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1 Cimmerian metamorphism and post Mid-Cimmerian
2 exhumation in Central Iran: insights from in-situ Rb/Sr
3 and U/Pb dating.

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15
16 **Abstract**

17
18 The pre-Alpine evolution of the Tethyan domains between Gondwana and Laurasia, and in
19 particular that of the Cimmerian continental blocks, remains poorly constrained. Central Iran is a
20 key area to constrain the closure of the Paleotethys and the collision of Laurasia with the Cimmerian

21 blocks drifted from Gondwana. The present study provides a combined metamorphic and
22 geochronologic approach focused on two areas of Central Iran: the Kashmar-Kerman Tectonic Zone
23 (KKTZ) and the Jandaq area, whose tectonometamorphic units are affected low- to middle-pressure
24 high-temperature metamorphism. We performed in-situ texturally constrained U/Pb dating on
25 titanite and Rb/Sr dating on mica to quantify the timing and intensity of burial and exhumation of
26 these metamorphic rocks. Results show that metamorphism in the central and eastern KKTZ (7-9
27 kbar; $\sim 700^{\circ}\text{C}$) is synchronous or slightly postdates the Cimmerian orogeny ($\sim 190\text{-}180$ Ma) and
28 relates to the collision following the Paleotethys closure, whereas metamorphism in the Jandaq
29 complex (12-13 kbar; $\sim 450^{\circ}\text{C}$) relates to the Paleotethys subduction. Based on paleogeographic
30 reconstructions, the KKTZ lied hundreds of kilometers south of the Paleotethys suture zone. As such,
31 we propose that metamorphism along the KKTZ results from the closure and shortening of a
32 rheologically weak domain located outboard of the main suture. In-situ Rb/Sr dating of biotite yields
33 precise and accurate cooling ages ($\pm 2\text{-}5$ Ma) ranging from 170 to 140 Ma. These cooling ages
34 document the exhumation of these terranes from the Mid-Cimmerian event onward (~ 170 Ma),
35 coeval with the widespread extension distributed across Iran and thought to reflect upper plate
36 extension above the Neotethyan subduction system.

37

38 **Keywords**

39 Central Iran; Collisional metamorphism; in-situ Rb/Sr dating; Cimmerian orogeny; Neotethys
40 subduction

41

42 **1 Introduction**

43
44 Plate kinematics since the Cambrian have been dominated by the diachronous drifting and
45 collision of Gondwana-derived continental blocks with Laurasia and the associated formation of
46 large-scale orogenic belts, with major implications for geodynamics (Şengör, 2012), paleoclimate
47 (Smith, 1979) and ore formation (Metcalf, 2021). The amalgamation of these blocks results from
48 the opening and closure of several Tethyan oceanic domains (Suess, 1895; Stille, 1958; Flügel, 1972;
49 Şengör, 1979): the Prototethyan oceans (which led to the Cadomian orogeny), the Paleotethys
50 ocean (Cimmerian and Indosinian orogenies), the Mesotethys or Rheic ocean (Variscan orogeny) and
51 the Neotethys ocean (Alpine and Himalayan orogenies). The mechanism driving this stepwise
52 fragmentation of the Gondwana margin and later accretion onto Laurasia, similar to a 'diffuse'
53 collision between Gondwana and Laurasia, likely relates to asthenospheric upwelling and horizontal
54 flow below Gondwana (Ziegler, 1992; Jolivet et al., 2016). Constraining its detailed processes and
55 exact timing is unfortunately frequently complicated by the Alpine overprint onto the geological
56 archive of earlier orogenic structures.

57
58 Central Iran is a key area to study this dynamic evolution (Fig. 1), as it preserves a mosaic of
59 Gondwana-derived 'Cimmerian' terranes thought to have once formed a composite continental
60 domain – or a set of continental blocks – between the Paleotethys and the Neotethys. Central Iran
61 is limited by several ophiolite belts (Fig. 1, e.g., Sistan, Sabzevar, Nain-Baft; Moghadam and Stern,
62 2015) and dissected by large-scale active faults (Walker et al., 2004), but largely escaped the
63 Paleogene-Neogene magmatism which affected most of Iran (Fig. 1). It exposes metamorphic and

64 magmatic rocks that preserve evidence of several major geological episodes: Panafrican/Cadomian
65 magmatism (Rossetti et al., 2015; Hassanzadeh et al., 2008; Moghadam et al., 2017), subduction and
66 closure of the Paleotethys, with subsequent Mesozoic tectonic activity during the Eo-, Mid- and Late-
67 Cimmerian events (Fürsich et al., 2009b; Zanchi et al., 2009b; Wilmsen et al., 2015; Rossetti et al.,
68 2017), and Mesozoic to present tectonics linked to the subduction of the Neotethys (Agard et al.,
69 2011; Mattei et al., 2015; Wilmsen et al., 2021).

70
71 Unravelling the tectono-metamorphic evolution of the region is essential to elucidate the
72 geodynamic setting and significance of these successive geological events. Metamorphic rocks,
73 however, are relatively sparsely exposed across desert areas and/or covered by Quaternary infill
74 (Fig. 1). Estimates of their pressure and temperature conditions, as well as absolute age data, are
75 mostly lacking. One example is the high temperature and medium pressure (HT-MP) metamorphic
76 event ascribed to the Cimmerian orogeny, around 220-185 Ma (Horton et al., 2008; Fürsich et al.,
77 2009a; Wilmsen et al., 2009a; Zanchi et al., 2009a, 2015; Wilmsen et al., 2015), which is documented
78 in Central and North-East Iran in the Kashmar-Kerman Tectonic Zone (Ramezani and Tucker, 2003;
79 Kargaranbafghi et al., 2012), as well as in the Jandaq (Bagheri and Stampfli 2008; Berra et al. 2017)
80 and Shotor-Kuh complexes (Rahmati-Ilkhchi et al., 2011; Malekpour-Alamdari et al., 2017). These
81 complexes are nevertheless located hundreds of kilometers apart and away from the Paleotethys
82 suture zone (Fig. 1), making this collisional metamorphism enigmatic. Furthermore, while the
83 absolute ages reported for these areas (Verdel et al., 2007; Bagheri and Stampfli 2008; Rahmati-
84 Ilkhchi et al., 2011; Masoodi et al., 2013; Tab. 1) are commonly related to the Mid-Cimmerian

85 tectonic event (~170-168 Ma; Fürsich et al., 2009b; Wilmsen et al., 2009b,a), the nature and
86 intensity of this event are still poorly constrained.

87
88 The present study aims to refine our understanding of the Cimmerian and subsequent
89 Jurassic events through a combined metamorphic and geochronologic approach focused on two
90 understudied areas of Central Iran: (1) the Kashmar-Kerman Tectonic Zone (KKTZ) and (2) the Jandaq
91 area. In order to obtain an extensive dataset of texturally constrained ages across a vast area, we
92 have calibrated and implemented the in-situ Rb/Sr dating on biotite and white mica combined with
93 titanite U-Pb geochronology. Results provide insights into the timing and intensity of burial and
94 exhumation of the investigated metamorphic rocks, as well as into the tectonics of the Cimmerian
95 orogeny.

96

97 **2 Geological context**

98

99 **2.1 Paleogeography of the Cimmerian blocks**

100

101 The Central-East-Iranian Microcontinent (CEIM) consists of the Yazd, Tabas, and Lut
102 continental blocks, which are separated by active strike-slip faults (Figs. 2,3). These faults are
103 thought to rework Palaeozoic and/or older structures (Berberian and King, 1981). The CEIM belongs
104 to the western Cimmerian blocks together with the Alborz, Sanand (or Sanandaj-Sirjan) and Afghan
105 blocks, which were part of the Gondwana passive margin during the Early Paleozoic.

106
107 During the Late Neoproterozoic and Early Cambrian, the Prototethys (locally termed Ran
108 ocean) is thought to have subducted beneath the northern part of Arabia, the western Cimmerian
109 blocks and northern India (Torsvik and Cocks, 2016; Fig. 3). This is documented by the presence of
110 extensive Cambrian andesites and trondhjemites in the western part of the CEIM (Ramezani and
111 Tucker, 2003) and by late Neoproterozoic–Early Cambrian granitoids in Iran (Hassanzadeh et al.,
112 2008). This active margin setting is associated with the Cadomian orogeny (Rossetti et al., 2015;
113 Moghadam et al., 2020).

114 During the Early Paleozoic, the western Cimmerian blocks lied close to Arabia based on their
115 similar paleomagnetic record (Wensink et al., 1979; Muttoni et al., 2009b) and on the nature of the
116 basement and of the overlying sedimentary succession (Stöcklin, 1968; Stöcklin, 1974; Berberian
117 and King, 1981; Wendt et al., 2005). A thick discontinuous and poorly deformed Cambrian to Triassic
118 succession records the passive margin history of these blocks (Torsvik and Cocks, 2009).

119
120 Following initial rifting during the late Silurian, the Paleotethys ocean opened during the
121 early Devonian, separating part of the Hunic terranes from the Gondwana margin (e.g., the
122 Karakum-Turan, Pamirs, Tarim, Qiangtang, North China terranes and the various terranes which now
123 make up southern Europe; Stampfli and Borel, 2002; Torsvik and Cocks, 2016). North of Gondwana,
124 remnants of a north-dipping subduction of the Paleotethys Ocean below Laurasia are documented
125 along northern Iran and in the Anarak-Jandaq zone in central Iran (Fig. 3a) :

126 (i) Around Mashhad (Fig. 1), remnants of an accretionary wedge marking Palaeotethys
127 closure has been extensively studied (Stöcklin, 1974; Alavi, 1991; Boulin, 1991; Ruttner, 1993; Alavi
128 et al., 1997; Sheikholeslami and Kouhpeyma, 2012).

129 (ii) Carboniferous eclogites occur in Shanderman and Rasht (Fig. 1) in the Talesh Mountains,
130 western Alborz (Zanchetta et al., 2009; Omrani et al., 2013; Rossetti et al., 2017).

131 (iii) In central Iran, the Carboniferous HP-LT metamorphism of the Anarak and Jandaq
132 Metamorphic Complexes (Bagheri and Stampfli, 2008; Zanchi et al., 2015), the Triassic forearc
133 succession of Nakhlak (Balini et al., 2009) and the arc detrital deposition of Godar-e-Siah (Berra et
134 al., 2017) can be framed in the same Paleotethys active margin setting (Fig. 2).

135 (iv) East of Mashhad, in the Fariman-Aghdarband region of NE Iran, the arc-related units of
136 Fariman and Darreh Anjir record active subduction during the Permian, as well as during the
137 Devonian and Carboniferous (Zanchetta et al., 2013; Moghadam et al., 2015).

138 (v) The arc-related succession of Aghdarband, exposed north of Fariman (Fig. 1), is also
139 consistent with Paleotethys subduction during the Early to Middle Triassic (Ruttner et al., 1991;
140 Zanchi et al., 2016).

141
142 The gradual opening of the Neotethys Ocean within the north-east rim of Gondwana during
143 the Early Permian (from about 275 Ma; Domeier and Torsvik, 2014) separated a series of
144 microcontinents and terranes (Sibumasu, Tibetan, Turkey, Alborz, Iran, Karakorum, Lut, Sanand,
145 Afghanistan, and Pakistan) from the northern Gondwana margin (Fig. 3b). Paleomagnetic and
146 sedimentary data indicate that these terranes, called the Cimmerian blocks, resided on
147 subequatorial paleolatitudes during the Late Permian-Early Triassic (Muttoni et al., 2009b). The

148 central Cimmerian terranes (e.g. central Afghanistan, Karakoram) were located northward with
149 respect to the other Cimmerian terranes (Angiolini et al., 2003; Angiolini, 2001; Campi et al., 2005,
150 Muttoni et al., 2009a). Whether the other Cimmerian terranes formed a single elongate continent
151 (referred to as Cimmeria) or several distinct microcontinental blocks remains unclear. Indeed, the
152 relative motion and potential paleolatitude difference between the western Cimmerian blocks
153 (Alborz v. CEIM v. Sanand v. Afghan) cannot be ascertained because palaeomagnetic results are
154 sparse and uncertainties too large (Muttoni et al., 2009b; Torsvik and Cocks, 2016).

155
156 In the Late Triassic (Fig. 3c), the closure of the Paleotethys ocean resulted in the Cimmerian
157 Orogeny, which was defined based on mid-Mesozoic convergent plate-margin activity stretching
158 from Bulgaria to southeastern Asia (Suess, 1895; Şengör, 1979). In Iran, it corresponds to the
159 collision of the western Cimmerian blocks with the active margin of the Turan Terrane (Berra et al.,
160 2007; Horton et al., 2008; Fürsich et al., 2009a; Zanchi et al., 2009a; Zanchetta et al., 2009). The
161 Cimmerian orogeny in Iran may have comprised several diachronous collisional events between the
162 different western Cimmerian blocks (Golonka, 2004). Inception of collision is proposed at 225 Ma
163 (Wilmsen et al., 2009b). The shift from Middle Triassic platform carbonates to the siliciclastic rocks
164 of the Shemshak Group (and equivalent successions) observed throughout Iran reflects the onset of
165 an Eo-Cimmerian deformation from approximately 220 to 185 Ma (Horton et al., 2008; Fürsich et
166 al., 2009a; Wilmsen et al., 2009a; Zanchi et al., 2009a, 2015; Wilmsen et al., 2015). The subduction
167 of the Neotethys below the Iranian terranes probably began in the Late Triassic or Early Jurassic, as
168 testified by arc magmatism (Berberian and King, 1981) and eclogite formation (Davoudian et al.,
169 2016) in the Sanandaj-Sirjan Zone. The formation of the extensional Nayband Basin (~210 Ma;

170 Fürsich et al., 2005; Wilmsen et al., 2009b) is thought to reflect reduced compression across the
171 Iranian blocks due to early back-arc extension after the initiation of this subduction.

172
173 The tectonic evolution of Iran after the main Cimmerian orogeny is marked by two discrete
174 tectonic phases: the Mid- and Late-Cimmerian events. Those events were first described in Europe
175 (Stille, 1924; Ziegler, 1975) and corresponding unconformities in the sedimentary record were later
176 described in Iran (Seyed-Emami and Alavi-Naini, 1990). In Iran, the Mid-Cimmerian event is defined
177 by an unconformity on top of the Shemshak Group and is apparently confined to the Bajocian (~170-
178 168 Ma; Fürsich et al., 2009b). This compressional event was characterized in the Alborz by rapid
179 uplift (60 m/Myr) followed by significant subsidence possibly marking the onset of spreading in the
180 South Caspian Basin (Fürsich et al., 2009b). This phase of rapid subsidence is observed across all of
181 northeast Iran and is interpreted as crustal extension across rotating blocks to explain the great
182 magnitude of relative deepening (Fürsich et al., 2009b; Wilmsen et al., 2009b,a).

