The composite Triassic–Eocene Poshteh pluton, eastern Iran, an Eo-Cimmerian element south of the main Paleotethys suture

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Abstract
The composite Poshteh Pluton, at the northeastern margin of the Central Iranian Microplate near Taybad in eastern Iran, is positioned at a critical tectonic junction, south of the inferred main Paleotethys suture and along the major regional Doruneh Fault system. It consists of two distinct intrusions. Quartz monzonite is dated in this study by zircon U–Pb ID-TIMS to 215.8 ± 0.5 Ma, an age that coincides with the time of closure of the Paleotethys during the late collisional stages of the Eo-Cimmerian Orogeny. It is geochemically very similar to coeval plutons present along and north of the Paleotethys suture, where they intruded Carboniferous-Permian arc sequences, ophiolites and flysch. The Poshteh quartz monzonite is located south of the suture in a position similar to the Anarak and related complexes further west, which previously have been interpreted as reflecting Mesozoic and Cenozoic disruption of the Eo-Cimmerian Orogen by extensional and transtensional processes. The Triassic quartz monzonite was subsequently invaded by granite at 41.23 ± 0.31 Ma. The emplacement was in part structurally controlled by the Doruneh Fault system and associated to hydrothermal alteration and Fe mineralization. The granite is thus a coeval member of a widespread late Eocene to Oligocene plutonic suite in the region, and likely the result of delamination and melting of the subcontinental lithosphere.

Keywords Eo-Cimmerian Orogeny · Paleotethys · Eocene · Pluton · Zircon · U–Pb

Introduction
Eastern Iran consists of a complex collage of tectonic elements. It includes late Precambrian to early Paleozoic terranes, derived from northern Gondwana, and variously affected, and eventually welded together, by two major orogenic cycles. The first in the Paleozoic, ending with the Triassic Eo-Cimmerian Orogeny, and the second in the Cenozoic corresponding to the Alpine Orogeny (e.g. Stöcklin 1977; Berberian and King 1981; Sengör 1984; Besse et al. 1998; Golonka 2004; Zanchi et al. 2009a). The late Precambrian processes involved mainly magmatic growth along the northern margin of Gondwana. Early Paleozoic rifting detached ribbon terranes and opened the Rheic Ocean (e.g. Stampfli and Borel 2002; Bagheri and Stampfli 2008; Torsvik and Cocks 2013; Domeier 2018; Ranjbar Moghadam et al. 2018). A subsequent stage of rifting in the Devonian caused deposition of clastic sediments and evaporites and was followed by marine sedimentation, gradually opening the Paleotethys (Bagheri and Stampfli 2008). This opening preceded diachronously from west to east, starting as back-arc related to subduction of the Rheic Ocean plate, and led to the separation from northern Gondwana of ribbon microcontinents (Galatian superterrane; Stampfli et al. 2013). Subduction of the Paleotethys plate started in the middle to late Devonian and continued through the Carboniferous, as recorded by ophiolites, by arc magmatism, and metamorphic stages at the margin of the Turan (Eurasian) continental plate (Rutten 1993; Bagheri and Stampfli 2008; Zanchetta et al. 2009, 2013; Shafaii Moghadam et al. 2015a). By the middle Triassic the Paleotethys plate was eventually completely consumed once the Cimmerian block, which had separated from Gondwana during the opening of the Neo- tethys in the Permian, collided with the Eurasian plate: this is referred to as the Eo-Cimmerian Orogeny. Late Triassic
to early Jurassic molasse deposits of the Shemshak Group define the final stages of this orogenic event (Zanchi et al. 2009a; Wilmsen et al. 2009). In Iran and Afghanistan the Paleoethys suture can be tracked mostly along the southern margin of the Turan Plate, but subsequent tectonic processes have contributed to its disruption, and some parts are now found south of the main suture. Of particular importance for the present paper are the relationships at the northern margin of the Central Iranian Microplate (CIM) south of the Doruneh Fault. Here the Anarak, Jandaq and Posht-e-Badam metamorphic complexes (Fig. 1) record several stages of the Paleoethys convergence and subduction history, including ophiolites and flysch deposits, late Carboniferous—early Permian metamorphism, and the final late Triassic Eo-Cimmerian collision (Fig. 1; Bagheri and Stampfli, 2008; Zanchi et al. 2009b, 2015).

The Neotethys had opened in the late Carboniferous to Permian (Stampfli et al. 2013) and eventually started to contract. In the Cenozoic it closed as a consequence of convergence and collision of Arabia with Central Iran. The tectonic processes in Iran were also directly affected by the larger scale convergence of India with Asia and the resulting lateral indentation (e.g. Gaina et al. 2015; Bagheri and Gol 2020). The interaction of the various tectonic elements through time has resulted in the very complex geological relationships we observe today.

