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Age and nature of 560–520 Ma calc-alkaline granitoids of Biarjmand, northeast Iran: insights into Cadomian arc magmatism in northern Gondwana

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ABSTRACT
The Biarjmand granitoids and granitic gneisses in northeast Iran are part of the Torud–Biarjmand metamorphic complex, where previous zircon U–Pb geochronology show ages of ca. 554–530 Ma for orthogneissic rocks. Our new U–Pb zircon ages confirm a Cadomian age and show that the granitic gneiss is ~30 million years older (561.3 ± 4.7 Ma) than intruding granitoids (522.3 ± 4.2 Ma; 537.7 ± 4.7 Ma). Cadomian magmatism in Iran was part of an approximately 100-million-year-long episode of subduction-related arc and back-arc magmatism, which dominated the whole northern Gondwana margin, from Iberia to Turkey and Iran. Major REE and trace element data show that these granitoids have calc-alkaline signatures. Their zircon O isotopes (~7.9 to +5.5; one point with εHf ~ -17.4) as well as bulk rock Nd isotopes (εNd(t) = –3 to –6.2) show that these magmas were generated via mixing of juvenile magmas with an older crust and/or melting of middle continental crust. Whole-rock Nd and zircon Hf model ages (1.3–1.6 Ga) suggest that this older continental crust was likely to have been Mesoproterozoic or even older. Our results, including variable zircon εHf(t) values, inheritance of old zircons and lack of evidence for juvenile Cadomian igneous rocks anywhere in Iran, suggest that the geotectonic setting during late Ediacaran and early Cambrian time was a continental magmatic arc rather than back-arc for the evolution of northeast Iran Cadomian igneous rocks.

1. Introduction
Peri-Gondwana terranes in the Alpine–Himalayan orogenic belt of Europe and Asia are characterized by Ediacaran–Cambrian ‘Cadmian’ arc-type igneous rocks (Fernandez-Suarez et al. 2002, 2003, 2014; Ustaomer et al. 2009, 2011). The Cadomian belt stretches from Iberia through central and southeast Europe into Turkey and Iran (Shafai Moghadam et al. 2015; Avigad et al. 2016) and may continue into the Qingtang terrane of Tibet (Wang et al. 2016). Formation of the Cadomian arc followed earlier collisional events that produced the supercontinent Gondwana (Powell et al. 1993; Dalziel 1997) and reflects subduction along its entire northern periphery to form the several-thousand-kilometre-long Cadomian–Avalonian arc (Kroner and Romer 2013; Garfunkel 2015). The Cadomian orogeny centred a marginal tectono-magmatic system of the Western Pacific style (Drost et al. 2011). Nearly all Cadomian crust rifted away from Gondwana during Permian–Triassic time, including that of Iran (Berberian and King 1981; Robertson et al. 1991; Garfunkel 2004) and accreted to the southern flank of Eurasia.

Cadmian igneous and metamorphic rocks comprise most of the basement of Iran. Late Neoproterozoic sedimentary rocks comprise a subordinate part of the basement. Exposures are documented from the west (Golpayegan), northwest (Khoy–Salmas, Zanjan–Takab), northeast (Torud–Biarjmand, Taknar), north (Lahijan granites), and central Iran (Saghand and Zaran) (Figure 1). This old crust comprises the core of the Palaeozoic ribbon continent ‘Cimmeria’ (Sengor 1991). Understanding the age, nature, and distribution of Cadomian crust is key to reconstructing the tectonic evolution of Iran. Our understanding of Cimmerian...
blocks and their Cadomian basement is advancing rapidly but remains incomplete, especially in Turkey and Iran. Reliable geochronological data for metagneous rocks of Cadomian terranes in Iran have only recently begun to be reported (Badr et al. 2013; Balaghi Einalou et al. 2014; Rossetti et al. 2015; Shafaii Moghadam et al. 2015).

In this study, we advance our understanding of the Cadomian basement of Iran by presenting and discussing the significance of new in situ LA-ICP-MS and SIMS zircon U–Pb ages as well as zircon Hf–O and bulk rock Sr–Nd isotope data from the Biarjmand granitoids of northeast Iran. Our previous study focused on the gneissic rocks (ortho-and paragneissic) from north of Sahad and north and south of Delbar. In this study we focus on poorly known granitic rocks and associated gneisses north of Dochah (Figure 2). To provide context for this study, we also compile all zircon U–Pb and Hf–O isotope data from Iranian Cadomian terranes and use these combined data to better understand late Neoproterozoic arc magma formation and evolution.

