Deciphering the Jurassic–Cretaceous evolution of the Hamadan metamorphic complex during Neotethys subduction, western Iran

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Abstract

The Hamadan high-grade metapelites in the northwestern part of the Sanandaj–Sirjan zone, Iran, show a polymetamorphic evolution with relics of a garnet-bearing metamorphic mineral assemblage (M1), a contact

24 metamorphic overprint (M2) related to the emplacement of the Middle to Late Jurassic Alvand composite pluton and

25 a Buchan-type regional metamorphic event (M3) marked by ^{40}Ar/^{39}Ar ages in the 80–70 Ma range that is associated

26 with penetrative ductile deformation producing a foliation and a thermal overprint onto the M2 assemblages. The

27 M1 event is exclusively preserved as small garnet grains and mineral inclusions contained therein, incorporated into

28 M2-stage cordierite porphyroblasts. Distinct metamorphic zones are developed over a region of ~600 km^2, which are

29 partly correlated with distance to the composite pluton: zones (1) cordierite + K-feldspar hornfels, and (2) andalusite

30 ± cordierite hornfels surround the Alvand composite pluton at a distance of up to 5 km. These two zones are clearly

31 related to M2 metamorphism associated with pluton emplacement. Zones (3) staurolite schist, (4) andalusite schist,

32 and (5) sillimanite schist are found outside of the contact aureole and are considered to be the result of regional M3

33 metamorphism in the eastern part distant to the Alvand composite pluton. Conventional thermobarometry shows that

34 temperatures in the area vary between ~560 to 660 °C for zones 1 and 2 and ~490 to 690 °C for zones 3–5. Phase

35 equilibria modelling in the MnNCKFMASHT system indicates two distinct isobaric prograde paths at low pressures,

36 at ~2.7 kbar for zones 1 and 2 and slightly higher pressures of around 3.5 to 5.5 kbar for zones 3 to 5. U–Th–Pb

37 monazite geochronology revealed overlapping ages of 168 ± 11 Ma and 149 ± 19 Ma in the hornfels (1 and 2) and

38 schistose (3–5) zones, respectively. These ages are similar to the intrusion age of the Alvand composite pluton

39 (153.3 ± 2.7 Ma to 166.5 ± 1.8 Ma) and are interpreted to reflect heating due to the emplacement of the composite

40 pluton (M2 contact metamorphic event). However, ^{40}Ar/^{39}Ar dating of white mica and amphibole yielded plateau

41 ages ranging from 80 to 69 Ma over the entire transect. The formation of schistosity in zones 3–5 postdates the

42 intrusion and is thus related to M3 metamorphism. The white mica fabric indicates formation of the foliation during

43 M3 garnet growth, which is followed by local retrogression of garnet to chlorite during exhumation. Consequently,

44 the ^{40}Ar/^{39}Ar white mica and amphibole ages likely indicate reheating during M3 to more than ca. 500 ± 25 °C

45 (argon retention temperature in amphibole). These data establish the occurrence of a Cretaceous, Buchan-style

46 regional metamorphic event that had not been firmly identified before. Subsequent Late Cretaceous exhumation of

47 the Hamadan complex with its high-grade metapelites is due to extension along the Tafrijan–Mangavi–Kandelan

48 fault, which represents a major ductile low-angle normal fault. Metamorphic temperatures coupled with mineral

49 ages from this and published work suggest a fast stage of cooling with a rate of ~6 °C/Ma during exhumation after

50 M3 metamorphism.

51

54

Introduction

56 Metamorphic belts generally occur in central sectors of collisional orogenic belts and provide evidence for a number

57 of superimposed tectonic processes like burial by subduction, subsequent heating and finally exhumation during

58 collisional orogeny, with interposed subduction- and/or collision-related plutons that may cause additional contact

59 metamorphic overprints. Therefore, the recognition of distinct metamorphic stages in such metamorphic terrains is

60 crucial for the interpretation of their geotectonic history. To achieve this aim, combining information about time

61 (ages and/or duration) to specific rock-forming processes and their physical conditions (e.g. temperature and

62 pressure) is fundamental (Mottram et al. 2014; Engi et al. 2017).

63 The Zagros orogen (Iran) is a segment of the Alpine–Himalayan collision zone that resulted from closure of the

64 Neotethys ocean and comprises an internal belt of collision-related metamorphic rocks (Stöcklin, 1968; Berberian

65 and King, 1981; Mohajjel and Fergusson, 2014; Shakerardakani et al., 2015; Hassandzadeh and Wernicke, 2016).

66 Within the Zagros orogen, the Sanandaj–Sirjan zone represents the metamorphic core of the Arabia-Eurasia collision

67 and contains a record of its magmatic and metamorphic evolution (Hassandzadeh and Wernicke, 2016). Among

68 debated issues is whether and to which way the metamorphic rocks are related to the collision process, and how the

69 plate tectonic imprint affected the formation of different metamorphic grades in this belt, especially where low

70 pressure metamorphism within a contact aureole and Buchan-type metamorphism interfere with each other.

71 The Sanandaj–Sirjan zone plays a crucial role in deciphering the evolution of the Arabia-Eurasia collision, but its

72 metamorphic history is still the least investigated aspect (Hassandzadeh and Wernicke, 2016). The Hamadan

73 metamorphic complex in the northwestern part of the Zagros collision zone is among the volumetrically most

74 significant exposures of metamorphic rocks, but ages of protoliths and metamorphic events are in dispute. The

75 Hamadan metamorphic complex exposes Middle to Late Jurassic granite and gabbro intrusions known as the Alvand

76 composite pluton, dated at 172–153 Ma based on U–Pb zircon and monazite ages (Shahbazi et al. 2010; Mahmoudi

et al. 2011; Sepahi et al. 2018), and is characterized by associated contact metamorphism. However, in some distance to the Alvand composite pluton, rocks of the Hamadan metamorphic complex are schistose and even a Late Jurassic granite gneiss, the Aliabad-e Damaq pluton, is schistose. Previous K–Ar mineral geochronology yielded, beside a single Early Cretaceous amphibole age, Late Cretaceous to Paleocene white mica and biotite ages for the metamorphic rocks as well as for the Alvand composite pluton itself (Valizadeh and Cantagrel 1975; Baharifar et al. 2004). The available observations and data do not allow a clear distinction whether the metamorphism in the Hamadan area is solely related to the emplacement of the plutonic rocks or whether several distinct metamorphic episodes occurred with distinct ages and pressure-temperature evolutions. These different interpretations imply either a long-lasting cooling history after intrusion of magmatic rocks or a more complex, polyphase metamorphic history.

Given the rather speculative metamorphic age assignments, the Hamadan metamorphic complex warrants a more detailed petrochronological study with the combined application of several petrological and geochronological methods to unravel superimposed metamorphic events. Consequently, the present study is aimed to clarify the petrology, P–T conditions and the age(s) of metamorphism in the Hamadan metamorphic complex. Our approach describes the field relationships, petrography and mineral chemical characteristics of high-grade metapelitic rocks from the Hamadan complex, as well as P–T results obtained by conventional thermobarometry and equilibrium assemblage diagrams using the THERIAK/DOMINO software (de Capitani and Brown, 1987; de Capitani and Petrakakis, 2010). We also present results of chemical U–Th–Pb monazite ages and a comprehensive data set of laser-probe $^{40}Ar/^{39}Ar$ muscovite and amphibole ages from different metamorphic zones allowing a better understanding of the petrochronological evolution of the Hamadan complex for which previously only K–Ar ages and fission track data existed (Baharifar et al., 2004; François et al., 2014).

In this study, we closely consider the relationships among distinct metamorphic events and the emplacement of Jurassic subduction-related plutons and develop a model with three stages of metamorphism ranging from pre-pluton emplacement (M1) via Middle/Late Jurassic contact metamorphism (M2) associated with emplacement of the composite Alvand pluton to a nearly pervasive Late Cretaceous thermal overprint (M3) partly associated with ductile deformation. The metamorphic stage M3 was not considered before for this region. We offer, therefore, a

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103 new model for the tectono-metamorphic evolution of the Hamadan complex where metamorphic stages M2 and M3

104 relate to distinct, separate stages during closure of the Neotethys.

105

Geological setting

107 Formation of the Zagros orogen results from the long-lasting convergence between Eurasia and Gondwana-derived

108 fragments during closure of the Neotethys Ocean (e.g. Agard et al., 2005; Agard et al., 2011). The Sanandaj–Sirjan

109 zone is in most parts a polymetamorphic zone within the interior part of the Zagros orogen that contains abundant

110 Mesozoic calc-alkaline plutons, which caused contact metamorphism of surrounding rocks (e.g., Stöcklin, 1968;

111 Berberian and King, 1981; Mohajjel et al., 2003; Agard et al., 2005; Agard et al., 2011; Chiu et al., 2013; François et

112 al., 2014; Shakerardakani et al., 2015; Hassanzadeh and Wernicke, 2016; Sepahi et al., 2018). The distribution of

113 metamorphic rocks is discontinuous and different metamorphic evolutions have been observed in different areas of

114 the Sanandaj–Sirjan zone (Berberian et al. 1981; Hassanzadeh and Wernicke, 2016).

115 Previous work (e.g. Shakerardakani et al., 2015; Hassanzadeh and Wernicke, 2016) interpreted the metamorphic

116 rocks within the Sanandaj–Sirjan zone to exhibit up to four stages of metamorphic evolution (Berberian et al., 1981)

117 and therefore representing a polymetamorphic terrain. In the central part of the Sanandaj–Sirjan zone in the area of

118 Shahrekord W of Esfahan, an early high-pressure metamorphic stage (max. 25 kbar at T = 600 °C; with $^{40}Ar/^{39}Ar$

119 phengite ages of ca. 184 Ma) likely relates to the accretion of a continental basement segment to Central Iran within

120 the Neotethyan subduction system (Davoudian et al., 2016). Another metamorphic stage is related to contact

121 metamorphism caused by Middle/Late Cretaceous plutons (Baharifar et al., 2004; Sepahi et al., 2004, 2019;

122 Mohajjel et al., 2006).

123 The Hamadan high-grade metamorphic rocks ("lower unit" in the following text) with an extension of ~600 km^2

124 occur mainly southeast- and southward to the Alvand composite pluton, which is one of the largest Jurassic

125 composite plutons in the Sanandaj–Sirjan zone (Fig. 1). The Hamadan complex contains potentially Triassic and

126 rare fossil-bearing Jurassic slates (e.g. bivalve *Posidonia alpina* (Dehghan, 1947)), phyllites and micaschists. The

127 slates are separated by a major normal fault (Tafrijan–Mangavi–Kandelan fault) from the amphibolite facies

5

128   metamorphic rocks next to the Alvand composite pluton (Fig. 2) as already recognized by Agard et al. (2005). This

129   unit contains various metapelitic schists, rare amphibolite lenses and some pegmatites. To the east of Hamadan, the

130   folded very low- to low-grade metamorphic rocks (but not the high-grade metamorphic rocks) are non-conformably

131   overlain by a Lower Cretaceous succession with a thin basal conglomerate overlain by andesitic lava and spilites,

132   sandstones with conglomerates, marls and Orbitolina limestones of Aptian–Albian age (ca. 125–100.5 Ma; Cohen et

133   al., 2013). Further on, we will refer to this structural unit as the "upper unit". The folded Aptian–Albian succession

134   (Fig. 2a, b) is folded, cut by an angular unconformity and then overlain by continental Oligocene–Miocene

135   sedimentary rocks. These structural relationships of low-grade metamorphic rocks and non-conformably overlying

136   Aptian–Albian strata imply that in the low-grade part metamorphism ceased not later than Early Cretaceous (at ca.

137   125 Ma). Similar Lower Cretaceous cover successions were reported from several areas of the Sanandaj–Sirjan zone

138   (Hassanzadeh and Wernicke, 2016 and references therein).

139   The Alvand pluton is composite consisting of gabbroic rocks, granites, including mylonitized varieties, and

140   leucocratic granitoids that occupy an area of ~800 km^2 (Shahbazi et al., 2010; Figs. 1, 2). A summary of existing age

141   data from the Alvand composite pluton is compiled in Table 1 and are shown on Fig. 2c. U–Pb zircon dating reveals

142   that the Alvand composite pluton formed during the Jurassic between 171.1 ± 1.2 and 153.3 ± 2.7 Ma with the main

143   intrusive phases between 166.5 and 163.9 Ma (Shahbazi et al., 2010; Mahmoudi et al., 2011; Chiu et al., 2013;

144   Zhang et al., 2018). U–Pb monazite ages of pegmatites give ages between 172.20 ± 0.91 Ma and 162.23 ± 0.77 Ma

145   (Sepahi et al., 2018). The schistose Aliabad-e Damaq granitic gneiss yields an upper intercept age of 164.2 ± 5.4 Ma

146   (Shahbazi et al., 2010) and is included within schistose metapelites. Valizadeh and Cantagrel (1975) reported biotite

147   and muscovite ages ranging from 68 Ma to 104 Ma (Rb–Sr) and 68–89 Ma (K–Ar) from the Alvand composite

148   pluton indicating that the pluton was affected by a Cretaceous metamorphic event or cooled through ca. 425 °C, the

149   Ar retention temperature of white mica.

150   The lower unit of the Hamadan metamorphic complex comprises variably deformed and metamorphosed pelitic,

151   semipelitic and psammitic metasediments as well as quartzite, amphibolite and hornblende garben schists and some

152   calc-silicate rocks (Baharifar et al., 2004; Sepahi et al., 2004). Fergusson et al. (2016) dated detrital zircons of an

153   andalusite-garnet-biotite schist from the Hamadan Phyllite (lower unit) from a locality 35 km southeast of Hamadan

154   and reported ^{206}Pb/^{238}U ages of 270 ± 12 Ma for the youngest zircons. Consequently, the maximum depositional age

155 of the Hamadan Phyllite is Middle Permian, which is consistent with the presence of Triassic and Lower Jurassic

156 fossils. A summary of existing age data from the Hamadan metamorphic complex is given in Table 1 and shown on

157 Fig. 2c. In the high grade lower unit of the Hamadan metamorphic complex, the timing of metamorphism is based

158 on K–Ar dating and gave three ages of 114.6 ± 0.3 Ma and 82.3 ± 2.4 Ma from amphibole in an amphibolite from

159 the staurolite zone, and 76.7 ± 1.9 Ma from muscovite of an andalusite-kyanite vein (Baharifar et al., 2004).

160 Baharifar et al. (2004) proposed that the Hamadan complex was influenced by tectono-metamorphic events during

161 the Jurassic to the mid-Cretaceous and by granitic intrusions in Late Cretaceous time although the Cretaceous

162 granites were not verified by later studies. However, the subsequent determination of Jurassic (171−153 Ma)

163 crystallization ages for the Alvand pluton and the large age range of almost 40 million years require a re-assessment

164 of significance of these ages for the metamorphic evolution.

165 Andalusite ± cordierite and cordierite + K-feldspar hornfelses are exposed in close proximity to the Alvand

166 composite pluton, whereas a large region of different, largely medium-grade metamorphic schists is exposed

167 adjacent to the eastern/south-eastern part of the composite pluton (Sepahi et al. 2004; Agard et al. 2005). Field

168 observation and textural evidence reveal that the hornfelses close to the Alvand granite grade into rocks with a

169 schistose fabric further to the east. The flat-lying schistosity is coeval with recrystallization of a garnet + staurolite +

170 biotite assemblage and refolded by an eastward-dipping crenulation schistosity (Agard et al., 2005). It is associated

171 with a predominant NNW–SSE trending lineation (Agard et al., 2005), which turns to nearly E–W orientation in the

172 region, where the schistose Aliabad-e Damaq granitic gneiss is exposed which probably represents a major apophyse

173 of the Alvand composite pluton.