183 A less well-constrained Late-Cimmerian event occurred during the Late Jurassic (~145 Ma),
184 mostly across Central Iran (Zanchi et al., 2009b; Wilmsen et al., 2015, 2021). This event is possibly
185 related to the opening and closure of the Iranian back-arc basins at the rear of the large-scale
186 subduction zone of the Neotethys (Rossetti et al., 2010; Agard et al., 2011). The opening and closure
187 of these small oceanic basins was accompanied by the activity of the Great Kavir-Doruneh Fault
188 (Javadi et al., 2013, 2015) and by significant lateral displacements and block rotations.

189
190 One of the consequences of these post-Cimmerian block movements/rotations could be the
191 large-scale translation (~500 km) of the Nakhlak, Anarak and Jandaq complexes from the

192 Paleotethys suture to the interior of Central Iran (Fig. 2; Bagheri and Stampfli 2008; Zanchi et al.
193 2015; Berra et al. 2017). These complexes, with Eurasian paleobiogeographic affinities and active
194 margin imprints, are now exposed between the Great Kavir Doruneh Fault and the Palaeozoic-
195 Triassic successions of Gondwana affinity of the Yazd block. Initial palaeomagnetic data indicated
196 that the CEIM reached its actual position after a 135° anticlockwise rotation (Davoudzadeh and
197 Weber-Diefenbach, 1987; Soffel et al., 1996). More recently, however, Mattei et al. (2015) proposed
198 a two-stage anticlockwise rotation of about 85° since the Jurassic, with a first stage occurring during
199 the Early Cretaceous and a second one after the Middle–Late Miocene. These rotations seem
200 confined to the CEIM and do not extend to the other tectonic provinces of Iran. The stratigraphy
201 and facies distribution show that the Yazd Block was emergent during most of the Jurassic period
202 and that the marine influence increased from the Tabas block to the Lut block (Fig. 3; Dercourt et al.
203 1986; Fürsich et al. 2003; Wilmsen et al. 2003, 2005, 2010). On this basis, Mattei et al. (2015)
204 proposed that the Yazd, Lut and Tabas blocks were oriented WSW–ENE during the Late Jurassic with
205 the Lut Block facing the Neotethys Ocean to the south and southeast (see Fig. 3c). Lastly, central
206 Iran was affected, from ~30 Ma onwards, by the Zagros orogeny marking the closure of the
207 Neotethys and resulting in overprinting deformation across Iran (e.g. Alborz and Kopeh Dagh; Agard
208 et al. 2011; Ballato et al., 2011; Jentzer et al. 2017).

209

210 **2.2 Studied areas**

211

212 This study focuses on the metamorphism related to the subduction and collision of the
213 Paleotethys. For the sake of clarity, geological contexts for the Rasht, Anarak and Shotor-Kuh

214 complexes were metamorphism related to the Paleotethys occurred are presented in the
215 supplementary section.

216

217 **2.2.1 Kashmar-Kerman Tectonic Zone**

218

219 The Kashmar–Kerman Tectonic Zone (KKTZ) separates the Yazd block from the Tabas block
220 (Haghipour et al., 1977a; Ramezani and Tucker, 2003). Unlike in those blocks, Neoproterozoic and
221 Lower Paleozoic rocks are well exposed in the KKTZ, including several metamorphic complexes
222 overlain by Mesozoic and Cenozoic sedimentary units. Ramezani and Tucker (2003) distinguished
223 three lithotectonic domains in the KKTZ separated by large strike-slip faults (Western, Central and
224 Eastern; Fig. 4):

225

226 A Western domain, bounded to the west by the Chapedony fault and to the east by the
227 Neybaz-Chatak fault, which comprises the Chapedony complex and several magmatic intrusions
228 (e.g. the Koshoumi granite and Daranjir Diorite) dated between 40 and 49 Ma (Ramezani and Tucker,
229 2003; Verdel et al., 2007). The Chapedony Metamorphic Complex includes high-grade gneisses,
230 migmatites and anatectic granites (Haghipour et al., 1977a) with peak metamorphism at
231 approximately 46 Ma (Ramezani and Tucker, 2003) and Ar-Ar ages for various minerals ranging from
232 40 to 48 Ma (Verdel et al., 2007; Kargaranbafghi et al., 2012, 2015). Late- to post-metamorphic
233 intrusion of granite-diorite plutons into the complex were dated at approximately 45 Ma (U-Pb
234 zircon; Ramezani and Tucker, 2003). Peak P–T conditions were estimated at 3 kbar and 650–750 °C
235 (Kargaranbafghi et al., 2015) using the Al-in-hornblende barometer (Anderson and Smith, 1995) and

236 the amphibole–plagioclase thermometer (Holland and Blundy, 1994). The Ar–Ar ages, together with
237 published U–Pb zircon and U–Th/He apatite and zircon ages (Tab. 1), imply rapid cooling from 750°C
238 to 60°C of the Chapedony complex between 49 and 30 Ma (Kargaranbafghi et al., 2015). The
239 Chapedony complex was interpreted as a core complex exhumed below the central domain by the
240 east deepening normal Neybaz-Chatak fault (Ramezani and Tucker, 2003; Verdel et al., 2007;
241 Kargaranbafghi et al., 2012, 2015; Fig. 4). The Chapedony core complex is interpreted as the result
242 of magmatic underplating and crustal extension associated with a regional magma flare-up during
243 the Eocene (Verdel et al., 2011; Kargaranbafghi et al., 2015).

244
245 A Central domain, between the Neybaz-Chatak and Posht-e-Badam faults (Fig. 4). It is
246 composed of the Posht-e-Badam complex, magmatic intrusions such as Chamgoo and Anarg
247 granodiorites or the Esmailabad Granite (Ramezani and Tucker, 2003), a Cambrian volcano-
248 sedimentary unit and Permian to Paleogene-Neogene sedimentary units. The Posht-e-Badam
249 Complex is made of an association of greenstones, schists, gneisses, amphibolites and marbles. The
250 complex is severely disrupted by thrusting, as also evidenced by large-scale folded marbles, and
251 intrusion of granitoid plutons. The Esmailabad Granite is dated at 218 ± 3 Ma (2σ) and the Chamgoo
252 Granodiorite at 213.5 ± 0.5 Ma (2σ). These intrusions, with distinctive peraluminous character and
253 high concentrations in LILE (Rb and Cs), were interpreted as anatectic or collisional granitoids
254 (Ramezani and Tucker, 2003) and ascribed to the Cimmerian orogeny. No Late Triassic magmatic
255 intrusions have been described in the KKTZ outside of this Central Lithotectonic Domain. The
256 medium-grade metamorphism within the Posht-e-Badam complex was dated at 219.2 ± 2.4 Ma (2σ ;
257 Ar–Ar hornblende; Kargaranbafghi et al., 2012) and also attributed to the Cimmerian orogeny.

258
259 An Eastern lithotectonic domain, between the Posht-e-Badam and the Kalmard Fault,
260 comprises the Tashk, Boneh-Shurow and Sarkuh complexes, magmatic intrusions (including the Ariz
261 and Polo granodiorites or the Zarigan, Douzakh-Darreh and Sefid trondhjemitic intrusions), a
262 Cambrian Volcano-sedimentary unit and Paleozoic and Mesozoic sedimentary units.

263 The Tashk Complex comprises weakly metamorphosed sedimentary and volcanoclastic rocks
264 (Haghipour et al., 1977a) deposited from the Late Neoproterozoic to Early Cambrian. Ramezani and
265 Tucker (2003) constrained the depositional age of the formation between 627 Ma (youngest zircon
266 population in a tuffaceous rock) and 533 Ma, which corresponds to the oldest known magmatic
267 intrusion (i.e. the Ariz Granodiorite). The Tashk Formation was deposited in a marginal marine
268 environment with volcanic influence and is unconformably overlain by Permian and Triassic shallow-
269 marine carbonates.

270 The Sarkuh Complex is composed of marble and medium grade metamorphic rocks such as
271 garnet-staurolite-andalusite schists, micaschists, amphibolites and metavolcaniclastics later
272 intruded by felsic porphyries. This complex is poorly studied and, to the authors' knowledge, no
273 geochronological data are available.

274 The Boneh-Shurow complex was divided in three areas for clarity, with the northern area
275 around the city of Posht-e-Badam, the central area in the Eskamblo mountain and the southern area
276 in the Posht-e-Shorkh mountain (Fig. 4). The Boneh-Shurow Metamorphic Complex is composed of
277 several lithologies such as:

278 (i) Mylonitic orthogneiss (Haghipour et al., 1977a) or protogneiss (Ramezani and Tucker,
279 2003), which is the dominant lithology in the northern area, yielding a U-Pb zircon age for the
280 protolith of 544 ± 7 Ma (2σ).

281 (ii) Micaschist, phyllite, slate, metasandstone, calcsilicate and carbonate rocks with detrital
282 U-Pb zircon ages from 602 Ma to 617 Ma (Ramezani and Tucker, 2003).

283 (iii) Garnet-amphibolite rocks including mainly hornblende, garnet, plagioclase and biotite.
284 The peak metamorphism was dated at 547.6 ± 2.0 Ma (2σ ; U-Pb zircon; Ramezani and Tucker, 2003).

285 (iv) Quartz-diorite intrusions emplaced at 547.0 ± 2.5 Ma (2σ ; U-Pb zircon; Ramezani and
286 Tucker, 2003).

287 (v) Mylonitic schists, dominant lithology in the southern area.

288 Kargaranbafghi et al. (2012) proposed that the Boneh-Shurow complex was affected by low-
289 grade metamorphism during a late stage of the Cimmerian orogeny (Ar-Ar white mica; 140.8 ± 0.6
290 Ma; 2σ). The formation of a Jurassic core-complex below an east-dipping normal fault was described
291 in the Central area (Masoodi et al., 2013; Soleimani et al., 2021) (Fig. 4).

292 Ramezani and Tucker (2003) proposed that the granite-tonalite intrusions and the Boneh-
293 Shurow granitic orthogneiss, based on their magmatic-arc affinity and similar ages (U-Pb zircon; 533
294 ± 1 and 542 ± 9 Ma respectively; 2σ), belong to a similar Late Neoproterozoic-Early Cambrian
295 magmatic event. The granite-tonalite intrusions are intrusive in the volcano-sedimentary Tashk
296 complex and may represent a shallow level of a volcanic arc. The Boneh-Shurow complex, which
297 comprises terrigenous, semi-pelitic and carbonaceous protoliths, is more likely a distal part of this
298 Late Neoproterozoic-Early Cambrian magmatic arc complex. The presence of the Cambrian Volcano-
299 Sedimentary Unit also supports the existence of a magmatic-arc setting dated at approximately 528

300 Ma (U-Pb zircon; Ramezani and Tucker, 2003). Finally, the trondhjemite intrusions of the Cambrian
301 Leucogranite Suite were dated at 525 ± 7 Ma (2σ ; U-Pb zircon) and show evidence for the Early
302 Cambrian subduction of young oceanic crust and possibly the termination of arc magmatism
303 (Ramezani and Tucker, 2003). This Neoproterozoic-Early Cambrian magmatic arc complex, which
304 was interpreted as marking the Prototethyan active margin of Gondwana (Ramezani and Tucker,
305 2003), was recently reappraised as the subduction zone of the Ran ocean (which predated the
306 Cadomian orogeny; Torsvik and Cocks, 2016). The impact of the Cimmerian orogeny on this domain
307 has been studied by Masoodi et al. (2013) who proposed three deformation stages: (1) a D1-1
308 dextral strike-slip event with an early Jurassic cooling phase (before 186 Ma, Eo-Cimmerian); (2) a
309 Middle Jurassic D1-2 extensional event with a top-to-NE sense of shear on low-angle normal faults
310 (Mid-Cimmerian); (3) a D2 reverse shear event occurring during the Early Cretaceous (Late-
311 Cimmerian).

312

313 **2.2.2 Jandaq complex**

314

315 The Jandaq Metamorphic Complex (JMC; Bagheri and Stampfli 2008; Berra et al. 2017) is
316 located immediately south of the Great Kavir Fault and includes large bodies of amphibolites,
317 garnet- staurolite-mica-bearing schists and gneiss intruded by pegmatites (Fig. 2; Berra et al. 2017).
318 The JMC is juxtaposed against the Arusan Ophiolite, which was intruded by Jurassic granitoids
319 (Aistov et al., 1984). Radiometric ages yield Carboniferous (333-318 Ma for Ar-Ar dating in white
320 mica) and Jurassic ages (163.86 ± 1.76 Ma and 156.56 ± 33.15 Ma for Ar-Ar dating in a muscovite

321 and in a hornblende respectively; 2σ), whereas the pegmatites have a Late Triassic age based on a
322 U-Pb single crystal zircon dating (215 ± 15 Ma; 2σ ; Bagheri and Stampfli, 2008).

323

324 Similarities in lithologies between the Posht-e-Badam complex and the Jandaq metamorphic
325 complex were pointed out by Bagheri and Stampfli (2008), and with the Boneh-Shurow Complex by
326 Ramezani and Tucker (2003).

327

328 **3 Analytical methods**

329

330 **3.1 Electron probe microanalysis (EPMA) and scanning electron microscope** 331 **(SEM) analysis**

332

333 EPMA and SEM were carried out at CAMPARIS (SU-IPGP, Paris, France) using the CAMECA
334 SX-100 and CAMECA SX-FIVE instruments and the data reducing method of Pouchou and Pichoir
335 (1991). Analytical conditions for spot analysis were 15 kV accelerating voltage and 10 nA specimen
336 current with a dwell time of 50 ms and a beam size of 2 μm . Fe_2O_3 , MnTiO_3 , diopside, Cr_2O_3 , K-
337 feldspar, anorthite and albite were used as standards. Mineral abbreviations are after Whitney and
338 Evans (2010).