The present study is focused on the composite Poshteh Pluton in eastern Iran, the product of two distinct magmatic events, the first in the Triassic and the second in the Eocene. This pluton is hosted by a metamorphosed supracrustal sequence and was originally mapped as an entirely Eocene intrusion (Ternet et al. 1980). The present documentation of the original Triassic history has special relevance for the paleogeography and the processes related to the closure of the Paleoethys and the Eo-Cimmerian Orogeny and the subsequent Cretaceous and Paleogene to recent tectonics in the region. An important question concerns the location of this Triassic intrusion to the south of the main Paleoethys suture, that mimics other occurrences further west, but also very similar in age and chemical composition to several coeval plutons north of the main Paleoethys suture zone (Figs. 1, 2). The Eocene granite corresponds instead to a very widespread magmatic province in eastern Iran. The paper documents the ages and the main compositional features of the distinct intrusions and discusses their regional tectonic significance in the context of the known tectonic framework.
Regional framework

The dominant geological elements of eastern Iran are the CIM and the Helmand (Afghan) block, which are separated by the Sistan suture zone (Eastern Iranian Orogen), and are succeeded farther north by a series of belts of different origins and ages (Fig. 1; Bagheri and Gol 2020). The CIM consists of the Lut, Tabas and Yazd blocks, which are separated from each other by a series of broadly N–S trending faults. The oldest units of the CIM are Ediacaran to Cambrian (e.g. Ramezani and Tucker 2003; Bagheri and Stampfli 2008), but they were variously affected by subsequent orogenies, especially in connection with the Devonian to Triassic Paleotethys events, and subsequently during several Mesozoic and Cenozoic orogenic stages. The CIM is bound to the north by the Doruneh Fault system, which has a multistage kinematic history. It acted as a dextral system in response to NW–SE transpression during the Eocene and Miocene, but later became a left-lateral system controlled by the anticlockwise rotation of the CIM (Javadi et al. 2013) and N–S compression (Mattei et al. 2012; Tadayon et al. 2017, 2019). Bagheri and Gol (2020) propose that the indentation of India into Eurasia played a role on the kinematics of these faults.

Geology across the Paleotethys suture south of Mashhad

Figure 2 displays the broad geological relationships across the Paleotethys suture in the region south of Mashhad in eastern Iran. A series of idealized cross sections illustrate the main structural relationships. The general structural grain is dominated by SE trending ridges, with exposures of pre-Jurassic rocks along the ridges, separated by bands of late Mesozoic and younger rocks with many faults and thrust faults, dominantly with top to the SW kinematics. These faults generally affect also the youngest Cenozoic strata, and are thus very young structures, but some are older and related to the various Mesozoic events that affected the region, especially the late Triassic-Eo-Cimmerian Orogeny.

Important evidence for the Paleotethys suture are the ophiolitic sequences preserved at Mashhad, the Fariman Complex and the Darreh Anji Complex (Fig. 2). The latter comprises a section of Devonian (ca. 380 Ma) gabbro and minor ultramafic rocks together with local felsic intrusives, which have been interpreted to indicate subduction initiation (Shafai Moghadam et al. 2015a). The gabbros are thrust on a sequence of basaltic lava flows with marble lenses and intercalations of radiolarian chert of Permian age (Zanchetta et al. (2013). The Fariman Complex consists of a lower unit of micaschists and calcareous phyllites with some meta-basalts, marbles and serpentine, and an upper unit including Permian limestone and basaltic to andesitic lava flows interfingering with volcanoclastics and clastic sedimentary rocks (Profile B in Fig. 2; Zanchetta et al. 2013). The rocks are Permian and the sheared contact between the two units predates deposition of the Kashaf Rud Formation in the Mid-Jurassic (Zanchetta et al. 2013). The latter authors suggest a back-arc origin of the Fariman complex. By contrast, Topuz et al. (2018), based on a geochemical investigation of basaltic rocks, concluded that the Fariman Complex originated by melting of a mantle plume and interpret it as a ‘fragment of an oceanic plateau, which escaped subduction and was accreted as exotic block in the Paleotethys suture zone.’ Ultramafic rocks, gabbridiorite and pillow basalt associated with chert, marble and turbiditic rocks characterize the northern extension of this complex at Mashhad (Shafai Moghadam et al. 2015a). The assemblage also includes a flysch sequence of deep sea turbidites which was interpreted as an accretionary prism by Alavi (1991).

Near Mashhad the above unit is unconformably overlain by the late Triassic–early Jurassic ‘Mashhad Phyllite’ (Fig. 2), which is considered the equivalent of the regionally widespread lower Shemshak Group (Sheikholeslami and Kouhpeyma 2012). These rocks were then covered unconformably by Jurassic clastic deposits, in part equivalent to the Kashaf Rud Formation, and subsequently by Cretaceous sediments.

In the Aghdarband Basin, late Triassic arc-related marine sediments (Sina Formation) were covered by continental deposits of the Miankuhi Formation (Ruttner 1993; Zanchi et al. 2016; Mazaheri-Johari et al. 2021, 2022). The Miankuhi Formation was deformed and intruded by the Torbat Jam granite at 217 ± 1.8 Ma (Zanchetta et al. 2013). The whole sequence was then covered by the late to post-tectonic Kashaf-Rud Formation. The Torbat Jam granite is part of a coeval suite of granitoid plutons intruding the Mashhad ophiolite between 217 and 200 Ma (Karimpour et al. 2010; Mirnejad et al. 2013; Deyhimi et al. 2019).

South of Fariman and west of Torbat Jam a series of basement units have been upthrust through the Cenozoic cover. These units include the Sibak Complex of Ediacaran metamorphic and intrusive rocks, locally cut by Ordovician gabbro (Ranjbar Moghadam et al. 2018), in addition to fragments of Devonian, Permian and Triassic units. The locally extensive Cretaceous strata in this area represent the eastern extension of the Sabzevar ophiolitic sequences, which developed in an internal ocean basin opened in this period (Rossetti et al. 2010; Shafai Moghadam et al. 2014).

The southernmost ridge shown on the map of Fig. 2 comprises some fragments of older metasediments and
sparse intrusives, which are the main subject of the present study. The main structural grain in the area trends to the SE and was likely controlled by the Doruneh Fault system, which in this area widens into a fan of multiple NE-dipping reverse faults (Farbod et al. 2011).