174

Previous P–T estimates

176 Sepahi et al. (2004) defined two possible clockwise P–T paths that suggest peak metamorphic conditions of up to

177 650 °C at 4 kbar in the high grade metamorphic lower unit of the Hamadan region. Based on Raman spectroscopy of

178 graphitic material, Agard et al. (2005) demonstrated temperatures ranging from 387 ± 9 to 567 ± 33 °C. Mohajjel et

179 al. (2006) proposed two main metamorphic events in alumosilicate-bearing schists accompanied by deformation

180 episodes based on mineral assemblages and thermobarometry. The results indicate that P–T conditions during the

181 second phase of prograde metamorphism reached a maximum of about 630 °C at a pressure of 4 kbar. The first

182 event presented by Mohajjel et al. (2006) pertains to oblique subduction of the Neotethys oceanic crust in Late

183 Jurassic to Early Cretaceous times, and the second stage is the Late Cretaceous event synchronous with the main

184 magmatism that increased the thermal gradient while sillimanite formed. However, no supporting age data for the

185 proposed Late Cretaceous metamorphic event are available. Sepahi et al. (2009) and Saki et al. (2012) calculated

186 peak metamorphic conditions of ~650–750 °C and ~2–4 kbar by means of conventional thermobarometry methods

187 in migmatites occurring adjacent to the Alvand composite pluton. Maximum P–T conditions of 700–750 °C at 5–6

188 kbars were recently presented for the high-grade metamorphic mafic rocks from the metamorphic aureole around the

189 pluton (Sepahi et al. 2013). Based on amphibole composition and hornblende-plagioclase thermometry,

190 thermobarometric calculations of Miri et al. (2016) provided P–T conditions of 492 to 508 °C and 4.3 to 4.9 kbar for

191 the cores of these minerals and 552 to 573 °C and 5.5 to 5.9 kbar for the outer rims, respectively. Considering the

192 methodological uncertainties in pressure (± 1.2 kbar) and temperature (±37 °C) determinations, Miri et al. (2016)

193 inferred an isobaric temperature increase related to thermal perturbation associated with Late Jurassic magmatism.

194 Recently, Sepahi et al. (2019) described migmatites along the northeastern part of the contact between the Alvand

195 composite pluton and the Hamadan metamorphic complex. These authors determined P–T conditions of 640–700 °C

196 and 3–5 kbar for migmatization, which was dated at ca. 170 Ma using the U–Pb zircon technique. These results

197 point to a causal link between the magmatic activity and the migmatization. A summary of existing P–T estimates

198 from the Hamadan complex is given in Table 2 and Fig. 2d.

199

Petrography

201 We collected and prepared thin sections of more than one hundred samples from different metamorphic zones within

202 the Hamadan complex for petrographic studies. The metamorphic units east of the Alvand composite pluton can be

203 texturally divided into two main zones: (i) an approximately 4 to 5 km wide hornfels zone, extending from the

204 contact of the pluton with the first appearance of spots of large cordierite to the last occurrence of cordierite in

205 andalusite-cordierite hornfels; (ii) a schistose zone comprising staurolite-bearing schists. This zone is followed

206 outward up to about 15 km distance from the contact by rocks with abundant porphyroblasts of andalusite and an

207   area marked by sillimanite replacing andalusite. The metapelitic rocks are locally interlayered with minor

208   amphibolite. Based on the studied samples, we distinguish three stages of metamorphism from textural and

209   deformational points of view:

210   (1) Relics of a first metamorphic stage (M1) are preserved as garnet inclusions in cordierite, which occur only within

211   the hornfels zone. These garnets also include fine-grained feldspar and chlorite of a lower grade metamorphic stage,

212   which are interpreted as early remnants of the M1 metamorphism (Fig. 3b).

213   (2) The contact metamorphic (hornfels) zones (M2) are not only characterized by typical porphyroblasts like

214   cordierite but also contain boudinaged pegmatite dykes surrounded by a weak foliation in the pelitic country rocks

215   (Fig. 3a, f, g) indicating a gradual transition between the contact metamorphic zones (M2) to schistose zones

216   dominated by M3 mineral assemblages.

217   (3) In the east, the schistose zones include also the Aliabad-e Damaq granite gneiss, which has a similar age as the

218   Alvand composite pluton. The area is characterized by a penetrative foliation and a NNW–SSE to N–S trending

219   mineral and stretching lineation, which is folded by later deformation stages (Agard et al., 2005; Mohajjel et al.,

220   2006). In the southeast, the stretching lineation is ca. E–W subparallel to the trend of the schistose Aliabad-e Damaq

221   granite gneiss.

222   In the following, we use these textural relationships to distinguish between mineral assemblages related to the M1,

223   M2 and M3 metamorphic stages. Based on a suitable mineralogy, textures and lack of alteration, we examined

224   several samples from the whole metamorphic area for thermobarometry. A list of selected samples investigated in

225   detail is presented in Table 3. Four samples were selected for chemical U–Th–Pb monazite dating and twelve

226   samples from all units of the Hamadan metamorphic complex were selected for ^{40}Ar/^{39}Ar dating of muscovite and

227   amphibole. Their geographical locations and mineral assemblages are given in Table 3 and Fig. 2a.

228   All mineral assemblages contain monazite and zircon, and most contain garnet, muscovite, biotite, plagioclase and

229   quartz. Graphite and ilmenite are present in low amounts in all samples. Chlorite and fine-grained white mica

230   ("sericite") are mostly present as secondary retrograde phases. In the following, we describe the zones from the

231   contact outward. The reactions are discussed in the text according to the numbering scheme in Table 6 and Fig 10.

232 **Hornfels zones**

233 *Cordierite + K-feldspar zone*

234 The cordierite + K-feldspar zone forms only a narrow band close to the Alvand composite pluton (Fig. 2). The

235 pelitic hornfels samples were taken within <1 km from the Alvand granite and display the mineral assemblage Crd +

236 Kfs + Bt + Grt + Pl + Qtz (mineral abbreviations are according to Kretz, 1983). Close to the contact, the samples

237 contain large well-preserved cordierite, rare small garnet and no andalusite or relics thereof. Pegmatite dykes

238 boudinaged by brittle faults are surrounded by a weakly developed foliation. The rocks develop prominent "spots"

239 (up to ~2 cm in size; Fig. 3a–c), corresponding to anhedral and roughly ovoid, inclusion-rich cordierite

240 poikiloblasts, which are set in a fine-grained granoblastic matrix of feldspar, biotite, muscovite, quartz and ilmenite.

241 The development of K-feldspar that occurs in anhedral, cryptoperthitic grains associated with quartz and the paucity

242 of muscovite at the highest grade close to the contact may be representative of leucosome development by partial

243 melting.

244 The peak metamorphic conditions of 620 ± 40 °C and 2.5 ± 0.5 kbar obtained in sample H101 (for details, see

245 below) are just below the wet granite solidus, but minor dehydration melting could have occurred by muscovite (Fig.

246 3d) dehydration and associated release of a hydrous fluid:

247 $Ms + Qtz = Kfs + And + H_2O$ (reaction #3)
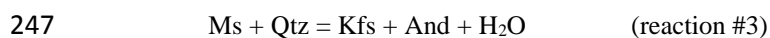
248 Garnet occurs as euhedral and rarely fractured and cracked crystals, within the matrix and in some cases as

249 inclusions in cordierite (Fig. 3b). The garnet inclusions are interpreted to be of an early-stage origin (M1), predating

250 the growth of cordierite, assuming that cordierite growth is due to the heat from the intrusion. The euhedral garnets

251 from the matrix are in equilibrium with cordierite and thus considered to be part of the hornfels stage (M2).

252 *Andalusite ± cordierite zone*

253 Following the narrow cordierite + K-feldspar zone, a 1–5 km wide And ± Crd zone extends outwards from the

254 contact with the pluton (Fig. 2). In this zone, metapelitic rocks mainly have a hornfelsic texture, a dark grayish color

255 (Fig. 3e), and typically contain the assemblage And ± Crd ± Kfs ± Grt ± Ms + Bt + Pl + Qtz. Andalusite generally

256 forms well-preserved, idioblastic, inclusion-poor crystals that occur in the matrix (Fig. 3e–g). Cordierite occurs as

257 ellipsoidal, inclusion-filled anhedral poikiloblasts with biotite and a low amount of muscovite in the matrix. A weak

258 foliation is observed in some samples (Fig. 3f, g). The zone is also marked by the existence of K-feldspar in some,

259 but not all rocks. In the outer parts of the andalusite $\pm$ cordierite zone, cordierite is gradually replaced by fine-

260 grained aggregates rich in chlorite and quartz while chiastolitic andalusite is present as large cm-sized idiomorphic

261 blasts with accumulations of graphite/opaque phases as inclusions and along its rectilinear grain edges (Fig. 3g). As

262 a whole, with increasing proximity to the composite pluton, the proportion of andalusite decreases and cordierite as

263 well as K-feldspar crystals become larger and more abundant. A possible reaction that involves And as reactant and

264 is forming Crd + Kfs is:

265     $Bt + And + Qtz = Crd + Kfs + H_2O$ (reaction #4)
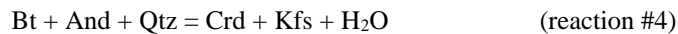
266 The garnets are present mostly as euhedral, partly fractured crystals and are considered to belong to the M2

267 metamorphic assemblage (Fig. 3h). In contrast M1 garnet is only found in the hornfels zone and occurs as relics

268 enclosed by cordierite.

269 **Schistose zones**

270 *Staurolite zone*

271 Further to the east, the staurolite zone follows the hornfels zone (Fig. 2). The dominant mineral assemblage in this

272 zone contains Grt + St + Bt + Ms + Pl + Qtz. The rocks are predominantly fine-grained, well-foliated schists (Fig.

273 4a, b), which display a typical S-C shear fabric.

274 Staurolite displays a pre- to syn-deformational fabric and is seen as euhedral to subhedral porphyroblasts up to ~3

275 cm length, or as small (~1 mm) blocky crystals generally replaced by chlorite at the rim (Fig. 4c, d). Some rare

276 staurolite grains seem to exhibit two stages of growth with an earlier growth phase defined by trails of inclusions,

277 followed by an inclusion-poor rim with ~2 mm thickness (Fig. 4d). The staurolite growth postdates partly the garnet

278 growth since small idioblastic garnet grains occur as inclusions within them (Fig. 4c). The inclusions in the

279 staurolite are parallel to foliation and indicate syn-deformative growth (Fig. 4d).

280   Garnet is nearly ubiquitous and locally occurs as perfect idiomorphic crystal up to ~6 mm in diameter. Texturally,

281   garnet mostly has a pre- to syn-deformative origin in all schistose metamorphic zones (Fig. 4e). Garnet grains

282   contain quartz, graphite and ilmenite inclusions and typically appear euhedral to subhedral in shape, although

283   corners and edges of crystals show typical retrograde features such as rounding and partial replacement by biotite

284   and chlorite (Fig. 4e). Biotite, muscovite and chlorite form tiny flakes, in preferred orientation defining the foliation.

285   Some biotite grains have been replaced by secondary chlorite (Fig. 4e).

286   Hornblende garbenschist and amphibolite interlayers within metapelites of the staurolite zone mainly contain

287   acicular and euhedral amphibole, and garnet in a fine-grained matrix of quartz, plagioclase, K-feldspar and the

288   accessories titanite, epidote and clinozoisite.

289   *Andalusite zone*

290   A large area of metapelitic schists adjacent to the staurolite zone displays the assemblage And ± St + Grt + Bt + Ms

291   + Chl + Pl + Qtz (Fig. 2). The andalusite porphyroblasts are up to ~10 cm in length (Fig. 4g) and typically display a

292   chiastolitic cruciform pattern of inclusions and accumulations of graphite and opaque phases along rectilinear grain

293   edges (Fig. 4h). Andalusite is slightly altered to fine- to coarse-grained "sericite", however, generally well-

294   preserved, idioblastic, inclusion-poor crystals are observed, too.

295   In samples, in which andalusite is present in larger amounts, muscovite reacted with staurolite to form andalusite

296   according to the following reaction (Fig. 4i, j):

297          $Ms + St + Qtz = And + Bt + H_2O$ (reaction #6)

298   Textural evidence indicates that staurolite growth preceded andalusite growth in some cases, while andalusite is

299   randomly pre- and syn-deformative. Some staurolite appears as poikiloblasts with sigmoidal inclusion trails

300   reflecting syn-deformative growth as well (Mohajjel et al., 2006; Fig. 4k).

301   Garnet frequently occurs as up to ~3 mm large porphyroblasts and few of them occur as euhedral to subhedral

302   inclusions in andalusite (Fig. 4h) suggesting that garnet predates the growth of andalusite while in other places

303 garnet is surrounded by the mica matrix. (Fig. 4l). Some garnets developed poikilitic grains with inclusions of

304 quartz, graphite and ilmenite (Fig. 4m).

305 *Sillimanite zone*

306 The onset of the sillimanite zone (Fig. 2), is marked by the appearance of fibrolite and rare fine-grained prismatic

307 sillimanite replacing andalusite. The diagnostic assemblage of this zone is Sil ± And + Grt + Bt + Ms + Qtz + Pl. In

308 this zone, no signs of coexisting K-feldspar and sillimanite or the development of migmatization is observed.

309 The sillimanite typically occurs in mats within the matrix and as fine needles intergrown with andalusite or coarse

310 muscovite (Fig. 4n, o), postdating the growth of andalusite (Agard et al. 2005). Garnet grains are commonly

311 scattered throughout the assemblage. Locally, sillimanite, staurolite and tourmaline were observed in shear bands in

312 this zone (Agard et al. 2005).

313 *Schistose granite gneiss of Aliabad-e Damaq*

314 The Jurassic granite gneiss of Aliabad-e Damaq shows a schistose fabric composed of biotite and subordinate white

315 mica, deformed K-feldspar (mostly microcline) and well recrystallized plagioclase. Quartz shows grain boundary

316 migration recrystallization (Fig. 4f), which typically forms close to amphibolite facies grade deformation (at ca.

317 500 °C according to Stipp et al. 2002).

318

Analytical techniques

320 The analytical techniques are described in the Appendix. These include: whole rock geochemical analyses of

321 metapelitic rocks, mineral chemical compositions determined by scanning electron microscope (EDX and WDX

322 spectrometer), Th–U–Pb monazite dating by electron microprobe and ^{40}Ar/^{39}Ar dating of amphibole and white mica.

323

Mineral chemistry

325   For garnet profiles, the largest grains were used since we confirmed they represent cuts through the center of the

326   garnet grain. Nearly all schist samples contain compositionally zoned *garnet* with dominantly almandine (62–82 %),

327   spessartine (2–23 %), minor pyrope (4–15 %) and subordinate grossular (1–10 %) (Table 5; Fig. 5). Only garnet

328   from the hornfels zone displays a homogeneous compositional zoning pattern indicating growth at constant PT

329   conditions (Fig. 5a). Garnet compositional profiles of schistose zones generally depict a rising trend in X_{alm} and a

330   decrease in X_{sps} from core-to-rim. X_{grs} and X_{prp} display mostly smoothly varying compositional trends (Fig. 5b-d).

331   These patterns are interpreted as typical trends for prograde garnet growth in metapelites.

332   In samples from both the hornfels and schistose zones, the minerals within the matrix show only a small

333   compositional variation or are chemically homogeneous. *Cordierite* is characterized by a minor variation in X_{Mg}

334   ranging from 0.43–0.50 (Table 5). *Biotite* is characterized by Al^{VI}, and Ti contents in the range of 0.27–0.43 and

335   0.06–0.21 a.p.f.u in the hornfels zones and in the range of 0.38–0.67 and 0.07–0.18 a.p.f.u in the schistose zones,

336   respectively. X_{Mg} of biotite varies between 0.35 and 0.51 in the hornfels zones and 0.38 and 0.55 in the schistose

337   zones. The F content of biotite is around 0.5 wt. % for nearly all selected samples (Table 5).

338   *Plagioclase* in the hornfels zones ranges in composition from almost pure albite $Ab_{91}An_7Or_0$ to andesine

339   $Ab_{69}An_{31}Or_0$ (Table 5). In schistose zone rocks, *plagioclase* varies from albite $Ab_{87}An_{13}Or_0$ to andesine

340   $Ab_{56}An_{43}Or_0$. *Alkali feldspar* is orthoclase $Ab_{10}Or_{89}An_0$ in the hornfels zones (Table 5). The chemical variations

341   reported here are not related to zoning but differ from sample to sample.

342   *Muscovite* has Si-contents ranging from 3.01 to 3.15 a.p.f.u in the hornfels zones, while these contents vary from

343   3.01 to 3.23 a.p.f.u. in the schistose zones. The K/(K+Na) ratio of muscovite is 0.94 to 0.96 in the hornfels zones

344   and 0.81 to 0.93 in the schistose zones (Table 5).

345   Finally, the chemistry of *staurolite* does not show systematic trends in the variation of X_{Fe} (0.77–0.89) and ZnO

346   contents (up to 0.71 wt. %) (Table 5). Although two generations of staurolite can be distinguished in the staurolite

347   zone, no significant chemical difference can be found.

348

349   **P–T determinations**

350   P–T conditions of metamorphism were obtained using conventional geothermobarometry as well as equilibrium

351   assemblage diagrams.

Conventional thermobarometry

353   For conventional thermobarometry the PET1.1 software (Dachs, 1998, 2004) was used. Appropriate matrix mineral

354   compositions, which are in equilibrium with each other and occur also in "domains" near garnet, were selected for

355   calculation of pressure and temperature conditions (Table 3).