339

340 **3.2 In-situ titanite U/Pb geochronology**

341

342 Titanite U-Pb petrochronology was carried out via laser ablation inductively coupled plasma
343 mass spectrometry (LA-ICP-MS/MS) in the Fipke Laboratory for Trace Element Research (FiLTER) at
344 the University of British Columbia, Okanagan using a Teledyne Analyte 193 nm Excimer laser coupled
345 to an Agilent 8900 triple quadrupole ICP-MS. All grains were analyzed in thin sections to preserve
346 textural relationships. Spot analyses were carried out across 5 analytical sessions using a 40-micron
347 diameter laser spot with a repetition rate of 6 Hz and fluence of 4.00 J/cm². The Ar sample gas (0.9
348 L/min) and He carrier gas (0.35 L/min) were mixed before the plasma using an in-house glass mixing
349 valve/signal smoothing device. The analytical setup was optimized for maximum signal using the
350 ²³⁸U/²³²Th ratio of the reference material 'NIST610' to within 3% of the certified value. Each spot
351 was pre-ablated with two laser bursts followed by a 25 second delay before 25 seconds of ablation.
352 Isotopic data from titanite were normalized to repeat analyses of the titanite reference material
353 'MKED' (²⁰⁶Pb/²³⁸U age of 1517.32 ± 0.32 Ma, Spandler et al., 2016) with 'Mount McClure' (²⁰⁷Pb/²³⁵U
354 age of 523.26 ± 1.27 Ma, Schoene and Bowring, 2006) used as a secondary reference material to
355 verify the analytical procedure. Down-hole element fractionation and instrument drift were
356 monitored based on the primary reference material and corrected for using the Lolite software
357 package v.4.5 (Paton et al., 2010, 2011). Analyses of the Mount McClure reference material as
358 unknowns yielded lower intercept dates of 526 ± 5 (MSWD = 1.0, n = 10/11), 525 ± 10 (MSWD = 1.7,
359 n = 7/9), 534 ± 7 (MSWD = 2.1, n = 8/10), 535 ± 4 (MSWD = 1.2, n = 12/13), 525 ± 5 (MSWD = 0.87,
360 n = 12/14) in Tera-Wasserburg space, all well within 2% uncertainty of the accepted value. Up to
361 1.5% additional uncertainty was added quadratically to the ²⁰⁶Pb-²³⁸U ratios and up to 0.5% to the
362 ²⁰⁷Pb-²⁰⁶Pb ratios of all analyses as indicated by the overdispersion of ratios from the secondary
363 reference materials from the same analytic sessions.

364

365 Trace element concentrations were measured with the U-Pb isotopes for each spot. A dwell
366 time of 10 ms was used for ^{29}Si , ^{31}P , ^{43}Ca and ^{90}Zr while a dwell time of 15 ms was used for ^{49}Ti , ^{89}Y ,
367 ^{93}Nb , ^{139}La , ^{140}Ce , ^{141}Pr , ^{146}Nd , ^{147}Sm , ^{153}Eu , ^{157}Gd , ^{159}Tb , ^{163}Dy , ^{165}Ho , ^{166}Er , ^{169}Tm , ^{172}Yb , ^{175}Lu , ^{178}Hf
368 and ^{181}Ta . Concentrations were calculated using Iolite v.4.5 (Paton et al., 2010, 2011) with NIST610
369 (GeoReM database, application version 27, <http://georem.mpch-mainz.gwdg.de>; Jochum et al.,
370 2005) as the primary reference material and NIST612 as the secondary reference material. Calcium
371 was the internal standard assuming stoichiometric concentrations. Measured trace element
372 concentrations of NIST612 are typically within 5% of expected values (GeoReM database,
373 application version 27, <http://georem.mpch-mainz.gwdg.de>; Jochum et al., 2005). All ages are
374 presented with 2σ uncertainties and data presented in supplementary material.

375

376 **3.3 In-situ mica Rb/Sr geochronology**

377

378 All analyses for Rb/Sr geochronology were performed in-situ in polished thin sections at the
379 ALIPP6 lab (ISTeP, Sorbonne University, Paris) using an Excimer 193 nm Analyte G2 Teledyne laser
380 ablation system coupled with an Agilent 8900 triple-quadrupole ICP-MS/MS coupled with a reaction
381 cell ORS⁴. The acquisition protocol is modified from Zack and Hogmalm (2016) and Hogmalm et al.
382 (2017), using N_2O as the reaction gas, to analyze relatively low Sr and Rb concentrations in
383 metamorphic mica. Analyses were performed in thin sections to preserve textural relationships with
384 $50\ \mu\text{m}$ spots with a repetition rate of 8 Hz and applying a fluence of ca. $4.46\ \text{J}/\text{cm}^2$ across 16 seconds
385 of ablation. The analytical setup was optimized for maximum sensitivity in gas mode for the targeted

386 mass isotopes in the reference material NIST610 (m/z set for the first and second quadrupoles
387 respectively at: 85 and 85; 86 and 102 then 88 and 104). Isotopic ratio from mica were normalized
388 to repeated analyses of the USGS reference material BCR-2G (Elburg et al., 2005) with ATHO-G
389 (Jochum et al., 2011) and BHVO-2G (Elburg et al., 2005) used as secondary USGS reference materials
390 to assess data quality.

391
392 We routinely measured a range of major and trace elements (Na, Mg, Al, Si, K, Ti, Fe during
393 0.1 ms and Nb, Ba, Cs during 2 ms) together with Rb and Sr isotopes (during 0.1 s) to gain information
394 on chemical variations of analyzed phases as well as detecting (and excluding) signals from inclusions
395 and/or alteration zones. Concentrations were calculated per formula unit considering 1 p.f.u. for K
396 and using BCR-2G as the primary reference material (GeoReM database, application version 27,
397 <http://georem.mpch-mainz.gwdg.de>; Jochum et al., 2005). Analyzed isotopic ratios of the secondary
398 reference material are within 5% uncertainty of the accepted value. Data reduction and instrument
399 drift correction is realized with a homemade Matlab program and isochron age calculation is
400 achieved with the IsoplotR software (Vermeesch, 2018). Data overdispersion is corrected by
401 additional uncertainty based on repeated analyses on reference material ATHO-G. All ages are
402 presented with 2σ uncertainties, see supplementary material for further details.

403

404 **3.4 Thermodynamic modelling**

405

406 Thermodynamic modelling was performed for four samples in the NCKFMASHTO system
407 using the Perple-X software (version 6.8.4; Connolly 1990, 2005). Their chemical composition was

408 estimated by averaging of thin-section surface quantitative composition scans using a FEG-SEM and
409 are as follow in oxide mass percentage: JA26 (SiO₂ 49.74, TiO₂ 1.00, Al₂O₃ 30.87, FeO 10.00, Fe₂O₃
410 1.77, Mn 0.00, MgO 1.18, CaO 0.32, Na₂O 0.83 and K₂O 3.82); SK1915g (SiO₂ 74.50, TiO₂ 0.86, Al₂O₃
411 13.99, FeO 5.05, Fe₂O₃ 0.89, Mn 0.07, MgO 1.19, CaO 0.24, Na₂O 0.81 and K₂O 2.30); PB1828a (SiO₂
412 57.13, TiO₂ 1.08, Al₂O₃ 21.57, FeO 8.08, Fe₂O₃ 0.00, Mn 0.29, MgO 1.84, CaO 0.45, Na₂O 0.53 and
413 K₂O 4.75) and BS1824C (SiO₂ 61.30, TiO₂ 0.87, Al₂O₃ 22.36, FeO 7.13, Fe₂O₃ 0.00, Mn 0.12, MgO 2.56,
414 CaO 0.58, Na₂O 0.56 and K₂O 4.13). Garnets in sample PB1828a display well-defined zoning with
415 distinct cores that were subtracted from the surface composition to calculate the bulk composition
416 relevant to infer peak equilibrium conditions. The thermodynamic dataset from Holland and Powell
417 (2011) was used with the following set of activity models for solid solutions: melt, chlorite, biotite,
418 orthopyroxene, garnet, white mica, chloritoid (White et al., 2014), ilmenite (White et al., 2000),
419 epidote (Holland and Powell, 2011) and feldspar (Holland and Powell, 2003). Pseudosection
420 calculation was performed with excess water owing to the abundance of hydrated phases. In the
421 absence of carbonates and negligible presence of organic matter in the studied sample, CO₂ was
422 neglected and a fixed water activity of 1 was used for the fluid. The influence of the redox state was
423 investigated by testing Fe³⁺/Fe_{TOT} mass ratios of 0, 0.15 and 0.25 following White et al. (2014) and
424 Ague (1991), and using the best fit to relative volume proportions.

425

426 **4 Results**

427 **4.1 Petrography and mineral chemistry**

428

429 The samples from six different metamorphic complexes, including five from the Kashmar-
430 Kerman Tectonic Zone, were studied and dated in order to constrain their P-T-t evolution. Our main
431 focus was on the Eastern domain of the KKTZ (Sarkuh, Tashk and Boneh-Shurow complexes), where
432 the metamorphism, which is poorly characterized, is considered Paleozoic based on one U-Pb zircon
433 age (Ramezani and Tucker, 2003). Only limited structural data exist for these tectonometamorphic
434 units (Ramezani and Tucker, 2003; Verdel et al., 2007; Masoodi et al., 2013; Soleimani et al., 2021),
435 such that their internal deformation and the relationships between them is still poorly understood,
436 particularly for the Sarkuh complex. In the Central domain, four samples were studied in the Posht-e-
437 Badam complex, whose timing of metamorphism is constrained by one Ar-Ar amphibole age
438 (Kargaranbafghi et al., 2012). In the Western domain, two samples were dated for further validation
439 of our method since radiometric and petrological data for the Chapedony complex are tightly
440 constrained (Ramezani and Tucker, 2003; Verdel et al., 2007; Kargaranbafghi et al., 2012, 2015).
441 Since the Jandaq complex is not only related to the subducted Anarak complex (Zanchi et al., 2009b;
442 Bagheri and Stampfli, 2008) but also shows a HT-MP metamorphism similar to that of the KKTZ
443 (Ramezani and Tucker, 2003; Bagheri and Stampfli, 2008), several samples were dated and their
444 metamorphic peak was estimated.

445

446 **4.1.1 Sarkuh complex**

447

448 Sampling in the Sarkuh complex, which exposes large-scale folds outlined by carbonate
449 horizons (Fig. 5a), was focused on the micaschists and mafic amphibolites. Mineral occurrences and
450 the composition of garnet, phengite, biotite and staurolite are presented in Table 2. The following

451 abbreviations are hereafter used to describe mineral chemistry (Fig. 6): $X_{Mg}=Mg/(Mg+Fe_{TOT})$;
452 $X_{Na}=Na/(Na+Ca+K)$ and $X_K=K/(Na+Ca+K)$.

453
454 Eleven micaschist samples (Table 2) contain garnet in a matrix of biotite, white mica, quartz
455 and plagioclase with minor oxide. Four of them contain large pluri-millimetric grains of staurolite
456 either rich in quartz inclusions (Fig. 5b; SK1908b and SK1911) or with conspicuous hourglass zoning
457 (though not related to major element zoning) and garnet inclusions, in an organic matter-rich matrix
458 (SK1808 and SK1912b; Fig. 5c). Five samples contain aluminosilicates with minor tourmaline (Fig. 5d;
459 $X_{Mg}=0.60-0.64$; $X_{Na}=0.83-0.85$; samples SK1806b, SK1809 and SK1814). Sample SK1806b contains
460 fibrous and prismatic sillimanite with minor K-feldspar, tourmaline and apatite. Sample SK1809
461 contains centimeter-large grains of andalusite, with staurolite and biotite inclusions and partly
462 replaced by biotite and sillimanite (Fig. 5e). Sample SK1814 (Fig. 5d) contains kyanite partly replaced
463 by both fibrous and prismatic sillimanite, white mica and biotite. Sample SK1915g contains kyanite
464 relicts partly pseudomorphed by biotite and sillimanite. Sample SK1923 shows cm-scale andalusite
465 crystals pseudomorphed by biotite, white mica, quartz and sillimanite. It also contains rounded pluri-
466 millimetric K-feldspar grains with quartz and biotite inclusions. Some samples neither contain
467 staurolite nor aluminosilicates (SK1812a and SK1919).

468
469 Two samples are mainly composed of quartz, plagioclase and biotite with minor garnet partly
470 retrogressed by chlorite ($X_{Mg} = 0.55-0.58$; SK1903b and $X_{Mg} = 0.27-0.65$; SK1910) and with minor
471 apatite. Sample SK1916b is a vein with large centimeter-scale grains of tourmaline ($X_{Mg} = 0.44-0.54$;
472 $X_{Na} = 0.79-0.88$) and white mica, within a matrix of quartz, plagioclase and minor biotite. Garnet is

473 found both in the matrix and as inclusion in tourmaline. Sample SK1806a is an undeformed
474 amphibolite-facies volcanoclastic rock with amphibole ($X_{Mg} = 0.56-0.55$; $X_{Na} = 0.21-0.22$; $X_K = 0.04-$
475 0.05), plagioclase, quartz, oxide and partly retrogressed biotite. Sample SK1913a is a deformed
476 amphibolite with garnet, amphibole ($X_{Mg} = 0.42-0.49$; $X_{Na} = 0.11-0.15$; $X_K = 0.07-0.10$), plagioclase,
477 quartz, biotite ($X_{Mg} = 0.48-0.51$; Ti pfu = 0.12-0.21) and oxide. Retrogression is shown by the
478 presence of garnet partly replaced by a complex mixture of biotite ($X_{Mg}=0.45-0.55$; Ti pfu=0.06-0.18),
479 chlorite ($X_{Mg} = 0.42-0.54$), epidote (Fe^{3+} pfu = 0.29-0.45), pumpellyite ($X_{Mg} = 0.37-0.70$) and apatite
480 (Fig. 5f).

481

482 **4.1.2 Boneh-Shurow complex**

483

484 The northern area of the Boneh-Shurow complex mainly consists of highly deformed
485 gneisses showing migmatitic textures (Fig. 7a) and abundant K-feldspar forming porphyroblasts or
486 distinct layers (Fig. 7b,c). Our four samples (BS1837, BS1947, BS1952e, BS1952g) are foliated
487 gneisses with alternating quartz-plagioclase-K-feldspar layers and amphibole ($X_{Mg} = 0.16-0.49$; $X_{Na} =$
488 $0.09-0.27$; $X_K = 0.04-0.20$)-epidote (Fe^{3+} pfu = 0.36-0.68)-garnet layers. Pluri-millimetric titanite
489 grains are present in both the matrix and as inclusion in garnet (Fig. 7d, e). In sample BS1947,
490 amphibole and biotite overgrowths, with minor chlorite, are found around the preexisting minerals
491 and in garnet fractures. In sample BS1952e, biotite is present in amphibole fractures. In sample
492 BS1952g, garnet and plagioclase host a few large biotite inclusions ($\sim 200 \mu m$), as well as some
493 retrograde chlorite.