**Setting and country rocks of the Posteh Pluton**

The study area is located at the northeastern edge of the CIM, in the Khorasan Razavi Province (Figs. 1, 2). It consists of a metamorphosed volcano-sedimentary sequence, intruded by the Posteh Pluton, and surrounded/covered by Eocene–Miocene sedimentary and volcanic assemblages (Figs. 2, 3).

The supracrustal sequence hosting the Posteh Pluton is non-fossiliferous. Based on regional considerations it was interpreted by Ternet et al. (1980) to be late Precambrian, but presently there is no direct evidence that would permit to verify this hypothesis. Another fault-bounded metasedimentary package further south contains sparse fossils indicating a Devonian age, but the sequence is lithologically different from that at Posteh (Ternet et al. 1980). The supracrustal rocks were subjected to regional metamorphism, which reached greenschist facies, with higher metamorphic grades superimposed along the contact metamorphic aureole of the Posteh Pluton (Ternet et al. 1980).

The metavolcanic rocks are mainly mafic and include meta-andesites and meta-basalts, with local amphibolites in the vicinity of the intrusion (Fig. 4b; Ternet et al. 1980). These rocks commonly occur as massive units, in part with thin layers or small green and greenish-grey pods. Primary textures are visible locally, with plagioclase as the major mineral, mostly replaced by secondary minerals. Other minerals are amphibole and pyroxene, pseudomorphically replaced by mica. Metamorphic minerals, including chlorite...
and epidote mostly replacing plagioclase and amphiboles, fill the vesicles in meta-basites and in crosscutting veins.

The metasedimentary rocks consist of mica-schist, marble, quartzite and conglomerate. The main components are silica-rich metapelites with interlayers of marble. The metapelites occur as lenses and thin layers with alternating bands dominated by quartz or by mica. Quartz and feldspar are completely recrystallized into a granoblastic texture oriented parallel to the foliation. The micaceous layers consist mainly of muscovite and some biotite with a lepidogranoblastic structure. There are pseudomorphs, possibly after garnet.

In the marbles and calc-silicates the minerals are mostly oriented (to the NW) and have two different grain sizes, fine- to medium-grained and coarse-grained. Quartz and feldspar are common but they are mostly fine-grained. Calcite, opaques and phlogopite are the major minerals, together with accessory quartz, olivine and amphibole. Large calcite crystals contain silicate inclusions, such as muscovite and pyroxene. White mica is generally oriented and locally surrounds calcite crystals with complex orientations.

The Poshteh Pluton

Lithologies and mineral compositions

The Poshteh Pluton is composed of two distinct intrusive units: older quartz monzonite (G1) and younger granite (G2) (Figs. 3, 4a).

The quartz monzonite (G1) consists of plagioclase (60%), K-feldspar (20–25%), quartz (5–10%), biotite (10%) and accessory zircon and apatite (Fig. 4c). Plagioclase is zoned, poly-synthetically twinned, and highly altered. The rock is approximately equigranular and locally undeformed away from the sheared domains.

Granite G2 is equigranular and massive. It consists of K-feldspar and plagioclase in amounts visually estimated at 50 and 20%, respectively, but the K-feldspar is perthitic and highly altered so that the exact proportions are difficult to determine. The norm suggests 30% orthoclase and 40% plagioclase. Quartz (about 30%) occurs as individual euhedral to subhedral grains frequently in association with clusters of green–brown hornblende (10%) and traces of brown biotite (Fig. 4d). Accessory minerals include apatite, epidote, zircon and ore minerals.

Dykes of granite G2, intruded into quartz monzonite G1, are oriented parallel to the NW-trending regional faults indicating a structural control during emplacement, but granite G2 is itself locally deformed, witnessing late reactivation stages. The shear zones are marked by a variety of mylonitic rocks ranging from protomylonitic to non-foliated cataclastic rocks, up to 100 m wide. Low-grade metamorphism in the shear zone formed sericite and chlorite along S-C structures.

Hydrothermal activity was associated with this shear zone when it was still at depth.

Small bodies of iron ore occur in the contact metamorphic aureoles of the pluton with the marbles and metavolcanic rocks. The geometry of these bodies suggests a link to the emplacement of the granite by hydrothermal metasomatism. However, iron ore veins also cut the overlying Miocene sedimentary cover, possibly reflecting late remobilization. Based on a study of fluid inclusions in the iron deposit Karimi et al. (2012) suggested an origin of the ore from hydrothermal fluids released during granite emplacement, mixing with meteoric water and precipitation during cooling.

Mineral chemistry

Chemical analyses of minerals were carried out using the CAMECA-SX-100 electron microprobe at the Iran mineral processing research center. Natural and synthetic oxides and silicate standards were used. Operating conditions were 12 kV and 15 nA with a counting time of usually 15 s. Biotite and plagioclase in the quartz monzonite (G1) (sample TR-117), and in the granite (G2) (TR-103) have been analyzed (totally 353 points, Table 1).

Biotite in both samples is primary magmatic (Fig. 5a). It has a Fe-rich composition with \( X_{Fe} = \frac{Fe^{2+}}{Fe^{2+} + Mg} = 0.75–0.1 \) and is more ferroan than magnesian (Tischendorf et al. 1997) (Fig. 5b). Biotite in G2 contains 2.26–2.96 wt% TiO\(_2\) (and so its colours change from brown to light brown. Crystallization pressures and temperatures of biotite in granitic magma can be estimated using the empirical geothermometer of Henry et al. (2005), for which the uncertainties are < 300 °C, and the Al-in-biotite geobarometer of Uchida et al. (2007), for which the uncertainties are < 1 kbar. The calculated temperature for G2 ranges between 652 and 674 °C (average 669 ± 6 °C (1σ), and the pressure between 2.6 and 3.1 kbar (average 2.9 ± 0.2 kbar (1σ)). Biotite in G1 indicates 682–694 °C (average 688 ± 8 °C (1σ)), and 2.9–3.6 kbar (average 3.5 ± 0.2 kbar (1σ)).