356   *Hornfels unit:* The Grt-Bt (Holdaway, 2000) and Grt-Crd geothermometers coupled with the GASP (Koziol, 1989)

357   and Grt-Pl-Ms-Bt (Hoisch, 1990; Wu, 2015) geobarometers were applied for assemblages from the cordierite + K-

358   feldspar and andalusite ± cordierite zones to calculate peak metamorphic temperatures of 620 ± 40 °C and 590 ±

359   30 °C, respectively (Table 3; Figs. 2d, 6a). In both samples, the pressure remains constant at around ~3.0 ± 1.0 kbar.

360   The average temperature increases towards the pluton.

361   *Schistose unit:* Mineral assemblages belonging to the staurolite zone yield peak P–T conditions of 590 ± 50 °C and

362   5.5 ± 1.5 kbar based on Grt-Bt (Holdaway, 2000) and Grt-St (Perchuk, 1969) thermometers and the Grt-Pl-Ms-Bt

363   barometer (Table 3; Figs. 2d, 6b). The Grt-Bt and Grt-St thermometers in combination with the GASP barometer

364   was applied for assemblages from the andalusite zone, yielding values around 600 ± 15 °C and 5.5 ± 0.7 kbar (Table

365   3; Figs. 2d, 6b). Finally, the Grt-Bt thermometer and the GASP barometer applied for the sillimanite zone shows

366   values of 670 ± 20 °C and 5.5 ± 0.5 kbar (Table 3; Figs. 2d, 6b). P–T results appear to exhibit a small decrease in

367   temperature from cordierite bearing rocks to andalusite bearing rocks in the hornfels unit and a small increase from

368   staurolite bearing rocks towards the andalusite and sillimanite bearing rocks in schistose zones (Fig. 7). A minor

369   increase in pressure from the contact outwards is also observed (Fig. 6).

Assemblage stability diagrams

371   Equilibrium assemblage diagrams were constructed in the chemical system MnO–Na_2O–CaO–K_2O–FeO–MgO–

372   Al_2O_3−SiO_2−H_2O–TiO_2 using the THERIAK/DOMINO software package from 03.01.2012 (de Capitani and Brown,

373   1987; de Capitani and Petrakakis, 2010) in conjunction with the thermodynamic database of Holland and Powell

374   (1998), updated to dataset 5.5 in 2003 that incorporates the Al_2SiO_5 triple point of Pattison (1992). Activity models

15

375   used in all calculations are those used in Tinkham and Ghent (2005) with the following exceptions: 1) margarite was

376   not considered as a component in white mica, and 2) the ternary feldspar model of Holland and Powell (1998), using

377   a molecular mixing model and asymmetric formalism, was used instead of separate plagioclase and pure orthoclase

378   or sanidine. Excess H_2O was assumed, and a melt phase was not considered in the modelling because only

379   subsolidus processes are discussed. Although graphite is observed in samples, the assumption of a pure H_2O fluid

380   phase does not result in large discrepancies in the position of phase boundaries (Connolly and Cesare, 1993; Pattison

381   et al., 2002; Pattison, 2006) and the presence of graphite probably resulted in a mixed C-O-H fluid (Connolly and

382   Cesare, 1993) during metamorphism.

383   Calculations were performed with samples containing representative mineral assemblages and bulk rock

384   compositions of the defined five zones (Table 4). The initial constraints on the peak P–T conditions recorded by

385   each sample are imposed by the highlighted field within the relevant assemblage stability diagram, which contains

386   the observed peak mineral assemblage identified in thin sections (Figs. 8, 9). Subsequently, two isobaric prograde

387   paths are displayed that pass through the mineral-assemblage sequences from hornfels and schistose areas and cross

388   the reaction boundaries on the equilibrium phase diagram, where minerals are expected to be either found or lost

389   from the mineral assemblage.

390   Figure 8 shows calculated assemblage stability diagrams for two metapelitic samples from the andalusite ±

391   cordierite and the cordierite + K-feldspar zones. Sample H101 from the cordierite + K-feldspar zone was taken

392   exactly at the contact (Fig. 2a). The observed mineral assemblage Crd + Grt + Bt + Kfs + Pl + Qtz + Ilm defines a

393   P–T field in the calculations diagram of 570 to 670 °C and <4 kbar. (Fig. 8b). Sample H48 from the andalusite +

394   cordierite zone is located ~5 km southeast of the contact (Fig. 2a) and comprises the stable mineral assemblage And

395   + Grt + Bt + Ms + Qtz + Pl + Ilm. The calculated phase stability field corresponds to lower pressure amphibolite

396   facies with T = 550 to 630 °C and P = 2.8 to 3.8 kbar (Fig. 8a). Texturally, the garnet inclusion in cordierite belongs

397   to an early stage of metamorphism (M1) in the hornfels zones. A temperature increase towards the intrusion is

398   clearly seen in these two samples, consistent with the higher average temperature in the cordierite + K-feldspar zone

399   derived from conventional geothermobarometry.

16

400 Mineral isograds in hornfelsic rocks can therefore be mapped around the intrusion and range from an outermost

401 occurrence of And-in up to Crd+Kfs-in and And-out close to the aureole contact. Pattison and Tracy (1991)

402 subdivided prograde sequences of the most common metapelitic assemblages into several facies series. The hornfels

403 isobaric path comply with the facies series 1/2a bathograds, that intersect the reactions 1, 2, 3 and 4 (Table 6; Fig.

404 10).

405 Sample H44 belongs to the staurolite zone and is located about 6 km east of the contact, just in the central part of the

406 zone (Fig. 2a). The mineral assemblage St + Grt + Bt + Ms + Qtz + Pl + Ilm is seen within a stability field extending

407 from 560 to 640 °C and 3.5 to 7.5 kbar in Figure 9a. Sample H30 east of the contact was chosen from the andalusite

408 zone (Fig. 2a). The calculated equilibrium assemblage diagram shows a P–T range of 540 to 640 °C and 2.5–4.0

409 kbar for the observed mineral assemblage And + Grt + Bt + Ms + Qtz + Pl + Ilm (Fig. 9b). Staurolite is considered

410 as a metastable phase. Several common reactions, such as St → Grt + Bt + ALS, or St + Chl → ALS + Bt, involve

411 staurolite as reactant and produce an aluminosilicate (ALS) polymorph. Pattison and Spear (2018) noted that the

412 width of zones of coexisting staurolite and andalusite in Buchan-type sequences is much wider than predicted by

413 equilibrium phase diagrams, which they attributed to disequilibrium processes related to sluggish dissolution of

414 staurolite and the lack of a thermodynamic driving force for the conversion of staurolite to andalusite. Although the

415 two samples from the staurolite (H30) and andalusite (H44) zones have different mineral assemblages, calculated

416 equilibrium temperatures are similar. In sample H44 the assemblage is stable over a wider P range and we conclude

417 that low P values, similar to H30, are likely. Sample H10 from the sillimanite zone has the largest distance to the

418 contact (Fig. 2a). The mineral assemblage Sil + Grt + Bt + Ms + Qtz + Pl + Ilm is stable over a wide P–T range of

419 540 to 770 °C and 3 to 8.5 kbar (Fig. 9c). Since andalusite is present in the sample, we argue that the presence of

420 sillimanite cannot significantly exceed the And-Sil reaction limiting the pressure to <4 kbar. The resulting T is about

421 600 to 650 °C

422 Observed spacing of staurolite, andalusite and sillimanite isograds in the schistose part persuade us to propose an

423 isobaric P–T path at a pressure less than 1.0 kbar higher than the hornfels bathograd. The schistose isobaric path

424 intersects reactions 5, 6 and 7 which conform with the facies series 2b/3 bathograd of Pattison and Tracy (1991)

425 (Table 6; Fig. 10). The onset of the Sil-in and St-out isograd might however place this isobaric path to higher

426 pressure through the intersection of the following reaction:

17

427         $Ms + St + Qtz = Sil + Bt + H_2O$
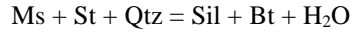
428   At pressures of 3–4 kbar, the St-out isograd appears at temperatures between 550 and 585 °C (Fig. 10).

429

430   **Geochronological results**

431   **U–Pb monazite data**

432   Monazites were analysed in four metapelitic samples (see Fig. 2a, c), one from the hornfels units (sample H101 from

433   the crd + K-fsp zone) and three from the schistose units (sample H54 from the st zone, sample H31 from the and

434   zone and sample H11 from the sil zone). The monazites have maximum grain sizes of 30 μm and elliptical to

435   rounded shapes. In all cases, the monazite grains occur in the fine-grained matrix, which is mainly composed of

436   muscovite, and the grains are in textural equilibrium with the matrix. The small sizes of monazite allowed only 2 or

437   3 spot analyses in each grain. A large variation of monazite ThO_2* from 2–18 wt. % in sample H101 marks an

438   isochron at 168 ± 11 Ma. The ThO_2* values show a broad cluster at around 6 wt. % (Table S1; Fig. 11a). Some

439   monazite analyses with ThO_2* <4 wt. % plot considerably below the isochron. This population matches data from

440   sample H54, where all monazites poorly define an errorchron of 47 ± 51 Ma, mostly at low ThO_2*. The samples

441   H11 with 162 ± 13 Ma and H31 with 149 ± 19 Ma display a second small monazite populations at around 64–70 Ma

442   along a poorly defined errorchron with PbO contents, which are just above the detection limit (Fig. 11a, b). The

443   small population of Late Cretaceous to Paleocene monazites is also obvious in a diagram age vs. Y_2O_3 (Fig. 11b)

444   and show a chemically slightly different environment of monazite growth. The younger monazite ages of 64–70 Ma

445   and 47 ± 51 Ma are similar in age, within error, to our Ar–Ar mineral ages (see below).

446   There is a compositional trend from the main Middle Jurassic monazite population with 1–3 wt. % Y_2O_3 toward the

447   younger population at around 1 wt. % Y_2O_3. Similar compositional trends from higher to lower ThO_2, UO_2 and CaO

448   with decreasing ages are observed (not shown). As both samples from the hornfels units (H11 and H101) and sample

449   H31 display similar isochrones at R^2 values between 0.85 and 0.93, the monazite ages show a broad unimodal

450   maximum at 160–180 Ma in the summarized histogram view (Fig. 11c). The minor Paleocene population appears

451 also in the histogram and in the related averages. This age potentially implies a younger, second, Late Cretaceous to

452 Paleogene thermal event.

$^{40}Ar/^{39}Ar$ muscovite data

454 White mica concentrates from eight samples were prepared to reveal potential age differences between hornfels and

455 schistose zones (see Fig. 2a, c for sample locations). For measurements, 7–10 grains per sample were used.

456 Analytical data are shown in Table S2. In the following, we describe the samples from various mineral assemblage

457 zones in a transect from west to east. Nearly all concentrates yield minor extraneous argon in the first one to three

458 steps and a plateau for major portions of the remaining steps of the experiment. An overview on significant

459 properties of all investigated samples is given in Figure 12.

460 Samples from the hornfels zones (H56 and H101) yield plateau ages of 80.21 ± 0.44 Ma and 76.61 ± 0.42 Ma,

461 respectively. The inverse isochron age of sample H56 is 81.8 ± 3.7 Ma with an initial $^{40}Ar/^{36}Ar$ ratio of 326 ± 69.

462 The inverse isochron age of sample H101 is 73.4 ± 2.6 Ma with an initial $^{40}Ar/^{36}Ar$ ratio of 435 ± 51 indicating the

463 presence of some excess argon significantly above the atmospheric $^{40}Ar/^{36}Ar$ ratio of 295.5. We interpret, therefore,

464 the plateau age of 76.61 ± 0.42 Ma as geologically significant, which excludes the first two steps with excess argon.

465 Both samples from the staurolite zone (H58 and H66) yield a weak staircase pattern with ages of 62.28 ± 7.00 Ma

466 and 60.18 ± 6.93 Ma of the first steps indicating some Argon loss. Sample H58 yields a plateau age of 72.65 ± 0.49

467 Ma, whereas for sample H66, the plateau age is 69.17 ± 0.83 Ma and the total gas age 71.33 ± 0.41 Ma. The inverse

468 isochron ages are 73.0 ± 1.3 Ma (H58) and 69.15 ± 0.97 (H66) with initial $^{40}Ar/^{36}Ar$ ratios of 231 ± 25 and 265 ± 37,

469 respectively. We consider the plateau ages of 72.65 ± 0.49 Ma (sample H58) and 69.17 ± 0.83 Ma (H66) as

470 geologically significant, whereas the first step ages indicate a later overprint and Ar loss.

471 Two samples from the andalusite zone (H32 and H68) give plateau ages of 73.50 ± 0.38 Ma and 73.66 ± 0.45 Ma.

472 The isochron ages are 72.32 ± 0.94 Ma (H32) and 70.6 ± 3.0 Ma (H68) with initial $^{40}Ar/^{36}Ar$ ratios of 322 ± 45 and

473 522 ± 240, respectively, the later potentially including some extraneous argon. Finally, samples H11 and H67 are

474 from the sillimanite zone. Sample H11 exhibits a plateau age of 70.59 ± 0.24 Ma, H67 a plateau age of 70.84 ± 0.26

475 Ma. The inverse isochron ages are 68.6± 3.8 Ma (H11) and 69.7 ± 9.0 Ma (H67) with initial $^{40}Ar/^{36}Ar$ ratios of 412

476 ± 90 and 436 ± 190, respectively.

477 In most cases, the inverse isochron ages (including all steps) are close to the plateau ages. Thus, we suggest that the

478 ages from all four investigated zones are geologically significant and represent the ages of cooling through the argon

479 retention in white mica, which is at 425 ± 25 °C in regional metamorphic areas (Harrison et al., 2009). In summary,

480 the two plateau ages of the hornfels zones are significantly older (80.21 and 76.61 Ma), whereas the ages of all

481 schistose zones are vary between 70.59 and 73.66 Ma within a relatively narrow age range.

482 **$^{40}Ar/^{39}Ar$ amphibole data**

483 We selected four amphibolite samples of the staurolite zone (H61, H114, H119, H120), which are not affected by

484 retrogression, for $^{40}Ar/^{39}Ar$ dating. Optically uniform amphibole grains free from macroscopically visible inclusions

485 were selected for dating. Ca. 10 grains were then finally used for the experiment and because of the relatively low

486 potassium content, ca. 7–9 steps were measured until fusion. All four samples show U-shaped argon release patterns

487 (Fig. 13) and exhibit excess argon in the first 2 to 3 steps, followed by ca. three relatively large steps with low

488 apparent ages and then an increase in ^{40}Ar content in the final steps. The $^{37}Ar_{Ca}/^{39}Ar_K$ ratio is between 5.8 to and 12

489 meaning relatively low potassium contents in all four amphibole concentrates. These central three steps yield similar

490 ages between minimum 62.6 ± 2.4 Ma (sample H120) and maximum 72.9 ± 1.1 Ma (sample H61). The total gas

491 ages (corresponding approx. K–Ar ages) range between 79.0 ± 1.1 Ma and 93.7 ± 1.4 Ma, reflect extraneous argon

492 incorporated in the sample and are geologically meaningless. Sample H61 also shows the highest K content with a

493 $^{37}Ar_{Ca}/^{39}Ar_K$ ratio between 5.8 and 6.7 and yields a weighted mean age of 73.9 ± 1.1 Ma. We consider this age as

494 geologically significant and to either reflect the maximum age of the third stage of metamorphism M3 or the age of

495 cooling through ca. 550 °C. The age is also very close to the age of white mica from nearby samples of the schistose

496 units. The other three amphibole samples are in full support of this Cretaceous age although the youngest age is

497 slightly younger than white mica ages from other samples.

498

499 **Discussion**

500 The metamorphic framework of the Hamadan complex presented in this study constrains three stages, M1–M3, in

501 metapelitic rocks which show a progressive change in mineral assemblages and mineral chemistry. Together with

502 age dating results and the regional geological framework, the results argue for a reinterpretation of the geological

503 and tectonic history of this specific part of the Sanandaj–Sirjan zone. We first discuss details of the metamorphic P–

504 T conditions, then the age dating results and finally we propose, based on our results, a new tectonic model for the

505 Hamadan metamorphic complex.

P–T conditions of the three metamorphic stages

507 Besides the overall stability field of inferred peak metamorphic assemblages, the P–T conditions can further be

508 constrained using the chemical composition of minerals.

509 The mineral assemblage related to M1 metamorphism is constrained exclusively by garnet grains enclosed in

510 cordierite of M2. The prograde metamorphic evolution of M1 can only be vaguely constrained by feldspar and

511 chlorite inclusions in garnet. The compositional zoning of M1 garnet in hornfels samples H101 is negligible and

512 thus no P–T variation from core to rim is envisaged. The measured composition is characterized by a high amount of

513 X_{alm} ranging from 0.78–0.81, X_{prp} of 0.09–0.12 and X_{sps} = 0.07–0.08 and X_{grs} = 0.02–0.03. Using the isopleths of the

514 grossular and pyrope components, we constrain the P–T conditions of M1 garnet growth to ~575–725 °C and ~3–5.5

515 kbar. However, the chlorite inclusions in this garnet generation favour P–T conditions closer to the lower range with

516 ~575 °C and ~3–5 kbar, just outside of the chlorite stability field (Fig. 14).