2. Geological setting

The Cadomian basement of Iran is characterized by both felsic igneous rocks – and their metamorphic equivalents – and detrital sedimentary successions. The oldest igneous rocks have Ediacaran–early Cambrian ages (520–600 Ma) (Ramezani and Tucker
Cadomian basement was deeply buried by younger sediments and exhumed during Cenozoic extension (Stockli 2004; Kargaranbafghi et al. 2015; Moghadam et al. 2016). Most Cadomian gneissic exposures show evidence of Eocene-Oligocene partial melting to form anatexite-diatectite and melt invasion and are often intruded by Eocene–Miocene plutonic rocks (Ramezani and Tucker 2003). Petrographic and geochemical data (bulk rock major and trace elements) show that most Cadomian igneous rocks in Iran are felsic plutons – with minor rhythmic extrusive rocks – and on a basis of trace element considerations belong to ‘volcanic arc granite’ suites (VAG) (Badr et al. 2013; Balaghie Einalou et al. 2014; Hosseini et al. 2015; Rossetti et al. 2015; Moghadam et al. 2015a). Moreover, bulk rock Nd and zircon Hf isotopic compositions of these rocks indicate that they formed by partial melting of older continental crust or by mixing between juvenile mantle-derived melts and continental crust, although exposures of this older crust are so far unknown (see Section 6). Interaction between juvenile melt and older crust is also suggested by Abbo et al. (2015) for isotopically evolved Cadomian igneous rocks. Remelting of Gondwana-derived sediments deposited in the back-arc basin is also suggested as an alternative process for the observed Hf isotope trends and Hf model ages in Cadomian igneous rocks of Turkey (Zlatkin et al. 2013). However, remelting of clay-and sand-rich graywackes is not confirmed by O and Hf isotopic composition of Cadomian zircons (see the next section).

In this article we contribute to our understanding of Cadomian crust formation in Iran by presenting new data for a representative example from the Torud–Biarjmand area of northeast Iran (Figure 1).

The study area is situated between the Alborz Mountains in the north and the Lut–Tabas block to the south. Torud–Biarjmand basement rocks are exposed over about 5000 km²; we refer to combined Cadomian basement exposures here as the Torud–Biarjmand complex. To the north-northwest, the Torud–Biarjmand complex is surrounded by Eocene magmatic rocks and by Palaeozoic phyllite, schist, and marble. The contacts between Palaeozoic rocks and Cadomian rocks are faulted and sometimes covered by Quaternary alluvium. The Torud–Biarjmand complex contains greenstreak amphibolite-facies meta-igneous and meta-sedimentary rocks and is overlain by un-metamorphosed Jurassic and Cretaceous sedimentary rocks. The age of metamorphism of the Biarjmand rocks is unclear, as metamorphic rims are absent in our zircons. However, Ar–Ar and K–Ar ages on muscovite and biotite grains from Torud orthogneiss yield ages of ca. 160 and 171 Ma, respectively (Rahmati-Ikhchi et al. 2010). The region was affected by four major deformation phases during Cadomian, Mid Jurassic, Cretaceous and Neogene (Rahmati-Ikhchi et al. 2010). The Torud–Biarjmand complex contains both psammitic to volcanogenic metasediments (paragneiss, mica schist, garnet-mica schist and amphibolite) and igneous felsic to mafic plutonic rocks (granitoids, gabbros and their metamorphic equivalent (orthogneiss)). Less
deformed granitic bodies intrude more deformed orthogneiss of broadly similar composition. Outcrops near Sahad in the south and Delbar in the east (Figure 2) have been studied by Maryam et al. (2014) and Moghadam et al. (2015b) (Figure 2). Here, U–Pb zircon dating of orthogneissic rocks yield late Neoproterozoic–early Cambrian (530–550 Ma) crystallization ages. Metamorphosed felsic and mafic (amphibolite) dikes and sills yield zircon U–Pb ages of ca. 534–554 Ma (Hosseini et al. 2015). Paragneissic rocks mostly consist of alternating quartzo–feldspathic and foliated biotite-rich layers. Garnet mica schists are similar to mica-rich paragneisses. Zircons from Biarjmand mica-schists show $^{206}\text{Pb}/^{238}\text{U}$ ages of 549–551 Ma (Maryam et al. 2014). In the Biarjmand region, fine-grained schists with marble intercalations are also common. Coarse-grained, non-metamorphosed granitic to granodioritic intrusions are abundant in the Biarjmand region (Figures 2 and 3A). The granitoids are either enriched in K-feldspar with minor ferromagnesian minerals or contain abundant amphibole and biotite. They show both coarse-grained proto-granular and/or mylonitic textures. Aplitic and dioritic dikes intrude the Biarjmand granitic pluton (Figure 3B). Garnet-rich granitic mylonites with quartz, feldspar, and biotite stretching lineations are abundant in the Torud–Biarjmand region. Mylonite is especially obvious at the granite contacts with metasedimentary host rocks. Granitic gneiss is common and characterized by stretching foliation of quartz, amphiboles and micas. Gabbros are a minor component of the Torud–Biarjmand complex and are petrographically classified as olivine gabbros (olivine + Cpx + plagioclase), mylonitic gabbros (Cpx + plagioclase ± K-feldspar augens) and garnet bearing gabbros (Cpx + plagioclase + garnet) (Figure 3D). Gabbroic rocks are cross-cut by younger granitic dikes (Figure 3C).

We studied orthogneiss and granitoids in a large exposure north of Dochah (Figure 2) in terms of major and trace element compositions, U–Pb geochronology, whole rock Sr–Nd and zircon Hf–O isotopes geochemistry.

![Figure 3](https://example.com/fig3.jpg)

**Figure 3.** (A) Hillside exposure of Biarjmand Cadomian granites. (B) Centimetric aplitic dikes injected into orthogneiss. (C) Close-up of gabbros with intruding granitic dikes in the Torud area. (D) Olivine + clinopyroxene and plagioclase in Torud gabbros. (E) and (F) Quartz + K-feldspar + biotite + plagioclase in the Biarjmand granodiorites.
3. Petrography and mineral assemblages

Because our area contains both metamorphic rocks with igneous and sedimentary protoliths and also non-metamorphosed igneous rocks, for clarity we modify IUGS QAP modal nomenclature for non-metamorphosed felsic igneous rocks (e.g. granite) and igneous prefix plus metamorphic terms (e.g. granitic gneiss) for our metamorphic rocks.