Plagioclase in both samples has essentially identical and very limited compositional ranges of An\(_{19.4–26.6}\) and Or\(_{0.6–1.7}\) (Table 1).

Whole rock geochemistry

Whole-rock major and trace element compositions (Table 2) were determined by X-ray fluorescence (Philips PW 1480) and inductively coupled plasma–mass spectrometry (ICP–MS) (PE 6100 DRC) at the Zanjan University geochemical laboratory. The REE were normalized to the chondritic values of Sun and McDonough (1989). For the spider diagram the trace element abundances were normalized to the primitive mantle values of Taylor and McLennan (1985).
Table 1  Mineral compositions of quartz monzonite (G1) and granite (G2) of Poshteh Pluton

|     | TR-103 (G2)       | TR-117 (G1)       |
|-----|-------------------|-------------------|
|     | Biotite            |                  |
| SiO₂ | 41.6              | 40.64             |
| TiO₂ | 2.25              | 2.66              |
| Al₂O₃ | 10.37             | 11.66             |
| FeO  | 28.72             | 26.06             |
| LiO  | 0.21              | 0.26              |
| MnO  | 0.7               | 0.02              |
| MgO  | 5.2               | 6.21              |
| CaO  | 0.06              | 0.15              |
| Na₂O | 0.28              | 0.12              |
| K₂O  | 9.15              | 8.12              |
| SrO  | BD                | BD                |
| Cr₂O₃| 0.01              | 0.01              |
| P₂O₅| 0.05              | 0.1               |
| Total| 98.6              | 98.2              |

|     | Aliv              |                  |
| Si  | 5.28              | 5.91              |
| Al²⁺| 2.72              | 2.16              |
| Al³⁺| 0.5               | 1.1               |
| Ti  | 0.27              | 0.22              |
| Cr  | BD                | BD                |
| Fe  | 3.77              | 3.51              |
| Li  | 0.08              | 0.02              |
| Mn  | 0.09              | 0.08              |
| Mg  | 1.22              | 1.17              |
| Ca  | 6.03              | 0.7               |
| Na  | 0.09              | 0.95              |
| K   | 1.83              | 2.03              |
| Y   | 5.91              | 5.7               |
| X   | 1.93              | 2.68              |
| Al total | 3.22           | 3.11              |
| Fe/Fe + Mg | BD                | BD                |
| Mn/Mn + Fe | BD                | BD                |
| Total Al | 3.22             | 3.11              |

|     | Plagioclase        |                  |
| SiO₂ | 61.28             | 62.06             |
| TiO₂ | 0.03              | BD                |
| Al₂O₃ | 23.87             | 23.87             |
| FeO  | 0.07              | 0.11              |
| CaO  | 6.03              | 2.44              |
| Na₂O | 9.29              | 8.96              |
| K₂O  | 0.18              | 0.26              |
| MgO  | BD                | BD                |
| MnO  | 0.01              | BD                |
| SrO  | BD                | BD                |
| Cr₂O₃| BD                | BD                |
| P₂O₅| 0.04              | BD                |
| Total| 100.8             | 100.77            |
The G1 quartz monzonite contains between 51 and 60% SiO₂, 11–13% Al₂O₃ and 4–6% K₂O + Na₂O. The unit is metaluminous, high-K calc-alkaline (Fig. 6) and manganous. Its chondrite normalized REE pattern is fractionated with enrichment of the LREE and almost no Eu anomalies (Fig. 7a). In the spider diagram the primitive mantle normalized pattern is quite uniform except for distinct negative anomalies for Ba, Nb and Ta and positive anomalies for Pb (Fig. 7b).

The younger G2 granite has 69–76% SiO₂, 10–13% Al₂O₃ and 5–10% K₂O + Na₂O. The rocks are ferroan, high-K calc-alkaline, and plot at the transition from metaluminous to peraluminous (Fig. 6). The REE pattern is parallel to that of G1, but at lower abundances and with a pronounced negative Eu anomaly (Fig. 7a). The data pattern in the spider diagram is also similar to that of G1, but it has lower abundances of all elements and stronger negative anomalies for Ba, Sr and Eu (Fig. 7b).

**Geochronology**

Zircon were separated from two samples by crushing, pulverizing and various enrichment stages. Before analysis the grains were chemically abraded following Mattinson (2005). The data were obtained by ID-TIMS U-Pb geochronology in Oslo, following the procedure of Krogh (1973), with details documented in Corfu (2004). Decay constants are those of Jaffey et al. (1971). Plotting was done with the program Isoplot (Ludwig 2009).
Table 2 Geochemical composition of quartz monzonite (G1) and granite (G2) of Poshteh Pluton

