. As detailed in Table 2, the initial $\varepsilon_{\text{Hf}}(t)$ values of these zircons range narrowly from -11.3 to -4.1 , with an arithmetic mean of -7.0 (Figure 7c). These correspond to t_{DM2} ages between 1661 Ma and 2089 Ma, with an arithmetic mean of 1832 Ma (Figure 7d).

4.4. Major and Trace Element Compositions of the Studied Syenogranites

The whole-rock geochemical data presented in Table 3 reveal that the barren syenogranite exhibits SiO_2 concentrations ranging from 74.05 to 76.72 wt.%. This rock is characterized as peraluminous, with an alumina saturation index (ASI; calculated as the molecular ratio $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$) between 1.03 and 1.14. Additionally, it displays notably low Mg# values (defined as molecular $\text{MgO}/(\text{MgO} + \text{Fe}_2\text{O}_3^{\text{T}})$), varying from 0.11 to 0.27. The syenogranite possesses elevated alkali content (total $\text{K}_2\text{O} + \text{Na}_2\text{O}$) ranging from 7.50 to 8.52 wt.% (Figure 8a), high K_2O content between 4.90 and 5.96 wt.% (Figure 8b), and high

K₂O/Na₂O ratios of 1.67–2.33, corresponding to the high-K calc-alkaline to shoshonitic granitic series (Figure 8b) [83].

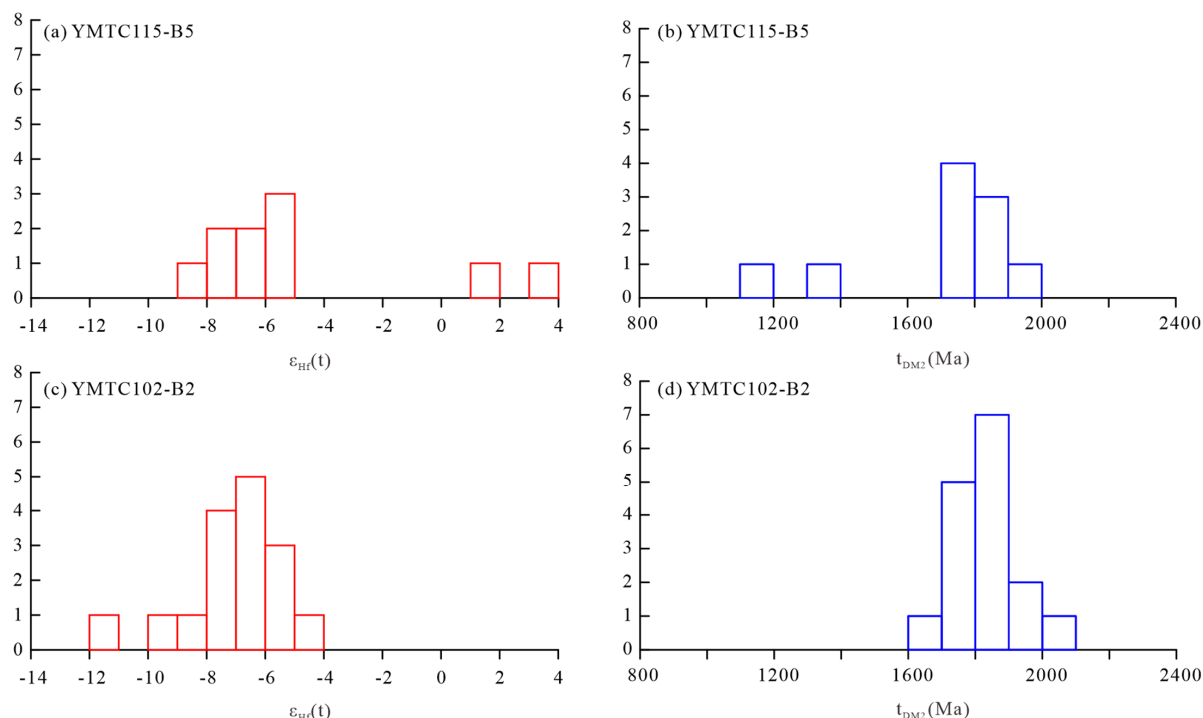


Figure 7. Histograms of $\epsilon_{\text{Hf}}(t)$ values (a) and T_{DM2} ages (b) of zircon from the barren syenogranite, and histograms of $\epsilon_{\text{Hf}}(t)$ values (c) and T_{DM2} ages (d) of zircon from the beryllium-bearing syenogranite in the Yama beryllium deposit.

Harker diagrams plotting major oxides against SiO₂ for the barren syenogranite show a robust decreasing trend in Al₂O₃ (Figure 8d), along with modest decreasing trends in CaO (Figure 8h), Na₂O (Figure 8i), and P₂O₅ (Figure 8j) with increasing SiO₂. Conversely, no significant correlations are observed between SiO₂ and K₂O (Figure 8b), TiO₂ (Figure 8c), Fe₂O₃^T (Figure 8e), MnO (Figure 8f), or MgO (Figure 8g).

The chondrite-normalized REE patterns of the barren syenogranite consistently display enrichment in total REE ($\Sigma\text{REE} = 88.30\text{--}131.74$), pronounced negative Eu anomalies ($\text{Eu}/\text{Eu}^* = 0.01\text{--}0.11$), and flat HREE patterns (Figure 9a). These granitic rocks exhibit slight enrichment in light REEs (LREEs) relative to HREEs ($\text{LREE}/\text{HREE} = 2.61\text{--}3.75$), and very low REE pattern slopes, as indicated by (La/Yb)_N values between 1.89 and 2.73 (Table 3).

Primitive mantle-normalized spider diagrams reveal that the barren syenogranites are significantly enriched in Rb, Th, and K, and strongly depleted in Sr, Ba, P, and Ti (Figure 9b).

