Title: Alteration of magmatic monazite in granitoids from the Ryoke belt (SW Japan): processes and consequences
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1   **Alteration of magmatic monazite in granitoids from the Ryoke belt (SW Japan):**

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3

4   **REVISION 1**

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16   **Running title**: Alteration of magmatic monazite

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18   With 7 figures and 2 tables

19

20   **ABSTRACT**

21   The alteration of magmatic monazite and its consequences for monazite geochronology are

22   explored in granitoids from the western part of the Ryoke belt (Iwakuni-Yanai area, SW

23   Japan). Biotite-granite samples were collected in two plutons emplaced slightly before the

24   main tectono-metamorphic event: one, a massive granite (Shimokuhara) adjoins schistose

25   rocks affected by greenschist facies metamorphism; and the second, a gneissose granite

26   (Namera) adjoins migmatitic gneiss that experienced upper-amphibolite facies conditions.

27   Despite contrasting textures, the granite samples have similar mineral modes and

28   compositions. Monazite in the massive granite is dominated by primary domains with limited

29   secondary recrystallization along cracks and veinlets. It is variably replaced by

30   allanite+apatite±xenotime±Th−U-rich phases. The outermost rims of primary domains yield a

31   weighted average $^{206}Pb/^{238}U$ date of 102 ± 2 Ma while the Th−U phases show Th−U−Pb dates

32   of 58 ± 5 and $15-14 \pm 2-3$ Ma. Monazite in the gneissose granite preserves sector- or

33   oscillatory-zoned primary domains cross-cut by secondary domains enriched in Ca, Y, U, P

34   and containing numerous inclusions. The secondary domains preserve concordant $^{206}Pb/^{238}U$

35   dates spreading from 102 ± 3 to 91 ± 2 Ma while primary domain analyses are commonly

36   discordant and range from 116 to 101 Ma.

37   Monazite alteration textures in the two granites chiefly reflect differences in their post-

38   magmatic histories. In the massive granite, monazite replacement occurred via a nearly

39   stoichiometrically-balanced reaction reflecting interaction with an aqueous fluid enriched in

40   Ca+Al+Si±F during hydrothermal alteration of the granitic assemblage, likely below 500 °C.

41   In the gneissose granite a small amount of anatectic melt, probably derived from the

42   neighboring metasedimentary rocks, was responsible for a pseudomorphic recrystallization of

43   monazite by dissolution-reprecipitation above 600 °C. Regardless of whether monazite

44   underwent replacement or recrystallization, primary monazite domains preserve the age of

45 magmatic crystallization for both plutons (102 ± 2 Ma and 106 ± 5 Ma). Conversely, the age

46 of monazite alteration is not easily resolved. Monazite replacement in the massive granite

47 might be constrained using the Th−U-rich alteration products; with due caution and despite

48 probable radiogenic Pb loss, the oldest date of 58 ± 5 Ma could be ascribed to chloritization

49 during final exhumation of the granite. The spread in apparently concordant ^{206}Pb/^{238}U dates

50 for secondary domains in the gneissose granite is attributed to incomplete isotopic resetting

51 during dissolution-reprecipitation, and the youngest date of 91 ± 2 Ma is considered as the age

52 of monazite recrystallization during a suprasolidus metamorphic event. These results reveal a

53 diachronous, ca. 10 Ma-long HT history and an overall duration of about 15 Ma for the

54 metamorphic evolution of the western part of the Ryoke belt.

55

56 **Keywords**: Monazite replacement, monazite recrystallization, dissolution-reprecipitation,

57 partial melting, LA-ICP-MS geochronology, Ryoke plutono-metamorphic belt

58   **INTRODUCTION**

59   Monazite, a lanthanide- and actinide-bearing phosphate, is a valuable petrological and

60   geochronological tool. In metamorphic rocks, it can be used to decipher the age and

61   Pressure−Temperature ($P-T$) conditions of distinct orogenic events (e.g. Williams et al. 1999;

62   Foster et al. 2004; Engi 2017) while in magmatic rocks it is useful for dating pluton

63   emplacement (e.g. Harrison et al. 1995; Be Mezeme et al. 2006), especially in cases where

64   zircon dates appear unreliable (Kusiak et al. 2004; Piechocka et al. 2017). However, monazite

65   is also known to be a versatile mineral (e.g. Catlos, 2013), so its petrochronological potential

66   must be carefully assessed.

67   In granitoids from the Ryoke belt (SW Japan), age dating of different samples from

68   the same pluton revealed a discrepancy between magmatic zircon and monazite dates. There

69   are cases of $^{206}Pb/^{238}U$ zircon dates being older than Th−U−Pb monazite dates (Skrzypek et al.

70   2016), and vice versa (Takatsuka et al. 2018a, b). When monazite dates are younger than

71   zircon's, it is commonly suspected that monazite reacted during post-magmatic processes, as

72   the mineral is prone to fluid-induced compositional modifications and/or age resetting (e.g.

73   Kelly et al. 2012; Seydoux-Guillaume et al. 2012; Villa and Williams 2013; Taylor et al.

74   2014). Thus, our objective was to identify monazite alteration textures and document possible

75   age discrepancies when zircon and monazite from the same sample are dated with the same

76   method.

77   The monazite-bearing granitoid samples described by Skrzypek et al. (2016) for zircon

78   dating (Iwakuni-Yanai area, western part of the Ryoke belt) represent a suitable material, as

79   they contain monazite with common alteration textures and different age populations. In this

80   contribution we document in detail the post-magmatic monazite alteration textures and

81   evaluate their impact on monazite's geochronological record. We focus on two granite

82   samples that have (1) crystallization ages slightly older (~5 Ma) than the main (regional)

83   tectono-metamorphic event recorded in the neighboring metasedimentary rocks, (2) similar

84   whole-rock compositions, and (3) contrasting monazite textures. The massive Shimokuhara

85   granite lies next to schistose rocks where regional conditions did not exceed ~2.5 kbar/550 °C

86   (Ikeda 2004), whereas the gneissose Namera granite adjoins migmatitic gneissose rocks where

87   regional metamorphism reached ~3 kbar/675 °C (Ikeda 2004; Fig. 1). As such, these two

88   granites provide an opportunity to explore the response of magmatic monazite to post-

89   emplacement metamorphism of varying grade. First, we report the mineralogy, texture and

90   mineral mode of biotite-granite samples from both plutons to assess their degree of similarity.

91   We then document their monazite textures, and date monazite *in situ* by laser ablation

92   inductively-coupled mass spectrometry (LA-ICP-MS). The results are used to identify the

93   processes involved in the alteration of magmatic monazite and the consequences for age

94   dating of magmatic/metamorphic events.