|    | G1  | G1  | G1  | G2  | G2  |
|----|-----|-----|-----|-----|-----|
| No | TR-107 | TR-115 | TR-117 | TR-103 | TR-105 |
| [%] |       |       |       |       |       |
| SiO₂ | 52.89 | 55.56 | 58.56 | 74.51 | 75.00 |
| TiO₂ | 1.18 | 0.76 | 0.20 | 0.09 | 0.11 |
| Al₂O₃ | 16.62 | 16.52 | 11.55 | 12.97 | 12.96 |
| Fe₂O₃ | 10.05 | 7.34 | 2.69 | 1.15 | 1.47 |
| BaO | 0.08 | 0.07 | 0.33 | 0.05 | 0.07 |
| CaO | 6.41 | 7.11 | 11.50 | 1.10 | 0.79 |
| K₂O | 2.50 | 1.88 | 3.03 | 4.40 | 4.79 |
| MgO | 4.26 | 4.41 | 0.51 | 0.15 | 0.24 |
| MnO | 0.21 | 0.16 | 0.08 | 0.03 | 0.20 |
| Na₂O | 3.61 | 3.16 | 3.08 | 4.03 | 3.70 |
| P₂O₅ | 0.21 | 0.16 | 0.06 | 0.03 | 0.03 |
| L.O.I | 1.94 | 2.86 | 8.42 | 1.44 | 0.81 |
| SUM | 99.96 | 99.99 | 100.01 | 99.95 | 99.99 |

[ppm]

|    |       |       |       |       |       |
|----|-------|-------|-------|-------|-------|
| Ag  | <0.1  | <0.1  | <0.1  | <0.1  | <0.1  |
| Al  | 80,812 | 82,940 | 58,678 | 65,153 | 63,286 |
| As  | 3.9   | 3.1   | 2.3   | 0.9   | 1.4   |
| Ba  | 574   | 498   | 2293  | 384   | 503   |
| Be  | 1.2   | 0.9   | 0.7   | 1.6   | 1.4   |
| Bi  | 0.3   | 0.3   | 0.2   | 0.2   | 0.2   |
| Ca  | 42,324 | 46,915 | 72,047 | 758   | 5181  |
| Cd  | <0.1  | <0.1  | <0.1  | <0.1  | <0.1  |
| Ce  | 22    | 21    | 20    | 40    | 47    |
| Co  | 18.9  | 16.6  | 3.6   | 2.4   | 2.1   |
| Cr  | 10    | 17    | 20    | 9     | 10    |
| Cs  | 1.9   | 1.8   | 1.2   | 1.9   | 2     |
| Cu  | 20    | 16    | 6     | 9     | 9     |
| Dy  | 3.78  | 2.53  | 0.8   | 1.83  | 1.48  |
| Er  | 2.07  | 1.46  | 1.31  | 0.86  | 0.66  |
| Eu  | 1.1   | 0.8   | 1.8   | 0.2   | 0.4   |
| Fe  | 56,974 | 43,624 | 17,739 | 7769  | 10,372 |
| Gd  | 1.6   | 1.19  | 1.16  | 1.74  | 1.65  |
| In  | <0.5  | <0.5  | <0.5  | <0.5  | <0.5  |
| K   | 15,410 | 14,620 | 19,091 | 27,769 | 26,487 |
| La  | 13    | 10    | 14    | 22    | 27    |
| Li  | 28    | 40    | 20    | 2     | 6     |
| Lu  | 0.31  | 0.2   | <0.1  | 0.14  | 0.12  |
| Mg  | 19,243 | >2%   | 3137  | 1164  | 1756  |
| Mn  | 1310  | 1090  | 516   | 213   | 186   |
| Mo  | 4.6   | <0.1  | 0.1   | <0.1  | <0.1  |
| Na  | 23,453 | 20,824 | 20,098 | 24,886 | 23,196 |
| Nb  | 7.5   | 4.7   | 3.4   | 7.3   | 6.8   |
| Nd  | 13    | 15.5  | 14.3  | 12.5  | 11.4  |
| Ni  | 5     | 5     | 4     | 2     | <1    |
| P   | 917   | 703   | 283   | 137   | 117   |
| Pb  | 9     | 2     | 4     | 7     | 9     |
| Pr  | 1.11  | 1.78  | 1.93  | 3.61  | 4.09  |
| Rb  | 78    | 80    | 41    | 147   | 123   |
| S   | 235   | 63    | 57    | 81    | 51    |
Sample SHR-104 represents the older quartz monzonite (G1) intrusion of the Poshteh Pluton. The zircon population consists largely of long prismatic crystals. Of the five analyses carried out (Table 3, Fig. 8a) two are concordant and two have some small amounts of older inherited Pb.
Table 3  U–Pb data—Poshteh Pluton