Studied granitoids and granitic gneisses show protogranular to mylonitic textures and contain alkali feldspar, plagioclase, quartz, and biotite with rare amphibole. Samples with more alkali feldspar are monzogranite (Figure 4), whereas granodiorite, tonalite and quartz monzodiorite contain more plagioclase and amphibole. Chlorite, epidote, clinozoisite, sericite, titanite and iron oxides are present as secondary minerals. Granitic gneisses contain large (perthitic) K-feldspar grains, highly deformed and recrystallized quartz, zoned plagioclase, biotite and allanite. Titanite + clinozoisite + epidote are secondary minerals (Figures 3E–F). Aplitic dikes have perthitic K-feldspar, quartz and minor biotite and plagioclase, whereas dioritic dikes contain plagioclase and hornblende with minor quartz, biotite and K-feldspar.

In the QAP diagram (Streckeisen 1976), the Biarjmand granitoids and granitic gneisses plot predominantly in the field of monzogranite with four exceptions: two granodiorites (BJ11–26 and BJ11–27), a tonalite (BJ11–22) and a quartz monzodiorite (BJ11–12) (Figure 4).

4. Analytical procedures

Major and trace elements were analysed at Actlabs, Canada (www.actlabs.com) using ICP-AES and ICP-MS. The uncertainty (1σ) is ~2% for major elements and 5–10% for trace elements (depending on concentration). Major and trace elements are presented in Supplementary Document 1 (see http://dx.doi.org/10.1080/00206814.2016.1166461 for supplementary documents).

Sr and Nd isotopic composition were determined at the Laboratório de Geologia Isotópica da Universidade de Aveiro, Portugal. The selected powdered samples were dissolved with HF/HNO3 in Teflon Parr acid digestion bombs at 200°C. After evaporation of the final solution, the samples were dissolved with 6 N HCl and evaporated to dry cake, then redissolved for chromatographic separation. The elements for analysis were purified using a conventional two-stage ion chromatography technique: (i) separation of Sr and REE elements in ion exchange column with AG8 50 W Bio-Rad cation exchange resin; (ii) purification of Nd from other lanthanide elements in columns with Ln resin (ELChroM Technologies) cation exchange resin. All reagents used in sample preparation were sub-boiling distilled, and pure water was produced by a Milli-Q Element (Millipore) apparatus. Sr was loaded with H3PO4 on a single Ta filament, whereas Nd was loaded with HCl on a Ta side filament in a triple filament arrangement. 87Sr/86Sr and 143Nd/144Nd isotopic ratios were determined using a multi-collector thermal ionization mass spectrometer (Timsic and Patterson 2014) VG Sector 54. Data were obtained in dynamic mode with peak measurements at 1–2 V for 86Sr and 0.5–1 V for 144Nd. Sr and Nd isotopic ratios were corrected for mass fractionation relative to 88Sr/86Sr = 0.1194 and 146Nd/144Nd = 0.7219. During this study, the SRM–987 standard gave a mean value of 87Sr/86Sr = 0.710255 ± 23 (N = 10; 95% CI) and the Nd–1 standard yielded 143Nd/144Nd = 0.5121009 ± 66 (N = 12; 95% CI). Nd model ages were calculated according to the procedure of DePaolo (1981). The one-stage model age (TDm) was calculated assuming Nd isotopic growth of the depleted mantle reservoir from εNd(t) = 0 at 4.56 Ga to εNd(t) = +10 at present, according to the following equation:

\[
T_{DM} = \frac{1}{\lambda} \ln \left( 1 + \frac{[^{143}\text{Nd}/^{144}\text{Nd}]_s - 0.51315}{[^{147}\text{Sm}/^{144}\text{Nd}]_s - 0.21377} \right)
\]

where \( \lambda = 6.54 \times 10^{-12} \text{ a}^{-1} \) and \( [^{143}\text{Nd}/^{144}\text{Nd}]_s \) and \( [^{147}\text{Sm}/^{144}\text{Nd}]_s \) are the measured ratios of samples. Bulk rock Sr–Nd isotopic data are presented in Supplementary Document 2.

Zircon U–Pb dating analysis was carried out using both LA-ICP-MS and SIMS. LA-ICP-MS analyses were done on an Agilent 7500a ICP-MS equipped with a 193 nm laser at the Institute of Geology and Geophysics, Chinese Academy of Sciences (IGG–CAS) using the analytical procedures described by Xie et al. (2008). Zircon 91,500 and GJ–1 were used as external calibration standards and the standard silicate glass NIST 610 was used to optimize the
Common Pb correction followed the method described by Andersen (2002), since the signal intensity of \(^{206}\text{Pb}\) was much lower than the other Pb isotopes and there is a large isobaric interference from Hg. Raw count rates for \(^{29}\text{Si}\), \(^{204}\text{Pb}\), \(^{206}\text{Pb}\), \(^{207}\text{Pb}\), \(^{208}\text{Pb}\), \(^{232}\text{Th}\) and \(^{238}\text{U}\) were collected for age determination. U, Th and Pb concentrations were calibrated using \(^{29}\text{Si}\) as an internal calibrate and NIST 610 as reference material. \(^{207}\text{Pb}/^{206}\text{Pb}\) and \(^{206}\text{Pb}/^{238}\text{U}\) ratios were calculated using the GLITTER program. Average \(^{207}\text{Pb}/^{206}\text{Pb}\), \(^{206}\text{Pb}/^{238}\text{U}\) and \(^{208}\text{Pb}/^{232}\text{Th}\) ratios in zircon 91,500 obtained during our analytical session were used to calculate correction factors. These correction factors were then applied to each sample to correct for both instrumental mass bias and depth-dependent elemental and isotopic fractionation. Ages were determined and concordia plots were constructed using ISOPLOT (Ludwig 2003). The zircon U–Pb age results are listed in Supplementary Document 3.