517 The results of conventional thermobarometry show a systematic increase in temperature toward the contact from

518 andalusite bearing rocks (zone 2) to cordierite bearing rocks (zone 1) in the hornfels area (M2 metamorphism, Fig.

519 7). There is no systematic trend in temperatures across the schistose region (zones 3–5, M3 metamorphism) toward

520 the contact. However, application of the GASP and GBMP barometers indicate slightly lower pressures for the

521 hornfels zones (Fig. 6). The equilibrium assemblage diagram approach allowed us to further constrain the observed

522 mineral assemblages to gradients in pressure and temperature within the Hamadan area. Although zones 1–2 and

523 zones 3–5 display typical low pressure, high temperature mineral assemblages we were able to identify a systematic

524 difference in pressure and thus propose two different prograde evolutions for hornfels and schistose rocks.

525 Consequently, we argue that the successive appearance and disappearance of andalusite and cordierite, contact

526 metamorphism (M2), is related to increasing temperature toward the Alvand composite pluton at isobaric conditions

527 of ~2.7 kbar (Fig. 10). Due to the large areal extension and the absence of plutonic rocks in the foliated metamorphic

528 rocks and irregular distribution of metamorphic field-gradients towards the Alvand composite pluton, contact

529 metamorphism is excluded to be responsible for the low P mineral assemblages in zones 3–5. We propose a

530 different, Buchan-style metamorphic event (M3) for the staurolite–andalusite–sillimanite sequence at slightly higher

531 pressures of 3.5 kbar (Fig. 10). The metamorphic overprint developed together with the deformation along the

532 eastern side of the composite pluton.

Age of M2 and M3 metamorphism

534 Although lithological and mineralogical features have been widely described, some controversies persist regarding

535 the metamorphic age of the Hamadan complex. The Alvand composite pluton formed during Middle Jurassic times,

536 as established by several U–Pb zircon ages. (Shahbazi et al. 2010; Mahmoudi et al. 2011; Sepahi et al. 2018)

537 However, for the Hamadan metamorphic complex, both the K–Ar muscovite and amphibole ages from the literature

538 (Baharifar et al. 2004; Valizadeh and Cantagrel 1975) and the new white mica and amphibole $^{40}Ar/^{39}Ar$ ages from

539 this study are much younger, Late Cretaceous, for both the hornfels and schistose zones. Consequently,

540 crystallization of the plutons pre-dates cooling after the latest peak metamorphic conditions by about 90 million

541 years.

542 Valizadeh and Cantagrel (1975) reported biotite and muscovite ages ranging from 68 Ma to 104 Ma (Rb–Sr) and

543 68–89 Ma (K–Ar) from the Alvand granitoid complex. Given the more recent U–Pb zircon and monazite ages of

544 153–172 Ma for various lithologies (gabbros, granitoids, pegmatites) of the Alvand pluton (Table 1), this implies

545 that the Alvand pluton was also affected by a Late Cretaceous metamorphic event, which is much younger than the

546 Jurassic age of intrusion. This is consistent with the observation that the schistose Aliabad-e Damaq granite gneiss

547 with an age of 164.2 ± 5.4 Ma belongs to the large Alvand composite pluton and is affected by formation of a

548 metamorphic foliation postdating the intrusion.

549 The monazite age of 168 ± 11 Ma from the cordierite + K-feldspar zone (sample H101) taken at the contact is within

550 error the same as the age of granite intrusion. The monazite age of 149 ± 19 Ma from the andalusite zone (sample

551 H31) overlaps within error with these ages. Experimental work suggests that closure temperatures for Pb in

552   monazite are typically in the range between 900 and 1100 °C, depending on cooling rate and grain size (Cherniak et

553   al., 2004). Hence, volume diffusion of Pb in monazite is very slow and inefficient (McFarlane and Harrison, 2006)

554   and unlikely to cause grain-scale resetting under the metamorphic conditions experienced by the Hamadan

555   metamorphic complex. Therefore, these monazite ages are interpreted to represent the age of the metamorphic event

556   M2 as they are similar, within error, to the U–Pb zircon intrusion ages of the pluton.

557   The ^{40}Ar/^{39}Ar white mica ages of the hornfels are between 80.2 and 76.6 Ma, whereas the white mica ages of all

558   metasedimentary rocks from the schistose zones are within a narrow range between 73.7 and 69.2 Ma (Campanian,

559   Late Cretaceous). Considering the fading of the schistosity towards the Alvand composite pluton and our new

560   ^{40}Ar/^{39}Ar white mica ages between 80 and 70 Ma and amphibole ages between 73 and 63 Ma, we argue that the M3

561   metamorphic event is much younger and occurred in the Late Cretaceous. Although imprecise due to the large

562   uncertainty, the monazite population of 64–70 Ma and 47 ± 51 Ma is similar in age to the Late Cretaceous

563   amphibole and white mica ^{40}Ar/^{39}Ar ages. These ages are similar to few K–Ar white mica ages reported by

564   Baharifar et al. (2004), which range between 68.7 ± 1.6 Ma and 76.7 ± 1.9 Ma. Baharifar et al. (2004) also reported

565   K–Ar amphibole ages of 114.6 ± 4.3 Ma and 82.3 ± 2.4 Ma as well as some K–Ar biotite ages ranging from 67.4 ±

566   1.7 to 59.4 ± 1.5 Ma. The oldest K–Ar amphibole age of 115 ± 4.3 Ma (Baharifar et al., 2004) is likely to have been

567   affected by extraneous argon, the presence of which was demonstrated by our argon release patterns, and is thus

568   considered geologically meaningless. The Late Cretaceous to Paleocene ages of all four amphibole concentrates

569   from the amphibolites (Fig. 13) argue that M3 exceeded 550 °C according to experimental calibration initially

570   proposed by Harrison et al. (1981). These new geochronological data establish that a distinct metamorphic event that

571   reached amphibolite-facies metamorphic conditions affected the Hamadan metamorphic complex in the Late

572   Cretaceous. Because of the similar, overlapping ^{40}Ar/^{39}Ar amphibole and white mica ages, the age of pervasive

573   deformation forming white mica during ductile deformation is considered to have been just after peak conditions of

574   M3. In addition, this Late Cretaceous thermal event affected not only the Hamadan complex but also the seemingly

575   undeformed Alvand composite pluton. Interestingly, a similar monazite age of 61 ± 8 Ma was reported for the

576   Soursat complex in the northwestern Sanandaj–Sirjan zone (Jamshidi Badr et al. 2010) implying a larger area

577   affected by this latest Cretaceous to earliest Paleogene metamorphism.

578   These relationships argue for the following interpretation: The latest regional metamorphism (M3) in the Hamadan

579   complex is younger than metamorphism in the upper unit east of the Tafrijan–Mangavi–Kandelan fault.

580   Unfortunately, there are no geochronological ages constraining the age of metamorphism in the upper unit, but it is

581   significantly lower grade. Agard et al. (2005) demonstrated, based on Raman spectroscopy, a temperature difference

582   of about 100 °C from 387 °C in the upper unit vs. 487 °C in the lower unit across the Tafrijan–Mangavi–Kandelan

583   fault. These authors found ca. NNW–SSE trending stretching lineation with top to the SSE motion, overprinted by

584   ESE-directed extensional shear bands. We therefore speculate that the Tafrijan–Mangavi–Kandelan fault represents

585   a ductile low-angle normal fault, which was active during Late Cretaceous and the cause for the exhumation and

586   cooling of the Hamadan metamorphic complex. Agard et al. (2005) already noted the special significance of this

587   fault representing a break in metamorphic conditions. However, future studies should focus on the age of

588   metamorphism in the upper unit and reveal more details on the kinematic evolution of the Tafrijan–Mangavi–

589   Kandelan fault.

590   **Cooling history of the Hamadan metamorphic complex**

591   Regional cooling after the latest regional metamorphic event (M3) can be inferred from the Ar retention temperature

592   of metamorphic amphibole (550 ± 25 °C) and muscovite (425 ± 25 °C) which occurred just after peak P–T

593   conditions. In Figure 15, we compile all thermochronological data from the region including our new Ar–Ar ages

594   and an apatite (U–Th)/He age (22.1 ± 0.70 Ma: Francois et al., 2014). The cooling path since ca. 80–75 Ma, after the

595   M3 metamorphic event, reached ca. 6 °C/Ma. The cooling through ca. 60°C, the He retention temperature in apatite,

596   occurred virtually contemporaneous with deposition of the Oligocene–Miocene succession at top (Fig. 2a, b). Final

597   cooling through the helium retention temperature in apatite is related to final exhumation and erosional denudation

598   after final closure of the Neotethyan suture (Agard et al., 2011).

599   **Tectonic implications**

600   The new data on P–T conditions and ages allow to propose a new geodynamic model (Fig. 16), which incorporates

601   data from the literature (e.g., Agard et al., 2011; Chiu et al., 2013; Shakerardakani et al., 2015; Hassanzadeh and

602   Wernicke, 2016) and also the M3 metamorphic event defined in this study. The short-lived Middle Jurassic pulse of

603   intrusions including the Alvand composite pluton, which was emplaced between 171.1 ± 1.2 and 153.3 ± 2.7 Ma, is

24

604   the result of Neotethyan subduction (e.g., Shahbazi et al. 2010; Mahmoudi et al. 2011; Sepahi et al. 2018; Fig. 16a).

605   This is the stage of contact metamorphism (M2). Production of calc-alkaline magmas ceased during Late Jurassic

606   and only few Early Cretaceous magmatic rocks formed. We argue that the formation of the schistosity in the

607   Hamadan Phyllite directly pertains to M3 and is the result of ductile deformation, which is related to the initial

608   exhumation of the Hamadan metamorphic complex. This interpretation explains why the schistose zonation fades

609   westwards to the Alvand composite pluton. The most straightforward interpretation is to relate foliation formation to

610   the Tafrijan–Mangavi–Kandelan fault and slow exhumation of the Hamadan metamorphic complex could be

611   explained by Late Cretaceous cooling and exhumation. This would need a major Late Cretaceous event of extension

612   within the Sanandaj–Sirjan zone potentially triggered by southwestward retreat of the subduction zone during

613   rollback and steepening of the Neotethyan ocean (Fig. 16b). Such an event is well known and resulted in the opening

614   of back-arc basins, e.g., Late Cretaceous Nain-Baft back-arc basin (e.g., Agard et al., 2011 and Hassanzadeh and

615   Wernicke, 2016 and references therein). Our new data argue that the main stage of regional metamorphism, M3, in

616   this sector of the Sanandaj–Sirjan zone is of Late Cretaceous age and is related with ductile low-angle normal

617   faulting and extension during opening of mentioned back-arc basins.

618

626

627   **Appendix: Analytical techniques**

628   **Whole rock-element measurements**

629 Whole-rock analyses were performed by using a Bruker Pioneer S4 X-ray fluorescence spectrometer at the Institute

630 of Earth Sciences, Karl-Franzens-University Graz. One gram of sample powder and seven grams of $Li_2B_4O_7$ flux

631 were fused to a glass bead for XRF measurement. About 80 international reference rock powders were used for

632 standardization, the reproducibility is better than 1% for major elements and better than 5% for trace elements.

633 Detection limits are for most elements 20 ppm. The whole rock chemistry of samples is shown in Table 4.

Mineral composition

635 Mineral compositions were determined at the NAWI Graz Geocenter - Institute of Earth Sciences, Karl-Franzens-

636 University Graz with a JEOL 6310 electron scanning microscope equipped with a LINK ISIS energy dispersive

637 spectrometer (EDS) and a MICROSPEC wavelength dispersive system (WDS). The emission current is operated

638 with 15-kV acceleration voltage and 6-nA beam current. The Phi-Rho-Z procedure was used for matrix correction.

639 Natural mineral standards were used for calibration: adular (Si, K and Al, EDS) garnet (Mg and Fe EDS), titanite (Ti

640 and Ca, EDS) on, Mn and Cl (EDS) on rhodonite and atacamite, respectively. Na and F were analyzed by WDS

641 using albite and synthetic F-phlogopite standards. Si, Al, Mg, and Fe were standardized on garnet in the case when

642 garnet is analyzed.

Electron microprobe (EMP) monazite dating

644 EMP dating was performed at Institute of Mineralogy, Technical University Freiberg, Germany and follows the

645 procedure of Schulz (2017). EMP–Th–U–Pb dating is based on the observation that common Pb in monazite

646 (LREE, Th) PO_4 is negligible when compared to radiogenic Pb resulting from the decay of Th and U (Montel et al.,

647 1996). Electron microprobe analysis of the bulk Th, U and Pb concentrations in monazite, at a constant $^{238}U/^{235}U$

648 ratio, allows for the calculation of a chemical model age (CHIME) with a considerable error (Suzuki et al., 1994;

649 Montel et al., 1996; Suzuki and Kato, 2008; Spear et al., 2009). The Mα1 lines of Th and Pb and the Mβ1 lines for U

650 of a PET crystal were selected for monazite analysis. Analytical errors of 2σ at 20 kV acceleration voltage, 100 nA

651 beam current, 5 µm beam diameter and counting times of 320 s (Pb), 80 s (U) and 40 s (Th) on peak have been

652 considered for the calculation of ages. For Pb, the error ranges typically from 0.016–0.024 wt. % for the given dwell

653 time, based on measurement on a reference monazite ("Madmon"; Schulz and Schüssler, 2013). Detection limits

654 (1σ) calculated from counting rates for elements with low concentrations in monazite are in element %: Si 0.004; Ca

26

655 0.005; Y 0.020; Pb 0.006; U 0.016; Sm 0.060. Orthophosphates of the Smithsonian Institution were used as

656 standards for REE analysis (Jarosewich and Boatner, 1991). Calibration of PbO was carried out on a crocoite

657 standard. The U was calibrated on a metal standard. The reference monazite labelled "Madmon" with special

658 ThO_2*–PbO characteristics (Schulz and Schüssler, 2013) was used for calibration and offline re-calibration of ThO_2

659 as well as for the control of data. Interference of YLγ on the PbMα line was corrected by linear extrapolation as

660 proposed by Montel et al. (1996). An interference of ThMγ on UMβ was also corrected. The number of single

661 analyses varies with the grain size of the monazites, e.g. 1–4 analyses in grains of <30 μm in diameter. Monazite

662 chemical ages were first calculated using the methods of Montel et al. (1996). Weighted average ages for monazite

663 populations calculated using Isoplot 3.0 (Ludwig, 2001) are interpreted as the time of closure for the Th–U–Pb

664 system of monazite during growth or recrystallization in the course of metamorphism. Ages were further determined

665 using the ThO_2*–PbO isochron method (CHIME) of Suzuki et al. (1994) and Montel et al. (1996) where ThO_2* is

666 the sum of the measured ThO_2 plus ThO_2 equivalent to the measured UO_2. This is based on the slope of a regression

667 line in ThO_2* vs. PbO coordinates forced through zero. In all analysed samples, the model ages obtained by the two

668 different methods agree exceptionally well.