95   **BACKGROUND: ALTERATION OF MAGMATIC MONAZITE**

96   The alteration of magmatic monazite follows two mechanisms: *replacement* by other phases

97   and/or *recrystallization* of new monazite. The most commonly observed monazite

98   replacement texture in (meta-)igneous rocks involves the peripheral growth of allanite, apatite

99   and minor Th−U-bearing phases following a nearly balanced reaction (e.g. Murata et al. 1957;

100   Broska and Siman 1998). Interestingly, monazite replacement by allanite and/or apatite is

101   reported in granitoids that underwent no metamorphism (Broska et al. 2005), greenschist

102   facies hydrothermal alteration (~340 °C; Ward et al. 1992; Poitrasson et al. 2000), and lower-

103   (5−6 kbar/550−600°C; Finger et al. 1998) to upper-amphibolite facies metamorphism (4−8

104   kbar/685−825°C; Broska et al. 2005). An extreme example of monazite replacement by

105   Aln−Ap at the magmatic stage is attributed to an increase in the Ca content of the melt (Dini

106   et al. 2004). Other cases of monazite replacement at low temperature (290−350 °C) involve

5   5

107   the growth of pyrite + brockite (Poitrasson et al. 2000) or thorite + fluorapatite + synchysite

108   (Didier et al. 2013) assemblages.

109   Monazite recrystallization corresponds to the formation of new domains with a

110   composition that differs from the initial magmatic grain. In this case a balanced reaction is

111   difficult to infer and it is only possible to document the relative element enrichment/depletion

112   with respect to primary monazite. Recrystallization may result in monazite evolving towards a

113   cheralitic [$Ca_{0.5}Th_{0.5}PO_4$], huttonitic [$ThSiO_4$], Heavy REE (HREE)-richer or U-richer

114   composition (Jefferies 1985; Ward et al. 1992; Poitrasson et al. 1996; 2000; Townsend et al.

115   2000; Didier et al. 2013). Again, monazite recrystallization is observed across a wide range of

116    $P-T$ conditions; it is ascribed to relatively low-T chloritization or greisenization (Jefferies

117   1985; Ward et al. 1992; Poitrasson et al. 1996; 2000; Didier et al. 2013) and high-T, late- to

118   post-magmatic fluid-mineral interaction (Hawkins and Bowring 1997). There is no evident

119   link between $P-T$ conditions and the composition of newly-formed monazite, but the extent

120   of recrystallization seems to increase with metamorphic grade. Townsend et al. (2000)

121   observed that monazite recrystallization occurs along micron-scale veins in low-grade

122   samples (~1.5 kbar/< 350−400 °C), whereas larger patchy domains are formed at higher grade

123   (~3−4 kbar/400−700 °C).

124   **GEOLOGICAL OUTLINE**

125   The Ryoke belt is an association of Cretaceous plutonic and metamorphic rocks located in SW

126   Japan (e.g. Nakajima 1994). The Iwakuni-Yanai area (W part of the Ryoke belt) exposes Late

127   Jurassic accretionary-wedge sediments metamorphosed along a low-P/high-T gradient and

128   intruded by abundant granitoids (Fig. 1; e.g. Brown 1998). From north to south, the weakly-

129   metamorphosed sediments transition to a low-grade schist zone and then a high-grade gneiss

130   zone (Fig. 1; Higashimoto et al. 1983). The tectono-metamorphic history involves an initial

131   contact metamorphic event followed by the growth of syn-tectonic parageneses that are used

132 to define the E−W trending, so-called regional metamorphic zones (Fig. 1; Ikeda 1998;

133 Okudaira et al. 2001; Skrzypek et al. 2016).

134   For metamorphic rocks, U−Pb zircon dating in the highest-grade Grt−Crd zone

135 brackets the timing of suprasolidus conditions between 103 and 97 Ma (Nakajima et al. 2013;

136 Skrzypek et al. 2016). Metamorphic monazite in the Kfs−Crd and Grt−Crd zones shows

137 features that indicate growth at mainly prograde to peak temperature conditions (600−800

138 °C), but U−Pb monazite dates ranging from ca. 100 down to 85 Ma are obtained, so that the

139 persistence of medium-T conditions (> 200−300 °C) until 85 Ma was proposed (Skrzypek et

140 al. 2018). U−Pb zircon ages for granitoids, in agreement with structural and petrological

141 observations, indicate pluton intrusion before (105−100 Ma), during (~100 Ma) and after (~96

142 Ma) the regional metamorphic peak (Fig. 1; Herzig et al. 1998; Nakajima et al. 2013;

143 Skrzypek et al. 2016). The present study focuses on the granitoids (Shimokuhara and Namera)

144 emplaced in the sediments before the metamorphic peak, and therefore affected by subsequent

145 regional metamorphism.

146   The Shimokuhara granite abuts schistose rocks of the Bt and Ms−Crd metamorphic

147 zones (Fig. 1). It is a massive, biotite-bearing granite that commonly shows biotite

148 chloritization, and locally contains K-feldspar phenocrysts, garnet, tourmaline and allanite as

149 primary phases (Murakami 1970; Higashimoto et al. 1983; Moutte and Iiyama 1984). The

150 contact with metasedimentary rocks gently dips to the east, and a metamorphic aureole

151 defined by andalusite porphyroblasts extends up to ~1 km east of the exposed contact

152 (Skrzypek et al. 2016). The pluton is thought to have intruded at a maximum depth of 2−3 km

153 (Murakami 1970). Peak $P-T$ estimates for regional metamorphism in the neighboring

154 schistose rocks range from 1 kbar/450 °C to 2.5 kbar/550 °C (Ikeda 2004). For a biotite-

155 granite sample, U−Pb dating of sector- and oscillatory-zoned zircon by LA-ICP-MS revealed

156 $^{206}Pb/^{238}U$ dates ranging from 108 to 102 Ma, with 105 ± 3 Ma being taken as the age of

157   granite intrusion (Skrzypek et al. 2016). Monazite dating of two granite samples by the

158   chemical isochron method (CHIME) yielded 87.3 ± 1.6 and 86.6 ± 2.1 Ma, and these results

159   were also regarded as the age of granite intrusion (Suzuki et al. 1996). K−Ar dating on biotite

160   from a two-mica granodiorite sample gave 87 Ma (data from Kawano and Ueda 1966,

161   recalculated by Higashimoto et al. 1983).

162   The Namera granite lies next to gneissose rocks of the Kfs−Crd zone (Fig. 1). It is

163   mainly a muscovite−biotite-bearing granite with a moderate gneissosity defined by aligned

164   biotite, but it locally shows a granodioritic or leucocratic facies. Accessory garnet and fibrous

165   sillimanite are reported, with pinitized cordierite observed in the leucocratic facies (Nureki

166   1974; Higashimoto et al. 1983). The contact with metasedimentary rocks gently dips towards

167   the east; isolated granite outcrops within the metasedimentary succession (Kfs−Crd zone; not

168   shown in Fig. 1) suggest that the Namera granite extends eastwards and lies a few meters

169   below the metasediments. Peak $P-T$ conditions of regional metamorphism in the neighboring

170   gneissic rocks are estimated at 2−3 kbar/590−675 °C (Ikeda 2004). For a biotite-granite

171   sample, U−Pb dating of oscillatory-zoned zircon by LA-ICP-MS yielded $^{206}Pb/^{238}U$ dates

172   between 107 and 100 Ma, with 105 ± 2 Ma being regarded as the age of granite intrusion

173   (Skrzypek et al. 2016). Monazite CHIME dates for three granite samples cluster at 91−89.9 ±

174   3.2 Ma and are also ascribed to granite intrusion (Suzuki et al. 1996). A K−Ar biotite date of

175   86.6 ± 4.3 Ma was obtained from a hornblende-biotite granite sample (Higashimoto et al.