SIMS U–Pb analyses were performed on a Cameca IMS–1280HR at IGG–CAS using standard operating conditions (7-scan duty cycle, ~8 nA primary \(\text{O}_2^–\) beam, 20 × 30 μm analytical spot size, mass resolution ~5400). U–Th–Pb ratios and absolute abundances were determined relative to the standard zircon Plešovic (337 Ma, Sláma et al. (2008)) and M257 (U = 840 ppm, Th/U = 0.27, Nasdala et al. (2008)), respectively. Measured Pb isotopic compositions were corrected for common Pb using \(^{204}\text{Pb}\). Average Pb of present-day crustal composition (Stacey and Kramers 1975) was used for the common Pb assuming that this was due to surface contamination during sample preparation. A long-term uncertainty of 1.5% (1 RSD) for \(^{206}\text{Pb}/^{238}\text{U}\) measurements of the standard zircon was propagated to the unknowns (Li et al. 2010). In order to monitor the external uncertainties of SIMS U–Pb measurements, analyses of an in-house zircon standard Qinghu were interspersed with unknowns. Nine analyses yield a weighted mean \(^{206}\text{Pb}/^{238}\text{U}\) age of 160.7 ± 1.8 Ma, identical within errors to the reported age of 159.5 ± 0.2 Ma (Li et al. 2013). Weight mean \(^{206}\text{Pb}/^{238}\text{U}\) age of 337.6 ± 2.3 Ma was also obtained for standard Plešovic. Uncertainties on individual analyses in the data table are reported at a 1σ level; mean ages for pooled U/Pb and Pb/Pb analyses are quoted with 95% confidence interval. Data reduction was carried out using an Isoplot/Ex v. 2.49 program (Ludwig 2003). Zircon U–Th–Pb isotopic data are presented in Supplementary Document 4.

In situ oxygen isotope analyses were conducted on previously dated zircons using the same Cameca IMS-1280 SIMS at IGG–CAS. After U–Pb dating, the sample mount was reground to ensure that any oxygen implanted in the surface from the \(\text{O}_2^–\) beam used for U–Pb analysis was removed. The \(\text{Cs}^+\) primary ion beam was accelerated to 10 kV, with an intensity of ~2 nA corresponding to a beam size of ~10 μm in diameter. A normal-incidence electron flood gun was used to compensate for sample charging during analysis. Negative secondary ions were extracted with an ~10 kV potential. Oxygen isotopes were measured using the multi-collection mode on two off-axis Faraday cups. One analysis consists of 16 cycles, with an internal precision generally better than 0.4‰ (2SE) on the \(^{18}\text{O}/^{16}\text{O}\) ratio. The detailed analytical procedures are similar to those reported by Li et al. (2010, 2013). Oxygen isotope ratios are expressed as delta notation \(\delta^{18}\text{O}\), representing deviation of measured \(^{18}\text{O}/^{16}\text{O}\) values from that of the Vienna Standard Mean Ocean Water (\(^{18}\text{O}/^{16}\text{O}\) VSMOW = 0.0020052) in parts per thousand. The internal precision from one spot analysis is typically better than 0.4‰ for \(^{18}\text{O}/^{16}\text{O}\) ratio (2SE). The results were corrected for instrumental mass fractionation factor (IMF), following the equation: \(\delta^{18}\text{O}_{\text{Corrected}} = \delta^{18}\text{O}_{\text{Measured}} - \text{IMF}\). IMF is monitored in terms of the difference between measured and recommended oxygen isotopic compositions of Penglai zircon standard with a \(\delta^{18}\text{O}\) value of 5.31‰ (Li et al. 2010). Zircon O isotopic data are presented in Supplementary Document 5.