The whole-rock geochemical data presented in Table 3 reveal that the beryllium-bearing syenogranite exhibits SiO₂ concentrations ranging from 71.61 to 77.61 wt.%. The rock is characterized as peraluminous, with ASI values of 1.02–1.09, and displays low Mg# values of 0.20–0.28. It has a high alkali content of 7.99–8.85 wt.% (Figure 8a), relatively high K₂O contents of 3.89–4.21 wt.% (Figure 8b), and high K₂O/Na₂O ratios of 0.80–1.09, consistent with the high-K calc-alkaline series (Figure 8b) [83].

Harker diagrams plotting major oxides against SiO₂ for the beryllium-bearing syenogranite demonstrate a pronounced decreasing trend in Al₂O₃ (Figure 8d), as well as modest decreasing trends in MgO (Figure 8g), CaO (Figure 8h), and Na₂O (Figure 8i) with increasing SiO₂. In contrast, no significant correlations are observed between SiO₂ and TiO₂ (Figure 8c), K₂O (Figure 8b), Fe₂O₃^T (Figure 8e), MnO (Figure 8f), or P₂O₅ (Figure 8j).

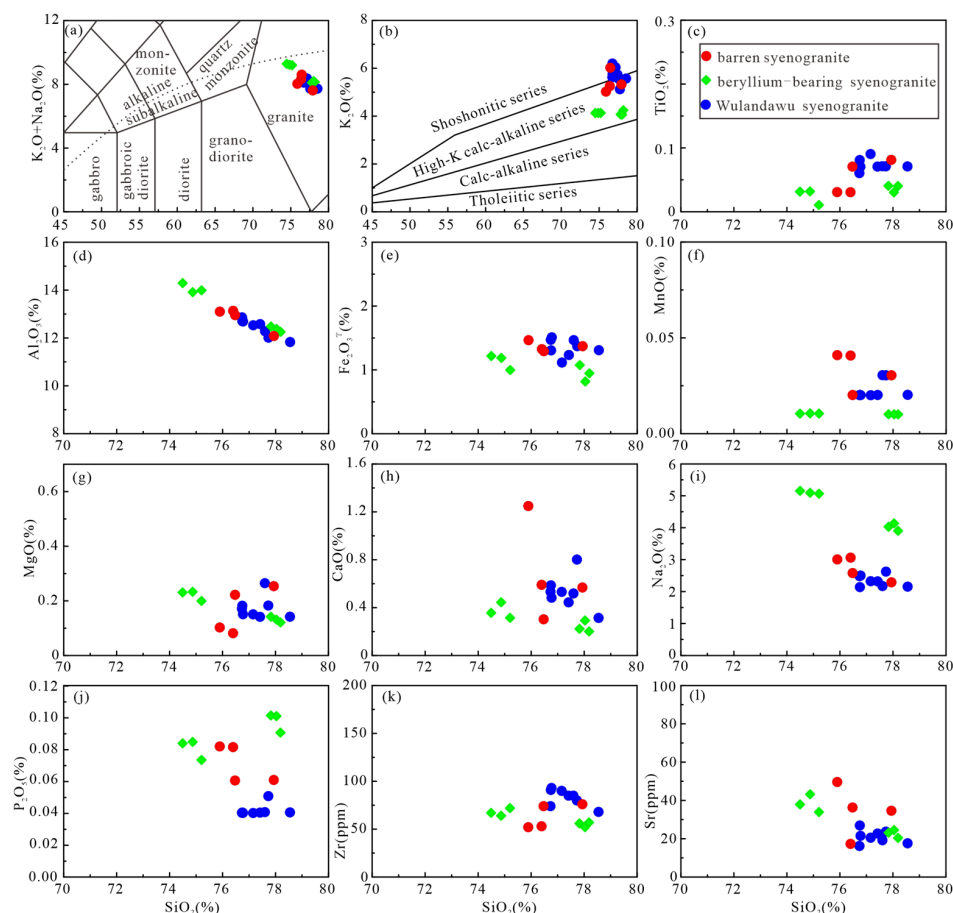


Figure 8. Harker diagrams for the studied syenogranites from the Yama beryllium ore district. (a) Total Alkali-Silica (TAS) diagram [84], Alkalic–subalkalic line adapted from Irvine and Baragar [85]. (b) K_2O vs. SiO_2 diagram [86]; (c–l) other oxides or elements vs. SiO_2 diagram. All major elemental data have been recalculated to 100% on a loss-on-ignition (LOI)-free basis. Data for the Wulandawu syenogranite are cited from Jiao et al. [18].

The chondrite-normalized REE patterns for the beryllium-bearing syenogranite consistently exhibit REE enrichment ($\Sigma REE = 96.07$ – 125.19), pronounced negative Eu anomalies ($Eu/Eu^* = 0.02$ – 0.05), and flat HREE patterns (Figure 9a). These granitic rocks are slightly enriched in LREEs relative to HREEs ($LREE/HREE = 1.93$ – 2.19), with very low REE pattern slopes ($(La/Yb)_N = 1.09$ – 1.14) (Table 3).

Primitive mantle-normalized spider diagrams reveal that the studied syenogranites are markedly enriched in Rb, Th, and K, and strongly depleted in Sr, Ba, P, and Ti (Figure 9b).

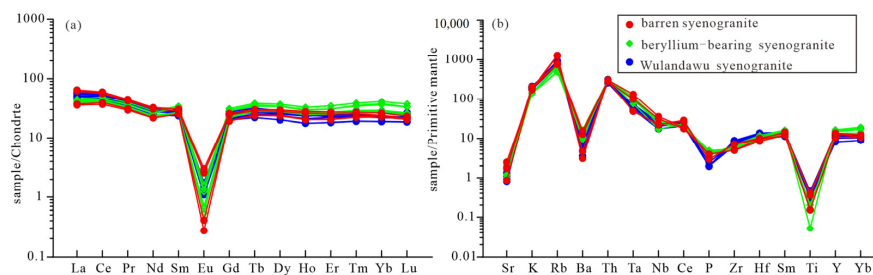


Figure 9. Chondrite-normalized REE patterns (a) and primitive mantle-normalized spider diagram (b) for the studied syenogranites from the Yama beryllium ore district in the Tataleng granitic batholith. The chondrite-normalized values are sourced from [79], while the primitive mantle normalization values are derived from [87]. Data for the Yama porphyritic syenogranite and Wulandawu syenogranite are cited from Jiao et al. [18].