176   1983).

177   **ANALYTICAL METHODS**

178   Two samples were collected for each granite and renamed for simplicity. The main samples

179   for the Shimokuhara granite (EY99A, renamed SHK) and Namera granite (EY117A, renamed

180   NAM) are the same as those described by Skrzypek et al. (2016) for U−Pb zircon dating.

181   They are used for both petrological descriptions and monazite geochronology. These samples

182   located close to the pluton margins were selected to explore the effects of post-emplacement

183   metamorphism whose $P-T$ conditions are recorded in the adjacent metamorphic rocks. The

184   additional samples (EY152A, renamed SHK2; EY140A, renamed NAM2) were collected

185   slightly farther from the pluton margin (100−300 m) and are used for comparing monazite

186   textures and compositions. Sample locations are indicated on Figure 1.

187   Mineral compositions were analyzed at Kyoto University and at the University of

188   Graz. In order to use a low beam current (500 pA), chlorite was measured on a scanning

189   electron microscope (Hitachi S3500H, Kyoto) equipped with an energy-dispersive system

190   (EDS) at 20 kV. All other minerals were measured with a JEOL JXA-8105 (Kyoto) and a

191   JEOL JXA-8530F Plus (Graz) electron microprobe (EMP) at 15 kV and with a 3μm beam

192   diameter. Analyses of all silicates except Aln were done at 10 nA and corrected with the ZAF

193   method, while those of REE±Th±U-bearing minerals were done at 150 nA and corrected

194   following the Bence and Albee (1968) method with the updated α-factor table of Kato (2005).

195   The analytical procedure for REE±Th±U-bearing minerals followed published guidelines (e.g.

196   Scherrer et al. 2000; Pyle et al. 2005; Suzuki and Kato 2008) and is available as Supplemental

197   Material together with the full set of monazite analyses. The REE measured for most

198   monazite analyses are La, Ce, Pr, Nd, Sm and Gd, but several analyses with the additional

199   HREE Tb to Yb were performed for Mnz−Xtm thermometry. Monazite [$A_4(TO_4)_4$] is

200   described according to A-site (Ca^{2+}, REE^{3+}, Y^{3+}, Th^{4+}, U^{4+}, Pb^{2+}) and T-site (P^{5+}, Si^{4+})

201   cations. Its composition is expressed according to the end-members X_{mnz} [($\Sigma_{REE}+Y$)/A], X_{cher}

202   [2Ca/A] and X_{hut} [(Th+U+Pb-Ca)/A], where A represents the sum of A-site cations. For

203   Mnz−Xtm thermometry, $X_{(HREE,Y)}$ in monazite is defined as [($\Sigma_{HREE}+Y$)/A], with Σ_{HREE}

204   including elements from Gd to Yb. For EMP age dating of the Th−U phases, the accuracy was

205   checked with several analyses of "ekanite" from Okkampitiya, Sri Lanka ($^{207}Pb/^{206}Pb$

9   9

206   age=562.1 ± 0.8 Ma; Nasdala et al. 2017) that yielded a weighted average Th−U−Pb age of

207   564 ± 6 Ma (n=5; MSWD=0.15).

208   Several calibrations are available for Mnz−Xtm thermomety. While the calibration of

209   Seydoux-Guillaume et al. (2002) considers the effect of Th, it is derived for a relatively high

210   temperature range (800−1100 °C). We prefer to use the calibrations of Heinrich et al. (1997)

211   and Pyle et al. (2001) as they rely on natural samples for a lower temperature range (400−700

212   °C). These two calibrations differ, probably because they are based on samples from different

213   metamorphic gradients, and we use both to obtain the range of likely temperature conditions

214   associated with the growth of Mnz−Xtm pairs in direct contact.

215   Electron microprobe element mapping of thin sections (~ 2 x 3 cm areas) was

216   performed at 15 kV, 60 nA, 10 μm beam diameter, 15 μm step size and 4 ms dwell time for

217   the CaKα, AlKα, PKα, FeKα, NaKα lines and the BSE signal. An in-house Matlab© routine

218   was used to identify the different minerals based on image analysis of the combined maps,

219   and to calculate modal abundances (Table 1). The granite mesonorm (Table 1) was calculated

220   following the procedure of Mielke and Winkler (1979) implemented in GCDkit (Janoušek et

221   al. 2006) using whole-rock analyses of Skrzypek et al. (2016) and with all Fe expressed as

222   FeO so as to exclude normative magnetite, since both granites belong to the ilmenite series of

223   Ishihara (1977).

224   Monazite dating by LA-ICP-MS was performed at Gakushuin University, Tokyo, on

225   separated grains mounted in epoxy, using a NWR-213 Nd:YAG laser with a 10 μm spot

226   diameter. The ablated material was carried to an Agilent 8800 single-collector tandem

227   quadrupole ICP-MS for monitoring ^{204}(Pb+Hg), ^{206}Pb, ^{207}Pb, ^{208}Pb, ^{232}Th^{16}O and ^{238}U^{16}O.

228   Fifteen analyses of unknowns were bracketed by five measurements of 44069 monazite

229   (Aleinikoff et al. 2006) and three of NIST SRM 610 glass (Pearce et al. 1997) for

230   normalization. Namaqualand monazite (1034.4 ± 5.8 Ma concordia age by LA-sector field-

231   ICP-MS; Gonçalves et al. 2016) was used as a consistency standard and concordant data

232   yielded the following weighted average ages (n=3 concordant analyses out of 5 sequences):

233    $^{206}Pb/^{238}U$ age=1061 ± 39 Ma (MSWD=1.40), $^{207}Pb/^{235}U$ age=1072 ± 47 Ma (MSWD=3.20),

234    $^{206}Pb/^{207}Pb$ age=1092 ± 91 (MSWD=5.20). Error calculations used the standard error (SE, 2σ

235   level) in order to take into account the large number of primary reference material analyses.

236   Considering the typical uncertainty on Pb/U ratios, monazite analyses with a concordance

237   ratio [U−U cc(%)=($^{206}Pb/^{238}U$ date)/($^{207}Pb/^{235}U$ date)*100] between 97 and 103% are referred

238   to as concordant. The results and full analytical procedure are available as Supplemental

239   Material. Data were plotted with Isoplot 3.76 (Ludwig 2004).

240   **SAMPLE DESCRIPTION**

241   **Samples SHK & SHK2**

242   The main Shimokuhara granite sample (SHK) was collected at the eastern margin of the

243   pluton, a few meters away from the contact with schistose rocks from the Bt zone (Fig. 1).