In situ zircon Lu–Hf isotopic analysis was carried out on a Neptune multi-collector ICP-MS equipped with a Geolas–193 laser-ablation system at IGG–CAS, and the analytical procedures were similar to those described by Wu et al. (2006). Lu–Hf isotopic analyses were obtained on the same zircon grains that were previously analysed for U–Pb and O isotopes, with ablation pits of 63 μm in diameter, ablation time of 26 seconds, repetition rate of 10 Hz and laser beam energy density of 10 J/cm². During laser ablation analyses, the isobaric interference of \(^{176}\text{Lu}\) on \(^{176}\text{Hf}\) was negligible due to extremely low \(^{176}\text{Lu}/^{177}\text{Hf}\) in zircons. Isobaric interference of \(^{176}\text{Yb}\) on \(^{176}\text{Hf}\) is corrected using independent mass bias factors for Hf and Yb for correction. During analysis, an isotopic ratio of \(^{176}\text{Yb}/^{172}\text{Yb}\) = 0.5887 was applied (Wu et al. 2006). Measured \(^{176}\text{Hf}/^{177}\text{Hf}\) ratios were normalized to \(^{179}\text{Hf}/^{177}\text{Hf}\) = 0.7325. The measured \(^{176}\text{Lu}/^{177}\text{Hf}\) ratios and the \(^{176}\text{Lu}\) decay constant of 1.865 × 10⁻¹¹ yr⁻¹ reported by Scherer et al. (2000) were used to calculate initial \(^{176}\text{Hf}/^{177}\text{Hf}\) ratios. The chondritic values of \(^{176}\text{Hf}/^{177}\text{Hf}\) = 0.0332 and \(^{176}\text{Lu}/^{177}\text{Hf}\) = 0.282772 reported by Bilchert-Toft and Albareda (1999) were used for the calculation of εHf values. The present-day \(^{176}\text{Hf}/^{177}\text{Hf}\) = 0.28325 and \(^{176}\text{Lu}/^{177}\text{Hf}\) = 0.0384 values (Griffin et al. 2004) were used to calculate the depleted mantle model age (TDM). Because zircons were formed in a granitic magma derived from crustal sources, we used the crustal residence model ages (T²DM). The mean
$^{176}$Lu/$^{177}$Hf ratio of 0.0093 for the upper continental crust was used to calculate the crustal residence model age ($T_{DM}$). The zircon Lu–Hf isotopic results are listed in Supplementary Document 5.

5. Results

5.1. Bulk rock major and trace elements geochemistry

Fifteen samples including granitoids (11 samples) and granitic gneisses (four samples) were selected for whole rock geochemistry (Supplementary Document 1). The rocks are similarly strongly felsic, with SiO$_2$ contents of 74.2–78.1 wt.% along with low MgO abundances (0.08–0.8 wt.%), except sample BJ11–12 (quartz monzodiorite) with SiO$_2$ = 53.9 wt.%, MgO = 5.4 wt.% and high LOI content (1.5 wt.%). K$_2$O content of these rocks is variable but mostly high; 0.5–5.1 wt.%. All samples are peraluminous A/CNK (molar Al$_2$O$_3$/CaO + Na$_2$O + K$_2$O) > 1 (Figure 5A). In the Rb vs. Nb + Y tectonic discriminant diagram (Pearce et al. 1984), the samples plot as volcanic arc granites (VAGs) (Figure 5B), except sample BJ11–24 (tonalite) with higher Nb and Y, perhaps reflecting abundant titanite and amphibole. Biarjmand granitoids are variably enriched in light rare earth elements (LREEs) relative to heavy rare earth elements (HREEs) – $La_{(n)}/Yb_{(n)}$ ~ 0.7–17.3 – in the chondrite-normalized diagram (Figure 6). Granitoids show conspicuous depletion in Eu, indicating equilibrium with feldspar in residue or cumulate. These rocks have positive Th, U, Rb anomalies and negative Nb, Ta and Ti anomalies compared to N-mid-oceanic ridge basalts (MORB) (Figure 6), further suggesting formation of these melts at a convergent plate margin.

Granitic gneisses share most trace element characteristics with granitic rocks, except that they show trace element affinities with adakites. They are more enriched in LREEs relative to HREEs ($La_{(n)}/Yb_{(n)}$ ~ 3.9–20.6) and lack Eu depletion. Granitic gneisses and granitoids are easily distinguished using Sr/Y vs. Y and $La_{(n)}/Yb_{(n)}$ vs. Yb$_{(n)}$ diagrams (Figure 5C and D). Gneissic rocks have low Y and Yb contents and overlap with adakites in their Yb$_{(n)}$ and $La_{(n)}/Yb_{(n)}$ relationships. They also show greater enrichments in Ba and Sr along with greater depletions in Nb–Ta–Ti, but both have trace element patterns that are consistent with formation at a convergent plate margin.

5.2. Bulk rock Sr–Nd isotopes

Sr–Nd isotopic analyses of the six Biarjmand samples – including three monzogranites, two granodiorites, and...
an orthogneiss – are presented in Supplementary Document 2 and plotted in Figure 7.

The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ($t = 520$ Myr) for Biarjmand gneiss and granitoids ranges between 0.703 and 0.714 (Supplementary Document 2), but this scatter may reflect the high Rb/Sr of these rocks, their low Sr contents (20–163 ppm), the choice of 520 Ma as the time of formation for calculating initial Sr isotope composition, and their subsequent deformation and alteration. The rocks yield a whole rock Rb–Sr errorchron age of 511 ± 14 Ma (MSWD = 2.0 and initial $^{87}\text{Sr}/^{86}\text{Sr} = 0.70412 ± 0.00077$).

Nd isotopes are more resistant to alteration and other concerns for Rb/Sr. The $\varepsilon$Nd of our samples varies between $-3$ and $-6.2$. Cadomian granitoids and gneiss of the Torud–Biarjmand complex have low $^{147}\text{Sm}/^{144}\text{Nd}$ (0.09–0.147) that are suitable for calculating Nd model ages, except for monzogranite sample BJ11–8. Except for BJ11–8, they give Nd model ages ($T_{DM}$) that range from 1.28 to 1.62 Ga (Supplementary Table 2). On a $^{143}\text{Nd}/^{144}\text{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$ plot (Figure 7), the Biarjmand granitoids are isotopically similar to other Cadomian gneissic rocks from both the Torud–Biarjmand complex (northeast Iran) and the Zanjan–Takab meta-igneous rocks (northwest Iran). The $\varepsilon$Nd ($t = 520$) values of Cadomian...
granites and gneissic rocks of the Torud–Biarjmand complex (−2.2 to −5.8) are also similar to Cadomian granites from the Bitlis massif (Turkey) (εNd = −1.2 to −2.9) (Ustaomer et al. 2009). Cadomian gneissic rocks are isotopically different from Cadomian ophiolitic lavas from Central Iran, which show highly radiogenic Nd values (Figure 7). They are also different from Cretaceous ophiolites and Eocene granitic rocks (Figure 7).