$^{40}Ar/^{39}Ar$ age-dating

670 $^{40}Ar/^{39}Ar$ analyses largely follow descriptions given in Rieser et al. (2006) and Cao et al. (2017). Preparation of the

671 samples before and after irradiation, $^{40}Ar/^{39}Ar$ analyses, and age calculations were carried out at the ARGONAUT

672 Laboratory of the Department of Geography and Geology at the University Salzburg. The samples were crushed

673 carefully with a hammer. The 200–250 μm white mica and amphibole fractions were hand-picked under the

674 microscope. Mineral concentrates were packed in aluminium-foil and placed in quartz vials. For calculation of the J-

675 values, flux-monitors were placed between each 4–5 unknown samples. The sealed quartz vials were irradiated in

676 the Rez reactor (Prague, Czechia) for 16 hours. Correction factors for interfering isotopes were calculated from 45

677 analyses of two Ca-glass samples and 70 analyses of two pure K-glass samples and are: $^{36}Ar/^{37}Ar$ (Ca) = 0.000225,

678 $^{39}Ar/^{37}Ar$ (Ca) = 0.000614, $^{38}Ar/^{39}Ar$ (Ca) = 0.011700 and $^{40}Ar/^{39}Ar$ (K) = 0.0266. Variations in the neutron flux

679 were monitored using the DRA1 sanidine standard for which a $^{40}Ar/^{39}Ar$ plateau age of 25.03 ± 0.05 Ma was

680 originally reported (Wijbrans et al., 1995). Here we use the revised value of 25.26 ± 0.05 Ma (van Hinsbergen et al.,

681   2008). $^{40}Ar/^{39}Ar$ analyses were carried out using a UHV Ar-extraction line equipped with a combined

682   MERCHANTEK^{TM} UV/IR laser system, and a VG-ISOTECH^{TM} NG3600 mass spectrometer. Stepwise heating

683   analyses of samples were performed using a defocused (~1.5 mm diameter) 25 W CO_2-infrared laser operating in

684   Tem$_{00}$ mode at wavelengths between 10.57 and 10.63 μm. The NG3600 is an 18 cm radius 60° extended geometry

685   instrument, equipped with a bright Nier-type source operated at 4.5 kV. Measurements were performed on an axial

686   electron multiplier in static mode. Peak-jumping and stability of the magnet was controlled by a Hall-probe. For

687   each increment the intensities of ^{36}Ar, ^{37}Ar, ^{38}Ar, ^{39}Ar, and ^{40}Ar were measured, the baseline readings on mass 34.5

688   were subtracted. Intensities of the peaks were back-extrapolated over 16 measured intensities to the time of gas

689   admittance either by a straight line or a curved fit, depending on the intensity and type of the pattern of the evolving

690   gas. Intensities were corrected for system blanks, background, post-irradiation decay of ^{37}Ar, and interfering

691   isotopes. Isotopic ratios, ages and errors for individual steps were calculated following suggestions by McDougall

692   and Harrison (1999) and Scaillet (2000) using decay factors reported by Steiger and Jäger (1977). Definition and

693   calculation of plateau ages was carried out using ISOPLOT/EX (Ludwig, 2003).

694

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847

848   **Figure captions:**

849   Figure 1. Simplified geological map of northern Sanandaj−Sirjan zone modified after Mahmoudi et al. (2011). The
850   Alvand composite pluton and Hamadan metamorphic complex are located in the center of the map.

851

Figure 2. Simplified geological map of the Hamadan metamorphic complex and Alvand composite pluton modified after Eghlimi (2000) and Baharifar et al. (2004). (a) A compilation showing the distribution of metapelitic zones and major faults in the Hamadan metamorphic complex. Sample numbers from east of the composite pluton are shown on the map. (b) Schematic cross section (AB) through the Hamadan region (after Sabzehei et al., 1977). Horizontal scale = vertical scale. (c) map with tagged locality distribution of previous and new geochronological data. (d) map with labelled previous and new P–T estimates. The illustrated path from the contact is in agreement with Fig. 7.

858

Figure 3. Photomicrographs and photographs showing the mineral assemblages and hornfelsic textures of cordierite + K-feldspar (a-d) and andalusite ± cordierite (e-h) zones. (a) The pelitic hornfels with a boudinaged pegmatite. Note a weak foliation in the hornfels and some small fringes around andalusite porhyroblasts parallel to the weak foliation. (b) and (c) Large cordierite poikiloblasts containing small garnet inclusions. (d) The occurrence of muscovite in the high-grade hornfels zone close to the contact aureole. (e-g) Fresh andalusite porphyroblasts is present in a fine-grained matrix. (f) Cordierite is replaced by fine-grained aggregates rich in chlorite and quartz. Note weak foliation overgrown by cordierite porphyroblast in (f) and (g). (h) Garnet occurs as euhedral crystal in the matrix. Mineral abbreviations are according to Kretz (1983).

867

Figure 4. Photomicrographs, field photographs and back-scattered electron image of schistose rocks in the staurolite (a-e), andalusite (g-m) and sillimanite (n-o) zones. (a) and (b) Garnet-staurolite bearing rocks are well-foliated schists which show a typical crenulation fabric. (c) and (h) Garnet crystals occur as inclusions in staurolite and andalusite. (d) Staurolite grain seem to show two stages of growth with the staurolite growth phase defined by trails of inclusions, followed by an inclusion poor rim in staurolite bearing rocks. Note chloritization of staurolite. (e) Biotite grains have been replaced by secondary chlorite in pressure fringes around garnet porphyroblasts. (f) Photomicrograph representing the foliation with grain boundary bulging of the schistose Jurassic granite gneiss of Aliabad-e Damaq. (g) and (h) Large andalusite porphyroblasts are typically present a chiastolitic cruciform pattern of inclusions and accumulations of graphite and opaque phases along rectilinear grain-edges. (i) Andalusite largely grew by replacement of muscovite. (j) Partial replacement of staurolite by andalusite. (k) Staurolite poikiloblast presenting sigmoidal inclusion trails reflecting syn-deformative growth in andalusite-staurolite bearing rocks. (l)

35

879 Microfabric of white mica in equilibrium with garnet. (m) Garnet is developed as poikilitic grains with inclusions of

880 quartz, graphite and ilmenite. (n) and (o) Sillimanite occur in mats within the matrix and as fine needles intergrown

881 with andalusite or coarse muscovite. Mineral abbreviations are according to Kretz, 1983.

882

883 Figure 5. (a-d) Compositional profiles across the largest garnet porphyroblasts observed in thin sections of the

884 samples investigated in this study. Garnet zonation patterns are consistent with a prograde metamorphism. (a) X_{prp}

885 varies in higher values of mole fraction than X_{sps} and X_{grs} in the flat compositional zoning of garnet. (b-d) The

886 increase of X_{alm} and concomitant decrease of X_{sps} at the transition from core to rim compositions. X_{grs} and X_{prp}

887 feature gentle compositional trends.

888

889 Figure 6. Intersections of selected thermometers and barometers of representative samples obtained from

890 conventional thermobarometry: (a) Hornfels zones; sample H101 is from the cordierite + K-feldspar zone, sample

891 H48 is from the andalusite ± cordierite zone. (b) Schistose zones; samples H44 and H54 are from the staurolite zone,

892 samples H28 and H30 are from the andalusite zone, and samples H10 and H16 are from the sillimanite zone. A

893 small systematic P increase from the contact outwards is observed, but a constant pressure may be possible in the

894 light of the uncertainties in the determination of peak pressures.

895

896 Figure 7. Calculated by PET (Dachs, 2004), the variation of temperatures of zones with distance from the contact is

897 shown. The highest temperature belongs to the cordierite-bearing rocks close to the composite pluton and sillimanite

898 bearing rocks far away from the contact.

899

900 Figure 8. Equilibrium assemblage diagrams for the bulk compositions of two hornfels samples. The field for the

901 appropriate peak assemblage identified in each sample is highlighted.

902

903 Figure 9. Equilibrium assemblage diagrams for the bulk compositions of three schistose samples. The field for the

904 appropriate peak assemblage identified in each sample is highlighted.

905

906   Figure 10. (a) Equilibrium assemblage diagram for the average Hamadan metapelite composition in the

907   MnNCKFMASHT system. (b) Calculated isopleths of X_{grs} and Mg# (Mg∕(Mg+Fe)) of garnet. The isobaric paths are

908   located on (a) and (b) representing the prograde evolution of M3 metamorphism and contact metamorphism in

909   different pressure. The location of the isobaric paths implies the most common assemblages of minerals in each

910   hornfels and schistose part. Numbered reactions are listed in Table 6.

911

912   Figure 11. (a) Th–U–Pb chemical model ages of monazite (Mnz). Total ThO_2* vs PbO (wt. %) isochron diagrams;

913   ThO_2* is ThO_2 + UO_2 equivalents expressed as ThO_2, after Suzuki et al. (1994). General minimal error 2σ on

914   monazite PbO analysis is shown in a bar. Regression lines with the coefficient of determination R^2 are forced

915   through zero as proposed by Suzuki et al. (1994) and Montel et al. (1996). Weighted average ages (Ma) with

916   MSWD and minimal error of 2σ are calculated from single analyses according to Ludwig (2001). (b) Y_2O_3 content

917   vs. age. (c) Age histogram from the few monazite samples dated by microprobe.

918

919   Figure 12. $^{40}Ar/^{39}Ar$ release patterns of white mica from the Hamadan metamorphic complex.

920

921   Figure 13. $^{40}Ar/^{39}Ar$ release patterns of amphibole from the Hamadan metamorphic complex.

922

923   Figure 14. (a) Estimated field of M1 garnet stability as constrained by compositional isopleths of X_{grs} and X_{prp}

924   corresponding to the garnet composition in Fig. 5a. (b) P–T phase diagram showing overlap of this domain with

925   stability field of mineral assemblages.

926

927   Figure 15. Time-integrated cooling rates of the Hamadan metamorphic complex, combining the results of $^{40}Ar/^{39}Ar$

928   age dating of investigated rock samples and previous geochronological data from the Hamadan complex.

929

930   Figure 16. Tectonic model for the geodynamic framework. (a) Middle–Late Jurassic subduction of the Neotethyan

931   oceanic lithosphere and formation of the Alvand composite pluton. (b) Late Cretaceous exhumation and cooling of

932   the Hamadan metamorphic complex triggered by extension due to retreat of the subduction zone and steepening of

933   the Neotethyan oceanic lithosphere.

934

935   Table 1. Summary of previous geochronological data related to metamorphism, subsequent cooling of metamorphic

936   rocks and the granitic and gabbroic rocks of the Hamadan metamorphic complex.

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938   Table 2. A summary of existing P–T estimates from the Hamadan metamorphic complex.

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940   Table 3. Locations and mineral assemblages used for P–T conditions, thermobarometry, U–Pb monazite and

941    $^{40}Ar/^{39}Ar$ age dating of investigated rock samples from the Hamadan metamorphic complex. Abbreviations: Crd -

942   Cordierite, St - Staurolite, And - Andalusite, Sil - Sillimanite, Grt - Garnet, Ms - Muscovite, Bt - Biotite, Qtz -

943   Quartz, Pl - Plagioclase, Kfs - K-feldspar, Amph - Amphibole, Ttn - Titanite, Czo - Clinozoisite, Ilm - Ilmenite.

944

945   Table 4. Whole-rock compositions used for phase diagram modelling.

946

947   Table 5. Representative electron microprobe analyses of the Hamadan metapelitic assemblages.

948

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953   error on 30 analyses of the "Madmon" monazite (Schulz and Schüssler 2013) is used. Age is in Ma.

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956 Table S2. Analytical results of $^{40}Ar/^{39}Ar$ mineral age dating from the Hamadan metamorphic complex.

Fig. 1   Click here to access/download;Figure;Fig. 1.tif ⬇

Fig. 1b

Fig. 2a & 2b

Quaternary

Marl, siltstone and sandstone (Oligo-Miocene)

Limestone (Oligo-Miocene)

Sandstone, conglomerate, limestone (Aptian-Albian)

Andesite (Aptian-Albian)

Slate and phyllite

Schistose zones

Staurolite zone

Andalusite zone

Sillimanite zone

Hornfels zones

Andalusite ± cordierite zone

Cordierite ± K-feldspar zone

Alvand pluton (granitoids)

Schistose granite (Aliabad granitic gneiss)

Pegmatite dike

● This study geochronological ages

○ Previous geochronological ages

Fault, inferred fault

N

12 km

Hamadan

Tuyserkan

Aliabad-e Damaq

C

pegmatites
162.23 ± 0.77 Ma & 172.20 ± 0.91 Ma (Mnz; U-Pb)
Sepahi et al., 2018

Alvand granitoid complex
39.5 Ma (Zrn; ZFT)
22.06 Ma (Ap; AHe)
23.1 Ma (Ap; AFT)
Francois et al., 2014

Alvand granitoid complex
157.1 ± 1.2 Ma - 171.1 ± 1.2 Ma (Zrn; U-Pb)
Zhang et al., 2018

Alvand granitoid complex
163.9 ± 1.8 Ma & 165.1 ± 2.0 Ma (Zrn; U-Pb)
Chiu et al., 2013

Alvand granitoid complex
165.0 ± 0.4 Ma & 165.2 ± 0.2 Ma (Zrn; U-Pb)
Mahmoudi et al., 2011

gabbro (Alvand batholith)
166.5 ± 1.8 Ma (Zrn; U-Pb)
Shahbazi et al., 2010

granite (Alvand)
161.7 ± 0.6 Ma & 163.9 ± 0.9 Ma (Zrn; U-Pb)
Shahbazi et al., 2010

leucocratic granitoid
153.3 ± 2.7 Ma & 154.4 ± 1.3 Ma (Zrn; U-Pb)
Shahbazi et al., 2010