244   There, the Shimokuhara pluton exhibits a leucocratic, biotite-bearing facies with some K-

245   feldspar phenocrysts (~2 cm in length), plagioclase and quartz aggregates with a noticeable

246   rounded shape (Fig. 2a). Centimeter-scale xenoliths of the neighboring metasediments and

247   small garnet (< 1 mm in diameter) are also observed in the granite. Sample SHK is massive

248   and contains Qtz, Kfs, Pl, Bt with accessory Grt, Ap, Zrn, Mnz, Xtm and secondary Ms, Chl,

249   Ep, Aln, Ttn.

250   In thin section, large (up to 1 cm in diameter) and rounded aggregates of quartz with

251   undulatory extinction dominate (Fig. 2a). K-feldspar phenocrysts ($Or_{87-93}Ab_{7-13}$) contain

252   albite rods and locally show the microcline twin. The core of euhedral and concentrically

253   zoned plagioclase ($An_{18-47}Ab_{52-81}Or_1$) is replaced by fine-grained brownish material, probably

254   damourite (Fig. 2a, c). Biotite flakes (X_{Fe}=0.68−0.71; F=0.27−0.68 wt%) are interstitial and

11   11

255   variably replaced by chlorite (brunsvigite, X_{Fe}=0.67−0.70), K-feldspar, muscovite

256   (F=0.27−0.40 wt%) and titanite (Fig. 2c). Garnet (up to 400 μm in diameter) is included in K-

257   feldspar phenocrysts and has virtually no inclusion. No mineral preferred orientation is visible

258   under the optical microscope (Fig. 2a).

259   Sample SHK2 was collected 300 m from the mapped contact with metasedimentary

260   rocks. It is a massive granite with Qtz, Kfs, Pl, Bt and accessory Chl, Ilm, Rt, Mnz, Xtm, Zrn,

261   Aln, Ap. Biotite chloritization appears to be less pervasive than in sample SHK.

262   **Samples NAM & NAM2**

263   The main Namera granite sample (NAM) was collected less than a hundred meters away from

264   the contact with migmatitic gneissose rocks from the Kfs−Crd zone (Fig. 1). There, the

265   Namera pluton shows a leucocratic, fine- to coarse-grained biotite-bearing facies and locally

266   contains muscovite, garnet, greenish pinitized cordierite and sillimanite. Several outcrops in

267   the eastern part of the pluton show a weak to moderate gneissosity defined by biotite

268   alignment, but an undeformed facies resembling the Shimokuhara granite is also observed

269   (Fig. 1). The granite is locally cut by cm-thick quartz veins, garnet-bearing aplitic veins and

270   Qtz−Kfs−Ms pegmatites, the latter being nearly parallel to the gneissosity. The outcrop of

271   sample NAM shows cm-thick, garnet-bearing leucocratic veins mostly parallel and locally

272   oblique to the gneissosity. A NE−SW striking fault is mapped in the vicinity of the sampling

273   locality, but no signs of it are visible on the outcrop. Sample NAM has a moderate gneissosity

274   (Fig. 2b) and contains Qtz, Kfs, Pl, Bt with accessory (pinitized) Crd, Grt, Ap, Zrn, Mnz, Xtm

275   and secondary Ms, Chl, Hem.

276   K-feldspar phenocrysts (Or$_{80-90}$Ab$_{10-20}$) exhibit a straight to wavy outline, are locally

277   perthitic and contain albite stringlets. Plagioclase (An$_{05-39}$Ab$_{60-94}$Or$_{01-02}$) is found as large

278   crystals (up to 0.5 cm in length), small polygonal grains (~100 μm in diameter) or in fine-

279   grained intergrowths with secondary muscovite. Plagioclase locally shows an irregular outline

12   12

280   and its core is replaced by skeletal muscovite and minor biotite (Fig. 2b). Quartz shows

281   undulatory extinction and forms slightly elongated ribbons parallel to the gneissose fabric

282   (Fig. 2b). Biotite (X_{Fe}=0.59−0.63; F=0.34−0.56 wt%) appears as equant flakes or elongate

283   laths that define the gneissosity, and is slightly replaced by chlorite (brunsvigite, X_{Fe}=0.63)

284   (Fig. 2b). Garnet (up to 200 μm in diameter) is included in plagioclase. Pools of fine-grained

285   plagioclase and Ms−Bt intergrowths surround minute relicts of pinitized cordierite (200 μm in

286   length). The textural position of both garnet and former cordierite suggests they are of

287   magmatic origin. Secondary muscovite (F=0.17−0.29 wt%) is preferentially developed inside

288   K-feldspar where it forms a network of Ms−Qtz veins (Fig. 2d). A few myrmekites are

289   observed at Kfs−Pl contacts.

290   Sample NAM2 is a gneissose granite from an outcrop where no leucocratic veins or

291   pegmatites are observed. It contains Qtz, Kfs, Pl, Bt and accessory Grt, pinitized Crd, Ms,

292   Chl, Mnz, Xtm, Zrn, Ap. The gneissose fabric is slightly stronger than in sample NAM.

293   **Comparison of mineral modes**

294   The whole-rock data published by Skrzypek et al. (2016) show that, despite a different

295   appearance, samples SHK and NAM have a similar major and trace element composition. For

296   further comparison, mineral modal abundances were calculated using whole-rock analyses

297   and EMP mapping of thin sections (Table 1). Granite mesonorm calculations indicate a

298   similar amount of Kfs, Qtz, Pl and Bt in SHK and NAM (Table 1). Element mapping of ~3.5

299   x 2 cm thin sections fails to accurately determine the modal amount of phenocrystic phases

300   (Kfs, Qtz, Pl), but can be used for the smaller phyllosilicates. It reveals that the modal amount

301   of Bt in NAM is almost equal to that of Bt plus Chl in SHK (Table 1). Since textural

302   observations indicate that chlorite forms at the expense of biotite, it can be deduced that the

303   original amount of biotite in both samples was identical, as suggested by the mesonorm

304   calculations.

13   13

305   **PETROLOGY OF MONAZITE AND ASSOCIATED PHASES**

306   **Sample SHK**

307   Monazite forms euhedral to skeletal grains (~50 to 250 µm in length) with generally few

308   inclusions (Fig. 3a). It is located inside biotite or more commonly at the boundary between

309   biotite and quartz/plagioclase. Monazite grains that were separated from the hand specimen

310   are mostly euhedral, and monazite observed in thin section tends to be more subhedral to

311   anhedral. Large euhedral monazite has micro-cracks along which allanite, hematite and

312   xenotime are observed (Fig. 3b, f), and is surrounded by apatite and a fringe of zoned allanite

313   that becomes REE-poorer towards the rim (Fig. 3b, c). Skeletal monazite is surrounded by

314   large allanite (~100 µm), small Th−U phases (< 30 µm in diameter) and minor xenotime (Fig.

315   3c). Elongate grains of oscillatory-zoned allanite are found at chlorite rims and are locally

316   separated from chlorite by a titanite film (Fig. 3d). In general, monazite occurs together with

317   allanite in the vicinity of chloritized biotite. Xenotime has two modes of occurrence; lone

318   matrix grains form euhedral overgrowths around zircon, whereas tiny anhedral xenotime (<

319   30 µm in length) occurs next to or in contact with altered monazite (Fig. 3b, c, f).

320   Primary and secondary monazite domains are distinguished in both *in situ* and

321   separated grains. Primary domains have sector- or oscillatory-zoned textures that occupy the

322   majority of most monazite grains (Fig. 3e). The primary domains may be subdivided into (1)

323   core domains with a juxtaposition of BSE-bright and BSE-dark sectors, and (2) thin (< 20 µm

324   in width) and BSE-dark outermost rim domains (Fig. 3e). The primary core and outermost rim

325   domains are differentiated by a sharp contrast on BSE images (Fig. 3e). Secondary domains

326   correspond to altered parts of the primary monazite zoning. They are generally associated

327   with the presence of cracks that cross-cut the primary zoning; on BSE images they appear as

328   darker zones up to ~20 µm in width, or as brighter and thin (< 5 µm) veinlets (Fig. 3f).