5.3. Zircon U–Pb geochronology

Three samples were dated: one granitic gneiss, granodiorite and monzogranite. These results are discussed below.

5.3.1. Sample BJ11–8 (monzogranite)
Zircons are euhedral to subhedral with long to short prismatic and even stubby shapes (av. ~70–150 μm). Zircons from this sample show evidence of inheritance in their cores. In cathodoluminescence images both new zircons and inherited cores have magmatic oscillatory zoning. Some inherited cores are unzoned. Results for 21 LA-ICP-MS analyses are listed in Supplementary Document 3. Twelve out of 21 analyses give and reflect inherited cores. In cathodoluminescence images both new zircons and inherited cores have magmatic oscillatory zoning. Core–rim structure is common. The cores show either complex zoning or are unzoned. Some zircon grains have dark discontinuous unzoned core. Data for 21 SIMS analyses on zircons are listed in Supplementary Document 4. These show moderately high Th/U ratios of 0.2–0.9, typical for magmatic zircons (Corfu et al. 2003; Corfu 2004). Rare inherited cores are present, and show overgrowth by wide magmatic rims. Data for 24 SIMS analyses of zircons are listed in Supplementary Document 3. These display high Th/U ratios of 0.4–0.9 for both cores and rims. In Terra–Wasserburg diagram (uncorrected for common Pb), zircons (excluding minor inherited cores with 207Pb/206Pb ages ~572, 589 and 627 Ma) are concordant with lower intercept at 561.3 ± 4.7 Ma (Figure 9).

5.3.2. Sample BJ11–28 (granodiorite)
Zircons from sample BJ11–28 are long to short prismatic, ~100–200 μm long. In CI images, they show magmatic concentric zoning. Core–rim structure is common. The cores show either complex zoning or are unzoned. Some zircon grains have dark discontinuous unzoned core. Data for 21 SIMS analyses on zircons are listed in Supplementary Document 4. These show moderately high Th/U ratios of 0.2–0.9, typical for magmatic zircons (Corfu et al. 2003; Corfu 2004). For three spots (points 10, 21 and 22) that show Pb loss, 13 analyses on zircons from coherent groups that are concordant and are close to the lower intercept with an age of 537.7 ± 4.7 Ma (MSWD = 1.12) (Figure 9).

5.3.3. Sample BJ11–19 (granitic gneiss)
Zircons from the Biarjmand granitic gneiss are short to long prismatic with variable length from 50 to >200 μm and aspect ratios of 1:2 to >1:4. Stubby-form zircons are also common. Zircons show oscillatory zoning, consistent with their magmatic origin (Corfu et al. 2003; Corfu 2004). Rare inherited cores are present, and show overgrowth by wide magmatic rims. Data for 24 SIMS analyses of zircons are listed in Supplementary Document 3. These display high Th/U ratios of 0.4–0.9 for both cores and rims. In Terra–Wasserburg diagram (uncorrected for common Pb), zircons (excluding minor inherited cores with 207Pb/206Pb ages ~572, 589 and 627 Ma) are concordant with lower intercept at 561.3 ± 4.7 Ma (Figure 9).

5.4. Zircon Hf and O isotopes

δ18O values for granitoid and granitic gneiss zircons (granodiorite BJ11–28 and BJ11–19) are moderately high and vary between 6.2‰ and 8.9‰. These δ18O values are similar to that of I-type granitic melt (5–8‰) (Li et al. 2007) and are lower than expected for supracrustal sources (Valley and Lackey 2005). The median δ18O values (7.5 ± 0.2‰) for Biarjmand zircons are similar to Cadomian granites from northwest Iran (~6.8–10.1; Moghadam et al. 2015d), although the latter may show more addition of crust into magma.

Thirty-six Lu–Hf analyses were obtained from two samples of Cadomian (520–560 Ma) zircons (Supplementary Document 5), one granitic gneiss (BJ11–19) and one granodiorite (BJ11–28). The initial Hf isotope ratios were calculated using measured 206Pb/238U ages of each analysis point. The Cadomian zircons (excluding inherited cores) show variable 176Hf/177Hf and hence εHf(t) values (Figure 10). Their εHf(t) varies from ~ −7.9 to +5.5 (average ~ −2.5).

Figure 8. LA-ICP-MS zircon U–Pb data for zircons from sample BJ11–8 (monzogranite).
excluding one point with εHf ~ –17.4 (point BJ11–28–17). These 176Hf/177Hf values for Cadomian zircons correspond to a mean 2T2DM age of ~ 1.4 Ga.

6. Discussion

In the following three subsections, our results provide new insights for three questions: (1) what was the timing of Cadomian magmatism in northeast Iran?; (2) what do Nd bulk rock and zircon Hf and O isotopic data reveal about crustal evolution in this region?; and (3) What are the tectonic implications of these data?