494

495 The central (BS1824C, BS1924, BS1925, BS1930) and southern (BS1817d, BS1931, BS1933B)
496 areas of the Boneh-Shurow complex mostly comprise micaschists showing either the transition from
497 andalusite to sillimanite (BS1925), kyanite to sillimanite (BS1824C), or andalusite to kyanite (Fig. 7h)
498 in a biotite + white mica + quartz + plagioclase matrix with minor oxide. Mafic amphibolite boudins
499 are also present (BS1930). Sample BS1824C is a micaschist with garnet, K-feldspar, prismatic
500 sillimanite and kyanite relics showing transformation to sillimanite. Garnet grains are highly
501 deformed yet preserve an helicitic texture with kyanite and/or sillimanite inclusions (Fig. 7f, g). K-
502 feldspar is present as large pluri-millimetric rounded grains with biotite and quartz inclusions, and
503 shows evidence of albite exsolution (Fig. 7i). Sample BS1924 is a garnet-, K-feldspar and kyanite-
504 bearing micaschist where K-feldspar is again pluri-millimetric and shows rounded grains with biotite
505 and quartz inclusions. Garnet grains are deformed with asymmetric pressure shadows and contain
506 biotite inclusions ($X_{Mg} = 0.21-0.47$; $Ti\ pfu = 0.090.23$). Biotite is partly retrograde and replaces garnet
507 and kyanite. A few tourmaline grains ($X_{Mg} = 0.60-0.64$; $X_{Na} = 0.84$) are present. Sample BS1925 is a
508 garnet micaschist with large centimeter-scale rounded grain of andalusite transforming to
509 sillimanite with white mica pressure shadows. Sample BS1931 is a garnet micaschist with pluri-
510 millimetric white mica. Sample BS1933b contains minor garnet and white mica in a biotite-rich
511 matrix. Sample BS1930 is an undeformed amphibolite with garnet, amphibole ($X_{Mg} = 0.34-0.40$; X_{Na}
512 $= 0.19-0.23$; $X_K = 0.06-0.08$), plagioclase, quartz and retrograde biotite. Titanite is present in the
513 matrix and in the garnet rims. Sample BS1817d is a deformed amphibolite with garnet, amphibole
514 ($X_{Mg} = 0.41-0.45$; $X_{Na} = 0.25-0.29$; $X_K = 0.02-0.03$), partly retrogressed biotite, plagioclase, quartz,
515 minor oxide and retrograde chlorite ($X_{Mg} = 0.42-0.59$).

516

517 **4.1.3 Tashk complex**

518
519 Sample TK1802a is a garnet micaschist with an organic matter-rich matrix made of white
520 mica, plagioclase, quartz and oxide. Intergrown biotite and chlorite ($X_{Mg} = 0.46-0.52$) are found in
521 garnet pressure shadows. Sample TK1802c is a deformed amphibolite with amphibole ($X_{Mg} = 0.53-$
522 0.61 ; $X_{Na} = 0.24-0.26$; $X_K=0.05-0.07$), plagioclase, quartz and biotite. Abundant titanite overgrowths
523 are found around ilmenite (Fig. 8a). Sample TK1802b is an undeformed metacarbonate with large
524 centimeter-scale, probably inherited, lozenge-shape K-feldspar porphyroblasts, phlogopite, as well
525 as minor oxide and titanite (Fig. 8b).

526

527 **4.1.4 Posht-e-Badam complex**

528
529 Sample PB1828a, PB1828b and PB1936c are micaschists with pluri-millimetric garnet grains
530 overgrown by staurolite within a matrix of white mica, biotite, sillimanite, quartz and plagioclase
531 with minor oxide and kyanite relics. Sillimanite has co-crystalized with white mica and biotite in
532 apparent equilibrium with the garnet rim (Fig. 8e). Staurolite growth postdates that of sillimanite
533 and of the garnet rim (Fig. 8d). Sample PB1938 is a micaschist with altered garnet and staurolite,
534 within a matrix of white mica, quartz, plagioclase and minor oxide. This sample is partly foliated,
535 with a biotite rich layer containing white mica, staurolite and a few tourmaline grains ($X_{Mg} = 0.45-$
536 0.57 ; $X_{Na} = 0.61-0.84$) which appear to postdate the formation of white mica (Fig. 8f).

537

538 **4.1.5 Chapedony complex**

539

540 Samples CH1823 is a foliated gneiss with alternating layers made of quartz and plagioclase
541 ($X_{Na} = 0.66-0.78$; $X_K = 0.01-0.03$) and layers hosting rounded grains of K-feldspar with biotite
542 inclusions ($X_{Mg} = 0.49-0.55$; Ti pfu = 0.15-0.19) within a quartz and biotite matrix ($X_{Mg} = 0.42-0.55$; Ti
543 pfu = 0.05-0.14) with minor tourmaline ($X_{Mg} = 0.70-0.72$; $X_{Na} = 0.70-0.75$). Sample CH1842 is a quartz-
544 plagioclase-biotite micaschist.

545

546 **4.1.6 Jandaq complex**

547

548 Sample JA17b is a micaschist with centimeter-scale, altered garnet partly replaced by biotite
549 and chlorite ($X_{Mg} = 0.33-0.40$) and staurolite porphyroblasts within a matrix of quartz, plagioclase
550 and white mica. Sample JA26 is a quartz-rich micaschist with co-crystallized chloritoid ($X_{Mg} = 0.11-$
551 0.12) and white mica, with minor plagioclase, epidote and oxide (Fig. 8c). Sample JA28b is a garnet
552 micaschist with minor staurolite in a matrix composed of quartz, plagioclase, white mica, with partly
553 retrograde biotite, chlorite and oxide.

554

555 **4.2 Thermobarometry**

556

557 We applied the amphibole-plagioclase thermobarometer of Holland and Blundy (1994) and
558 Molina et al. (2015). Results with standard deviation are $696 \pm 30^\circ\text{C}$ and 10.5 ± 1.8 kbar (SK1913a)
559 and $760 \pm 24^\circ\text{C}$ and 8.5 ± 0.5 kbar (SK1806) for the Sarkuh complex; $641 \pm 9^\circ\text{C}$ and 7.3 ± 0.6 kbar
560 (BS1952g) and $646 \pm 12^\circ\text{C}$ and 7.7 ± 1.2 kbar (BS1930) for the Boneh-Shurow complex; $655 \pm 14^\circ\text{C}$
561 and 4.3 ± 0.7 kbar (TK1802c) for the Tashk complex.

562

563 The formation temperature of dated titanite is estimated with the Zr content using the
564 thermobarometer of Hayden et al. (2008). In samples BS1837, BS1952e and BS1947, the activity of
565 SiO_2 is considered as 1 because of the presence of quartz and the activity of TiO_2 is considered
566 between 0.5 and 1 because of the absence of rutile (Hayden et al., 2008). By varying the activity of
567 TiO_2 and the pressure between 6.5 and 9.5 kbar, we obtain a similar range of formation temperature
568 in samples BS1837 (674-768°C), BS1952e (664-769°C) and BS1947 (667-789°C). For sample BS1930
569 and TK1802c, the formation temperature is poorly constrained because of the presence of
570 impurities and the absence of quartz leading to an activity of SiO_2 between 0.5 and 1 (Hayden et al.,
571 2008). The temperature range is 618-823°C for BS1930 (with pressure between 6.5 and 9.5 kbar)
572 and 602-886°C for TK1802c (with pressure between 3.5 and 5 kbar).

573
574 Thermodynamic modelling was conducted on HT-MP samples from the Sarkuh (SK1915g),
575 Boneh-Shurow (BS1824C) and Posht-e-Badam complexes (PB1828a). The estimated peak P-T area is
576 defined by the Grt-Sil-Bt-Ph-Qz paragenesis (Figs. 8e, 9b, c) and by the composition of biotite in
577 equilibrium with sillimanite. The whole-rock composition of sample PB1828a (Posht-e-Badam),
578 where garnets are centimeter-scale and abundant (10-15%), was corrected by subtracting the mean
579 composition of prograde garnet cores, which may no longer have been part of the rock chemical
580 system at peak conditions. This appeared unnecessary for the other samples, since modal
581 proportions of garnet in SK1915g and BS1824c are low (<5%) and since it shows no preserved
582 prograde zoning. Removal of garnet cores also has a negligible impact on biotite composition.
583 Thermodynamic modelling was also conducted for the Jandaq complex (sample JA26; Fig. 9d). In the

584 four pseudosections, the predicted Ti-bearing phase (rutile for BS1824C, JA26, PB1828a and ilmenite
585 for SK1915g) and mineral modal compositions correspond to the thin section observations.

586
587 The pseudosection for the Sarkuh complex yields P-T conditions around 6.5-8.5 kbar and 685-
588 755°C, based on the X_{Mg} (0.32-0.39) and Ti content (0.11-0.22 pfu) of biotite, in the range of potential
589 melting. In the estimated P-T area, garnet composition (Alm_{79-74} Sps_{2-10} Grs_{2-6}) matches that of
590 microprobe data (Alm_{80-83} Sps_{7-5} Grs_{1-2}). Predicted modal (volume percent) abundances, i.e. 1-5%
591 garnet, 8-15% biotite, 8-10% sillimanite and 5-15% white mica, are in agreement with observations.
592 The pseudosection for the Posht-e-Badam complex gives P-T conditions around 6.5-7.5 kbar and
593 650-695°C, based on the X_{Mg} (0.41-0.48) and Ti content (0.09-0.13 pfu) of biotite. Predicted modal
594 abundances are ~6% garnet, ~17% biotite, ~8% sillimanite and ~35% white mica. Garnet gives a
595 composition (Alm_{74-75} Sps_{4-6} $Grs_{4.5-5}$) again similar to that of microprobe data (Alm_{65-81} Sps_{23-5} Grs_{4-6}).
596 Ilmenite postdates rutile, although both crystals can be observed together in some areas. The
597 pseudosection for the Boneh-Shurow complex gives P-T conditions around 7.5-9.5 kbar and 700-
598 790°C, based on the X_{Mg} (0.43-0.56) and Ti content (0.15-0.24 pfu) of biotite. This paragenesis likely
599 formed past the solidus (8-9.5kbar and 750-790°C) to explain the abundance of K-feldspar (i.e. melt
600 in the pseudosection, Fig. 7i). Predicted modal abundances are 5-10% garnet, 15-25% biotite, 10-
601 15% sillimanite and 15-25% white mica. Garnet gives a composition (Alm_{68-74} $Sps_{2-4.5}$ Grs_{5-8}) similar
602 to that of microprobe data (Alm_{72-78} Sps_{16-5} Grs_{2-3}) but with a higher grossular content. The
603 pseudosection for the Jandaq complex gives P-T conditions around 11.6-13 kbar and 410-480°C,
604 based on the X_{Mg} of chloritoid (0.10-0.12) and the Si content of coexisting white mica (3.08-3.12 pfu),
605 and on the presence of paragonite, epidote, ilmenite, rutile and chlorite (Fig. 9d, 8c). The P-T

606 estimated area predicts realistic mineral abundances for white mica (37-39%) and chloritoid (28-
607 30%).

608
609 In order to further assess the retrograde P-T path, thermobarometric calculations were
610 performed on sample PB1828a using Thermocalc V. 3.21 (Holland and Powell, 1998, Holland and
611 Powell, 2011). The H₂O activity was set to 1 based on the lack of significant halogen and carbon
612 contents in hydrous minerals. Mineral activities were determined using the AX software (Holland
613 and Powell, 1998; updated in 2011). The retrograde assemblage of staurolite ($\text{Mg}_{0.534} \text{Fe}^{2+}_{2.921} \text{Al}_{18.232}$
614 $\text{Si}_{7.459} \text{O}_{44} (\text{OH})_4$), white mica ($\text{K}_{0.866} \text{Na}_{0.172} \text{Al}_{2.932} \text{Mg}_{0.045} \text{Fe}^{2+}_{0.065} \text{Si}_{2.970} \text{O}_{10} (\text{OH})_2$), biotite ($\text{K}_{0.902} \text{Na}_{0.037}$
615 $\text{Al}_{1.777} \text{Mg}_{1.114} \text{Fe}^{2+}_{1.253} \text{Si}_{2.639} \text{Ti}_{0.107} \text{O}_{10} (\text{OH})_2$), plagioclase ($\text{K}_{0.003} \text{Na}_{0.607} \text{Ca}_{0.423} \text{Al}_{1.478} \text{Si}_{2.523} \text{O}_8$) and
616 garnet rim ($\text{Ca}_{0.215} \text{Mn}_{0.105} \text{Mg}_{0.292} \text{Fe}^{2+}_{2.269} \text{Fe}^{3+}_{0.179} \text{Al}_{2.037} \text{Si}_{2.894} \text{O}_{12}$) gives a P-T estimate with
617 standard deviation of 629 ± 27 °C and 6.5 ± 1.1 kbar and a sigfit of 0.97. This estimate is close to the
618 estimated peak condition (6.5-7.5kbar and 650-695°C) and coincides, in the pseudosection, with the
619 destabilization of sillimanite to staurolite (Fig. 9) at the beginning of the retrograde path.

620
621 We also applied the methods of Henry et al. (2005) and Wu and Chen (2015), with the latter
622 taking pressure into account, to estimate temperature based on peak biotite composition (Tab. 3).
623 Results show some discrepancy between the Wu and Chen (2015) method (using pressure
624 estimation of the pseudosections) and the Henry et al. (2005) method: 650 ± 16 °C at 10.5 kbar versus
625 544 ± 22 °C (sample SK1913a); 522 ± 25 °C at 4.3 kbar versus 525 ± 33 °C (TK1802c); 594 ± 16 °C at 7
626 kbar versus 404 ± 38 °C (PB1828a); 692 ± 26 °C at 9 kbar versus 538 ± 31 °C (BS1824C). Using the
627 pressure calculated from the pseudosection, the revised method of Wu and Chen (2015) for the Ti

628 in biotite give higher formation temperatures, consistent with our other temperature estimates. The
629 method of Henry et al., 2005 is used to index all biotite dates without a pressure estimation.

630

631 **4.3 Titanite U/Pb geochronology**

632

633 We dated three gneisses from the northern Boneh-Shurow complex. All give similar lower
634 intercept dates (Fig. 10): 178.0 ± 1.2 Ma (BS1952e); 179.6 ± 1.0 Ma (BS1947) and 187.4 ± 0.9 Ma
635 inclusions and those from the matrix. However, titanite from BS1837 is mostly in garnet core, which
636 could explain a partly prograde older date. Another sample from the central part Boneh-Shurow
637 complex gives a less well-constrained date: BS1930 (191.3 ± 4.0 Ma), due, in part, to smaller titanite
638 that contain significant impurities, and have lower radiogenic/common Pb ratios. This date,
639 nevertheless, overlaps with that obtained for BS1837 in the northern part.

640 One sample from the Tashk complex (TK1802c; Fig. 10), with titanite filled with impurities
641 and growing around ilmenite, yielded a poorly constrained date (225 ± 12 Ma), which while
642 significantly different than those in the Boneh-Shurow complex, is nonetheless compatible with the
643 Cimmerian orogeny.

644

645 **4.4 Mica Rb/Sr geochronology**

646

647 Results of Rb/Sr dating are presented in Table 3 and Figure 11 for biotite, and in
648 supplementary data for white mica. The obtained Rb/Sr dates may represent crystallization or
649 cooling ages, and are therefore compared to the Si content of white mica and to the Ti-in-biotite
650 temperature estimated using the thermometer of Henry et al. (2005). Commonly reported closure

651 temperatures for the Rb–Sr geochronometer range from 500°C to >600°C for white mica (e.g.,
652 Blanckenburg et al. 1989; Freeman et al. 1997; Glodny et al. 1998, 2008; Purdy and Jäger 1976) and
653 between 300°C–450°C for biotite (Armstrong et al., 1966; Jager et al., 1967; Verschure et al., 1980;
654 Del Moro et al., 1982; Jenkin et al., 2001).