Furthermore, beryllium concentrations in the three analyzed beryllium-bearing syenogranite samples exceed the detection threshold (>1000 ppm). When expressed as BeO content, values reach up to 0.28%, thereby directly qualifying these samples as beryllium ore. Additionally, the beryllium-bearing syenogranites exhibit significantly lower Fe₂O₃ and MnO concentrations but higher Na₂O abundance than the barren syenogranites, likely due to alteration processes.

5. Discussion

5.1. Geochronology of the Studied Syenogranites

We previously reported an age of 441.2 ± 5.0 Ma for the Wulandawu syenogranite, located near the southern segment of the fluorite-bearing fault within the Yama beryllium deposit [18]. In this study, we determined a consistent and reliable zircon mean $^{206}\text{Pb}/^{238}\text{U}$ age of 437.7 ± 3.3 Ma for the barren and an age of 432.0 ± 2.3 Ma for the beryllium-bearing syenogranite, both situated near the northern segment of the beryllium-bearing fault in the same district. These results collectively suggest that the emplacement of both the barren and beryllium-bearing syenogranites likely occurred during the Early Silurian. The ages obtained are concordant within analytical uncertainties, likely indicating the emplacement age of the syenogranite in the Yama beryllium deposit. Zircon analyses from the barren syenogranite samples revealed several other discrete ages, including both younger and older values. Given that extensive geochronological data indicate granitoids within the Tataleng granitic batholith were emplaced during O₃–S₁, it is reasonable to infer that the magmatic zircon ages in the barren syenogranite should not be younger than the Early Silurian. Two notably younger $^{206}\text{Pb}/^{238}\text{U}$ ages, 319.0 ± 6.6 Ma and 244.5 ± 7.6 Ma (Table 1), are unlikely to represent subsequent magmatic events, as geological evidence supports a single primary granitic intrusion episode for these samples. Consequently, these younger ages are interpreted as either laboratory contamination during zircon separation or as altered zircons that have undergone replacement by later hydrothermal fluids, both of which lack geological significance regarding the emplacement timing of the barren syenogranite. Conversely, the significantly older $^{206}\text{Pb}/^{238}\text{U}$ ages in the barren are considered to represent inherited or mixed magmatic zircon grains. Furthermore, a precise zircon mean $^{206}\text{Pb}/^{238}\text{U}$ age of 432.0 ± 2.3 Ma was determined for the beryllium-bearing syenogranite within the Yama beryllium deposit. This age, consistent within analytical uncertainties with two previously reported ages, further corroborates the emplacement timing of the syenogranite in this area. Consequently, it is proposed that the syenogranite in the Yama beryllium district was intruded during the Early Silurian, considering that 443.8 Ma marks the boundary between the Ordovician and Silurian according to the International Commission on Stratigraphy's (ICS) International Chronostratigraphic Chart 2022/10 [88]. The age range of 432.0 Ma to 441.2 Ma for the syenogranite in the Yama beryllium deposit falls within the published age range for the Tataleng granitic batholith, as previously noted. Additionally, other significantly older $^{206}\text{Pb}/^{238}\text{U}$ ages of zircon from beryllium-bearing syenogranites are interpreted as inherited or mixed magmatic zircon grains, analogous to those observed in barren syenogranite samples.

In summary, both barren and beryllium-bearing syenogranites in the Yama beryllium deposit have emplacement ages ranging from 432.0 to 437.7 Ma. Since beryllium mineralization is spatially associated with beryllium–fluorite-bearing fault zones within the syenogranite host rocks, geological and geochronological evidence suggests the timing of beryllium mineralization corresponds to, or postdates, these emplacement ages.

5.2. Petrogenetic Types of the Studied Syenogranites

Previous studies on granites from the Tataleng granitic batholith have reported considerable variability in petrogenetic types, predominantly S-type [18,23,31,32,35], with minor A-type [30] and I-type granites [22,36]. In the present study, both barren and beryllium-bearing syenogranites plot within the granite field on the total alkali versus silica (TAS) diagram (Figure 8a). Based on this classification and their mineralogical assemblages, these are identified as barren medium- to coarse-grained syenogranite and altered beryllium-bearing medium- to coarse-grained syenogranite, respectively. Geochemically, the two types exhibit strong similarities and also share characteristics with the Wulandawu syenogranite [18]. Both are essentially peraluminous, with ASI values of 1.03–1.14 (average = 1.11) and 1.02–1.09 (average = 1.06), respectively (Table 3, Figure 10a). These values do not strongly support classification as either I-type or S-type granites based on the criteria of [89]. It should be noted that certain geochemical diagrams based on major element compositions may be unsuitable for classifying the altered beryllium-bearing syenogranite due to alteration processes that can cause elemental mobility. However, both barren and beryllium-bearing syenogranites display nearly identical rare earth element (REE) patterns (Figure 9a) and spidergrams (Figure 9b). Consequently, trace element concentrations, including REEs and high field strength elements (HFSEs), are less influenced by hydrothermal alteration and can be used to understand the petrogenesis of the syenogranite in the Yama area. Therefore, diagrams utilizing specific major and/or trace elements (such as Al, Fe, Mg, Ga, Zr, Nb, Ce, and Y) are relatively robust against low to moderate alteration and thus remain effective for both unaltered and altered granitic rocks [90]. The samples analyzed herein exhibit low concentrations of Zr + Nb + Ce + Y (Figure 10b,c), suggesting that these rocks are more consistent with fractionated I- or S-type granites rather than A-type granites.