329   Monazite-(Ce) compositions are dominated by a huttonite-type substitution [REE^{3+} +

330    $P^{5+} \leftrightarrow (Th,U)^{4+} + Si^{4+}$] (Fig. 4a). The analyses define a trend with an evolution towards the

331   monazite end-member from the primary core (X_{hut}=0.02−0.13; X_{cher}=0.05−0.10;

332   X_{mnz}=0.78−0.92; 48 analyses on 36 grains) to the outermost rim (X_{hut}=0.01−0.03;

333   X_{cher}=0.05−0.07; X_{mnz}=0.91−0.93; 9 analyses on 9 grains) domains (Fig. 4a, Table 2). This

334   evolution is due to Si + U + Th decrease and P + Light REE (LREE) increase while Y and

335   HREE remain constant. Analyses of the secondary domains follow the huttonitic trend

336   defined by primary domain analyses, with some excess in U ± Th ± Pb being responsible for a

337   slight deviation (Fig. 4a). Secondary domain analyses define two subgroups based on Y and U

338   contents; one shows markedly low Y (Y_2O_3=0.55−1.03 wt%) and U (UO_2 < 0.15 wt%),

339   whereas the other has Y (Y_2O_3=1.65−3.27 wt%) and U (UO_2=0.32−1.22 wt%) contents in the

340   range of primary domain analyses (Fig. 4b). With respect to primary domains, the Y-poor

341   subgroup of secondary domains is depleted in Y, U, Th, Ca and enriched in both LREE and

342   HREE.

343   Allanite analyses yield low oxide sums (94−97 wt%) that are ascribed to

344   metamictization, since Th-richer spots tend to have lower totals. The composition is mainly

345   that of allanite-(Ce) with recalculated Fe^{3+}/Fe_{tot} ratios of 0.04−0.32 (normalization to 25

346   negative charges) pointing to some feriallanite component, and Al_{tot} (1.99−2.27 a.p.f.u.)

347   indicating some proportion of the Ep-Czo component. The analyses of BSE-bright cores show

348   Σ_{REE} between 0.71 and 0.80 a.p.f.u., negligible actinide contents (< 0.02 a.p.f.u.) and about

349   0.08−0.14 wt% fluorine. One spot from a darker rim overgrowth has a Σ_{REE} of 0.54 a.p.f.u.,

350   actinide abundances below detection limits and about 0.05 wt% fluorine.

351   Lone matrix xenotime (observed in thin section) is enriched in U and Er, Tm, Yb (i.e.

352   HREE with an ionic radius close to that of U), whereas tiny xenotime occurring in monazite

353   alteration textures (separated monazite grains) is enriched in Gd, Tb, Dy (i.e. HREE with an

15   15

354 ionic radius close to that of Th) and slightly in Ce and Th. When in direct contact with

355 monazite, xenotime is always adjacent to the BSE-dark, Y-poor secondary domains (Fig. 3f).

356 Six analyses of Y-poor secondary monazite domains in contact with xenotime show a nearly

357 constant $X_{(HREE,Y)}$ (0.060 ± 0.006) and yield temperature estimates of 471 and 562 °C with the

358 calibration of Pyle et al. (2001) and Heinrich et al. (1997), respectively.

359   The tiny Th−U phases (< 30 μm in diameter) are not easily quantified but valid

360 analyses (not influenced by the surrounding minerals) yield a constant oxide sum of about 90

361 wt% (Table 2). They correspond to Th-rich (ThO$_2$=51.67−62.77 wt%) silicates with

362 significant amounts of U (UO$_2$=2.14−12.82 wt%), P (P$_2$O$_5$=0.71−9.21 wt%) and Y

363 (Y$_2$O$_3$=0.84−3.40 wt%). Their formula approaches (Th, U, REE)(Si, P)O$_4$ with the missing 10

364 wt% being ascribed to free H$_2$O (Table 2); they can be considered as a monoclinic (huttonite;

365 note that monazite is also monoclinic) or tetragonal (thorite) polymorph of (hydrous) ThSiO$_4$

366 (Fig. 4c). Electron microprobe age dating of the Th−U phases was attempted due to their high

367 actinide content. The cores of euhedral grains (~30 μm in diameter) included in a K-feldspar

368 phenocryst yield Th−U−Pb dates of 369 ± 6 and 262 ± 7 Ma, whereas the rims show much

369 younger dates of 23−21 ± 3 Ma. These inclusions are, however, not related to monazite and

370 their ages are not discussed further. The other, generally smaller (< 20 μm) Th−U phases

371 occur in the matrix and are always associated with allanite around anhedral monazite; one

372 grain preserves a Th−U−Pb date of 58 ± 5, three grains yield 15−14 ± 2−3 Ma and four other

373 grains have PbO below the detection limit.

374 **Sample SHK2**

375 Large (100 μm in diameter) and euhedral monazite grains surrounded by a thin fringe of

376 Aln−Ap with local xenotime and Th−U phases are common, but skeletal aggregates

377 associated with more abundant apatite are also observed (see Supplemental Material for BSE

378 images). Monazite mostly preserves a sector zoning pattern that is affected by micro-cracks

16   16

379   and locally riddled with tiny holes. One euhedral and zoned allanite (~ 500 µm in length) is

380   entirely included in a plagioclase grain within a Kfs phenocryst and is regarded as primary in

381   origin. Other allanite occurrences correspond to thin (~ 50 µm in diameter) laths next to

382   monazite or along the cleavages of chloritized biotite. Large matrix xenotime (~ 100 µm in

383   diameter) shows oscillatory zoning cross-cut by BSE-dark zones. Overall, the textures are

384   similar to those observed in SHK, with more occurrences of large euhedral monazite that

385   might be correlated with the lesser degree of chloritization.

386   Monazite domains similar to those distinguished in SHK are identified (Fig. 4a, b).

387   The primary domains show a dominant huttonitic substitution and an increase in the monazite

388   component from the core (X_{hut}=0.04−0.15; X_{cher}=0.03−0.05; X_{mnz}=0.80−0.92) to the outermost

389   rim (X_{hut}=0.04−0.07; X_{cher}=0.03−0.04; X_{mnz}=0.89−0.94). A few BSE-dark secondary domains

390   could be analyzed and their composition mostly falls in a subgroup with markedly low Y

391   (Y_2O_3 < 0.77 wt%) and U (UO_2 < 0.05 wt%) contents. Only one valid analysis of a Th−U

392   phase could be obtained. Like in SHK, it is a Th-rich silicate (Fig. 4c) with a somewhat higher

393   amount of REE (ΣREE_2O_3=15.30 wt%) and PbO below the detection limit.