| Properties | Weight [ug] | U [ppm] | Th/U | Pbc [pg] | 206/204 | 207/235 | 2 sigma [abs] | 206/238 | 2 sigma [abs] | rho [abs] | 207/206 [abs] | 2 sigma [abs] | 206/238 [abs] | 2 sigma [abs] | 207/206 [abs] | 2 sigma [abs] |
|------------|-------------|---------|------|----------|---------|---------|---------------|---------|---------------|-----------|---------------|---------------|---------|---------------|---------------|---------|---------------|---------------|
| SHR-104 Quartz monzonite (G1), Poshteh Pluton (34°43′37″N / 60°19′40″E) | | | | | | | | | | | | | | | | | |
| Z eu lp-fr CA [2] | 55 | 1519 | 0.62 | 0.5 | 6724 | 0.25357 | 0.00115 | 0.03478 | 0.00014 | 0.94 | 0.05288 | 0.00008 | 220.4 | 0.9 | 229.5 | 0.9 |
| Z eu lp-fr CA [2] | 43 | 1258 | 0.36 | 0.8 | 3453 | 0.23939 | 0.00101 | 0.03418 | 0.00012 | 0.89 | 0.05080 | 0.00010 | 216.6 | 0.8 | 217.9 | 0.8 |
| Z eu lp CA [1] | 19 | 543 | 0.52 | 0.3 | 3481 | 0.23731 | 0.00109 | 0.03407 | 0.00012 | 0.85 | 0.05051 | 0.00012 | 216.0 | 0.7 | 216.2 | 0.9 |
| Z eu lp CA [1] | 21 | 579 | 0.53 | 0.6 | 2025 | 0.23706 | 0.00150 | 0.03402 | 0.00014 | 0.72 | 0.05054 | 0.00022 | 215.7 | 0.9 | 216.0 | 1.2 |
| Z eu lp CA [1] | 39 | 1143 | 0.43 | 2.5 | 954 | 0.23682 | 0.00126 | 0.03375 | 0.00010 | 0.67 | 0.05089 | 0.00020 | 214.0 | 0.6 | 215.8 | 1.0 |
| SHR-106 Granite (G2), Poshteh Pluton (34°43′57″N/60°19′50″E) | | | | | | | | | | | | | | | | | |
| Z fr flat CA [1] | 1 | 151 | 0.61 | 0.6 | 1380 | 0.04176 | 0.00021 | 0.00644 | 0.00002 | 0.65 | 0.04702 | 0.00019 | 41.38 | 0.11 | 41.54 | 0.21 |
| Z fr flat CA [1] | 1 | 204 | 0.69 | 1.0 | 4038 | 0.04158 | 0.00014 | 0.00641 | 0.00002 | 0.87 | 0.04707 | 0.00008 | 41.17 | 0.10 | 41.37 | 0.14 |
| Z fr flat CA [1] | 1 | 147 | 0.47 | 1.1 | 474 | 0.04225 | 0.00047 | 0.00640 | 0.00002 | 0.42 | 0.04785 | 0.00050 | 41.15 | 0.10 | 42.02 | 0.46 |
| Z fr flat CA [2] | 1 | 258 | 0.53 | 1.2 | 5096 | 0.03518 | 0.00013 | 0.00541 | 0.00001 | 0.78 | 0.04716 | 0.00011 | 34.78 | 0.08 | 35.11 | 0.13 |

(a) Z = zircon; eu = euhedral, lp = long prismatic; fr = fragment; CA = zircon treated with chemical abrasion (Mattinson 2005); [2] = number of grains analyzed
(b) weight and concentrations are known to better than 10%
(c) Th/U model ratio inferred from 208/206 ratio and age of sample
(d) Pbc = total common Pb in sample (initial + blank)
(e) raw data, corrected for fractionation and spike
(f) corrected for fractionation, spike, blank (206/204 = 18.07; 207/204 = 15.57) and initial 230Th disequilibrium, assuming Th/U (magma) = 4 (Schärer 1984); error calculated by propagating the main sources of uncertainty; The U–Pb ratio of the spike used for this work is adapted to 206Pb/238U = 0.015660 for the ET100 solution as obtained with the ET2535 spike at NIGL
together defining a discordia line with a lower intercept age of 215.8 ± 0.5 Ma (MSWD = 0.36). The line projects towards an upper intercept age of 2.4 Ga. A fifth data point plots below the line, probably because of some Pb loss.

Sample SHR-106 is a granite representing G2. The extracted zircon grains were mostly fragments of crystals, with some euhedral faces. Three of the analyses are clustered near the Concordia curve and yield an average 206Pb/238U age of 41.23 ± 0.31 Ma (Fig. 8b). There is some scatter (MSWD = 6.0), which, however, is propagated into the proportionally large uncertainty. A fourth analysis provides a much younger apparent age, which is ascribed to Pb loss.

Discussion

Quartz monzonite (G1) and the Eo-Cimmerian Orogeny

The Paleotethys opened in the Paleozoic, diachronously from west to east (e.g. Bagheri and Stampfli 2008; Stampfli et al. 2013; Torsvik and Cocks 2013; Jamei et al. 2020). Northward subduction underneath the Turan Plate (Eurasia) consumed the Paleotethys Ocean in the Carboniferous and Permian. This process is recorded by arc sequences, by ophiolites and by sedimentary rocks along, and north of the Paleotethys suture zone in northern Iran, and further east in Afghanistan and central Asia (Alavi 1991; Natal’ in and Sengör 2005; Zanchetta et al. 2013; Shafaii Moghadam et al. 2015a; Siehl 2017).

Closure of this ocean and collision of the Central Iranian terranes with the Eurasian margin occurred during the Eo-Cimmerian Orogeny. In the Aghdarband Basin (Fig. 2) continental sediments of the late Triassic Miaankuhi Formation, which lie unconformably on late Triassic marine shales (top of Sina Formation), were the youngest ones affected by this event (Mazaheri-Johari et al. 2021, 2022). Rocks and structures formed during this orogenic phase are cut by late tectonic plutons such as the 217–200 Ma Mashhad granitoids (Karimpour et al. 2010; Mirnejad et al. 2013; Deyhimi et al. 2019) and the coeval Torbat Jam granite about 100 km further to the SE (Fig. 2; Zanchetta et al. 2013). These plutons postdate major deformation and thrusting and are, thus, clearly late-tectonic. The deformed orogen was then covered by the Shemshak Group, a regionally widespread molasse-type deposit formed in the concluding stages of the Eo-Cimmerian Orogeny, in the latest Triassic and early Jurassic (Zanchi et al. 2009a; Wilmsen et al. 2009).