Additional geochemical features, including low CaO contents (Table 3; Figure 8h), high K₂O/Na₂O ratios (Table 3), and the presence of Al-enriched accessory minerals such as muscovite, further indicate that the studied syenogranites align more closely with S-type granites as defined by [91,92]. This interpretation is supported by the relatively low calculated T_{Zr} temperatures of 699–739 °C for the barren syenogranite and 703–722 °C for the beryllium-bearing syenogranite (Table 3), which fall within the typical temperature range for both I-type and S-type granites (<800 °C; e.g., ref. [93]). The relatively lower K₂O/Na₂O ratios observed in the beryllium-bearing syenogranite compared to the barren syenogranite may be attributed to minor albitization, wherein potassium-bearing alkali feldspar is partially replaced by sodium-rich albite. Furthermore, similar to the Wulandawu syenogranite [18], trace element variations, such as the colinear increases in the Rb/Sr (Figure 10d) and Rb/Ba (Figure 10e) ratios with decreasing Zr + Y + Ce in the studied syenogranites, are interpreted as signatures of extensive crystal fractionation [90], reinforcing the classification of these rocks as fractionated S-type granites. Negative correlations between Rb and Th (Figure 10f) observed along the evolutionary trajectory are also characteristic of S-type granite differentiation [94].

In summary, both barren and beryllium-bearing syenogranites are best classified as fractionated S-type granites, with the beryllium-bearing variety exhibiting a greater degree of alteration.

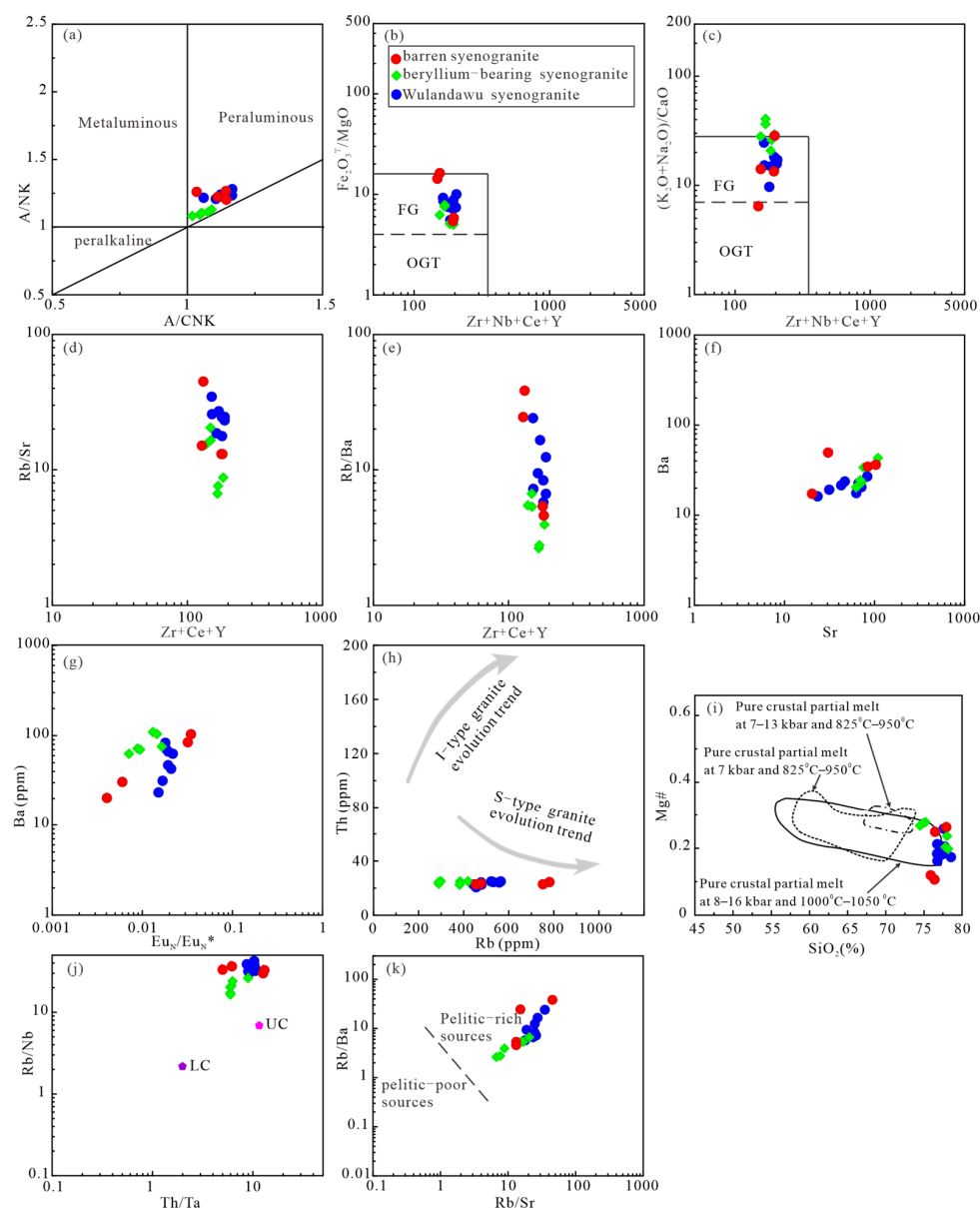


Figure 10. (a) A/CNK vs. A/NK diagram [95]; (b) Zr + Nb + Ce + Y vs. $\text{Fe}_2\text{O}_3^{\text{T}}/\text{MgO}$ and (c) Zr + Nb + Ce + Y vs. $(\text{K}_2\text{O} + \text{Na}_2\text{O})/\text{CaO}$ diagrams [90], FG-fractionated felsic granites; OGT-unfractionated M-, I- and S-type granites; (d) Zr + Ce + Y vs. Rb/Sr and (e) Zr + Ce + Y vs. Rb/Ba diagrams [90]; (f) Ba vs. Sr and (g) Eu anomaly vs. Ba [96]; (h) Rb vs. Th diagram [94]; (i) SiO_2 vs. Mg# diagrams for the studied syenogranites from the East Elikendagou area. Also shown are fields of pure crustal partial melts obtained from experimental studies on the dehydration melting of low-K basaltic rocks at 8–16 kbar and 1000–1050 °C [97], moderately hydrous (1.7–2.3 wt.% H_2O) medium- to high-K basaltic rocks at 7 kbar and 825–950 °C, and pelitic rocks at 7–13 kbar and 825–950 °C [98]; (j) comparison of the Rb/Nb and Th/Ta ratios, with lower crust (LC) and upper crust (UC) values from [99]; (k) Rb/Sr vs. Rb/Ba diagram [100] for the studied syenogranites from the East Elikendagou area. Data for the Wulandawu syenogranite are cited from Jiao et al. [18].