655
656 With our Rb/Sr in-situ dating method, we obtained constrained dates for biotite with 2σ
657 between 1.4 and 43 Ma. Constrained dates were obtained even for texturally heterogeneous or
658 small biotite crystals, such as in sample TK1802a, where biotite is interstratified with chlorite, or
659 sample BS1824C, where biotite appears as small inclusions in K-feldspar (Fig. 7i). In white mica,
660 however, the spread of $^{87}\text{Rb}/^{86}\text{Sr}$ ratios is restricted, leading to poorly constrained isochron slopes
661 and dates (2σ between 15 and 210 Ma). Using another mineral to constrain the origin of the isochron
662 can significantly refine its slope (i.e. SK1916b with tourmaline and TK1802b with K-feldspar; Fig. 11).
663 However, this can only be done assuming a closed system behavior and/or that minerals have co-
664 crystalized. The date of TK1802b, in particular, must be interpreted with caution because the
665 phlogopite and K-feldspar may have been inherited (Fig. 8b).

666 In the four samples from the Posht-e-Badam complex, biotite shows similar dates (158–162
667 Ma) and consistent Ti-in-biotite temperatures ($\sim 353\text{--}435^\circ\text{C}$). White mica and biotite co-crystalized
668 but two of the white mica dates are significantly older (184–270 Ma; they also have a low Si content
669 ~ 3.0 pfu).

670 In the northern Boneh-Shurow, biotite, mainly retrograde, gives dates ranging from 129 to
671 154 Ma, with similar Ti-in-biotite temperatures ($\sim 435\text{--}485^\circ\text{C}$). In the central Boneh-Shurow, biotite,
672 largely retrograde too, shows dates between 141 and 163 Ma, with similar Ti-in-biotite

673 temperatures ($\sim 487\text{-}539^\circ\text{C}$; samples BS1924, BS1925 and BS1930). Biotite from sample BS1824C,
674 whether from the matrix or as inclusion in K-feldspar, gives the same date (152-159Ma) and the
675 same Ti-in-biotite temperature ($\sim 537\text{-}542^\circ\text{C}$). White mica dates with Si content of $\sim 3.05\text{-}3.12$ pfu
676 overlap with biotite dates. In the southern Boneh-Shurow, retrograde biotite in BS1817d has a
677 poorly constrained date (189 ± 25 Ma). Sample BS1933b has a biotite date (159 ± 2 Ma) and Ti-in-
678 biotite temperature ($\sim 515^\circ\text{C}$) similar to others. Sample BS1931 gives a similar biotite date (150 ± 4
679 Ma) and Ti-in-biotite temperature ($\sim 424^\circ\text{C}$); its white mica date is however significantly older (234
680 ± 22 Ma).

681 In the Sarkuh complex, white micas (with Si content $\sim 3.04\text{-}3.08$ pfu) generally have poorly
682 defined dates that broadly overlap with biotite dates, save for the vein sample SK1916b, which has
683 a significantly older white mica and tourmaline date (198 ± 15 Ma) than that of biotite (<180 Ma).
684 Biotite in samples SK1806a, SK1812a, SK1814, SK1915g, SK1923, SK1910, SK1809 and SK1913a
685 formed near peak and have Ti-in-biotite temperature of $\sim 466\text{-}617^\circ\text{C}$ with dates spreading from 157
686 to 175 Ma. Biotite in samples SK1806b, SK1919, SK1908b, SK1912b and SK1808, which is post-peak
687 (with Ti-in-biotite temperature of $\sim 286\text{-}549^\circ\text{C}$), shows a date spread from 149 to 177 Ma. Late
688 retrograde biotite overgrowths in samples SK1903b and SK1911 have dates of 104 ± 23 and $122 \pm$
689 11 (with Ti-in-biotite temperatures of $\sim 389^\circ\text{C}$ and $\sim 319^\circ\text{C}$, respectively). Metasomatic biotite from
690 sample SK1913a is associated with chlorite, pumpellyite and apatite and has a poorly constrained
691 date of 140 ± 43 Ma with a Ti-in-biotite temperature of $\sim 519^\circ\text{C}$ (similar to other biotite in this
692 sample).

693 In samples from the Jandaq complex, white mica yields significantly older dates (468-182
694 Ma) than biotite (213-142 Ma), consistent with the fact that biotite formed after white mica in both

695 samples (JA17b and JA28b). However, sample JA17b has a significantly older biotite date (202 ± 11
696 Ma; $\sim 360^\circ\text{C}$) than sample JA28b (148 ± 6 Ma; $\sim 390^\circ\text{C}$). White mica from sample JA26 shows a
697 significantly higher Si content (~ 3.12 pfu) than in all other samples.

698 In samples from the Tashk complex, biotite shows a relatively large spread of date (150-182
699 Ma). The temperature of biotite Ti-in-biotite is similar for TK1802b and TK1802c (~ 460 and $\sim 526^\circ\text{C}$
700 respectively). The Ti content of biotite in sample TK1802a could not be measured due to
701 interstratification with chlorite.

702 In the two samples from the Chapedony complex, biotite has the same date (40-44 Ma) and
703 Ti-in-biotite temperature ($\sim 555^\circ\text{C}$).

704

705 **5 Discussion**

706

707 **5.1 Metamorphic evolutions in Central Iran**

708

709 The widespread occurrence of sillimanite or of the Grt-St-Bt paragenesis in metapelites from
710 the Sarkuh, Boneh-Shurow and Posht-e-Badam complexes is indicative of MP-MT to MP-HT
711 metamorphism (Tab. 2). In the Sarkuh and Boneh-Shurow complexes, some samples record the
712 transition from andalusite to sillimanite (BS1925, SK1809, SK19323) and some from kyanite to
713 sillimanite (BS1824c, BS1924, SK1814, SK1915g; Fig. 5). While broadly similar peak P-T conditions
714 were obtained for both types, further work is needed to understand if these mineralogical
715 evolutions represent somewhat different metamorphic grades and prograde paths, and therefore
716 distinct tectonometamorphic units in these complexes.

717 The migmatitic texture observed in the gneisses from the northern Boneh-Shurow complex
718 (Figs. 7a,b) supports the existence of melting. In some samples from the Sarkuh and Boneh-Shurow
719 complexes, rounded, pluri-millimetric K-feldspar grains are found with rounded biotite and quartz
720 inclusions and albite exsolution (samples BS1824C, BS1924, SK1806b and SK1923; Fig. 7i; Tab. 2).
721 Biotite inclusions have similar composition and date as matrix biotite and, therefore, K-feldspar
722 likely formed after biotite by local breakdown of white mica. The perthitic nature of the feldspar
723 indicates it formed above albite exsolution, i.e. $T > 700^{\circ}\text{C}$ (Tuttle and Bowen, 1958). These
724 observations indicate that the Sarkuh and Boneh-Shurow complexes reached the wet solidus for
725 metapelites, contrary to the Posht-e-Badam complex.

726 White mica and biotite in the Posht-e-Badam complex have respectively lower Si content
727 ($\sim 3.03\text{-}3.18$ pfu) and Ti content ($\sim 0.7\text{-}0.14$ pfu) than in the Boneh-Shurow and Sarkuh complexes
728 (Fig. 6). These compositional trends hint at lower temperature and pressure conditions for the
729 Posht-e-Badam complex than for the Boneh-Shurow and Sarkuh complexes, and to a somewhat
730 different P-T path. This difference is confirmed by its older metamorphic age (~ 219 Ma;
731 Karagaranbafghi et al., 2012), when compared to that obtained for the Boneh-Shurow complex with
732 titanite ($\sim 190\text{-}180$ Ma; Fig. 13). In the Posht-e-Badam complex samples, staurolite is observed
733 overgrowing garnet (Fig. 8d). Our P-T estimation for staurolite formation ($629 \pm 27^{\circ}\text{C}$ and 6.5 ± 1.1
734 kbar) indicates that it formed near-peak at the beginning of the retrograde path. This retrogression
735 in the staurolite stability field is not documented in the Boneh-Shurow and Sarkuh complexes, where
736 staurolite only appears as part of the prograde or peak paragenesis with garnet and biotite.

737 The pseudosections for these three complexes (Fig. 9) allow to estimate the peak P-T
738 conditions (Grt-Sil-Bt-Ph) experienced by the Sarkuh (7-8.5 kbar, $685\text{-}755^{\circ}\text{C}$), Boneh-Shurow (8-9.5

739 kbar, 750-800°C) and Posht-e-Badam complexes (7-7.5 kbar, 670-690°C), confirming a slightly lower
740 metamorphic peak for the latter (Fig. 14). All indicate MT- to HT-MP conditions typical of collisional
741 gradients and are similar to the P-T estimate inferred for the Cimmerian collisional metamorphism
742 of the Shotor-Kuh complex (Fig. 12; Rahmati-Ilkhchi et al., 2011).

743 In samples from the Sarkuh and Boneh-Shurow complexes, the estimated peak conditions
744 are partly past the solidus (<30% melt, Fig. 9), as corroborated by the presence of migmatitic
745 textures in gneisses and of rounded K-feldspar grains with albite exsolution in metapelites. However,
746 since white mica is still preserved and no major migmatitic texture is observed in the studied
747 metapelite samples, partial melting was probably limited (<10-20% melt). The P-T estimate for
748 BS1824c must be taken with caution since the composition of the chemical system could have been
749 partly modified by melting, but is nonetheless consistent with our other P-T estimates (Fig. 12).

750
751 In the Jandaq complex, the presence of both HP-LT (Ph+Cld; Fig. 8a) and MP-MT (Grt+St+Bt)
752 paragenesis reveals the existence of two different metamorphic gradients and probably two distinct
753 metamorphic units. We tentatively relate the first paragenesis to the Paleotethys subduction and
754 the second one to a collisional context related to the Cimmerian orogeny. White mica from the
755 Jandaq, Sarkuh and Boneh-Shurow complexes has similar silica contents (~3.1-3.3 pfu). Biotite from
756 Jandaq shows a narrower range and mainly lower Ti content (~0.7-0.14 pfu) than biotite from the
757 Boneh-Shurow and Sarkuh complexes (~0.5-0.22 pfu), confirming a higher P/T gradient for the
758 Jandaq complex compared to the Kashmar–Kerman Tectonic Zone (KKTZ).

759 Thermodynamic modelling indicates peak conditions of 11.6-13 kbar, 410-480°C for the
760 Jandaq HP-LT paragenesis, hence comparable with the HP-LT gradient of the Rasht and Anarak

761 metamorphic complexes ascribed to the subduction of the Paleotethys (Fig. 12; Rossetti et al., 2015;
762 Zanchetta et al., 2018). In the Tashk complex, by contrast, the absence of high pressure or high
763 temperature diagnostic minerals indicates that it only experienced lower grade metamorphic
764 conditions.

765
766 The peak estimates obtained from pseudosection modeling, amphibole-plagioclase and Zr in
767 titanite thermobarometry (Holland and Blundy 1994; Hayden et al. 2008; Molina et al. 2015, Fig. 12)
768 are mutually consistent. Estimates for the temperature of biotite formation using Henry et al. (2005)
769 are somewhat lower (Fig. 12) and generally underestimated because pressure is not considered.

770

771 **5.2 Interpretation of geochronological data**

772

773 **5.2.1 Interpretation of in-situ titanite, biotite and white mica dates**

774

775 Calculated U/Pb titanite dates are considered here to date peak metamorphism due to the
776 high closure temperature of titanite (~800°C; Kohn, 2017) and to the absence of significant
777 differences in age or composition between matrix titanite and titanite inclusions in garnet.

778 Calculated Rb/Sr dates are considered to mark the time when diffusion of Sr in the crystalline
779 network and across mineral boundaries becomes insignificant. They are therefore commonly
780 interpreted as "cooling ages", i.e. dating cooling of the sample below the system closure
781 temperature (Jager et al., 1967). Aside temperature, however, several parameters may influence
782 the closure of the system, such as the cooling rate, recrystallisation, fluid-rock interactions, grain

783 size and interaction with adjacent minerals. Some authors have estimated that fluid- and
784 deformation-enhanced recrystallisation is more efficient than diffusive re-equilibration (Villa, 2010).

785 In the studied samples, biotite formed at peak or near peak conditions, at temperatures
786 significantly higher than its assumed typical closure temperature (300-450°C). Therefore, the Rb/Sr
787 dates for biotite most likely reflect cooling ages (or later resetting through fluid infiltration or
788 recrystallization). The potential higher closure temperature of white mica (500°C to >600°C) makes
789 the interpretation of Rb/Sr for this mineral more ambiguous: they could correspond to either
790 crystallization or cooling ages. Dates obtained for white mica in this study are rather poorly
791 constrained because of relatively low Sr ratios (Fig. 11), which is best explained by the relative
792 enrichment in Sr compared to Rb in white mica (Bebout et al., 2007). In high grade rocks, the
793 coexistence of white mica and biotite induces a partitioning of Sr and Rb with Sr favoring white mica
794 (Yang and Rivers, 2000). The in-situ Rb/Sr technique applied here therefore provides optimal
795 constraints, through biotite, on the cooling history of metamorphic rocks, but is not conveniently
796 suited to constrain the peak or early cooling stages.

797

798 **5.2.2 An exhumation history punctuated by Cimmerian events?**

799

800 Almost all Rb/Sr ages are Jurassic and fit in the range of the Mid- (~170 Ma) to Late (~140
801 Ma) Cimmerian tectonic events (Fürsich et al., 2009b; Wilmsen et al., 2015; Fig. 13; Tab. 3), with
802 three main populations of biotite cooling ages:

803 (i) ~170 Ma, for the Tashk and Sarkuh complexes.

804 (ii) ~160-150 Ma for the Boneh-Shurow and Posht-e-Badam complexes, with a narrow age
805 range for the latter. In the Boneh-Shurow complex, ages tend to decrease from 160 to 140 with
806 decreasing formation temperatures for biotite (Fig. 13), with the exception of one age in the
807 southern part. This may reflect gradual exhumation associated with biotite formation. No such trend
808 is observed in the Sarkuh complex, whose ages range mainly between 170 and 150 Ma.

809 (iii) A subordinate population ranging between 150 and 140 Ma at Jandaq, Sarkuh and
810 Boneh-Shurow complexes. Some biotite ages between 140 and 100 Ma, in the Sarkuh complex,
811 stand out as significantly younger and may be related to reequilibration during final exhumation.