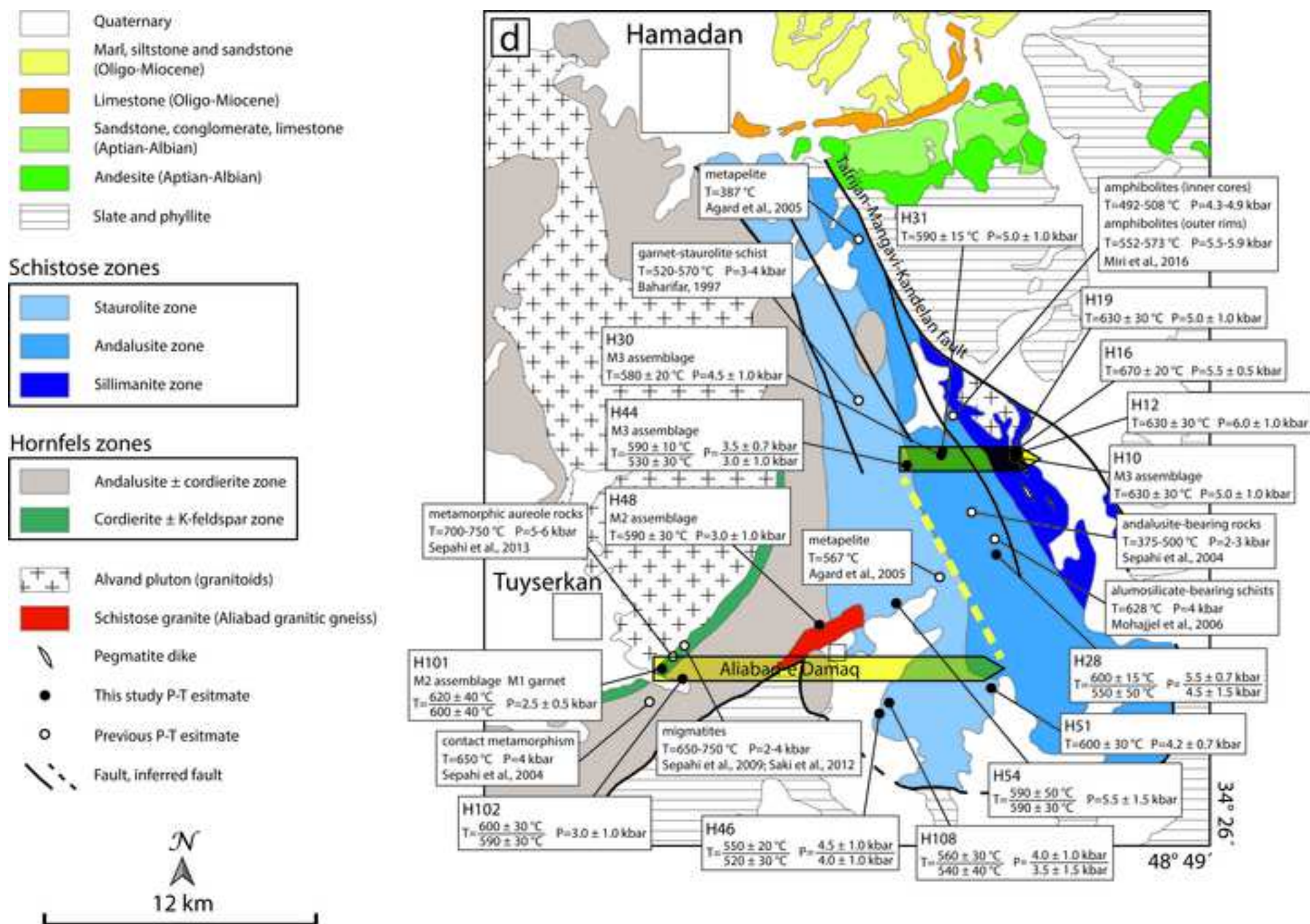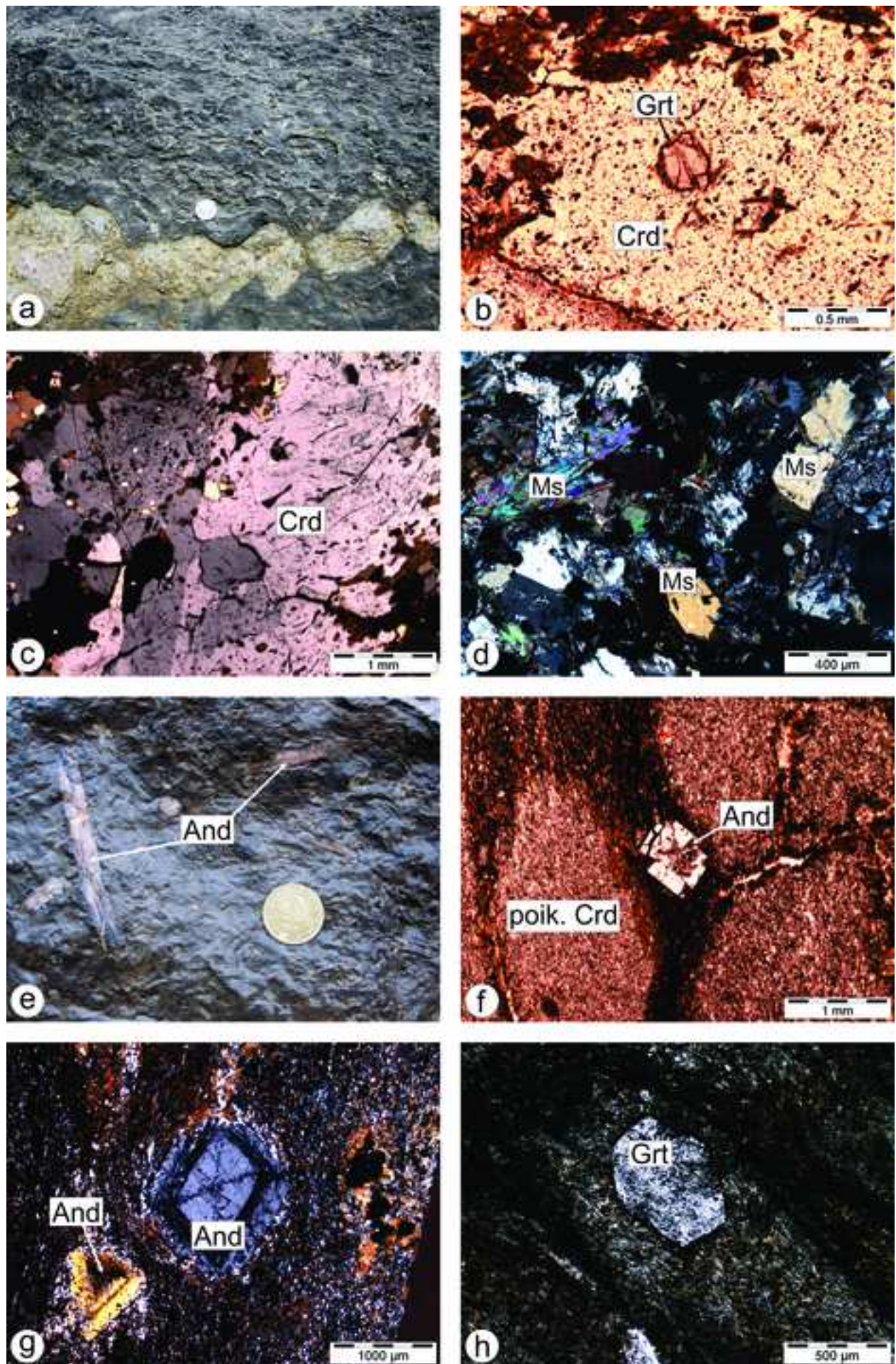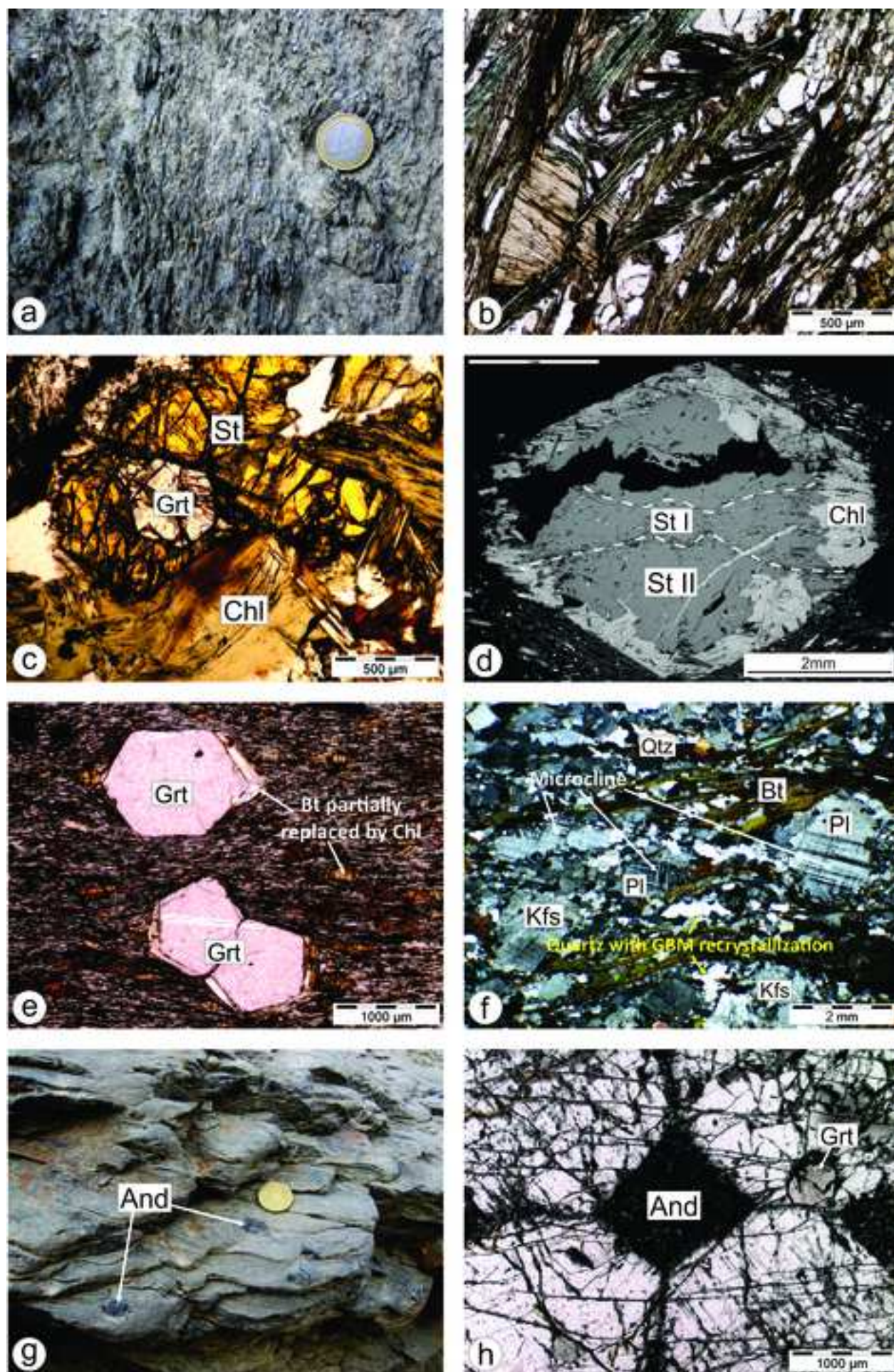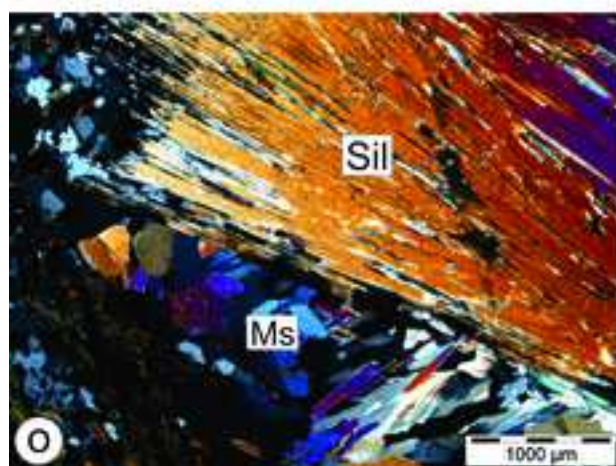mafic rocks (Alvand batholith)
89.1 ± 3 Ma (Mu & Bt; K/Ar)
78.5 - 88.5 Ma (Mu & Bt; Rb/Sr)
Valizadeh and Cantagrel, 1975

acidic rocks (Alvand granitoid complex)
63 - 80 Ma (Mu & Bt; K/Ar)
65 - 68 Ma (Mu & Bt; Rb/Sr)
Valizadeh and Cantagrel, 1975

H66
69.17 ± 0.63 Ma (Mu; Ar/Ar)

migmatite
173.60 ± 3.1 Ma (Zrn; U-Pb)
Sepahi et al., 2018

H61
72.90 ± 1.10 Ma (Amph; Ar/Ar)

amphibolite in staurolite schist
82.3 ± 2.4 Ma (Amph; K/Ar)
Baharifar et al., 2004

H31
149 ± 19 Ma (Mnz; U-Pb)

H32
73.50 ± 0.38 Ma (Mu; Ar/Ar)

H11
162 ± 13 Ma (Mnz; U-Pb)
70.59 ± 0.24 Ma (Mu; Ar/Ar)

H67
70.84 ± 0.26 Ma (Mu; Ar/Ar)

H68
73.66 ± 0.45 Ma (Mu; Ar/Ar)

andalusite and kyanite vein
76.7 ± 1.9 Ma (Mu; K/Ar)
Baharifar et al., 2004

Tabrian-Mangan-Kandelan fault

H101
168 ± 11 Ma (Mnz; U-Pb)
76.61 ± 0.42 Ma (Mu; Ar/Ar)

H56
80.21 ± 0.44 Ma (Mu; Ar/Ar)

amphibolite in staurolite schist
114.6 ± 0.3 Ma (Amph; K/Ar)
Baharifar et al., 2004

HS4
< 50 Ma (Mnz; U-Pb)

H120
62.60 ± 2.40 Ma (Amph; Ar/Ar)

HS8
72.65 ± 0.49 Ma (Mu; Ar/Ar)

H119
65.50 ± 2.50 Ma (Amph; Ar/Ar)

H124
65.80 ± 2.50 Ma (Amph; Ar/Ar)

34° 26'

48° 49'

Fig. 2d_revised

Fig. 5   Click here to access/download;Figure;Fig. 5.tif

Fig. 6

Fig. 6

Fig. 7 — Click here to access/download;Figure;Fig. 7.tif

Fig. 7

Andalusite ± cordierite zone

(a) H48

All assemblages + Qtz + H₂O

Cordierite + K-feldspar zone

(b) H101

All assemblages + Qtz + H₂O

Fig. 8

Fig. 9

Fig. 11

Fig. 12

H32, Mu, Andalusite zone
data-point error ellipses are 2σ
Age = 72.32±0.94 Ma
Initial $^{40}Ar/^{36}Ar$ =322±45
MSWD = 0.28

H68, Mu, Andalusite zone
data-point error ellipses are 2σ
Age = 70.6±3.0 Ma
Initial $^{40}Ar/^{36}Ar$ =522±240
MSWD = 0.96

H11, Mu, Sillimanite zone
data-point error ellipses are 2σ
Age = 68.6±3.8 Ma
Initial $^{40}Ar/^{36}Ar$ =412±90
MSWD = 2.0

H67, Mu, Sillimanite zone
data-point error ellipses are 2σ
Age = 69.7±9.0 Ma
Initial $^{40}Ar/^{36}Ar$ =436±190
MSWD = 4.1