5.3. Origins and Petrogenesis of the Studied Syenogranites

The ~437.7 Ma magmatic zircons in the barren syenogranite and the ~432.0 Ma magmatic zircons in the beryllium-bearing syenogranite yielded $\varepsilon_{\text{Hf}}(t)$ values of -8.5 to $+3.4$ (mean -5.7) and -11.3 to -4.1 (mean -7.0), respectively (Figure 7a,c), suggesting that the magmas of both syenogranite types were derived predominantly from reworked crustal sources. Moreover, both barren and beryllium-bearing syenogranites display low Mg#

values of 0.11–0.27 and 0.20–0.28, respectively, which are comparable to those observed in pure crustal partial melts (Figure 10g). This similarity supports the interpretation that these granitic magmas were generated directly from crustal material without significant interaction with mantle peridotite or mixing with mantle-derived melts (e.g., refs. [18,23,101,102]). Additionally, ratios of large ion lithophile elements (LILE) to high field strength elements (HFSE), such as Rb/Nb, Th/Ta, La/Ta, and Rb/Zr, serve as useful proxies for identifying potential crustal source reservoirs. These ratios are generally preserved or enriched in melts produced by anatexis [103]. The LILE/HFSE ratios (e.g., Rb/Nb, Th/Ta) observed in both barren and beryllium-bearing syenogranites, which are comparable to those of the Wulandawu syenogranite, approximate the average upper crustal values reported by Rudnick and Gao [99] (Figure 10g). This correspondence further supports an origin from upper crustal sources for the studied syenogranites.

The Al_2O_3 concentration in melts is observed to be below 15 wt.% when the pressure is less than 16 kbar, characterizing these as low- Al_2O_3 partial melts. The residual restite in such cases typically consists of amphibole, plagioclase, and orthopyroxene [97]. Both barren and beryllium-bearing syenogranites exhibit relatively low Al_2O_3 contents of 11.89–12.88 wt.% and 12.16–13.62 wt.%, respectively, suggesting that the source regions of these syenogranites were likely situated at relatively shallow crustal levels. This inference is supported by their comparatively low calculated T_{Zr} temperatures of 699–739 °C and 703–722 °C, respectively (Table 3). Additionally, these syenogranites display relatively elevated HREEs (i.e., $\text{Yb} > 4.70$ ppm and $\text{Yb} > 5.79$ ppm, respectively) and Y contents (>46.00 ppm and >54.60 ppm, respectively), along with flat HREE patterns (Figure 9a), precluding garnet as a residual phase and further implying the interpretation of a shallow source. Consequently, it is posited that these syenogranites predominantly originated from relatively shallow Paleoproterozoic reworked crust, a conclusion reinforced by Hf isotopic data and average T_{DM2} ages of approximately 1748 Ma and 1832 Ma (Figure 7b,d). Geochemically, the studied syenogranites are primarily peraluminous (Figure 10a), indicating the involvement of metasedimentary rocks during partial melting processes (e.g., refs. [101,102,104]). Their elevated K_2O contents, high Rb/Sr ratios of 13.11–45.09 and 6.69–20.54 and high Rb/Ba ratios of 4.60–38.42 and 2.64–6.67 further imply derivation from a pelitic-rich source (Figure 10i) [100], consistent with characteristics reported for the Wulandawu syenogranite [18].

Both barren and beryllium-bearing syenogranites exhibit significant chemical variations, as demonstrated by the Harker diagrams (Figure 8). The general decrease in Al_2O_3 (Figure 8d), CaO (Figure 8h), and Na_2O (Figure 8i) with increasing SiO_2 suggests the fractionation of feldspar and Ti-bearing phases during crystallization [105]. Additionally, both types of syenogranites exhibit comparable negative anomalies in Sr, Ba, P, and Ti on the spider diagrams (Figure 9b), further supporting the fractional crystallization of feldspar, apatite, and Ti-bearing phases [105] during their formation. Furthermore, pronounced Sr depletion (Figure 9b), strong negative Eu anomalies (Figure 9a), and strong negative correlations between Rb/Sr and $\text{Zr} + \text{Y} + \text{Ce}$ (Figure 10d), as well as between Rb/Ba and $\text{Zr} + \text{Y} + \text{Ce}$ (Figure 10e), along with positive correlations between Sr and Ba (Figure 10f) and between Eu anomalies and Ba (Figure 10g), indicate clear feldspar fractionation. In conclusion, the geochemical characteristics and variations observed in the studied syenogranites are consistent with fractional crystallization processes involving feldspar, ferrous minerals, and titanium-bearing minerals following partial melting.