394   **Sample NAM**

395   Monazite occurs as subhedral, rounded or anhedral grains (~20 to 250 µm in length) with a

396   variable amount of inclusions (Fig. 5a). Monazite is found inside biotite, plagioclase and at

397   the boundary between major minerals. Tiny zircon in contact with monazite is common, and

398   only one relatively large (~100 µm in diameter) and oscillatory-zoned xenotime is found in

399   textural equilibrium with monazite in the matrix. Other xenotime occurrences are scarce,

400   minute inclusions in monazite. Micro-cracks are rare and are not associated with extensive

401   alteration of the original zoning (Fig. 5c, f).

402   Two contrasting domains are observed in both *in situ* and separated monazite. All

403   grains exhibit a primary sector- or oscillatory-zoned part that is randomly cross-cut and

404   replaced by a secondary BSE-dark and inclusion-rich domain (Fig. 5c, e, f). The proportion of

405   primary to secondary domains varies; only a few grains fully preserve their primary zoning

406   (Fig. 5d), most grains exhibit a juxtaposition of both domains, and some grains are dominated

407   by secondary domains in which only patches of primary domains remain (Fig. 5c, e, f). In

408   general, the secondary domains are preferentially developed in monazite core (Fig. 5e) but

409   examples of replacement occurring at the rim are also found (Fig. 5f).

410   Primary domains are nearly inclusion-free, whereas secondary domains are riddled

411   with inclusions that range from a few microns up to ~50 μm in diameter (Fig. 5a, e, f). The

412   inclusions occur as single grains or more commonly form multi-phase aggregates with a

413   rounded or tubular shape (Fig. 5g, h). A survey of 906 inclusions in 84 grains reveals the

414   presence of, in order of decreasing abundance: Zrn, Ms, Kfs, Qtz, Hem, Bt, Po, Pl and scarce

415   Chl, Ap, Xtm and Ant (Fig. 6; Supplemental Material). Among these inclusions, Zrn, Ms, Kfs,

416   Qtz, Bt, Po, Pl, Ap and Xtm are thought to have formed during the magmatic or metamorphic

417   stage while Hem, Chl and Ant are considered as later alteration products. The most common

418   assemblages are Ms−Qtz and Qtz−Kfs−Pl±Bt. A significant number of Si-rich inclusions is

419   also observed (Fig. 5g, h); they appear as orange to brownish blebs under the optical

420   microscope, and their EDS spectrum shows a prominent Si peak with additional Al, K, Na, Ca

421   or Fe peaks (Fig. 6a). They are easily damaged during Raman analysis and their spectrum

422   does not reveal any discernible vibration band outside that of the monazite host (Fig. 6b).

423   Primary and secondary domains are clearly distinguished based on their compositions

424   (Fig. 4, Table 2). Analyses from primary domains follow a dominant huttonitic substitution

425   (X_{hut}=0.01−0.07; X_{cher}=0.06−0.14; X_{mnz}=0.80−0.94; 41 analyses on 29 grains), whereas those

426   from secondary domains cluster on a narrow cheralite−monazite exchange trend (X_{hut}=0−0.02;

427   X_{cher}=0.10−0.15; X_{mnz}=0.84−0.89; 29 analyses on 23 grains; Fig. 4a). With respect to primary

428   domains, secondary domains are depleted in Si, LREE and enriched in Ca, Y, U, P, HREE

429   while Th remains constant. The strongest contrast is seen in the Y and U contents (Fig. 4b);

430   they are on average low in primary domains (Y_2O_3=0.91 wt%; UO_2=0.19 wt%) and

431   significantly higher in secondary ones (Y_2O_3=2.30 wt%; UO_2=1.43 wt%).

432   Compared to the large matrix xenotime grain, minute xenotime included in monazite

433   is enriched in La, Ce and depleted in U, Er, Tm, Yb. Xenotime inclusions in monazite, like

434   other minerals, always occur inside the secondary domains. Five analyses of secondary

435   monazite domains in contact with included xenotime show a slightly varying $X_{(HREE,Y)}$ (0.089

436   ± 0.018) and yield temperature estimates of 584 and 664 °C with the calibration of Pyle et al.

437   (2001) and Heinrich et al. (1997), respectively.

438   **Sample NAM2**

439   Monazite is found inside Bt, Kfs or between major minerals; it mostly has a euhedral and

440   slightly rounded shape, but skeletal grains are locally present (see Supplemental Material for

441   BSE images). Monazite shows oscillatory or sector zoning cross-cut by BSE-dark zones with

442   tiny holes or inclusions. Only a few inclusions in monazite could be identified in this thin

443   section; these are Zrn, Po, Ms, Xtm, Chl, Hem, Bt, Qtz and K−Al-bearing Si-rich phases. Tiny

444   (< 30 μm in length) grains of skeletal allanite are locally found in between the major minerals

445   or along biotite cleavages. Euhedral to anhedral matrix xenotime is common and shows sector

446   or oscillatory zoning cross-cut by BSE-dark zones next to chloritized biotite. Monazite

447   textures are globally similar to those observed in sample NAM, but one difference is the

448   limited occurrence of tiny allanite in NAM2.

449   Monazite analyses reveal the same contrast between primary and secondary domains

450   as observed in NAM (Fig. 4a, b). Primary domain analyses follow a huttonitic exchange

451   (X_{hut}=0.02−0.08; X_{cher}=0.02−0.06; X_{mnz}=0.86−0.97) and have relatively low Y and U contents

452   (Y_2O_3=0.42−1.90 wt%; UO_2=0.04−0.38 wt%). Conversely, secondary domain analyses

19   19

453   indicate a cheralitic exchange (X_{hut}=0−0.04; X_{cher}=0.01−0.11; X_{mnz}=0.85−0.99) and are

454   commonly enriched in Y and U (Y_2O_3=1.49−3.03 wt%; UO_2=0.15−3.26 wt%).

MONAZITE GEOCHRONOLOGY

455

Sample SHK

456

457   Twenty-six analyses were performed on fifteen grains, with two spots (16.4/16.6) located on

458   cracks being rejected due to severe discordance. The remaining 24 analyses are located on

459   primary core (15 spots, U/Th ≤ 0.03) and outermost rim (9 spots, U/Th > 0.03) domains, and

460   13 are concordant. Concordant core analyses show $^{206}Pb/^{238}U$ dates spreading from 116 ± 3 to

461   98 ± 3 Ma (106 Ma on average; Fig. 7a, b) with no clear maximum in the probability density

462   distribution. The three concordant spots from the outermost rim parts yield a weighted

463   average $^{206}Pb/^{238}U$ date of 102 ± 2 Ma (MSWD=1.13, n=3), which is similar to the lower

464   intercept date tentatively calculated with all outermost rim analyses in an inverse Concordia

465   diagram (102 ± 2 Ma, MSWD=2.2, n=9).