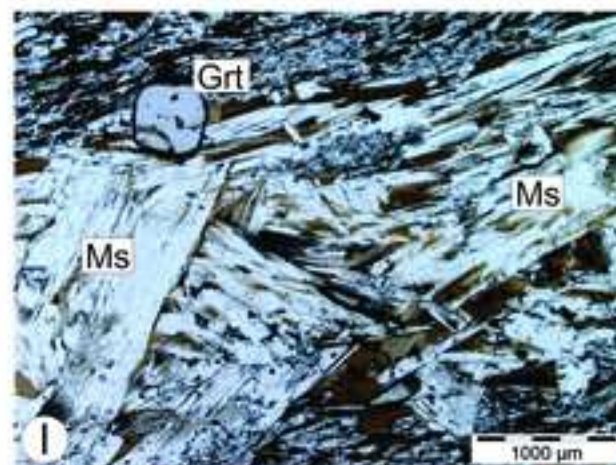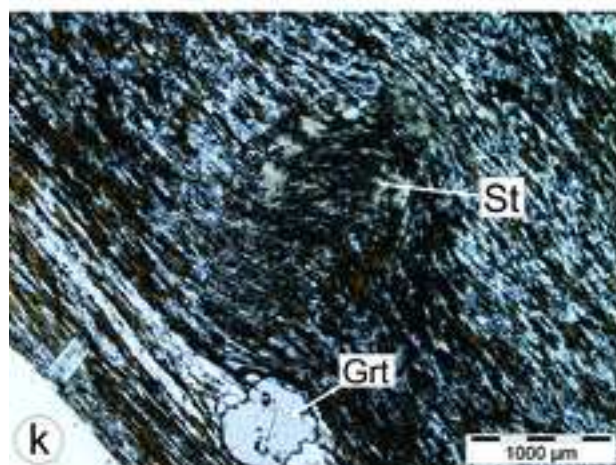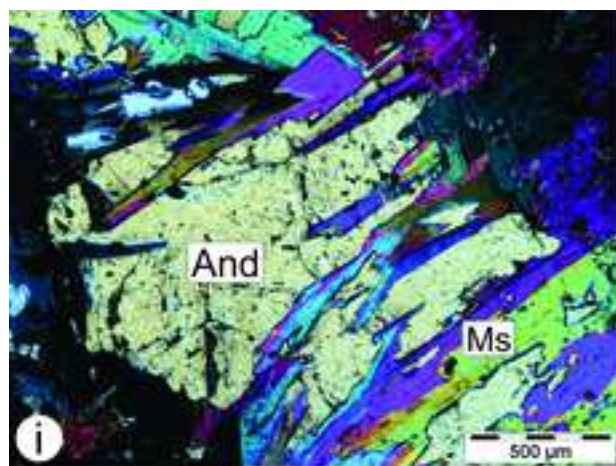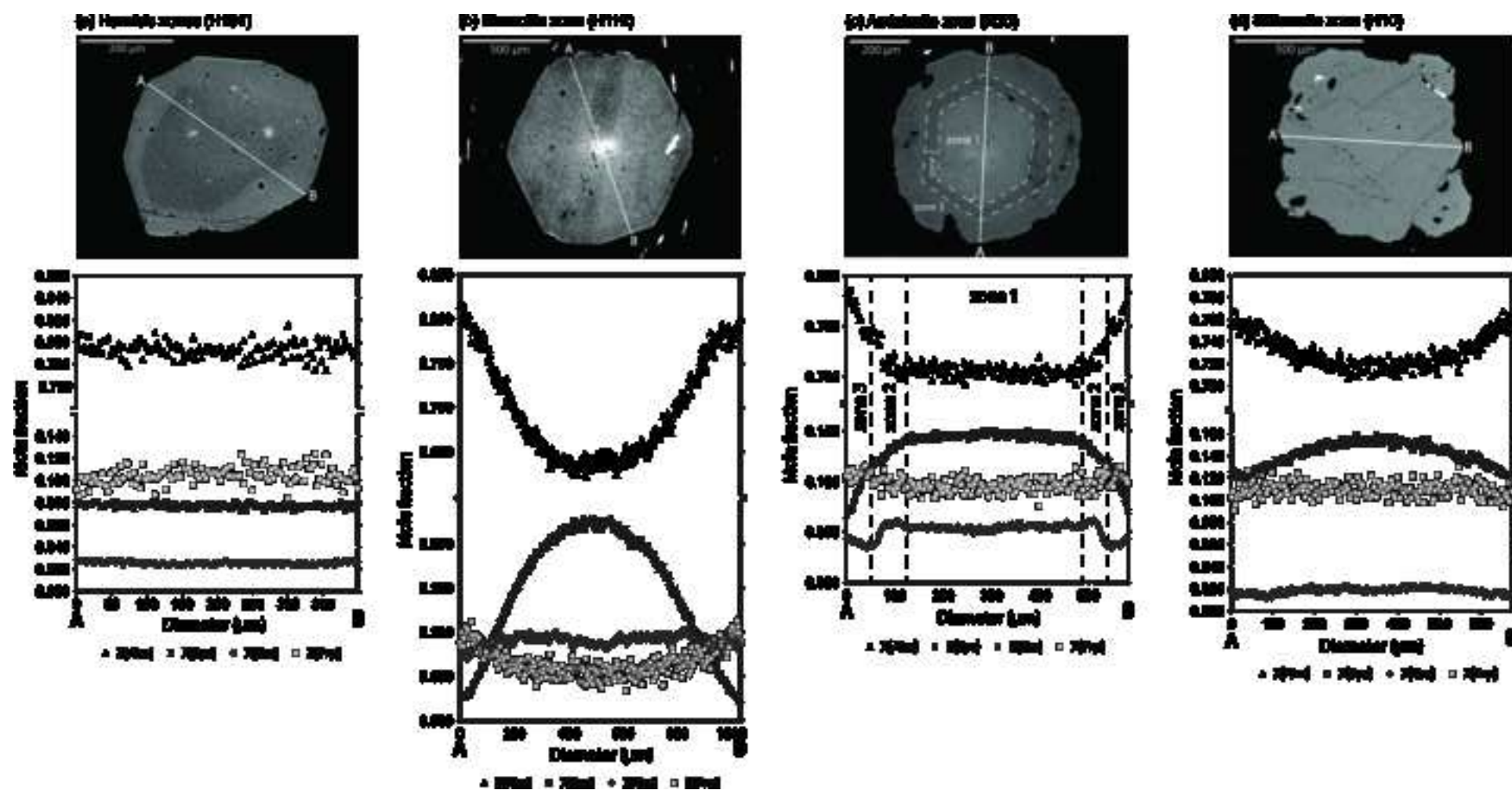Fig. 13   Click here to access/download;Figure;Fig. 13.tif ±
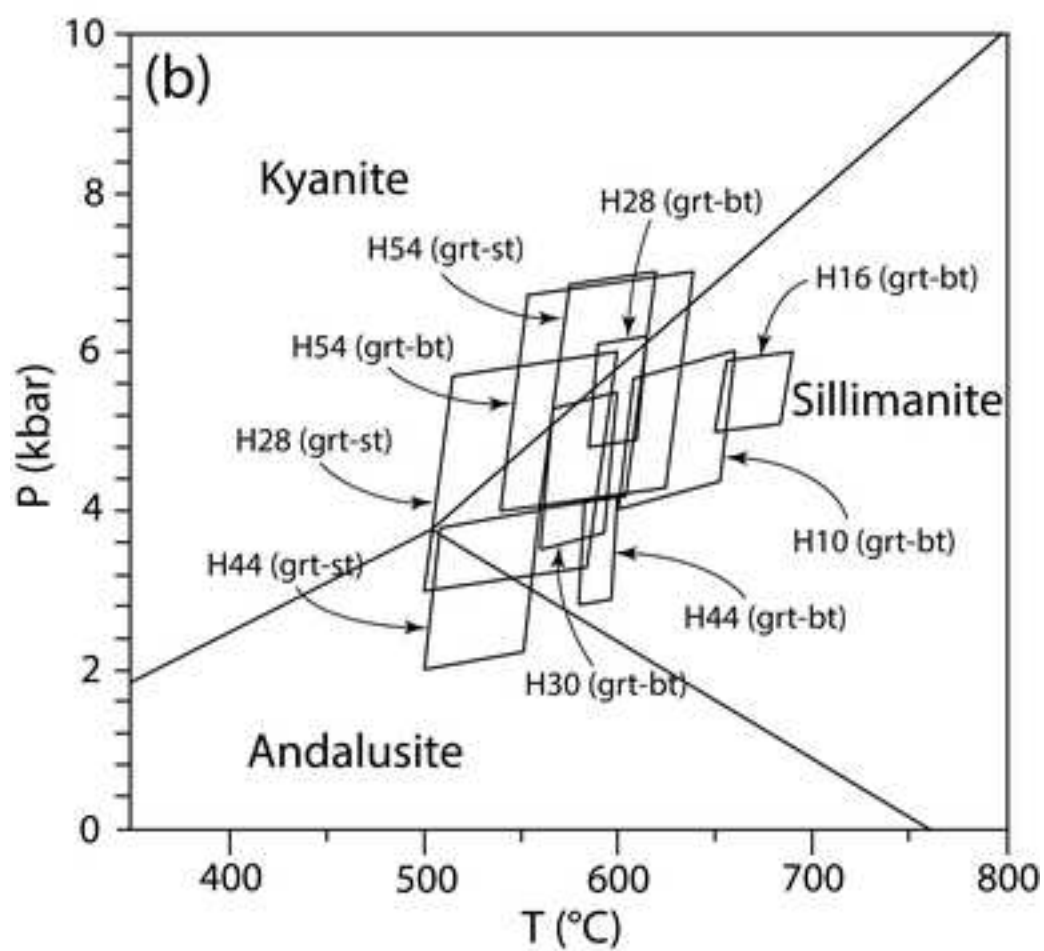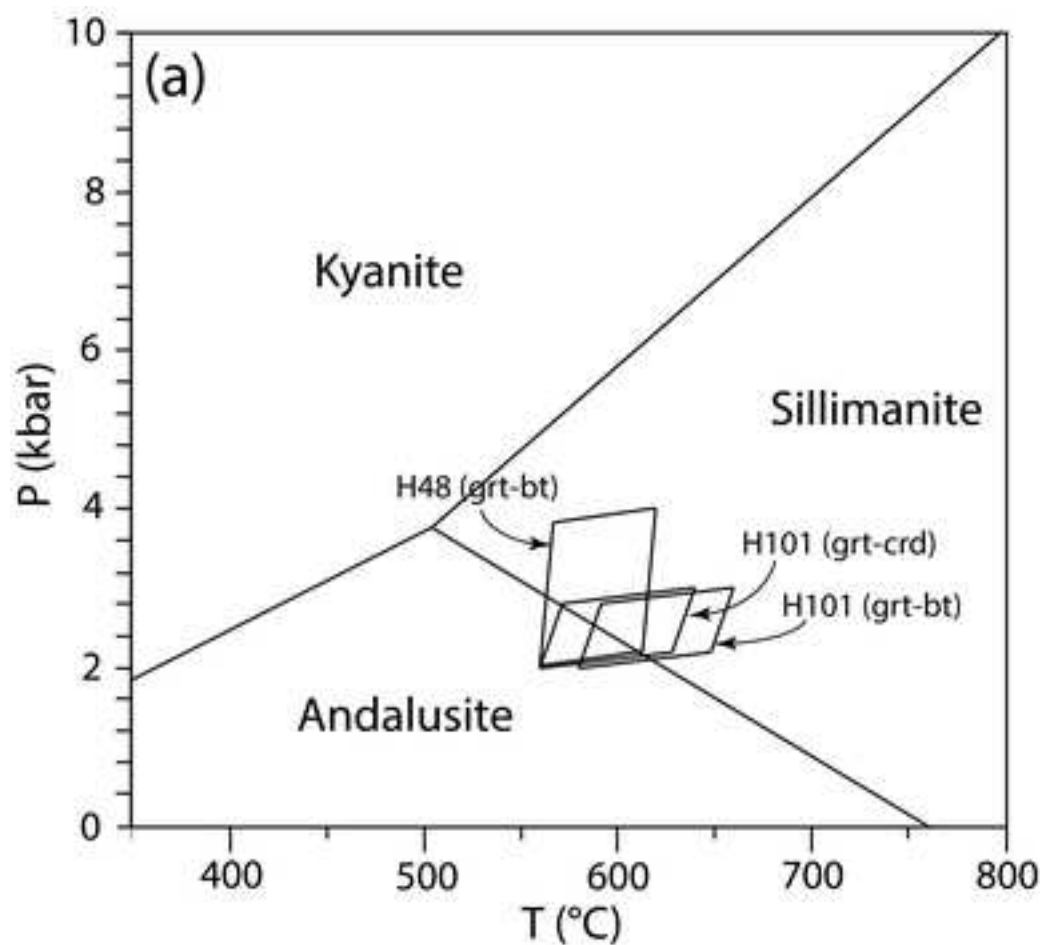
Fig. 14   Click here to access/download;Figure;Fig. 14.tif ±

Fig. 14

Figure legend:

Amph: ⁴⁰Ar/³⁹Ar amphibole (this study)
Ms: ⁴⁰Ar/³⁹Ar muscovite (this study)
Bt: K/Ar biotite (Valizadeh and Cantagrel, 1975)
ZFT: Apatite fission track (Francois et al., 2014)
AFT: Zircon fission track (Francois et al., 2014)
AHe: Apatite (U-Th)/He (Francois et al., 2014)

Fig. 16
Click here to access/download;Figure;Fig. 16.tif ⬇

(a) ~ 165 Ma (Late Jurassic). SW — trench retreat, Neotethys. Labels: contact aureole, batholith, stratovolcano, SSZ crust, mantle lithosphere, asthenosphere. NE.

(b) Late Early Cretaceous ~ 100 - 80 Ma. trench retreat. Labels: Tafrijan-Mangavi-Kandelan fault (ductile low angle normal fault), Lower Cretaceous andesite and sediments, very low to low grade metamorphic rocks, upper unit.

Table 1

Table 1. Summary of previous geochronological data related to metamorphism, subsequent cooling of metamorphic rocks and the granitic and gabbroic rocks of the Hamadan metamorphic complex.

Area	Rock type	Method	Dated mineral	Age ± error (Ma)	Authors
Hamadan metamorphic complex	andalusite and kyanite vein	K/Ar	muscovite	76.7 ± 1.9	Baharifar et al., 2004
	amphibolite in staurolite schist	K/Ar	amphibole	82.3 ± 2.4	
	amphibolite in staurolite schist	K/Ar	amphibole	114.6 ± 0.3	
Alvand plutonic complex	acidic rocks (Alvand granitoid complex)	K/Ar	muscovite and	63 - 80	Valizadeh and Cantagrel, 1975
		Rb/Sr	biotite	65 - 68	
	mafic rocks	K/Ar	muscovite and	89.1 ± 3	
		Rb/Sr	biotite	78.5 - 88.5	
	gabbro	$^{206}Pb/^{238}U$	zircon	166.5 ± 1.8	Shahbazi et al., 2010
	granite (Alvand)	$^{206}Pb/^{238}U$	zircon	163.9 ± 0.9	
		$^{206}Pb/^{238}U$	zircon	161.7 ± 0.6	
	leucocratic granitoid	$^{206}Pb/^{238}U$	zircon	154.4 ± 1.3	
		$^{206}Pb/^{238}U$	zircon	153.3 ± 2.7	
	Alvand granitoid complex	$^{206}Pb/^{238}U$	zircon	165.2 ± 0.2	Mahmoudi et al., 2011
		$^{206}Pb/^{238}U$	zircon	165.0 ± 0.4	
		$^{206}Pb/^{238}U$	zircon	165.1 ± 2.0	Chiu et al., 2013
		$^{206}Pb/^{238}U$	zircon	163.9±1.8	
		ZFT	zircon	39.5	Francois et al., 2014
		AHe	apatite	22.06	
		AFT	apatite	23.1	
		$^{206}Pb/^{238}U$	zircon	157.1 ± 1.2	Zhang et al., 2018
		$^{206}Pb/^{238}U$	zircon	167.7 ± 1.5	
		$^{206}Pb/^{238}U$	zircon	171.1 ± 1.2	
		$^{206}Pb/^{238}U$	zircon	159.3 ± 1.0	
		$^{206}Pb/^{238}U$	zircon	167.2 ± 1.3	
		$^{206}Pb/^{238}U$	zircon	167.3 ± 2.7	
	pegmatites	$^{206}Pb/^{238}U$	monazite	162.23 ± 0.77	Sepahi et al., 2018
		$^{206}Pb/^{238}U$	monazite	172.20 ± 0.91	
	migmatite	$^{206}Pb/^{238}U$	zircon	173.60 ± 3.1	Sepahi et al., 2019

Table 2-revised

Table 2. A summary of existing P–T estimates from the Hamadan metamorphic complex.

Rock Type	T (°C)	P (kbar)	Authors
garnet-staurolite schist	520-570	3-4	Baharifar, 1997
low-grade andalusite-bearing rocks	375-500	2-3	Sepahi et al., 2004
contact metamorphism	650	4	Sepahi et al., 2004
metapelite	387-567	-	Agard et al., 2005
aluminosilicate-bearing schists	628	4	Mohajjel et al., 2006
migmatites	650-750	2-4	Sepahi et al., 2009; Saki et al., 2012
metamorphic aureole rocks	700-750	5-6	Sepahi et al., 2013
amphibolites (inner cores)	492-508	4.3-4.9	Miri et al., 2016
amphibolites (outer rims)	552-573	5.5-5.9	Miri et al., 2016

Table 3-revised

Table 3. Locations and mineral assemblages used for P–T conditions, thermobarometry, U–Pb monazite and $^{40}Ar/^{39}Ar$ age dating of investigated rock samples from the Hamadan metamorphic complex. Abbreviations: Crd - Cordierite, St - Staurolite, And - Andalusite, Sil - Sillimanite, Grt - Garnet, Ms - Muscovite, Bt - Biotite, Qtz - Quartz, Pl - Plagioclase, Kfs - K-feldspar, Amph - Amphibole, Ttn - Titanite, Czo - Clinozoisite, Ilm - Ilmenite.

Zone	Rock Type	Sample no.	Lat. (N)	Long. (E)	Assemblage	Assemblage-forming event	T (°C)	P (bar)	gt	gb	(U-Th)-Pb monazite ages (Ma)	Ms	Amph
Cordierite + K-feldspar zone													
	metapelitic hornfels	H101	34°34′12″	48°25′11″	Crd+Grt+Bt+Kfs+Pl+Qtz+Ilm	M2 event, M1 garnet	620 ± 40	2.5 ± 0.5	grt-bt	grt-pl-wm-bt	168 ± 11	76.61 ± 0.42	-
							600 ± 40	2.5 ± 0.5	grt-crd			-	-
Andalusite ± cordierite zone													
	metapelitic hornfels	H56	34°31′06″	48°29′34″	And+Crd+Grt+Bt+Ms+Kfs+Pl+Qtz+Ilm		-	-	-	-	-	80.21 ± 0.44	-
	metapelitic hornfels	H102	34°31′25″	48°28′45″	And+Crd+Grt+Bt+Ms+Kfs+Pl+Qtz+Ilm		600 ± 30	3.0 ± 1.0	grt-bt	GASP	-	-	-
							590 ± 30	3.0 ± 1.0	grt-crd			-	-
	metapelitic hornfels	H48	34°32′26″	48°35′26″	And+Grt+Bt+Ms+Pl+Qtz+Ilm	M2 event	590 ± 30	3.0 ± 1.0	grt-bt	GASP	-	-	-
Staurolite zone													
	metapelitic schist	H46	34°30′20″	48°36′24″	St+Grt+Bt+Ms+Pl+Qtz+Ilm		550 ± 20	4.5 ± 1.0	grt-bt	grt-pl-wm-bt	-	-	-
							520 ± 30	4.0 ± 1.0	grt-st				-
	metapelitic schist	H108	34°30′29″	48°36′41″	St+Grt+Bt+Ms+Pl+Qtz+Ilm		560 ± 30	4.0 ± 1.0	grt-bt	grt-pl-wm-bt	-	-	-
							540 ± 40	3.5 ± 1.5	grt-st				-
	metapelitic schist	H54	34°33′19″	48°38′23″	St+Grt+Bt+Ms+Pl+Qtz+Ilm		590 ± 50	5.5 ± 1.5	grt-bt	grt-pl-wm-bt	47 ± 51	-	-
							590 ± 30	5.5 ± 1.5	grt-st				-
	metapelitic schist	H44	34°37′13″	48°38′25″	St+Grt+Bt+Ms+Pl+Qtz+Ilm	M3 event	590 ± 10	3.5 ± 0.7	grt-bt	grt-pl-wm-bt	-	-	-
							530 ± 30	3.0 ± 1.0	grt-st				-
	metapelitic schist	H51	34°31′14″	48°41′04″	Grt+Bt+Ms+Pl+Qtz+Ilm		600 ± 30	4.2 ± 0.7	grt-bt	grt-pl-wm-bt	-	-	-
	metapelitic schist	H58	34°29′36″	48°40′14″	Grt+Bt+Ms+Pl+Qtz+Ilm		-	-	-	-	-	72.65 ± 0.49	
	metapelitic schist	H66	34°45′11″	48°33′24″	St+Grt+Bt+Ms+Pl+Qtz+Ilm		-	-	-	-	-	69.17 ± 0.63	-
	metapelitic schist	H116	34°30′20″	48°37′06″	St+Grt+Bt+Ms+Pl+Qtz+Ilm		-	-	-	-	-	-	-
	Amphibolite	H61	34°40′18″	48°35′10″	Amph+Pl+Qtz+Ttn+Ep+Czo		-	-	-	-	-	-	72.90 ± 1.10
	Amphibolite	H119	34°31′03″	48°36′35″	Amph+Pl+Qtz+Ttn+Ep+Czo		-	-	-	-	-	-	65.50 ± 2.50
	Amphibolite	H120	34°31′03″	48°36′36″	Amph+Pl+Qtz+Ttn+Ep+Czo		-	-	-	-	-	-	62.60 ± 2.40
	Amphibolite	H124	34°31′06″	48°36′37″	Amph+Pl+Qtz+Ttn+Ep+Czo		-	-	-	-	-	-	65.80 ± 2.50
Andalusite zone													
	metapelitic schist	H32	34°37′28″	48°39′28″	And+St+Grt+Bt+Ms+Pl+Qtz+Ilm		-	-	-	-	-	73.50 ± 0.38	-
	metapelitic schist	H68	34°35′12″	48°41′32″	And+St+Grt+Bt+Ms+Pl+Qtz+Ilm		-	-	-	-	-	73.66 ± 0.45	-
	metapelitic schist	H28	34°34′33″	48°41′33″	And+St+Grt+Bt+Ms+Pl+Qtz+Ilm		600 ± 15	5.5 ± 0.7	grt-bt	GASP	-	-	-
							550 ± 50	4.5 ± 1.5	grt-st			-	-
	metapelitic schist	H30	34°37′26″	48°39′28″	And+Grt+Bt+Ms+Pl+Qtz+Ilm	M3 event	580 ± 20	4.5 ± 1.0	grt-bt	GASP	-	-	-
	metapelitic schist	H31	34°37′28″	48°39′29″	And+Grt+Bt+Ms+Pl+Qtz+Ilm		590 ± 15	5.0 ± 1.0	grt-bt	GASP	149 ± 19	-	-
Sillimanite zone													
	metapelitic schist	H19	34°37′25″	48°41′57″	Sil+And+Grt+Bt+Ms+Pl+Qtz+Ilm		630 ± 30	5.0 ± 1.0	grt-bt	GASP	-	-	-
	metapelitic schist	H16	34°37′21″	48°41′26″	Sil+Grt+Bt+Ms+Pl+Qtz+Ilm		670 ± 20	5.5 ± 0.5	grt-bt	GASP	-	-	-
	metapelitic schist	H12	34°37′20″	48°42′09″	Sil+Grt+Bt+Ms+Pl+Qtz+Ilm		630 ± 30	6.0 ± 1.0	grt-bt	GASP	-	-	-
	metapelitic schist	H10	34°37′19″	48°42′09″	Sil+Grt+Bt+Ms+Pl+Qtz+Ilm	M3 event	630 ± 30	5.0 ± 1.0	grt-bt	GASP	-	-	-
	metapelitic schist	H11	34°37′20″	48°42′09″	Sil+Grt+Bt+Ms+Pl+Qtz+Ilm		-	-	-	-	162 ± 13	70.59 ± 0.24	-
	metapelitic schist	H67	34°37′20″	48°41′34″	Sil+Grt+Bt+Ms+Pl+Qtz+Ilm		-	-	-	-	-	70.84 ± 0.26	-