5.4. Tectonic Significance of the Studied Syenogranites

The geochemical properties of the studied syenogranites closely resemble those of the Wulandawu syenogranite, as documented by Jiao et al. [18]. Both barren and beryllium-

bearing syenogranites are interpreted to have originated from relatively shallow, reworked crustal sources, as evidenced by their comparable geochemical compositions and negative $\varepsilon_{\text{Hf}}(t)$ values. Furthermore, on the Yb vs. Ta (Figure 11a) and Nb vs. Y (Figure 11b) diagrams, the majority of the analyzed syenogranite samples plot within the within-plate granite (WPG) field, although they are close to the syn-collisional granites (syn-COLG) field. Moreover, in the Y + Nb vs. Rb plot (Figure 11c), most samples fall within the syn-collisional granite (syn-COLG) or post-collisional granite (post-COLG) fields, although a few beryllium-bearing syenogranite samples are positioned within the WPG field. Due to alteration processes such as albitization affecting the beryllium-bearing syenogranite samples, the R1 vs. R2 plot, which relies on specific major element ratios, is deemed unsuitable for these samples. Nevertheless, in the R1 vs. R2 plot, all barren syenogranite samples, along with three beryllium-bearing syenogranite samples, plot within the syn-collision to post-orogenic granitic field (Figure 11d). The remaining three beryllium-bearing syenogranite samples fall within an undefined field. Overall, combined with the fact that most granitoids of the same age in the region are syn-collisional granites (e.g., refs. [18,22,23]), we conclude that the data derived from the studied syenogranites in the Yama beryllium deposit predominantly indicate a syn-collisional tectonic setting during the early Silurian period (432.0–441.2 Ma). This tectonic setting appears to be closely linked to the syn-collisional stage, i.e., the closure of the South Qilian Ocean, in the South Qilian Belt (SQB) between 450 and 418 Ma, as previously concluded by Jiao et al. [23]. After 418 Ma, the South Qilian Belt entered the post-collisional stage.

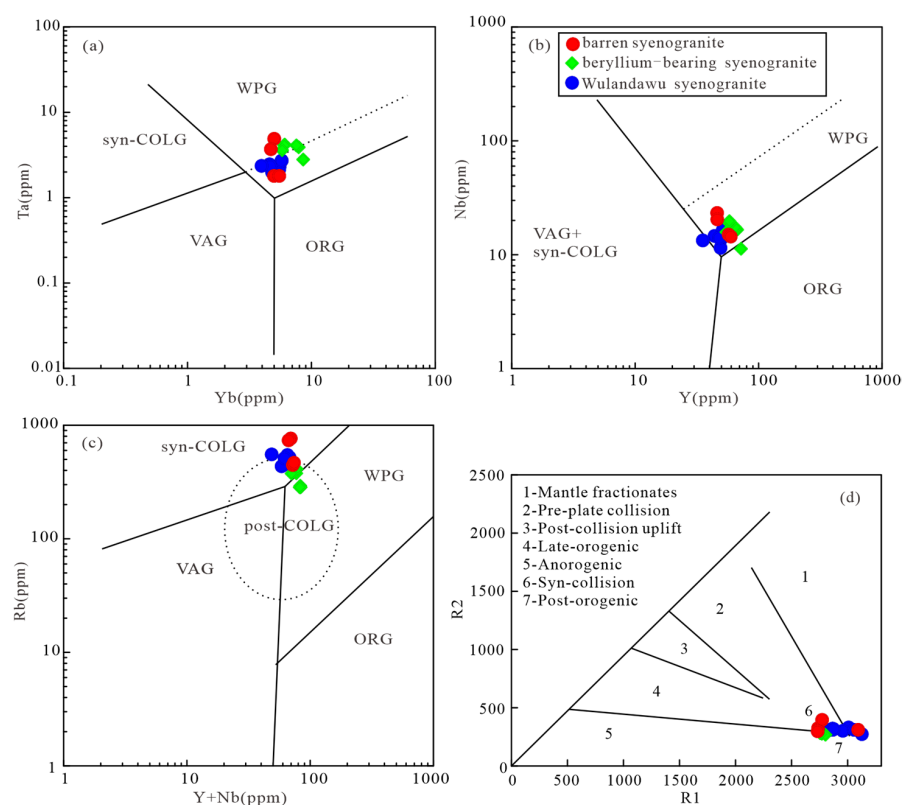


Figure 11. (a) Yb vs. Ta, (b) Y vs. Nb, and (c) (Yb + Nb) vs. Rb diagrams [106] for the studied syenogranites from the Yama beryllium ore district. WPG: within-plate granites, VAG: volcanic arc granites, syn-COLG: syn-collisional granites, post-COLG: post-collisional granites, and ORG: ocean ridge granites. (d) R1 vs. R2 diagram [107] for the studied syenogranites from the Yama beryllium ore district. $R1 = 4Si - 11(Na + K) - 2(Fe + Ti)$, $R2 = 6Ca + 2Mg + Al$. Data for the Wulandawu syenogranite are cited from Jiao et al. [18].

5.5. Redox State and Mineralization Potential of the Studied Syenogranites

The development of fractionated granites is often linked to significant interactions between melt and fluid, which typically occur prior to the mineralization of rare metals such as W, Sn, Nb, Ta, Li, Be, Sb, and REEs [108,109]. These interactions are commonly identified by low K/Rb ratios below 150, Nb/Ta ratios below 6, and Zr/Hf ratios below 26 [108,110]. Consequently, whole-rock Zr/Hf ratios ranging from 18 to 46 and Nb/Ta ratios below approximately 5 in granites have been proposed as potential indicators of Sn–W–(U)-related granites [111]. However, beryllium-bearing syenogranite samples are unsuitable for analysis using the diagram in Figure 12a because alteration processes affect their whole-rock compositions. Similar to the Wulandawu syenogranite, the barren syenogranite from the Yama beryllium deposit shows low K/Rb ratios between 55.51 and 105.00, relatively low Nb/Ta ratios ranging from 4.82 to 8.44, and low Zr/Hf ratios between 19.63 and 27.41 (Figure 12a), plotting in the field between Sn–W–(U)-related granites and Sn–W-rare-metal barren granites (Figure 12a, ref. [111]). This suggests that some melt–fluid interactions occurred during the formation of these syenogranites, indicating certain potential for the formation of Sn–W–(U) deposits.