The G1 quartz monzonite of the Poshteh Pluton, dated here at 215.8 ± 0.5 Ma, is thus coeval with the Mashhad and Torbat Jam plutons, suggesting that it has a similar tectonic origin related to the Eo-Cimmerian collision. There are also geochemical similarities between these units. The REE pattern of the Torbat Jam granodiorite (Zanchetta et al. 2013; Ghavi et al. 2018) is parallel to that of G1, with just slightly higher abundances, and the same similarity is evident for the PM-normalized trace elements (Fig. 7a, b). The REE patterns reported by Karimpour et al. (2010) and Mirnejad et al. (2013) for various phases of the Mashhad plutons (not plotted) are also similar to those of the Poshteh quartz monzonite, with essentially no Eu anomalies and similar to slightly lower overall abundances. Mirnejad et al. (2013) distinguish an earlier phase of I type affinity related to subduction and a subsequent phase of S-type affinity related to crustal melting. Karimpour et al. (2010) concluded that the Mashhad granitoids are S-type plutons, also consistent with the presence of abundant xenocrystic zircons. Some xenocrystic zircon is also present in the Poshteh quartz monzonite, with essentially no Eu anomalies and similar to slightly lower overall abundances. Mirnejad et al. (2013) distinguish an earlier phase of I type affinity related to subduction and a subsequent phase of S-type affinity related to crustal melting. Karimpour et al. (2010) concluded that the Mashhad granitoids are S-type plutons, also consistent with the presence of abundant xenocrystic zircons. Some xenocrystic zircon is also present in the Poshteh quartz monzonite, reflecting the presence of crustal components, but the geochemistry does not conform to the S-type criteria.

The trace element plots in Fig. 9 also reveal some differences between the various Triassic plutons, implying slight diverging tectonic conditions. The plots of Rb vs. Ta + Yb and Ta vs. Yb (Fig. 9a, c) place Poshteh firmly in the volcanic arc sector whereas several of the Mashhad and Torbat Jam intrusions have higher abundances of Rb and especially of Ta, which place them close to, or inside the
syn-collisional field. The plot of Th/Yb vs. Nb/Yb (Fig. 9e) also shows a distinct deviation between Poshteh and the Mashhad and Torbat Jam intrusions, with a stronger affinity to continental crust of the latter. These trends likely reflect variations in geographical terms with respect to the tectonic suture, possibly combined with slight temporal trends related to the progress of the convergence and collision.

The anomalous position of the Postheh Pluton
south of the main Paleotethys suture

An Eo-Cimmerian syn-collisional origin of the Poshteh G1 quartz-monzonite is thus supported by the timing and composition of this unit, and similarities with the plutons in the Mashhad–Torbat Jam area. There is, however, a major difference in the setting. The Mashhad–Torbat Jam granitoids were emplaced in an assemblage of Devonian to Permian ophiolites, are volcanic rocks and a major flysch accretionary complex, all located inside and north of the Paleotethys suture (Fig. 2). By contrast, the Postheh Pluton is located south of the Paleotethys suture where it intruded a sequence of supracrustal rocks presumed to be Proterozoic by Ternet et al. (1980) but still undated due to the lack of fossils.

A potential explanation could be that the analogy to the Mashhad–Torbat Jam rocks is just accidental, and that the Poshteh quartz monzonite developed in a different setting altogether, within the colliding plate but still as an expression of the same orogenic process.

The alternative, and probably more realistic, interpretation is that the Postheh quartz monzonite and surrounding rocks represent a tectonically disrupted allochthonous fragment of the Eo-Cimmerian Orogen (Fig. 10). This explanation has been proposed for the Anarak, Jandaq and Poshteh-Badam metamorphic complexes, which are located in the northwestern part of the CIM south of the Doruneh Fault (Fig. 1; Bagheri and Stampfli 2008). Another example is the coeval late Triassic suite of tonalite–granite documented in the Saghand area on the western side of the CIM (Fig. 1; Ramezani and Tucker 2003). The Anarak Metamorphic Complex and associated units comprise a range of lithologies including metapelites, marbles, mafic volcanic rocks, locally with pillow lavas and ultramafic rocks. Units of the Anarak Metamorphic Complex preserve a record of Carboniferous and Permian magmatism and tectonism and locally high pressure metamorphism (Bagheri and Stampfli 2008; Zanchi et al. 2009a, 2015). Carboniferous orogenic activity
has been documented from the Jandaq segment (Berra et al. 2017).

Based on paleomagnetic and geological data Davoudzadeh et al. (1981) deduced that the CIM had undergone a 135° anticlockwise rotation since the Triassic. They suggested that the rotation could have sliced off a margin of the Turanian Plate and its orogenic cover and moved it to its present location at Nakhlak. Muttoni et al. (2009), however, expressed a note of caution on the magnitude of this rotation. More recent paleomagnetic studies (Mattei et al. 2012, 2015) confirm the occurrence of counterclockwise rotations in the CIM, both in the early Cretaceous and the Miocene, although of lesser magnitude than previously suggested. The position of the exotic Eo-Cimmerian crustal fragments has been interpreted by Bagheri and Stampfli (2008) as the consequence of Cretaceous extension in the Neotethys ocean system, which opened a back-arc basin, forming the Sabzevar and related ophiolites and detach- ing Eo-Cimmerian fragments from the margin of the Turan plate. The displacement was further enhanced by fragmentation and faulting between the various blocks of the CIM and considerable anti-clockwise rotation of the block reflecting the stress caused by the larger scale interaction between the Eurasian, Arabian and Indian plates. This events would have transferred fragments of the Paleotethys suture. Zanchi et al. (2021) expand this concept by linking the two stage counterclockwise rotations in the CIM, both in the early Cretaceous and the Miocene, forming a common Sabzevar-Nain ophiolite. The rotation also disrupted elements of the Paleotethys suture and carried them westward along the margin of the CIM. A second stage of counterclockwise rotation in the Cenozoic split the Nain from the Sabzevar ophiolite, moving it southwestward to its present position. and also transferred the Anarak and Nakhlak Paleotethys elements.