6.1. Timing of Cadomian magmatism in northeast Iran

The Cadomian orogen as preserved in rocks of southern Eurasia encompassed magmatic, tectonic and metamorphic events and sedimentary responses that spanned the period from mid-Neoproterozoic (~750 Ma, Cryogenian–Ediacaran) to the early Cambrian (~540–520 Ma) that happened along the periphery of the supercontinent Gondwana (peri-Gondwana) (Linnemann et al. 2004, 2014) (Linnemann et al. 2008). However, it seems that Cadomian events encompassed a shorter time in eastern peri-Gondwana (Turkey and Iran), where it appears to have lasted from Ediacaran (~600 Ma) to early-middle Cambrian (~500 Ma) time, with the most intense activity occurring around 550 Ma (Figure 12A) (Ustaomer et al. 2009; Badr et al. 2013; Balaghi Einalou et al. 2014; Abbo et al. 2015; Rossetti et al. 2015; Avigad et al. 2016). The U–Pb ages presented in this study as well as previous published data (Maryam et al. 2014; Moghadam et al. 2015b) on the Torud–Biarjmand region support the conclusion that Cadomian magmatism lasted a few tens of millions of years (~40 million years) in northeast Iran.

Our compiled data (Figure 1, Figure 12A) shows that the Cadomian magmatism in Iran lasted from ~600 to <500 Ma. The only known exceptions are the Chahdegan rocks from west Iran, which show U–Pb zircon ages of ~637 Ma (Nutman et al. 2014). Interestingly, zircons with older ages than >600 Ma occur as inherited cores in Iran Cadomian rocks. It seems that igneous rocks that are older than
The involvement of older crust is strongly suggested by the distinctive zircon Hf model ages (1.0–2.2 Ga, mostly ~1.4 Ga) and bulk rock Nd model ages (1.3–1.6 Ga) of Biarjmand orthogneiss and granitoids. This could indicate anatexis of Mesoproterozoic crust or mixing of juvenile Cadomian melts with Palaeoproterozoic or Archaean crust. The lack of known crust of pre-Cadamian crust in Iran and dearth of inherited zircons of either age makes this question unresolvable at present. Nevertheless, the isotopic information indicates that the Cadomian arc in Iran was a continental arc, similar to the modern Andes. Moderately high zircon O (6.2–8.9‰) but variably low εHf values (~7.9 to +5.5) and bulk rock εNd values (~2.2 to −5.8) are also consistent with a continental arc setting (Figure 10). On the δ18O against εHf(t) diagram (Figure 11), Torud–Biarjmand complex rocks differ from Cadomian juvenile granites and gabbros from east Iran and also from Salmas (northwest Iran) rocks, which show high δ18O and constant εHf(t) values.

U–Pb dates and Hf isotope compositions indicate an important episode of arc granitoid production and recycling of older continental crust during Cadomian time along the whole of northern Gondwana. Regional considerations suggest a geotectonic setting starting with an Andean-type marginal continental arc that changed to a Western Pacific style continental arc and a back-arc basin developed on a thinned and stretched continental crust (Linnemann et al. 2014).

Cadamian crustal evolution (ca. 600–500 Ma) of Iran was dominated by recycling of continental crust as suggested by the abundance of granitoids with zircons having negative or low values of εHf(t) (Figures 10 and 12). During the ~100-million-year long Cadomian magmatic arc activity, juvenile arc magmas interacted with older continental crust or sediments derived from these. None of the εHf(t) values for the zircons sits on the depleted mantle array, suggesting that these have been affected by crustal contamination. However, our compiled Hf isotope data (Figures 10 and 12C) reveal that Cadomian rocks from the Taknar and Zanjan–Takab complexes (east and northwest Iran, respectively) are more juvenile than other Cadomian rocks of Iran, suggesting the presence of juvenile Neoproterozoic arc crust (new crust in Figure 10) in eastern Iran. Cadomian (630–540 Ma) juvenile rocks are inferred elsewhere in West Gondwana (Anti-Atlas belt, Morocco) from detrital zircon survey (Avigad et al. 2012). Also, Cadomian ophiolitic basalts with high bulk rock Nd isotopic compositions are known from central Iran (Posht-e-Badam = Saghand) (Figure 7). Crustal contamination could be related to the crustal thickening along the Cadomian arc. In this scenario, long-lasting magmatic

6.2. Isotope perspective on the crustal evolution of northeast Iran

Geochemical evidence discussed in the next section indicates that the rocks we studied formed in an arc.
pulses (approximately 100 million years) led to crustal thickening, therefore enhancing MASH processes (Melting, Assimilation, Storage and Homogenization). This crustal thickening can be responsible for adakitic signature of the approximately 560-million-year-old granitic gneisses. However, isotopic signatures of granitic gneisses and granitoids are quite similar (although orthogneiss has lower mean $\varepsilon_{Hf(t)}$ values), and the La$_{(n)}$/Yb$_{(n)}$ ratio for gneissic rocks is not too high. This may show that the adakitic signature could be inherited by more crustal addition to the parent melt, or even by amphibole fractionation in the source area.