Moreover, the redox state of magmatic melts in granites is often linked to the genetic types of associated magmatic–hydrothermal ore deposits [109,112–114]. Oxidized calc-alkaline granitoids are generally associated with the mineralization of copper, molybdenum, and gold [74,115,116], whereas reducing granitic intrusions are typically connected to the mineralization of tungsten, tin, and other related metals [117,118]. Based on zircon oxybarometry results, such as predominantly negative ΔFMQ values (Figure 12b), low $f\text{O}_2$ values mostly plotting below the FMQ buffer (Figure 12c), and low $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratios (Figure 12d) in primary zircon, both barren and beryllium-bearing syenogranites exhibit relatively reducing conditions, similar to those observed in the Wulandawu syenogranite [18]. This suggests a potential for mineralization of tungsten, tin, and other related rare metals like beryllium. Furthermore, most zircon grains from the barren syenogranite studied, as well as some from the beryllium-bearing syenogranite, fall within or near the magmatic field on the La vs. $\text{Sm}_\text{N}/\text{La}_\text{N}$ (Figure 12e) and $\text{Sm}_\text{N}/\text{La}_\text{N}$ vs. Ce/Ce^* (Figure 12f) diagrams. However, a few zircon grains from the beryllium-bearing syenogranite plot within the hydrothermal transitional fields in these figures, suggesting they have been slightly affected by hydrothermal processes. This implies that certain hydrothermal fluids may have caused varying degrees of alteration to some magmatic zircon, likely contributing to beryllium mineralization in the studied area. Petrographic observations reveal that the bleached beryllium-bearing syenogranite exhibits distinct bleached colors and alteration features, including the presence of albite, quartz, muscovite, and sericite, compared to the unaltered barren syenogranite. These alterations primarily result in the altered beryllium-bearing syenogranite having significantly lower Fe_2O_3 and MnO concentrations but significantly higher Na_2O and Be abundances than the barren syenogranite.

In particular, within the bleached yet typically altered syenogranite ores of the MBe 3 beryllium ore body, which has not experienced significant fracturing or faulting, the concentration of beryllium significantly exceeds the cutoff grade for BeO. This observation suggests that post-magmatic alterations may play a direct role in the genesis of magmatic–hydrothermal beryllium mineralization. During these alterations, beryllium was likely precipitated and enriched in the altered syenogranite, forming the bleached beryllium-bearing syenogranite that hosts major beryllium ores with BeO grades exceeding 0.28% (Table 3). Considering the geological and geochemical characteristics of the Yama beryllium deposit described above, it appears that the studied syenogranite directly formed the initial stage of magmatic-related beryllium mineralization, and that bertrandite is of magmatic–hydrothermal origin. However, identifying the source of the ore-forming fluids responsible

for the alteration assemblage of muscovite, sericite, quartz, and albite, as well as the beryllium mineralization in the Yama beryllium deposit, is challenging because traditional stable isotope analyses (e.g., H, O, and C isotopes) cannot be directly performed on the beryllium-bearing syenogranite. For instance, the beryllium ores mainly consist of fault breccia or altered syenogranite rather than hydrothermal vein types. Additionally, the quartz in the ores may be of both magmatic and hydrothermal origin, making it impossible to separate hydrothermal quartz for hydrogen and oxygen isotope analysis.