Sample NAM

466

467   Twenty-six analyses were performed on nineteen grains, with two severely discordant spots

468   (19.8/19.13) being discarded as they are affected by the presence of an inclusion or located at

469   a grain edge. The remaining 24 analyses are clearly differentiated based on their U/Th ratio;

470   they belong to primary (11 spots, U/Th≤0.01) and secondary (11 spots, U/Th≥0.10) domains,

471   two spots (20.14/20.15) with an intermediate U/Th ratio (0.02−0.03) being considered as

472   analyses of mixed domains. Primary domain analyses show a significant degree of

473   discordance and their $^{206}Pb/^{238}U$ dates spread between 116 ± 6 and 101 ± 6 Ma, with only one

474   concordant spot (19.15) at 106 ± 5 Ma (Fig. 7d). The latter result agrees with the weighted

475   average $^{206}Pb/^{238}U$ date of 109 ± 2 (MSWD=2.00, n=11) calculated with all primary domain

476   analyses. Secondary domain analyses have a better concordance; 10 spots out of 11 are

20   20

477 concordant and their ^{206}Pb/^{238}U dates spread from 102 ± 3 to 91 ± 2 Ma (Fig. 7d). No clear

478 maximum arises from this spread, and no correlation between ^{206}Pb/^{238}U date and U/Th ratio

479 is apparent. For example, the two spots with the highest U/Th ratios yield the oldest (19.3:

480 102 ± 2 Ma, U/Th=0.21) and youngest (19.11: 91 ± 2 Ma, U/Th=0.21) ^{206}Pb/^{238}U dates of the

481 distribution (Fig. 7c).

482 **MONAZITE ALTERATION PROCESSES**

483 The SHK and NAM granite samples have, despite contrasting textures (Fig. 2a, b), nearly

484 similar mineral modes (Table 1), whole-rock compositions, U−Pb zircon ages (Skrzypek et al.

485 2016) and U−Pb monazite ages for primary domains (Fig. 7b, d). These similarities are used

486 to demonstrate that the different monazite alteration textures are not due to initial

487 compositional differences, but to the respective $P-T$ and chemical conditions of the

488 metamorphic overprintings that affected the granites.

489 **Monazite replacement at low to medium temperature (Shimokuhara granite)**

490 Monazite in SHK and SHK2 mostly consists of primary domains for which sector or

491 oscillatory zoning and compositional variations along a single exchange trend suggest a

492 magmatic origin (Figs. 3e and 4a). However, numerous observations indicate alteration of

493 primary monazite. The change from a euhedral to a skeletal shape points to a variable degree

494 of monazite dissolution. Allanite, apatite, xenotime and Th−U phases − i.e. secondary

495 products that partly use elements originally contained in monazite − develop directly adjacent

496 to altered monazite, along its margins or cracks (Fig. 3b, c, f). Euhedral allanite is also found

497 farther from monazite along chlorite cleavages (Fig. 3d), indicating some REE mobility since

498 monazite is the main, primary LREE-carrier in the rock. Monazite replacement by Aln−Ap is

499 dominant, but recrystallization is additionally observed in the secondary, porous domains that

500 developed in cracks or at the rim of some grains (Fig. 3b, f).

21   21

501   The formation of hydrous phases such as allanite and chlorite next to altered monazite,

502   and indications of (limited) element transport strongly suggest that monazite replacement

503   involved the action of an aqueous fluid. The secondary domains also bear similarities with the

504   textures reported for monazite alteration experiments in the presence of fluids (e.g.

505   Hetherington et al. 2010). In several of these experiments the composition of secondary

506   monazite shows LREE enrichment and Y, Ca, Th, U depletion (Harlov et al. 2011; Williams

507   et al. 2011; Grand'homme et al. 2018), as observed for the Y-poor subgroup of secondary

508   domains (Fig. 4b). The effects of aqueous fluid on monazite dissolution were investigated for

509   a wide range of $P-T$ conditions and fluid compositions (e.g. Oelkers and Poitrasson 2002;

510   Schmidt et al. 2007; Hetherington et al. 2010; Budzyń et al. 2011; 2015; 2017; Harlov et al.

511   2011; Williams et al. 2011; Grand'homme et al. 2018; Trail 2018). The main outcome of

512   experimental studies is that alkali-bearing fluids (NaOH, KOH) more efficiently lead to

513   dissolution and recrystallization, while observations of natural samples insist on the role of

514   halogens in enhancing monazite dissolution (e.g. Förster 2001; Didier et al. 2013).

515   In addition to H_2O, a source of Ca (±Al, Si) has to be identified to explain the growth

516   of Aln−Ap after monazite (Fig. 3b, c). In a similar case of magmatic monazite replacement in

517   S-type granitoids, Broska et al. (2005) proposed plagioclase (for Ca) and biotite (for Al, Si, F)

518   as likely sources. In sample SHK biotite and the Ca-rich core of large plagioclase are

519   extensively altered (Fig. 2a, c), and biotite chloritization could have released some F

520   (~0.27−0.68 wt%) that was subsequently incorporated in allanite (~0.08−0.14 wt%). To

521   explain monazite replacement by Aln−Ap in granitoids, Broska and Siman (1998) proposed a

522   reaction that involves anorthite breakdown but not biotite chloritization. Following the

523   minerals observed in monazite alteration textures, we use the mineral compositions measured

524   in SHK to present a stoichiometrically-balanced reaction that tries to conserve most elements

525   (except HREE−Y, Th−U and F; see Supplemental material for the calculation procedure). The

526   main uncertainty in our approach lies in the elements that potentially arrived and escaped with

527   the fluid (+Ca, Si; -F, Na, K). The proposed reaction is: 1 Mnz + 5.462 Pl + 4.041 Bt + 3.828

528   Qtz + 3.078 H_2O + 1.148 Ca^{2+} → 1.095 Aln + 0.316 Ap + 1.317 Chl + 1 Ttn + 3.268 Kfs + 1

529   Ms + 2.780 Ab (+ 0.013 HREE−Y + 0.089 Th−U + 0.300 F). The slight excess of HREE−Y

530   and Th−U can be accounted for by the occurrence of tiny xenotime and Th−U phases in the

531   monazite alteration textures (Fig. 3b, c, f), while the high proportions of Kfs and Ab

532   compared to those observed in the chloritization textures suggest that some K and Na are

533   instead stored in damourite or were lost in the fluid together with excess F. The major

534   alteration process can therefore be explained by a nearly balanced replacement reaction that

535   leaves magmatic monazite as a metastable relict.

536   The formation of the observed replacement textures requires an increase in Ca activity

537   and H_2O influx. At the magmatic stage, the first condition can be met and lead to monazite

538   replacement by Aln−Ap (Dini et al. 2004), but contemporaneous chlorite growth is unlikely.

539   Conversely, both conditions are satisfied by hydrothermal alteration for which plagioclase

540   resorption and biotite chloritization are commonly reported (e.g. Alderton et al. 1980;

541   Yuguchi et al. 2015). The results of Mnz−Xtm thermometry suggest that the rare secondary

542   domains in SHK formed at 471−562 °C. This range agrees with temperature estimates for the

543   regional overprint in the neighboring schist zone (450−550 °C; Ikeda 2004), but the upper

544   temperature limit conflicts with the widespread occurrence of chlorite. This could suggest that

545   limited monazite recrystallization actually occurred at lower-amphibolite facies conditions (>

546   500 °C), whereas later monazite replacement was due to a more pervasive hydrothermal

547   alteration of the granite at greenschist facies conditions (< 500 °C). Similar monazite

548   replacement textures were reported in rocks that experienced peak metamorphic temperatures

549   of 340 °C (Poitrasson et al. 2000), 550−600 °C (Finger et al. 1998) and up to 685−825 °C

550   (Broska et al. 2005). Although the $P-T$ conditions at which monazite replacement occurs are