812 Some older Rb/Sr ages are also obtained (Fig. 13). Two white mica ages in the Posht-e-Badam
813 complex (228 ± 42 and 214 ± 30 Ma), and one from the Boneh-Shurow complex (234 ± 22 Ma), are
814 significantly older than the biotite ages obtained for the same samples, and are synchronous with
815 the Eo-Cimmerian event. The biotite Rb/Sr ages of this study are consistent with previously
816 published cooling ages for the Posht-e-Badam, Boneh-Shurow and Jandaq complexes (Fig. 13, Tab.
817 2) and are altogether similar. They are also internally consistent with this study titanite ages (177-
818 237 Ma; Figs. 10,13). These ages stand in contradiction with previous suggestions of an early
819 Paleozoic metamorphism for the Sarkuh, Tashk and Boneh-Shurow complexes, which were based
820 on a single U-Pb Zircon age of 547.6 ± 2 Ma (Ramezani and Tucker, 2003). Our results instead support
821 the Cimmerian Ar-Ar hornblende age obtained for one sample of the Posht-e-Badam complex (219.2
822 ± 1.2 Ma; Karagaranbafghi et al., 2012). In the Jandaq complex, apart from one poorly constrained
823 Carboniferous Rb/Sr age, other Rb/Sr ages lie between 180 and 260 Ma, which would either coincide
824 with the Cimmerian orogeny or a late stage of Paleotethyan subduction (Tab. 3, Figs. 12, 13). Given
825 that the P-T conditions experienced by this sample (JA26) align on the same HP-LT gradient as the

826 subducted Anarak or Rasht complexes (Fig. 12), this age more likely relates to a subduction stage,
827 possibly partial reset during the Cimmerian orogeny.

828 In the Tashk complex, the presence of inherited biotite and K-feldspar could explain the older
829 titanite age obtained for the TK1802b sample. The other biotite ages (168 ± 3 and 153 ± 3 Ma) are
830 in the range of biotite ages of the neighboring Sarkuh complex. Verdel et al (2007), however,
831 obtained older Ar-Ar biotite ages between 218 and 295 Ma. Furthermore, the onlap of Permian and
832 Triassic shallow-marine carbonates onto the Tashk complex demonstrate its exposure before the
833 onset of the Cimmerian orogeny. Considering our U-Pb titanite age (225 ± 12 Ma) and P-T estimate
834 ($655 \pm 15^\circ\text{C}$; 4-5 kbar; Fig. 12), these observations collectively indicate shallow burial and heating of
835 the Tashk complex during the Cimmerian Orogeny and its subsequent exhumation after the Mid-
836 Cimmerian event. Further investigation of the tectonic relationships between the Tashk complex
837 and the adjacent higher-grade complexes of the Boneh-Shurow and Sarkuh complexes is needed to
838 explain the detailed stacking of the various tectonometamorphic units. At the other end of the
839 spectrum, the two biotite ages obtained for the Chapedony massif (42.5 ± 2.0 and 42.7 ± 1.8 Ma)
840 confirm its cooling and exhumation as a core-complex during the Eocene (Kargaranbafghi et al.
841 2015).

842
843 As discussed above (section 5.2.1), biotite ages are interpreted to reflect cooling of the
844 metamorphic terranes below $300\text{-}450^\circ\text{C}$ during the Jurassic and the Cretaceous. During the Jurassic,
845 the Yazd block and the KKTZ were exposed (Figs. 3,14), whereas the Tabas and Lut Blocks formed a
846 submerged extensional basin with large tilted fault blocks (Figs. 3,14; Wilmsen et al. 2009b; Salehi
847 et al. 2018). Provenance studies of the Lower Jurassic Tabas sedimentary record point to a dominant

848 recycled-orogen source, possibly from the KKTZ (Salehi et al., 2018). This indicates significant erosion
849 of the KKTZ during the lower Jurassic and before the Mid-Cimmerian event. After the Mid-
850 Cimmerian event, Upper Jurassic sedimentation shows increased subsidence and development of a
851 carbonate platform in the Tabas Block (Wilmsen et al., 2009a), similar to what is observed in the
852 Alborz (Fürsich et al., 2009b). This may be indicative of a prolonged extensional setting responsible
853 for the exhumation of the Sarkuh and Boneh-Shurow complexes. This would be consistent with
854 claims of core-complex formation along the KKTZ at ~165 Ma (Masoodi et al., 2013; Soleimani et
855 al., 2021). However, while the Sarkuh and Boneh-Shurow complexes were interpreted as sharing the
856 same deformation and exhumation history (Masoodi et al., 2013), the 10 Myr difference of biotite
857 cooling ages (Fig. 13) points to somewhat diachronous exhumation for the two sectors.

858
859 Given that the biotite ages obtained here are mostly younger than 170 Ma, we propose that
860 cooling and exhumation started and/or accelerated during the Mid-Cimmerian event, probably as a
861 result of renewed extension. No compressive event is indeed observed in the KKTZ at 170 Ma, nor
862 any magmatic stage which may have reset ages through heating. The data also mark a diachronous
863 exhumation of the different complexes of the KKTZ, with earlier exhumation in the east than in the
864 west. This may relate to the large-scale fault bounding the KKTZ to the east, which accommodated
865 the Jurassic rotation of the Tabas block (Wilmsen et al., 2009a; Fig. 3).

866

867 **5.3 Geodynamic implications**

868

869 We report the existence of a collisional MP-HT metamorphic imprint in the KKTZ, in the
870 Boneh-Shurow, Sarkuh and Posht-e-Badam complexes, with peak burial to depths of 20-25 km at
871 180-190 Ma for the Boneh Shurow complex and possibly somewhat earlier, during the Eo-
872 Cimmerian event, for the Posht-e-Badam complex (Fig. 12; ~210 Ma; Karagaranbafghi et al. 2012).
873 Shortening accompanied the tectonic stacking of these units, as shown by the juxtaposition of the
874 Sarkuh complex onto the lower pressure Tashk complex (Fig. 4). Rb/Sr geochronology reveal a main
875 cooling and exhumation stage ~170-160 Ma for all units, hence during or immediately postdating
876 Mid-Cimmerian times.

877 Their exhumation likely took place through extensional tectonics accompanied by the
878 formation of core-complexes (Masoodi et al., 2013; Soleimani et al., 2021). On a broader scale, the
879 exhumation of the Boneh-Shurow, Sarkuh, Posht-e-Badam and Tashk complexes coincided with
880 distributed crustal-scale extension in Iran (Fig. 14, Fürsich et al. 2009b; Wilmsen et al. 2009b,a),
881 marked by pronounced subsidence, normal faulting and block rotation in the Tabas block (Wilmsen
882 et al., 2009b; Salehi et al., 2018) and by the onset of seafloor spreading in the South Caspian Basin
883 (Fürsich et al., 2009b). Meanwhile, widespread and profuse arc magmatism related to the
884 subduction of the Neotethys is recorded in the transtensional Sanandaj-Sirjan zone (Agard et al.,
885 2011; Hassanzadeh and Wernicke, 2016), as well as arc-related magmatic activity and HT-LP
886 metamorphism in the Deh-Salm and Anjul metamorphic complexes to the north of the Lut block
887 (Bröcker et al., 2016). During the Middle Jurassic, before the anticlockwise rotation of the CEIM
888 (Mattei et al., 2015), the Lut block was located, like the Sanandaj-Sirjan Cimmerian block, in the
889 upper plate magmatic arc setting of the Neotethyan subduction zone (Figs. 3, 14; Esmaeily et al.
890 2005; Bröcker et al. 2014; Mahmoudi et al. 2010). This period between 170 and 160 Ma therefore

891 corresponds to an important geodynamic evolution, from Paleotethys closure and Eo-Cimmerian
892 orogeny to a prevalence of Neotethys subduction dynamics with back-arc basin formation from the
893 Middle Jurassic onwards (Agard et al., 2011; Moghadam and Stern, 2015). We propose that the
894 prevalence of cooling ages between 170 and 160 Ma reflects an increase of exhumation rates in the
895 KKTZ during that period.

896
897 In contrast, the origin of the collisional metamorphism documented here is enigmatic: while
898 peak burial appears broadly coeval with the Eo-Cimmerian orogeny associated with Paleotethys
899 closure (220-185 Ma; Wilmsen et al., 2009b), the KKTZ is located several hundred kilometers away
900 from the suture zone, with no evidence for significant shortening and thickening in between. A
901 similar collisional MP-HT metamorphism of Cimmerian age is observed in three distinct areas (KKTZ,
902 Jandaq and Shotor Kuh; Rahmati-Ilkhchi et al., 2010, 2011) located along large-scale strike-slip
903 systems, which were active throughout the Mesozoic (Figs. 1, 2; KKTZ fault system: Masoodi et al.
904 2013; Doruneh fault: Berra et al. 2017; Malekpour-Alamdari et al. 2017; Naini-Kalmard fault:
905 Wilmsen et al. 2021). Formation and exposure of these different complexes may therefore
906 tentatively be related to the accommodation of transcurrent deformation following Paleotethys
907 closure and subsequent exhumation. There is no petrological or radiochronological evidence to
908 support that this metamorphism might have occurred earlier (e.g., during the Panafrican or
909 Paleozoic) and been overprinted during the Cimmerian event.

910
911 The KKTZ and the Yazd block were mainly emergent during the Jurassic period with no
912 significant sedimentary burial throughout the Jurassic (Dercourt et al. 1986; Fürsich et al. 2003;

913 Wilmsen et al. 2003, 2005, 2010). Additionally, Cimmerian deformation and metamorphism appear
914 absent from the Yazd block: the Triassic rocks are weakly deformed (Salehi et al., 2018) and the
915 molasse-type sediments marking the closure of the Paleotethys are metamorphosed in the
916 greenschist facies at most (i.e., the Shemshak formation; Zanchi et al., 2015). The observed
917 collisional metamorphism appears restricted to the KKTZ and did not significantly affect the rest of
918 the CEIM.

919
920 Two possible tectonic settings can therefore be envisioned for the genesis of this collisional
921 metamorphism: (i) crustal shortening and burial occurred as a result of the main collisional
922 movements, in the vicinity of the Paleotethys suture zone, and the metamorphosed units were later
923 displaced and offset along large-scale transcurrent faults; (ii) crustal shortening and burial took place
924 across a weak zone/domain located outboard of the main collision zone, between the Yazd and
925 Tabas block (I.e. to the south of the Alborz and Central Iran blocks; Fig. 3, 14).

926
927 The first hypothesis implies a several hundred kilometer large displacement of the KKTZ
928 relative to the Yazd block after the Eo-Cimmerian to reach the present-day CEIM configuration. No
929 significant differential rotation within the CEIM is however documented by paleomagnetic data from
930 the Triassic onwards (Soffel et al., 1996; Mattei et al., 2015). Furthermore, the Paleotethys suture
931 zone is continuously described between Iran and Afghanistan (Fig. 1) and it is unlikely that the KKTZ
932 was situated along the suture, east of the Yazd block in place of the Afghan block. Therefore, we
933 favor the second hypothesis and suggest that the polyphase “Cimmerian” metamorphism and
934 magmatism in the KKTZ (and in the Shotor Kuh area) marks the closure and shortening of

935 rheologically weak domains, possibly small basins with thinned crust, which were still separating the
936 Western Cimmerian blocks at the time, e.g. between the Yazd and Tabas blocks.

937

938 **6 Conclusions**

939

940 This study places constraints on the timing and intensity of burial and exhumation of
941 Cimmerian metamorphic rocks in Central Iran based on combined metamorphic and geochronologic
942 data:

943 (1) In-situ Rb/Sr dating of biotite yields precise and accurate cooling ages ($\pm 2-5$ Ma), mostly
944 in the range 170-150 Ma, that are consistent between the different complexes. In contrast, white
945 mica gives poorly constrained ages due to Sr partitioning between biotite and white mica.

946 (2) Results show that the central (Posht-e-Badam) and eastern (Boneh-Shurow, Sarkuh)
947 domains of the Kashmar–Kerman Tectonic Zone (KKTZ) were buried along a Barrovian metamorphic
948 gradient down to ~ 25 km (8-9 kbar; $\sim 750^\circ\text{C}$), during the Jurassic. This stands in contradiction with
949 former studies advocating for Paleozoic metamorphism of the eastern complexes. The higher
950 pressure and lower temperature metamorphism of the Jandaq complex occurred earlier, during
951 subduction of the Paleotethys.

952 (3) This Barrovian metamorphism is coeval with or slightly younger than the Cimmerian
953 orogeny, as shown by titanite crystallization ages between 190 and 180 Ma. Based on
954 paleogeographic reconstructions showing that the KKTZ lied hundreds of kilometers south of the
955 Paleotethys suture zone, we propose that this metamorphism results from the closure and

956 shortening of a rheologically weak domain located outboard of the main suture, which separated
957 the Yazd and Tabas blocks (such as a small basin with thinned crust).

958 (4) Exhumation of the KKTZ metamorphic rocks occurred between 170 and 140 Ma, just after
959 the Mid-Cimmerian event (~170 Ma). This is coeval with the widespread extension documented
960 throughout Iran, which is thought to result from upper plate extension above the Neotethyan
961 subduction system.