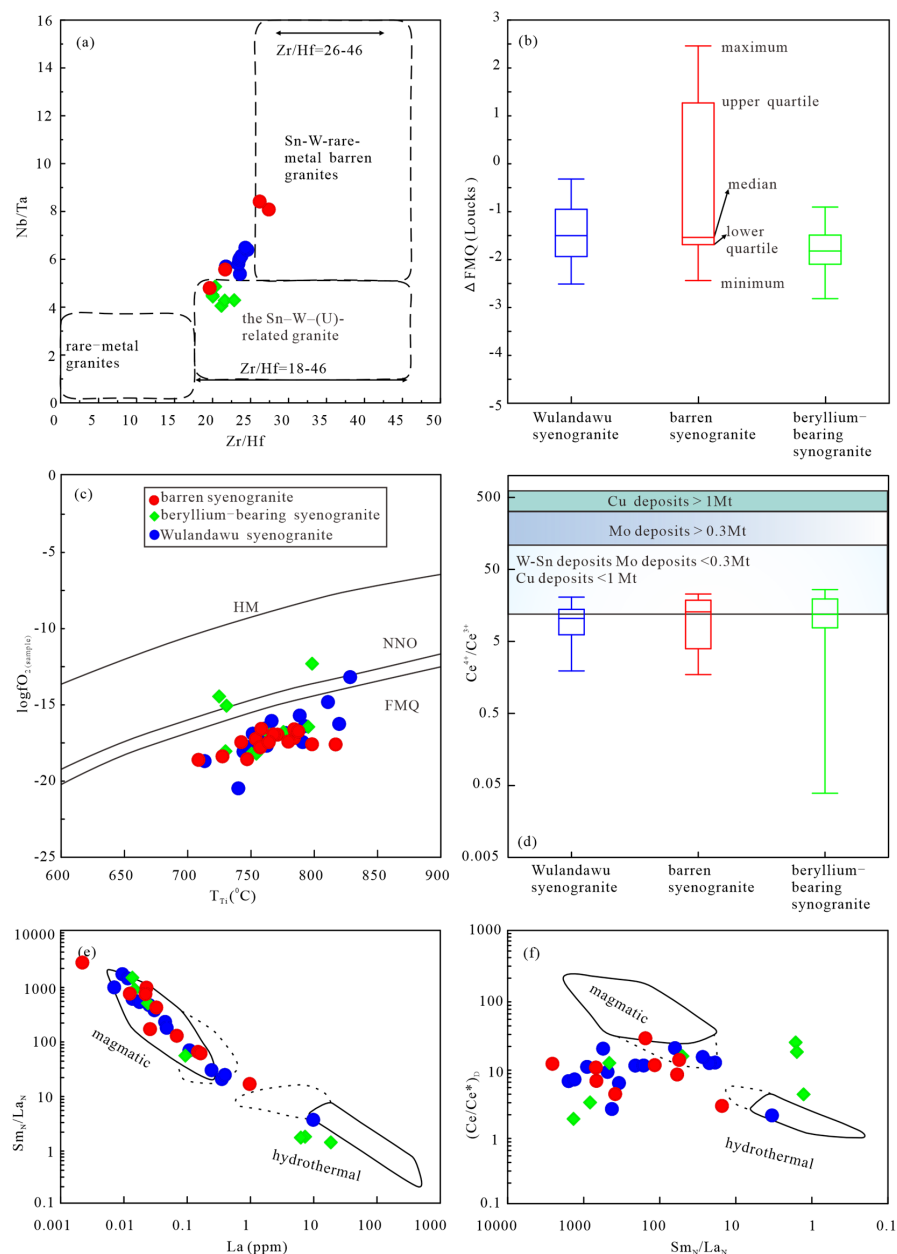


Figure 12. (a) Diagram of Nb/Ta vs. Zr/Hf [111] for the studied syenogranites; (b) box plot of zircon ΔFMQ values, calculated using the method described by Loucks et al. [73]; (c) diagram of calculated zircon saturation temperatures (T_{Zr} , °C) vs. oxygen fugacity ($f\text{O}_2$) values. MH, magnetite-hematite buffer [119]; NNO, Ni–NiO buffer [120]; FMQ, fayalite–magnetite–quartz buffer [72]; (d) box plot of zircon $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratios of the studied syenogranites, compared with the range of zircon $\text{Ce}^{4+}/\text{Ce}^{3+}$ ratios from zircon from granites related to W–Sn, Mo, and Cu deposits, as proposed by Yang et al. [109]; (e) chondrite-normalized Sm/La ratio vs. La (ppm) and (f) Ce anomaly $(\text{Ce}/\text{Ce}^*)_D$ vs. Sm_N/La_N [121]. The chondrite-normalized values are sourced from [79]. Data for the Wulandawu syenogranite are cited from Jiao et al. [18].

Moreover, within the fluorite–beryllium-bearing faults, both beryllium ore bodies were composed predominantly of fault breccia ores, and a small amount of fluorite ore bodies were present. According to previous dating results from the Yama fluorite mining area, the likely ore-forming age of fault-controlled hydrothermal vein-type fluorite mineralization is approximately 155 Ma [28]. It can be inferred that during the Late Jurassic (155 Ma) in the Yama area, medium- to low-temperature, low-salinity fluorine–beryllium hydrothermal fluids formed at depth. In these fluids, beryllium was likely transported as fluoride complexes, which exhibit strong solubility and transport capabilities. As these late-stage fluorine–beryllium hydrothermal fluids ascended along ore-controlling faults within the studied barren syenogranite, fluorine precipitated first through fluorite crystallization, creating low-temperature hydrothermal vein-type fluorite mineralization controlled by the faults. Subsequently, as the availability of free fluorine gradually declined, the stability of beryllium–fluorine complexes decreased, causing beryllium to transition from a mobile to a precipitated state. This process likely led to a second stage of beryllium enrichment within fault zones and fractured syenogranite, resulting in fault-controlled hydrothermal beryllium mineralization closely along the faults.

In summary, the initial stage of beryllium mineralization likely formed through a magmatic–hydrothermal process involving the studied syenogranites. Subsequently, late-stage fluorite deposition contributed to fault-controlled hydrothermal beryllium mineralization associated with ore-controlling faults. Therefore, exploration efforts for beryllium deposits in the Qilian Orogen should prioritize targeting ore-controlling faults within fractionated S-type syenogranite bodies.

6. Conclusions

The integrated results from zircon U–Pb dating, zircon Hf isotopic and trace element analyses, and whole-rock geochemical data for both barren and beryllium-bearing syenogranites lead to the following conclusions:

1. High-precision LA–ICP–MS zircon U–Pb dating constrains the emplacement ages of the barren and beryllium-bearing syenogranites to 437.7 ± 3.3 Ma and 432.0 ± 2.3 Ma, respectively. These ages are consistent with previously reported zircon U–Pb ages of 441.2 ± 5.0 Ma for the Wulandawu syenogranite, located near the southeastern segment of the fluorite-bearing fault zones.
2. Both barren and beryllium-bearing syenogranites are characterized as peraluminous with elevated alkali contents, relatively low $(K_2O + Na_2O)/CaO$ and $Fe_2O_3^T/MgO$ ratios, low Zr + Nb + Ce + Y concentrations, strongly negative Eu anomalies, high Rb/Sr and Rb/Ba ratios, and negative correlations between Rb and Th, indicating fractionated S-type syenogranites. The mineralogical assemblage, geochemical data, and zircon Hf isotopic compositions (mean $\epsilon_{Hf}(t)$ values of -5.7 and -7.0 for barren and beryllium-bearing syenogranites, respectively) collectively suggest derivation from the partial melting of a pelitic-rich source within the relatively shallow Paleoproterozoic reworked lower crust. This magmatic event occurred during an early Silurian syn-collisional tectonic regime associated with the closure of the South Qilian Ocean.
3. Zircon from the studied syenogranites exhibits low oxygen fugacity (i.e., negative ΔFMQ values) and low Ce^{4+}/Ce^{3+} ratios, indicating comparatively reducing conditions. When combined with the genetic classification of the syenogranite as a fractionated S-type granite, evidenced by low whole-rock K/Rb ratios, relatively low whole-rock Nb/Ta ratios, and a few zircon grains with relatively high La contents and Sm_N/La_N ratios, the syenogranite demonstrates certain potential for beryllium mineralization. It is proposed that the initial stage of beryllium mineralization formed through a magmatic–hydrothermal process following the crystallization of the studied

syenogranite. Subsequently, the fault-controlled fluorite deposition contributed to late-stage-associated beryllium mineralization within faults in the Yama beryllium deposit. Therefore, exploration efforts for beryllium deposits in the Qilian Orogen should prioritize targeting ore-controlling fault structures within fractionated S-type syenogranite bodies.

Supplementary Materials: The following supporting information can be downloaded at: <https://www.mdpi.com/article/10.3390/min16100977/s1>, Table S1. Trace element composition of zircon from the syenogranite in the Yama beryllium deposit. Note: Different types of ΔFMQ values were calculated using the methods of Trail et al. [122] or Loucks et al. [73], based on the corresponding T_{Zr} and T_{Ti} temperatures.

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