Alavi et al. (1997) proposed an alternative hypothesis whereby the exotic orogenic fragments such as Nakhlak were originally part of allochthonous nappes thrust onto the CIM plate, and subsequently further disrupted by plate rotations and faulting.

For the exotic Postheh occurrence dated in this study it is not possible to be certain on the preferred mechanism that was responsible for the present location south of the Paleotethys suture. The quartz monzonite and host rocks are entirely isolated in the expanse of post-Cretaceous rocks, which were themselves also extensively disrupted by faulting and thrusting (Fig. 2). The Alavi hypothesis, which combines Eo-Cimmerian nappe thrusting and later strike slip faulting, would seem to be the more plausible process. In its simplest terms, the alternative model of Zanchi et al. (2021), which proposes the transport of Paleotethys suture fragments to their present locations piggy-back on slices of the CIM during two stages of post-Triassic anticlockwise rotation, would imply an origin of Postheh south of the Paropamisus belt (Fig. 1). This zone, representing the Paleotethys suture in Afghanistan, comprises a row of late Triassic plutons, continuing the trend set in the Mashhad region (Siehl 2017). The domains south of the Paropamisus are highly fragmented and faulted. The Doruneh Fault system widens eastward and appears to terminate at the important N–S fault near the boundary to Afghanistan (Fig. 1). In the Ferdows thrust and fold system, west of Poshteh (Fig. 1), Mattei et al. (2020) document local clockwise rotation. This suggests that combinations of left- and right-lateral faulting in the eastern Doruneh Fault system may have controlled the translation of Poshteh but the potential links remain speculative. Detailed structural analyses of the eastern Doruneh Fault termination should provide key information on these processes.

### Granite (G2) and the Eocene magmatic stage

Emplacement of the younger granite G2 at 41.23 ± 0.31 Ma coincides with a period of intense magmatism across the entire region, and especially in the Sistan suture zone (Camp and Griffis 1982; Shafaroudi et al. 2013; Golmohammadi et al. 2015; Mohammadi et al. 2016; Javidi Moghaddam et al. 2020; Shahbazi et al. 2021; Rezaei-Kakhae et al. 2021). There are geochemical similarities between the G2 granite of the Poshteh Pluton and similar coeval intrusions such as at Sangan (Fig. 1), for example, the gently sloping REE pattern with distinct negative anomalies and negative anomalies of Nb and Sr (Fig. 7; Golmohammadi et al. 2015). In the plots of Rb vs. Ta + Yb and Ta vs. Yb (Fig. 9b, d) the Postheh granite corresponds broadly to the position of other coeval intrusions though the region, albeit somewhat displaced toward lower Yb and Ta + Yb abundances, still in the volcanic arc field but already leaning towards a syncollisonal character. The plot of Th/Yb vs. Nb/Yb (Fig. 9f) indicates a pronounced affinity with continental crust, similar but more extreme than the other comparable intrusions.

Golmohammadi et al. (2015) concluded that the chemical composition of the granitoids indicates an origin related to subduction. Zarrinkoub et al. (2012), however, pointed out that a subduction origin of this suite was not possible, because other evidence suggested closing of the Sistan Ocean already in the late Cretaceous. They proposed instead that this magmatic phase was related to orogenic collapse and detachment of thickened lithosphere, an analogous conclusion as reached by Pang et al. (2013), and by Rossetti et al. (2014) for central Iran. A process involving lithospheric delamination and asthenospheric upwelling causing widespread melting of the sub-contontental lithospheric mantle is also favoured by Omidianfar et al. (2020).
In the Poshteh Pluton the G2 intrusions were to some degree controlled by the northwest structure related to the Doruneh system. suggesting a close connection between regional deformation and magmatism. Bagheri and Gol (2020) related oroclinal buckling in the East Iranian Orogen to westward escape of the Afghan (Helmand) block as a consequence of the collision of India with Eurasia. These authors interpret the sequence of tectonic and magmatic events to result from lithospheric delamination, which caused the widespread regional Eocene to Miocene magmatic activity. The processes observed in the Poshteh Pluton can well be explained by this model.

Conclusions

The Poshteh area near Taybad, at the northeastern margin of the CIM, contains a fragment of the Eo-Cimmerian Orogen marked by late tectonic quartz monzonite (G1), which intruded at 215.8 ± 0.5 Ma during the collisional stage of the orogeny after closing of the Paleotethys. The quartz monzonite is geochemically similar to coeval plutons near Mashhad and Torbat Jam. The latter intruded Carboniferous-Permian arc sequences, ophiolites and flysch, which formed at the Paleotethys convergent margin. The Poshteh quartz monzonite (G1) is instead located south of the main Paleotethys suture. It may represent an Eo-Cimmerian element developed within the colliding upper plate, but then tectonically transferred to the lower plate, in a similar way as the occurrences farther west at Anarak and Saghard.

The younger G2 granite in the Poshteh Pluton formed at 41.23 ± 0.31 Ma. Its emplacement was also accompanied by widespread hydrothermal alteration that developed locally important Fe mineralization in the country rocks. The geochemical composition of this granite is comparable to that of coeval and abundant late Eocene to Oligocene plutons in the region. It can be attributed to widespread melting of the subcontinental lithosphere related to delamination, presumably linked to the formation of the Eastern Iranian Orocline as suggested by Bagheri and Gol (2020).

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