962

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964

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975

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- 1353

1354 **Figure Captions**

1355

1356 *Figure 1: Simplified geological map of Iran showing the main tectonic subdivisions and locations*
1357 *discussed in the text. Insert showing the Iranian cimmerian blocks and blue lines showing oceanic*
1358 *sutures. Ophiolites are shown in pink and intrusive and volcanic rocks in red. Metamorphic complex*
1359 *related to the Paleotethys closure in green areas. KKTZ=Kashmar-Kerman Tectonic Zone,*
1360 *AMC&JMC=Anarak and Jandaq Metamorphic complexes. Modified from Paul et al. 2010.*

1361

1362 *Figure 2: Geological map of central Iran. Modified from Zanchi et al. (2015).*

1363

1364 *Figure 3: a,b,c) Configuration of the western Cimmerian blocks from the Cambrian to the Jurassic*
1365 *modified from Torsvik and Cocks (2016) and Wilmsen et al. (2009). d) Paleogeographic situation of*
1366 *the CEIM after the Cimmerian orogeny modified from Salehi et al. (2018). The timeline (Ma) shows*
1367 *the major orogeneses and oceans affecting the Cimmerian blocks as well as the metamorphic events*
1368 *affecting Central Iran (References in text).*

1369

1370 *Figure 4: Simplified geological map of the Kashmar-Kerman Tectonic Zone with localization of*
1371 *samples and cross sections. Modified from Ramezani and Tucker (2003) and Haghypour et al. (1977).*

1372

1373 *Figure 5: (a) Panorama of part of Sarkuh complex showing large folded marble layers.*
1374 *Microphotographs of samples from the Sarkuh complex of (b) millimetric garnet and staurolite with*
1375 *biotite in their pressure zones, (c) centimetric zoned staurolite with garnet inclusion and biotite in*

1376 *pressure zone (d) millimetric prismatic sillimanite associated with biotite, (e) centimetric andalusite*
1377 *grains partly replaced by sillimanite, biotite and garnet and (f) zoned garnet replaced by pumpellyite*
1378 *and biotite.*

1379
1380 *Figure 6: (a) Composition of garnets in a Grs vs Sps diagram and in a Sps+Grs-Alm-Prp ternary*
1381 *diagram. (b) Composition of white micas in a 2Al vs SiFeMg diagram. (c) Composition of biotites in a*
1382 *XMg vs Ti diagram.*

1383
1384 *Figure 7: Boneh-Shurow complex. (a) Deformation in gneisses with migmatization from the area 1.*
1385 *(b) Augen gneiss with large phenocrystals of K-feldspar from the area 1. (c) K-feldspar rich layers in*
1386 *the gneiss from area 1. SEM pictures of gneiss samples from the area 1 of (d) garnet with titanite*
1387 *inclusions and (e) Bt+Qz+Pl and Amp+Ttn+Grt layers. (f,g) SEM pictures of sample IC1824 from area*
1388 *2 showing a highly deformed garnet with helicoidal texture and sillimanite inclusions. (h)*
1389 *Pseudomorphosis of Andalusite into kyanite from area 2. (i) SEM picture of a large round grain of K-*
1390 *feldspar from sample IC1824c with inclusions of biotite and quartz and albite exsolutions.*

1391
1392 *Figure 8: SEM pictures and microphotographs of samples of (a) an amphibolite with biotite and*
1393 *titanite overgrowing ilmenite from the Tashk complex and (b) potentially inherited phlogopite and*
1394 *automorph K-feldspar in a carbonate; (c) cocrystalized chloritoid and white mica from the Jandaq*
1395 *complex; (d) cocrystalized biotite and sillimanite in equilibrium with garnet, (e) staurolite and biotite*
1396 *overgrowing garnets and (f) late euhedral and zoned tourmaline from the Posht-e-Badam complex.*

1397

1398 *Figure 9: pseudosection of samples IC1915g, IC1828a, IC1824c and AN26d from the Sarkuh, Poshte-*
1399 *Badam, Boneh-Shurow and Jandaq complexes respectively.*

1400
1401 *Figure 10: U-Pb isochrons of titanites with 2σ uncertainties and Zr content in ppm with data plotted*
1402 *on the colour gradient scale. Tera-Wasserburg plots of titanite petrochronological data. Lower*
1403 *intercept ages are calculated using the robust regression of Powell et al. (2020) and reported at 2SE*
1404 *with s the spine width. Diagrams generated using the ChrontouR package (Larson 2020;*
1405 *doi:10.17605/OSF.IO/P46MB) written for the open R software environment.*

1406
1407 *Figure 11a: Biotite Rb/Sr dates with 2σ uncertainties from the Jandaq, Chapedony, Poshteh-Badam*
1408 *and Boneh-Shurow complexes.*

1409
1410 *Figure 11b: Biotite Rb/Sr dates with 2σ uncertainties from the Boneh-Shurow, Tashk and Sarkuh*
1411 *complexes.*

1412
1413 *Figure 11c: Biotite Rb/Sr ages with 2σ uncertainties from the Sarkuh complex.*

1414
1415 *Figure 12: Summary of P-T peak conditions from this study with peak conditions from Shotor-Kuh*
1416 *(Rahmati-Ilkhchi et al., 2011); Anarak (Zanchetta et al., 2018) and Rasht blueschists (Rossetti et al.,*
1417 *2015).*

1418

1419 *Figure 13: This study ages with 2σ bar from the Jandaq and KKTZ metamorphic complexes associated*
1420 *with biotite temperature of crystallization from Ti in biotite (Henry et al., 2005). Same order of*
1421 *samples as in Table 3, poorly constrained ages of IC1812a, IC1915g and IC1913a are excluded. Ages*
1422 *from the literature are also summarized on the left part, see text and table 1 for further details.*

1423
1424 *Figure 14: Geodynamic model for Central-East Iran during the Jurassic, with the Cimmerian main*
1425 *orogeny ~200 Ma, followed by back-arc rifting ~180 Ma and development of the Caspian Sea ~165*
1426 *Ma. Large scale strike slip movements later brought together the MP-HT rocks of the Jandaq and the*
1427 *KKTZ complexes.*

1428

1429 **Tables**

1430 *Table 1: Geochronological data (2σ) from the Kashmar-Kerman Tectonic Zone in Central Iran.*

1431

1432 *Table 2: Samples nature, mineralogy (+quartz, +feldspar, + Fe-oxides), localization and mineral*
1433 *composition. Mineral abbreviations are after Whitney and Evans (2010).*

1434

1435 *Table 3: Rb-Sr dating with biotite temperatures based on Ti content (Henry et al. 2005), Si content*
1436 *of white mica and micas textural relationships with mineral abbreviations after Whitney and Evans*
1437 *(2010).*

1438

1439

1440

Rock	Method and dated mineral	Age and error (Ma)	Interpretation	Author(s)
Western domain				
Daranjir diorite	U-Pb zircon TIMS	43.4±0.2	Intrusion	Ramezani and Tucker (2003)
Khoshoumi granite	U-Pb zircon TIMS	44.3±1.1	Intrusion	Ramezani and Tucker (2003)
Chapedony porphyroblastic gneiss	U-Pb zircon TIMS	52±9	Intrusion	Ramezani and Tucker (2003)
Chapedony biotite-gneiss	U-Pb zircon TIMS	46.8±2.5	Intrusion	Ramezani and Tucker (2003)
Chapedony migmatite	U-Pb zircon TIMS	46.3±1.7	Intrusion	Ramezani and Tucker (2003)
Khoshoumi volcanic rocks	Ar-Ar biotite	41.2±2.4	Lava extrusion	Verdel et al. (2007)
Khoshoumi dacite	Ar-Ar plagioclase	42.0±1.8	Lava extrusion	Verdel et al. (2007)
Khoshoumi dacite	Ar-Ar K-feldspar	40.5±2.5	Lava extrusion	Verdel et al. (2007)
Neybaz Mt. Mylonitic gneiss	U-Pb zircon TIMS	49.0±0.1	Intrusion	Verdel et al. (2007)
Chapedony granitic dyke	Ar-Ar white mica	42.8±0.6	Cooling age	Kargarabafghi et al. (2012)
Chapedony hornblende granite	Ar-Ar hornblende	46.4±0.4	Cooling age	Kargarabafghi et al. (2012)
Chapedony hornblende granite	Ar-Ar K-feldspar	43.8±1.4	Cooling age	Kargarabafghi et al. (2012)
Chapedony mylonitic granite	Ar-Ar biotite	45.0±0.6	Cooling age	Kargarabafghi et al. (2012)
Chapedony mylonitic granite	Ar-Ar K-feldspar	42.7±0.2	Cooling age	Kargarabafghi et al. (2012)
Chapedony mylonitic gneiss	Ar-Ar biotite	45.6±1.4	Cooling age	Kargarabafghi et al. (2012)
Chapedony sample	Ar-Ar biotite	49.1±1.0	Cooling age	Kargarabafghi et al. (2012)
Chapedony subvolcanic dyke	Ar-Ar hornblende	45.4±0.4	Cooling age	Kargarabafghi et al. (2012)
Chapedony subvolcanic dyke	Ar-Ar K-feldspar	44.0±0.4	Cooling age	Kargarabafghi et al. (2012)
Chapedony hornblende granodiorite	Ar-Ar biotite	40.0±1.2	Cooling age	Kargarabafghi et al. (2012)
Chapedony biotite gneiss	Ar-Ar biotite	41.9±0.8	Cooling age	Kargarabafghi et al. (2012)
Chapedony biotite gneiss	Ar-Ar hornblende	47.6±1.0	Cooling age	Kargarabafghi et al. (2015)
Chapedony mylonitic granodiorite	Ar-Ar biotite	43.1±1.8	Cooling age	Kargarabafghi et al. (2015)
Chapedony mylonitic granodiorite	Ar-Ar hornblende	45.3±1.2	Cooling age	Kargarabafghi et al. (2015)
Chapedony mylonitic granodiorite	Ar-Ar K-feldspar	42.7±1.6	Cooling age	Kargarabafghi et al. (2015)
Kuh-e-Neybaz mylonite	Ar-Ar white mica	43.1±0.8	Shear zone activity	Kargarabafghi et al. (2015)
Kuh-e-Neybaz augen gneiss	Ar-Ar biotite	41.3±1.4	Cooling after metamorphism	Kargarabafghi et al. (2015)
Kuh-e-Neybaz gneiss	Ar-Ar biotite	46.7±1.6	Cooling after metamorphism	Kargarabafghi et al. (2015)
Khoshoumi granite	Ar-Ar K-feldspar	44.4±0.6	Cooling after intrusion	Kargarabafghi et al. (2015)
Central domain				
NW Kuh-e-Chamgou metarhyolite	K-Ar whole rock	170 ±11	Cooling after metamorphism	Aistov et al. (1984)
Chamgo granodiorite	U-Pb zircon TIMS	213.5±0.5	Intrusion	Ramezani and Tucker (2003)
Esmail-Abad granite	U-Pb zircon TIMS	218.0±3.2	Intrusion	Ramezani and Tucker (2003)
Esmail-Abad ophiolitic melange	Ar-Ar hornblende	187.6±1.8	Intrusion	Bagheri and Stampfli (2008)
Posht-e-Badam silicate marble	Ar-Ar white mica	220.5±0.5	Cooling after metamorphism	Kargarabafghi et al. (2012)
Posht-e-Badam silicate marble	Ar-Ar white mica	45.5±0.6	Cooling after metamorphism	Kargarabafghi et al. (2012)
Posht-e-Badam retrogressed gneiss	Ar-Ar white mica	55.4±0.6	Cooling after metamorphism	Kargarabafghi et al. (2012)
Posht-e-Badam granodiorite	Ar-Ar hornblende	219.2±2.4	Metamorphism	Kargarabafghi et al. (2012)
Posht-e-Badam quartz phyllonite	Ar-Ar white mica	180.9±1.4	Cooling after metamorphism	Kargarabafghi et al. (2012)
Posht-e-Badam complex phyllite	Ar-Ar white mica	93.6±1.6	Cooling through ca. 400 °C	Kargarabafghi et al. (2015)
Eastern domain				
Tashk formation volcanoclastic tuff	U-Pb zircon TIMS	627±19	Younger detrital zircon	Ramezani and Tucker (2003)
Zarigan leucogranite (S)	U-Pb zircon TIMS	529±16	Intrusion	Ramezani and Tucker (2003)
Zarigan leucogranite (S)	U-Pb zircon TIMS	784±69	Intrusion	Ramezani and Tucker (2003)
Douzak-Darreh leucogranite (S)	U-Pb zircon TIMS	525.7±1.0	Intrusion	Ramezani and Tucker (2003)
Rhyolite of Cambrian volcanosedimentary unit	U-Pb zircon TIMS	528.2±0.8	Lava extrusion	Ramezani and Tucker (2003)
Dacite-porphry of Cambrian volcanosedimentary unit (S)	U-Pb zircon TIMS	527.9±1.0	Lava extrusion	Ramezani and Tucker (2003)
Polo Mt. granodiorite	U-Pb zircon TIMS	530±21	Intrusion	Ramezani and Tucker (2003)
Ariz Mt. Granodiorite	U-Pb zircon TIMS	533±1	Intrusion	Ramezani and Tucker (2003)
Boneh Shurow granitic gneiss (Saghand)	U-Pb zircon TIMS	544±7	Intrusion	Ramezani and Tucker (2003)
Boneh Shurow garnet-amphibolite	U-Pb zircon TIMS	547.6±2.0	Metamorphism	Ramezani and Tucker (2003)
Boneh Shurow quartz-diorite	U-Pb zircon TIMS	547.0±2.5	Intrusion	Ramezani and Tucker (2003)
Boneh Shurow granitic gneiss	U-Pb zircon TIMS	544.4±6.7	Intrusion	Verdel et al. (2007)
Boneh Shurow schist	Ar-Ar biotite	149.75±0.5	Cooling after metamorphism	Verdel et al. (2007)
Boneh Shurow metapelitic schist	Ar-Ar biotite	156.9±0.6	Cooling after metamorphism	Verdel et al. (2007)
Boneh Shurow	Ar-Ar biotite	159.6±0.6	Cooling after metamorphism	Verdel et al. (2007)
Tashk formation	Ar-Ar biotite	218.3±0.5	Cooling after metamorphism	Verdel et al. (2007)
Tashk formation graywacke	Ar-Ar biotite	295.4±1.1	Cooling after metamorphism	Verdel et al. (2007)
Tashk formation	Ar-Ar biotite	281.3±1.2	Cooling after metamorphism	Verdel et al. (2007)
Boneh Shurow Reato-Liassic phyllite	Ar-Ar white mica	140.8±0.6	Low grade metamorphism	Kargarabafghi et al. (2012)
Boneh Shurow granodiorite	Ar-Ar biotite	120.1±2.4	Cooling age	Kargarabafghi et al. (2012)
Boneh Shurow granodiorite	Ar-Ar hornblende	355.5±5.2	Minimum cooling age	Kargarabafghi et al. (2012)
Boneh Shurow garnet-amphibolite	Ar-Ar hornblende	175.8±2.0	Cooling after metamorphism	Masoodi et al. (2013)
Boneh Shurow granitic gneiss	Ar-Ar white mica	165.5±1.0	Deformation related cooling age	Masoodi et al. (2013)
Boneh Shurow garnet-amphibolite	Ar-Ar biotite	149.0±1.0	Cooling after metamorphism	Masoodi et al. (2013)
Boneh Shurow meta-quartz-diorite	Ar-Ar K-feldspar	115.9±2.6	Cooling age	Masoodi et al. (2013)
Boneh Shurow granitic gneiss	Ar-Ar K-feldspar	129.1±0.8	Cooling age	Masoodi et al. (2013)
Posht-e-Sorkh mylonitic schist	Ar-Ar white mica	168.6±1.2	Shear zone cooling age	Masoodi et al. (2013)
Posht-e-Sorkh mylonitic schist	Ar-Ar white mica	167.7±1.0	Shear zone cooling age	Masoodi et al. (2013)
Posht-e-Sorkh mylonitic biotite gneiss	Ar-Ar biotite	230.4±1.8	Shear zone cooling age	Masoodi et al. (2013)
Posht-e-Sorkh mylonitic garnet-biotite gneiss	Ar-Ar biotite	168.1±1.6	Shear zone cooling age	Masoodi et al. (2013)
Posht-e-Sorkh mylonitic schist	Ar-Ar biotite	182.9±1.6	Shear zone cooling age	Masoodi et al. (2013)
Posht-e-Sorkh mylonitic biotite gneiss	Ar-Ar biotite	197.2±1.6	Shear zone cooling age	Masoodi et al. (2013)
Posht-e-Sorkh mylonitic biotite gneiss	Ar-Ar K-feldspar	191.1±1.2	Shear zone cooling age	Masoodi et al. (2013)
Posht-e-Sorkh mylonitic biotite gneiss	Ar-Ar K-feldspar	186.1±1.6	Shear zone cooling age	Masoodi et al. (2013)
Posht-e-Sorkh mylonitic garnet-biotite gneiss	Ar-Ar K-feldspar	167.9±1.4	Shear zone cooling age	Masoodi et al. (2013)