Table 4

Table 4. Whole-rock compositions used for phase diagram modelling.

	H101 Crd + Kfs zone	H48 And ± crd zone	H116 St zone	H44 St zone	H30 And zone	H10 Sil zone	Average Hamadan composition
SiO_2	61.87	62.72	45.40	64.61	64.52	63.55	63.45
TiO_2	0.86	0.85	1.39	0.77	0.83	0.92	0.85
Al_2O_3	19.80	20.20	25.87	18.31	19.38	18.34	19.21
Fe_2O_3	7.39	6.80	10.85	6.52	5.99	7.82	6.90
MnO	0.18	0.19	0.26	0.08	0.12	0.15	0.14
MgO	2.21	2.12	3.10	2.16	1.83	2.18	2.10
CaO	0.42	0.28	1.05	1.26	0.97	0.51	0.69
Na_2O	1.53	1.28	1.15	1.54	1.24	1.22	1.36
K_2O	4.28	3.91	5.30	3.22	3.21	3.32	3.59
P_2O_5	0.20	0.16	0.21	0.19	0.15	0.21	0.18
LOI	1.38	2.08	4.99	2.02	2.27	2.15	1.98
Total	100.13	100.59	99.57	100.67	100.50	100.38	100.45

Table 5

Table 5. Representative electron microprobe analyses of the Hamadan metapelitic assemblages.

	Cordierite + K-feldspar zone (Sample H101)							Andalusite ± cordierite zone (Sample H48)					
	Crd	Grt core	Grt rim	Bt	Ms	Pl	Kfs	And	Grt core	Grt rim	Bt	Ms	Pl
SiO_2	48.34	37.07	37.43	35.39	48.49	63.84	64.94	35.65	36.95	36.96	33.03	46.17	62.32
TiO_2	0.00	0.00	0.03	2.32	0.30	0.00	0.05	0.00	0.01	0.06	3.77	0.58	0.00
Al_2O_3	31.95	21.01	20.75	18.32	36.29	23.36	18.99	64.04	21.07	21.04	18.48	37.87	23.96
FeO	11.17	34.91	34.81	20.75	1.22	0.00	0.00	0.00	33.43	33.87	19.98	1.17	0.10
MnO	0.59	3.44	3.25	0.00	0.00	0.08	0.00	0.02	4.07	4.34	0.16	0.09	0.02
MgO	5.81	2.45	2.63	9.00	0.86	0.00	0.00	0.12	3.62	3.09	8.27	0.63	0.00
CaO	0.00	0.84	0.91	0.00	0.01	4.36	0.01	0.02	0.28	0.39	0.04	0.00	5.13
Na_2O	0.37	0.00	0.00	0.16	0.34	8.79	1.10	0.00	0.00	0.00	0.08	0.34	8.32
K_2O	0.02	0.02	0.04	9.28	10.35	0.22	14.86	0.00	0.03	0.03	9.13	9.83	0.17
F	-	-	-	0.52	0.16	-	-	-	-	-	0.70	0.23	-
Total	98.24	99.74	99.85	95.73	98.00	100.66	99.95	99.87	99.46	99.78	93.64	96.92	100.02
Oxygen	18	12	12	11	11	8	8	20	12	12	11	11	8
Si	5.058	3.007	3.026	2.709	3.122	2.801	2.984	0.963	2.989	2.992	2.603	3.009	2.758
Ti	0.000	0.000	0.002	0.134	0.015	0.000	0.002	0.000	0.001	0.004	0.223	0.028	1.250
Al	3.940	2.008	1.977	1.653	2.754	1.208	1.028	2.039	2.009	2.007	1.716	2.909	0.000
Fe2+	0.977	2.368	2.354	1.328	0.066	0.000	0.000	0.000	2.250	2.291	1.317	0.064	0.000
Fe3+	0.000	0.000	0.000	0.000	0.000	0.001	0.000	0.006	0.000	0.000	0.000	0.000	0.004
Mn	0.052	0.236	0.223	0.000	0.000	0.003	0.000	0.000	0.279	0.298	0.011	0.005	0.001
Mg	0.906	0.296	0.317	1.027	0.083	0.000	0.000	0.005	0.437	0.373	0.972	0.061	0.000
Ca	0.000	0.073	0.079	0.000	0.001	0.205	0.001	0.001	0.024	0.034	0.003	0.000	0.243
Na	0.075	0.000	0.000	0.024	0.042	0.748	0.098	0.000	0.000	0.000	0.012	0.043	0.714
K	0.003	0.002	0.004	0.906	0.850	0.012	0.871	0.000	0.003	0.003	0.918	0.817	0.010
Cl	-	-	-	0.000	0.000	-	-	-	-	-	0.000	0.000	-
F	-	-	-	0.126	0.033	-	-	-	-	-	0.174	0.047	-
Total	11.011	7.990	7.982	7.907	6.966	4.978	4.984	3.014	7.992	8.002	7.949	6.983	4.979
Xab	-	-	-	-	-	0.775	0.101	-	-	-	-	-	0.738
Xan	-	-	-	-	-	0.212	0.001	-	-	-	-	-	0.252
Xor	-	-	-	-	-	0.013	0.899	-	-	-	-	-	0.010
Xalm	-	0.797	0.792	-	-	-	-	-	0.753	0.765	-	-	-
Xprp	-	0.100	0.107	-	-	-	-	-	0.146	0.124	-	-	-
Xgrs	-	0.025	0.027	-	-	-	-	-	0.008	0.011	-	-	-
Xsps	-	0.079	0.075	-	-	-	-	-	0.093	0.099	-	-	-
XMg	0.481	0.111	0.119	0.430	-	-	-	-	0.163	0.140	0.425	-	-
K/(K+Na)	-	-	-	-	0.952	-	-	-	-	-	-	0.950	-

Table 5 (Continued)

Table 5. (Continued)

	Staurolite zone (Sample H46)							Andalusite zone (Sample H28)								Sillimanite zone (Sample H19)						
	St core	St rim	Grt core	Grt rim	Bt	Ms	Pl	And	St core	St rim	Grt core	Grt rim	Bt	Ms	Pl	Sill	And	Grt core	Grt rim	Bt	Ms	Pl
SiO_2	27.16	28.77	36.41	37.21	38.01	48.47	60.42	36.52	27.81	27.99	36.65	36.89	35.19	47.12	61.82	36.48	36.50	36.59	36.93	34.88	47.20	62.05
TiO_2	0.48	0.55	0.08	0.06	1.47	0.39	0.04	0.07	0.55	0.48	0.00	0.02	1.56	0.48	0.00	0.00	0.07	0.01	0.02	3.00	0.87	0.04
Al_2O_3	53.43	54.50	20.89	20.49	19.33	36.99	25.12	63.15	53.33	52.83	21.70	20.11	18.55	37.06	24.26	63.39	63.37	20.71	20.53	19.14	36.98	24.06
FeO	13.62	14.14	27.68	35.50	16.32	0.84	0.00	0.00	14.15	13.82	32.26	35.83	19.06	1.34	0.00	0.00	0.00	32.97	33.05	20.34	0.92	0.00
MnO	0.00	0.06	9.86	1.34	0.03	0.04	0.03	0.05	0.22	0.11	5.41	2.33	0.03	0.04	0.03	0.00	0.00	6.17	6.12	0.06	0.00	0.08
MgO	1.90	2.39	1.22	2.30	10.47	0.54	0.00	0.32	1.70	1.59	2.74	2.47	9.46	0.54	0.02	0.00	0.18	2.59	2.53	7.89	0.51	0.00
CaO	0.02	0.01	3.11	3.06	0.09	0.02	7.02	0.03	0.00	0.00	1.72	1.29	0.18	0.04	5.61	0.02	0.00	0.83	0.74	0.00	0.08	5.43
Na_2O	0.00	0.01	0.00	0.00	0.21	0.98	7.37	0.00	0.03	0.00	0.00	0.00	0.22	1.20	8.53	0.00	0.02	0.00	0.00	0.20	0.77	8.49
K_2O	0.00	0.03	0.03	0.01	8.30	8.37	0.05	0.00	0.01	0.00	0.00	0.01	8.32	9.91	0.03	0.01	0.00	0.02	0.03	9.03	9.92	0.04
ZnO	0.26	0.43	-	-	-	-	-	-	0.52	0.25	-	-	-	-	-	-	-	-	-	-	-	-
Cl	-	-	-	-	0.01	0.01	-	-	-	-	-	-	0.11	0.01	-	-	-	-	-	-	0.01	0.02	-
F	-	-	-	-	0.52	0.14	-	-	-	-	-	-	0.00	0.00	-	-	-	-	-	0.51	0.16	-
Total	96.87	100.89	99.28	99.97	94.76	96.80	100.05	100	98	97	100	99	93	98	100	99.90	100.13	99.89	99.95	95.07	97.44	100.19
Oxygen	46	46	12	12	11	11	8	20	46	46	12	12	11	11	8	5	20	12	12	11	11	8
Si	7.621	7.761	2.979	3.008	2.838	3.121	2.685	0.983	7.720	7.834	2.938	3.023	2.737	3.050	2.732	0.984	0.983	2.970	2.998	2.683	3.057	2.747
Ti	0.101	0.112	0.005	0.004	0.083	0.019	0.001	0.001	0.115	0.101	0.000	0.001	0.091	0.023	0.000	0.000	0.001	0.001	0.001	0.174	0.042	0.001
Al	17.670	17.328	2.014	1.952	1.701	2.808	1.316	2.004	17.449	17.427	2.050	1.942	1.700	2.827	1.264	2.015	2.011	1.981	1.965	1.735	2.823	1.256
Fe^{2+}	3.196	3.190	1.873	2.376	1.019	0.045	0.000	0.000	3.285	3.235	2.096	2.445	1.240	0.073	0.000	0.000	0.000	2.161	2.208	1.309	0.050	0.000
Fe^{3+}	0.000	0.000	0.000	0.000	0.000	0.000	0.004	0.006	0.000	0.000	0.000	0.000	0.000	0.000	0.001	0.005	0.005	0.000	0.000	0.000	0.000	0.003
Mn	0.000	0.014	0.683	0.092	0.002	0.002	0.001	0.001	0.052	0.026	0.367	0.162	0.002	0.002	0.001	0.000	0.000	0.424	0.421	0.004	0.000	0.003
Mg	0.795	0.961	0.149	0.277	1.165	0.052	0.000	0.013	0.704	0.663	0.327	0.302	1.097	0.052	0.001	0.000	0.007	0.313	0.306	0.905	0.049	0.000
Ca	0.006	0.003	0.273	0.265	0.007	0.001	0.334	0.001	0.000	0.000	0.148	0.113	0.015	0.003	0.266	0.001	0.000	0.072	0.064	0.000	0.006	0.258
Na	0.000	0.005	0.000	0.000	0.030	0.122	0.635	0.000	0.016	0.000	0.000	0.000	0.033	0.151	0.731	0.000	0.001	0.000	0.000	0.030	0.097	0.729
K	0.000	0.010	0.003	0.001	0.791	0.688	0.003	0.000	0.004	0.000	0.000	0.001	0.825	0.818	0.002	0.000	0.000	0.002	0.003	0.886	0.820	0.002
Zn	0.054	0.086	-	-	-	-	-	-	0.107	0.052	-	-	-	-	-	-	-	-	-	-	-	-
Cl	-	-	-	-	0.001	0.001	-	-	-	-	-	-	0.014	0.001	-	-	-	-	-	0.001	0.002	-
F	-	-	-	-	0.123	0.029	-	-	-	-	-	-	0.000	0.000	-	-	-	-	-	0.124	0.033	-
Total	29.443	29.470	7.979	7.975	7.760	6.888	4.979	3.009	29.452	29.338	7.926	7.989	7.754	7.000	4.998	3.005	3.008	7.924	7.966	7.851	6.979	4.998
Xab	-	-	-	-	-	-	0.653	-	-	-	-	-	-	-	0.732	-	-	-	-	-	-	0.7372
Xan	-	-	-	-	-	-	0.344	-	-	-	-	-	-	-	0.266	-	-	-	-	-	-	0.2605
Xor	-	-	-	-	-	-	0.003	-	-	-	-	-	-	-	0.002	-	-	-	-	-	-	0.0023
Xalm	-	-	0.629	0.789	-	-	-	-	-	-	0.713	0.809	-	-	-	-	-	0.728	0.736	-	-	-
Xprp	-	-	0.050	0.092	-	-	-	-	-	-	0.111	0.100	-	-	-	-	-	0.105	0.102	-	-	-
Xgrs	-	-	0.092	0.088	-	-	-	-	-	-	0.050	0.037	-	-	-	-	-	0.024	0.021	-	-	-
Xsps	-	-	0.229	0.031	-	-	-	-	-	-	0.125	0.054	-	-	-	-	-	0.143	0.140	-	-	-
XMg	-	-	0.074	0.104	0.533	-	-	-	-	-	0.135	0.110	0.469	-	-	-	-	0.127	0.122	0.409	-	-
K/(K+Na)	-	-	-	-	-	0.849	-	-	-	-	-	-	-	0.844	-	-	-	-	-	-	0.894	-

Table 6

Table 6. Mineral reactions related to assemblages in the Hamadan metapelites. Numbered reactions are shown in Fig. 10 and discussed in the text.

	Reaction number in P–T phase diagram	Reaction label	Simplified reaction in KFMASH
Hornfels Zones	1	And-in (from Chl)	Ms + Chl + Qtz = And + Bt + H2O
	2	And to Crd (Chl-free)	Bt + And + Qtz + H2O = Ms + Crd
	3	And+Kfs-in	Ms + Qtz = Kfs + And + H2O
	4	And to Crd+Kfs-in	Bt + And + Qtz = Crd + Kfs +H2O
Schistose Zones	5	St-in (from Chl)	Ms + Chl = St + Bt + Qtz + H2O
	6	St to And (Chl-free)	Ms + St + Qtz = And + Bt + H2O
	7	Sil-in (from And)	And = Sil

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