6.3. Tectonic implications

The geochemical data summarized in Figures 5 and 6 suggest that Cadomian rocks from the Torud–Biarjmand complex of northeast Iran formed in a magmatic arc above a subduction zone. The isotopic data discussed in the previous section indicate that this was a continental arc. This magmatic system evolved quickly and we see evidence in the Biarjmand gneiss that early-formed arc
rocks (represented by 561 ± 4 Ma gneiss) were deformed and uplifted before late-formed arc magmas (represented by 538 ± 5 and 522 ± 4 Ma granitoids) intruded (Figure 13). This observation is consistent with conclusions from farther west, where the different pulses of magmas intrude the older gneissic rocks and metasediments. Following the assembly of the Gondwana supercontinent, which started in the Cryogenian (Collins and Pisarevsky 2005), Cadomian arc-type magmatism formed thickened continental arc in the peri-Gondwana region including Iran, Turkey and Bohemian, Armorican, and Iberian Massifs (Linnemann et al. 2008, 2014; Ustaomer et al. 2009; Pereira et al. 2011; Yilmaz Şahin et al. 2014; Orejana et al. 2015; Moghadam et al. 2015b). Cadomian arc magmatism evolved to a marginal Western Pacific style orogenic system at the end of the Ediacaran (e.g. Drost et al. 2004, 2011; Linnemann et al. 2008). In fact there are

Figure 13. Summary of the tectonic evolution of Cadomian terranes within Iran (and North Gondwana) is shown in a series of panels (see text for explanations).
two alternatives for the evolution of the Cadomian arc rocks. For the first scenario, Linnemann et al. (2014) considered an arc–back-arc basin model for the evolution of peri-Gondwana during Ediacaran–Cadamian time. In this scenario, the back-arc basin was floored by thinned continental crust and flanked by a magmatic arc to the north and by intact Gondwana to the south (Figure 13). According to Linnemann et al. (2014), the Cadomian back-arc spreading started at ca. 570 Ma and produced pillowled E–MORB-like lavas, andesites, and calc-alkaline meta-basalts, all associated with black cherts (Buschmann et al. 2001). Nance and Murphy (1994) proposed a model in which oblique subduction beneath the arc led to strike–slip shearing and back-arc basin opening.

The distribution of Cadomian ophiolites is taken as evidence of one or more back-arc basins (Kounov et al. 2012; Von Raumber et al. 2015). Suspected Cadomian ophiolites in Iran are identified in two regions, Posht-e-Badam (Saghand, central Iran) and Zanjan–Takab (northwest Iran; Figure 1). There are not yet any detailed studies or U–Pb zircon ages for these ophiolites but their association with known Cadomian metamorphic rocks is taken as evidence that they are also Cadomian. In Posht-e-Badam, suspected Cadomian ophiolitic rocks include metamorphosed peridotites, MORB-like metavolcanic rocks, marble, and rodingites. The whole rock geochemical compositions of these lavas are similar to those found in back-arc basins (Moghadam et al., submitted) and isotopically these lavas are juvenile (Figure 7). In Zanjan–Takab, the ophiolite contains metamorphosed peridotites and metamorphosed magmatic rocks (tremolite-actinolite schists). Further studies are needed to determine the relationship between these inferred Cadomian ophiolites and Cadomian continental arc igneous rocks like those of Biarjmand. In the second scenario, which was recently proposed by Avigad et al. (2016), the back-arc basin was opened as early as 570 Ma, as graywacke sediments were deposited, starting ca. 570 Ma. This is confirmed by the youngest detrital zircons in Eastern Mediterranean (Tauride block) Neoproterozoic sediments (Abbo et al. 2015). Avigad et al. (2016) considered an approximately 50-million-year time interval for the cycle of back-arc sedimentation, metamorphism and later granitic magmatism, which were responsible for the construction of the Tauride block. They concluded that Tauride (and other Cadomian terranes) were finally accreted to Afro-Arabia during the late Ediacaran–early Cambrian. This model best explains the existence of younger granitic intrusions (~540 Ma) intruding the Tauride metasediments which were deposited in a back-arc basin subsequent to ~580 Ma. However, the trace of the older Cadomian arc north of the back-arc basin isn’t clear. Have the older Cadomian rocks been totally uplifted and eroded and therefore supplied the back-arc sediments? Furthermore, the presence of ca. 580 Ma granulites shows crustal thickening beneath an evolved arc during that time (Koralay 2015), whereas exhumation of eclogites at 535 Ma shows that subduction beneath north Gondwana was active during the early Cambrian (Candan et al. 2015).

The presence of gabbros with ages of ~556–585 Ma in northeast Iran (Moghadam et al., submitted), which do not intrude metasediments, are another matter of debate. These are clearly older than the sedimentation age of ca. 570 Ma. We suggest that the oldest metasediments were deposited in the hinterland near the Afro-Arabian margin, and the back-arc basins opened in the latest Ediacaran. These back-arc basins continued to open to form the Rheic Ocean beginning ~530 Ma, allowing the western arc fragments (including Iberia) to migrate north where they accreted to the southern margin of Laurussia.

7. Conclusions

Bulk rock geochemistry reveals that granitoids and gneiss of the Torud–Biarjmand complex in northeast Iran are arc-related calc-alkaline rocks that formed in a continental arc. U–Pb zircon geochronology shows that these rocks have ages of ca. 560–520 Ma and are part of Cadomian igneous activity in peri-Gondwana. Zircon Hf–O and bulk rock Nd isotope compositions indicate the involvement of Mesoproterozoic or older continental crust. Our geochemical and geochronological data are consistent with late Ediacaran–early Cambrian arc magmatism along the active continental margin of north Gondwana, and we suggest that evidence of a Cadomian back-arc basin may exist in central and northwest Iran.

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