1441 Table 1

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Zone	Samples	Lithology	Mineral													GPS coordinates		Grt (core-rim)					Ph			Bt		St					
			Grt	St	Bt	Ph	Amp	Ep	Clid	And	Ky	Sil	Or	Ttn	Pmp	Chl	Cb	Ap	LONGITUDE	LATITUDE	Alm	Prp	Sps	Grns	XMg	X _{Mg}	Si	pfu	X _{Mg}	Ti	pfu	X _{Mg}	
Jandaq	JA17b	Micaschist	x	x	x	x												54.21025	33.82339	83-81	11-12	1-2	5	0.11-0.12	0.45-0.57	2.97-3.07	0.36-0.39	0.08-0.11	0.07-0.09				
	JA26	Micaschist				x			x	x							54.57128	34.00672						0.27-0.35	3.08-3.13								
	JA28b	Micaschist	x	x	x	x									x		54.56811	33.96711	79-84	8-	8-1	7-5	0.09-0.11	0.37-0.56	3.04-3.10	0.41-0.49	0.03-0.17	0.09-0.12					
Chapedony	CH1823	Gneiss			x						x	x					55.09092	32.52186									0.42-0.55	0.05-0.19					
	CH1842	Micaschist			x							x					55.10728	32.50481									0.44-0.50	0.16-0.21					
Posht-e-Badam	PB1828a	Micaschist	x	x	x	x				x	x						55.37764	32.95903	65-81	7-10	23-5	4-6	0.09-0.10	0.39-0.49	2.95-3.05	0.41-0.48	0.09-0.13	0.16-0.19					
	PB1828b	Micaschist	x	x	x	x				x							55.37764	32.95903	81-83	9-11	5-2	5-6	0.11-0.12	0.47-0.54	2.97-3.04	0.47-0.49	0.07-0.13	0.17-0.19					
	PB1936c	Micaschist	x	x	x	x				x							55.37833	32.95467	73-82	7-10	16-2	5-6	0.09-0.11	0.37-0.46	2.98-3.03	0.43-0.47	0.10-0.13	0.17-0.19					
	PB1938	Micaschist	x	x	x	x						x			x		55.37286	32.96125	83-84	10	2-1	5	0.11	0.38-0.47	2.97-3.03	0.35-0.43	0.09-0.18	0.15-0.17					
Boneh-Shurow N	BS1837	Gneiss	x					x				x					55.59769	33.04028	-	-	-	-	-				0.43-0.46	0.08-0.18					
	BS1947	Gneiss	x	x		x	x						x		x		55.59883	33.03994	55-53	12-15	6-1	31-33	0.16-0.22						0.44-0.49	0.12-0.16			
	BS1952e	Gneiss	x	x		x	x						x		x		55.60722	33.04192	-	-	-	-	-				0.46-0.53	0.05-0.15					
	BS1952g	Gneiss	x	x		x	x				x		x		x		55.60722	33.04192	57-54	3-8	0-1	41-37	0.04-0.10						0.46-0.53	0.05-0.15			
Boneh-Shurow C	BS1824c	Micaschist	x	x	x					x	x	x					55.41244	32.60558	72-85	12-17	16-2	2-3	0.13-0.18	0.51-0.70	3.05-3.15	0.41-0.62	0.13-0.28						
	BS1924	Micaschist	x	x	x					x	x	x	x				55.41292	32.60611	69-71	11-14	16-11	3-4	0.14-0.16	0.54-0.62	3.02-3.07	0.39-0.43	0.11-0.21						
	BS1925	Micaschist	x	x	x				x	x							55.41403	32.60664	76-79	18-17	4-3	1-2	0.20-0.18	0.44-0.66	3.01-3.11	0.41-0.55	0.12-0.17						
	BS1930	Amphibolite	x	x		x							x				55.40803	32.60503	57-62	6-7	10-5	28-26	0.10-0.11						0.37-0.46	0.13-0.20			
Boneh-Shurow S	BS1931	Micaschist	x	x	x												55.35733	32.34019	70-74	14-15	7-3	9-8	0.17-0.18	0.28-0.34	3.01-3.05	0.37-0.56	0.03-0.13						
	BS1933b	Micaschist	x	x	x												55.36078	32.33297	65-68	12-13	15-12	7-9	0.16	0.31-0.34	3.02-3.06	0.50-0.58	0.12-0.15						
	BS1817d	Amphibolite	x	x	x	x							x				55.35856	32.38917	69-70	17	3-2	10-12	0.19-0.20						0.31-0.48	0.08-0.11			
Tashk	TK1802a	Schiste	x	x	x								x				55.51464	32.26028	72-83	5-9	15-0	7-8	0.06-0.10	0.40-0.57	2.96-3.04	0.40-0.42	0.04-0.08						
	TK1802b	Carbonate		x							x	x		x			55.51464	32.26028									0.97-0.98	0.02-0.05					
	TK1802c	Amphibolite		x									x				55.51464	32.26028									0.57-0.59	0.12-0.19					
Sarkuh	SK1806a	Amphibolite		x													55.55126	32.21813									0.60-0.61	0.18-0.21					
	SK1806b	Micaschist	x	x	x					x	x	x			x		55.55126	32.21813	76-77	12-13	5-3	6-8	0.13-0.15	0.50-0.53	3.08-3.11	0.41-0.51	0.13-0.20						
	SK1808	Micaschist	x	x	x	x											55.56086	32.18825	76-77	9-11	10-7	4-5	0.10-0.13	0.47-0.64	3.07-3.13	0.50-0.59	0.05-0.11	0.14-0.18					
	SK1809	Micaschist	x	x	x	x				x	x	x					55.57389	32.18708	75-83	8-9	16-6	1-3	0.09-0.10	0.36-0.55	3.02-3.11	0.34-0.43	0.06-0.18	0.14-0.16					
	SK1812a	Micaschist	x	x	x												55.59192	32.24531	79-82	12-11	7-5	2	0.13-0.11	0.44-0.51	3.09-3.10	0.40-0.43	0.15-0.21						
	SK1814	Micaschist	x	x	x					x	x	x					55.58650	32.23917	73-77	13-12	12-7	3-4	0.15-0.14	0.46-0.57	3.02-3.08	0.42-0.46	0.10-0.20						
	SK1903b	Micaschist	x	x											x		55.53892	32.24044	69-75	14-9	11-2	5-13	0.11-0.16						0.54-0.55	0.09-0.10			
	SK1908b	Micaschist	x	x	x	x											55.55831	32.19300	79-82	6-11	8-0	6-8	0.09-0.12	0.45-0.61	2.99-3.13	0.42-0.56	0.05-0.11	0.16-0.20					
	SK1910	Micaschist		x											x	x	55.56392	32.19297									0.40-0.60	0.07-0.27					
	SK1911	Micaschist	x	x	x	x											55.56519	32.19225	76-79	10-13	6-1	8-10	0.11-0.14	0.45-0.60	3.05-3.10	0.53-0.58	0.05-0.11	0.18-0.21					
	SK1912b	Micaschist	x	x	x	x											55.56553	32.19028	71-79	6-13	18-0	5-9	0.08-0.14	0.50-0.61	3.01-3.13	0.51-0.61	0.06-0.11	0.17-0.22					
	SK1913a	Amphibolite	x	x		x	x							x	x	x	55.59917	32.25169	61-54	9-7	17-5	14-33	0.13-0.11						0.45-0.55	0.06-0.21			
	SK1915g	Micaschist	x	x	x						x	x					55.59092	32.24481	80-83	12-10	7-5	1-2	0.13-0.11	0.39-0.48	2.99-3.11	0.26-0.42	0.09-0.22						
	SK1916b	Vein	x	x	x									x			55.58825	32.24294	76-73	8-7	13-17	4-3	0.09-0.08	0.47-0.50	3.05-3.09								
SK1919	Micaschist	x	x	x												55.57292	32.23033	82-84	14-10	3-2	2-4	0.10-0.14	-	-				0.27-0.31	0.11-0.19				
SK1923	Micaschist	x	x	x						x	x	x				55.56042	32.22797	67-74	9-10	21-11	3-6	0.13-0.12	0.51-0.57	3.01-3.08	0.39-0.42	0.09-0.21							

1443 Table 2

1444 Table 3

Zone	Samples	biotite ages $\pm 2\sigma$ (Ma)	MSWD	Ti T°C	std	white mica ages $\pm 2\sigma$ (Ma)	MSWD	Si pheng	std	Micas textural relationships
Jandaq	JA17b	202.0 \pm 11.5	2.6	307	31	382.3 \pm 86.0	2	3.03	0.08	Ph in pressure zones St, Bt replacing Ph
	JA26					220.3 \pm 41.8	1.9	3.12	0.02	Syn Clid
	JA28b	147.9 \pm 5.7	2.3	390	87	198.2 \pm 16.4	2.4	3.08	0.02	Ph in pressure zones Grt, Bt replacing Ph and Grt
Chapedony	CH1823	42.5 \pm 2.0	2.6	554	18					Growth with foliation
	CH1842	42.7 \pm 1.8	1.7	555	19					Retrograde
Posht-e-Badam	PB1828a	160.7 \pm 2.0	2.9	404	39	227.7 \pm 42.3	1.9	2.98	0.02	Bt and Ph syn rim Grt
	PB1828b	159.2 \pm 1.9	2.6	353	47	131.8 \pm 126.8	1.1	3.01	0.04	Bt and Ph syn rim Grt
	PB1936c	159.4 \pm 1.5	2.3	426	27	214.2 \pm 30.1	2.1	3.01	0.02	Bt and Ph syn rim Grt
	PB1938	160.5 \pm 1.5	2.4	435	26					Bt and Ph post Grt
Boneh-Shurow N	BS1947	141.4 \pm 2.3	1.1	467	43					Retrograde
	BS1952e	151.3 \pm 2.8	2.7	485	14					Retrograde
	BS1952g	134.7 \pm 6.2	2.4	435	56					Retrograde
Boneh-Shurow C	BS1824c	158.0 \pm 1.4	2.7	538	30	154.8 \pm 32.9	1	3.12	0.02	Composing the matrix, prograde or peak ?
	BS1824c	155.5 \pm 3.0	1.5	542	33					Inclusion in K-Feldspar
	BS1924	160.7 \pm 2.4	1.2	506	44	198.6 \pm 33.2	2.1	3.05	0.01	Composing the matrix, partly retrograde
	BS1925	149.6 \pm 1.7	2.5	487	28	170.9 \pm 45.8	2.1	3.07	0.03	Ph in pressure zone And, Bt in pressure zone Grt
	BS1930	149.3 \pm 7.9	2.8	539	28					Retrograde
Boneh-Shurow S	BS1931	150.1 \pm 4.0	2.4	424	40	233.5 \pm 21.7	1.3	3.03	0.01	Bt in pressure zone Grt, Ph pre Grt
	BS1933b	159.0 \pm 1.7	2	515	14					Bt post Grt, Ph pre Grt
	BS1817d	189.0 \pm 25.4	1.2	254	41					Retrograde
Tashk	TK1802a	153.3 \pm 3.0	2.6							In pressure zones Grt
	TK1802b	180.5 \pm 1.4	2	460	36					Inherited ?
Sarkuh	TK1802c	168.3 \pm 2.9	1.3	526	33					Syn Amp et Qz
	SK1806a	172.5 \pm 2.4	2.2	617	8					Post/syn Amp
	SK1806b	169.8 \pm 2.1	1.7	549	19					In pressure zone Grt
	SK1808	152.7 \pm 4.0	2.7	336	37					In pressure zone Grt et St
	SK1809	161.3 \pm 3.5	2.2	466	33					In pseudomorph And and Grt and pre/syn Sil
	SK1812a	159.9 \pm 2.9	2.7	544	20	455 \pm 210	2.1	3.08	0.04	Composing the matrix, prograde or peak ?
	SK1814	168.6 \pm 1.7	2.5	540	34	155.7 \pm 23.3	1.9	3.04	0.02	Bt and Ph post Ky and pre/syn Sil
	SK1903b	104.0 \pm 23.5	3	389	24					Retrograde
	SK1908b	169.7 \pm 7.0	1.2	332	68					In pressure zone Grt et St
	SK1910	169.7 \pm 3.2	2.2	549	31					Composing the matrix, prograde or peak ?
	SK1911	122.4 \pm 11.2	2.5	319	53	125.0 \pm 36.7	2.5	3.07	0.02	Bt pressure zone Grt et St, Ph matrix
SK1912b	161.5 \pm 7.9	1.2	286	49					Pressure zone St	
SK1913a	160.2 \pm 4.9	2.8	544	22					Syn Amp	
SK1913a	139.6 \pm 42.9	1.5	519	24					Retrograde	
SK1915g	172.8 \pm 2.4	1.9	534	22	270.2 \pm 151.7	2.2	3.04	0.04	Bt and Ph post Ky and pre/syn Sil	
SK1916b					197.1 \pm 15.0	2.4	3.06	0.01	Vein	
SK1919	154.5 \pm 4.1	2.9	473	36					Post Grt	
SK1923	169.4 \pm 2.6	2	553	19					In pseudomorph And and syn Sil	

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1449 **Highlights**

1450 - Coupled in-situ Rb/Sr and U/Pb dating allows to constrain metamorphic terranes evolution.

1451 - Iran was affected by a Jurassic collisional metamorphism related to the Paleotethys closure.

1452 - Metamorphic terranes are exhumed from ~170 Ma onward in a context of upper plate
1453 extension.

1454

1455 This work is original, has not been published previously and is not under consideration for
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1458

1459 Authors contributions are as follow:

1460

1461 Thomas Gyomlai : Conceptualization, Supervision, Investigation, Methodology, Formal analysis,
1462 Writing - Original Draft, Visualization.

1463 Philippe Agard : Conceptualization, Supervision, Writing - Review & Editing, Resources, Funding
1464 acquisition.

1465 Laurent Jolivet : Supervision, Writing - Review & Editing, Resources, Funding acquisition.

1466 Tiphaine Larvet : Writing - Review & Editing, Formal analysis, Investigation.

1467 Guillaume Bonnet : Writing - Review & Editing, Validation.

1468 Jafar Omrani : Resources, Funding acquisition.

1469 Kyle Larson : Writing - Review & Editing, Formal analysis, Resources.

1470 Benoit Caron : Methodology, Resources, Validation.

1471 Julie Noël : Formal analysis, Validation.

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1474 **Declaration of interests**

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1476 The authors declare that they have no known competing financial interests or personal
1477 relationships that could have appeared to influence the work reported in this paper.

1478

1479 The authors declare the following financial interests/personal relationships which may be considered as
1480 potential competing interests:

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