551   likely controlled by the whole-rock Ca and Al contents (e.g. Spear, 2010), we propose that the

552   textures can hint at a low- or high-grade origin. The presence of chlorite, fractured monazite

553   and allanite growing farther from monazite would indicate the dominance of pervasive fluid

554   circulation at greenschist facies conditions, whereas the formation of Aln−Ap coronas should

555   reflect the increasing role of diffusion at higher metamorphic grade (Broska et al. 2005;

556   Finger et al. 1998; 2016; Ondrejka et al. 2012).

557   **Monazite recrystallization at high temperature (Namera granite)**

558   Monazite is the dominant LREE-rich phase observed in samples NAM and NAM2. It is

559   composed of sector- and oscillatory-zoned primary domains whose composition lies on the

560   same huttonitic trend as that defined by primary domain analyses from sample SHK (Figs. 4a

561   and 5d). This supports the magmatic origin of monazite in both samples. A significant part of

562   monazite is also occupied by secondary domains enriched in Ca, Y, U and P (Fig. 4). The

563   formation of secondary domains generally preserves the shape of the original magmatic grain,

564   and it is necessary to clarify whether this pseudomorphic recrystallization occurred at the

565   solid state or under the action of a fluid phase.

566   Primary and secondary domains are separated by a sharp boundary on BSE images,

567   and their composition is clearly distinct (Figs. 4 and 5c, e, f). This stands against

568   intracrystalline diffusion, which should result in more gradual changes in terms of texture (see

569   the diffusion zone imaged by Gardés et al. 2006) and composition (e.g. Cherniak et al. 2004).

570   In addition, BSE images show that recrystallization does not necessarily start from the rim

571   and proceed inwards, but that a preferential recrystallization of the monazite core is dominant

572   (Figs. 5e and 7c). The latter observation suggests that a mobile medium was able to make its

573   way up to the more unstable core, possibly through precursor cracks. Thus, the action of a

574   fluid phase is inferred to explain the selective dissolution of primary monazite. Dissolution

575   was selective; it is striking that the primary domains have a narrow compositional range,

576   although a larger spread might be expected by analogy with the composition of primary

577   monazite in sample SHK, which has the same bulk composition as sample NAM (Fig. 4a).

578   The primary domains that are presently observed must have been the most robust parts during

579   dissolution, whereas we suppose that the originally Th−U-richer monazite core was more

580   reactive, more porous to allow fluid circulation, hence preferentially dissolved. The different

581   solubilities of the initial domains could have been due to a higher degree of disequilibrium

582   with respect to the fluid, and/or to denser radiation damage in the supposedly Th−U-rich core.

583   Dissolution of the primary domains had to be followed by precipitation of secondary

584   monazite at exactly the same place in order to explain the pseudomorph texture (Fig. 5f). Both

585   characteristics are accounted for by the dissolution-reprecipitation process detailed by Putnis

586   (2002). A key aspect of this process is that a fluid-mineral interface has to be maintained so

587   that recrystallization can proceed (Putnis and Putnis 2007). In samples NAM and NAM2,

588   recrystallization corresponds to the transformation of primary huttonitic monazite into

589   secondary cheralitic monazite (Fig. 4a). According to the unit cell parameters of natural

590   monazite-(Ce) (Ni et al. 1995), huttonite (Taylor and Ewing 1978) and $Ca_{0.5}Th_{0.5}PO_4$ (Terra et

591   al. 2008), this transformation should result in a unit cell volume decrease of at least ~0.14%.

592   Although limited, the volumetric decrease indicates that a porosity could be maintained and

593   progressively filled by inclusions, the latter being distinctively abundant in the secondary

594   domains (Fig. 5a, e, f).

595   The major factor controlling the spatial and temporal extent of the dissolution-

596   reprecipitation process is the fluid composition (Putnis 2009). In the case of NAM and NAM2

597   samples, it was a fluid able to initially dissolve monazite and to reprecipitate not only new

598   monazite but also the whole range of inclusions that are found in the secondary domains (Zrn,

599   Ms, Kfs, Qtz, Bt, Po, Pl, Ap, Xtm). Similar inclusions are also reported in zircon from the

600   same sample (Qtz, Pl, Kfs, Ms, Bt, Hem, Ap, Xtm; Skrzypek et al. 2016). Because of these

25   25

601   constraints, the fluid medium is thought to be anatectic melt. Dissolution, transport and

602   reprecipitation of zircon, apatite or xenotime in the presence of alkaline fluids has been

603   reported (e.g. Tomaschek et al. 2003; Rubatto et al. 2008; Harlov 2011; Broom-Fendley et al.

604   2016), but at low pressure the solubility of both monazite and zircon is higher in granitic melt

605   than in aqueous fluid (e.g. Montel 1986; Watson 1979; Ayers and Watson 1991). In addition,

606   the rounded or tubular shape (Fig. 5g, h) and the nature of the most abundant multi-phase

607   inclusions in recrystallized monazite (Si-rich (melt?), Qtz−Kfs−Pl, Qtz−Ms assemblages)

608   strongly suggest that a hydrous granitic melt crystallized in the pores. A similar texture with

609   numerous and tiny multi-phase inclusions is reported for metamorphic monazite that grew in

610   the presence of leucogranitic melt (Bea and Montero, 1999).

611   The origin of the anatectic liquid that is thought to have favored monazite

612   recrystallization is debatable. Although small amounts of partial melt are difficult to

613   recognize, the Ms−Qtz network texture observed in thin section (Fig. 2d) might be evidence

614   of internal or external melt percolation. On the one hand, regional metamorphism in the

615   neighboring Kfs−Crd zone peaked at 2−3 kbar/590−675 °C (Ikeda 2004) and locally reached

616   the range of H_2O-present melting reactions for granitic compositions (e.g. Weinberg and

617   Hasalová 2015). The Namera granite might therefore have experienced a small degree of *in*

618   *situ* wet melting, which produced a liquid that could dissolve monazite, zircon and xenotime.

619   The latter mineral, which is scarcely observed in the matrix and relatively U-rich, represents a

620   likely source for U and Y enrichment in recrystallized monazite. On the other hand,

621   metasedimentary rocks in the Kfs−Crd zone show leucocratic layers that are ascribed to

622   partial melting (Brown 1998; Kawakami and Ikeda, 2003), and this external melt could

623   equally have infiltrated the Namera granite. The presence of undeformed Grt-bearing

624   leucocratic veins and Qtz−Kfs−Ms pegmatite on NAM outcrop supports this view, but a

625   similar observation is unfortunately not made on NAM2 outcrop. Melt provided by the

626   metasedimentary rocks would also have the advantage of being in greater disequilibrium with

627   the granitic assemblage, thereby enhancing monazite dissolution. It is therefore proposed that

628   monazite recrystallization occurred at 584−664 °C and was facilitated by the infiltration of

629   melt possibly derived from the neighboring metasedimentary rocks.

630   **CONSEQUENCES FOR MONAZITE GEOCHRONOLOGY**

631   **Preservation of magmatic age**

632   In monazite from sample SHK, the primary core and outermost rim domains show no

633   difference in ^{206}Pb/^{238}U date, and concordant rim analyses yield a weighted average ^{206}Pb/^{238}U

634   date of 102 ± 2 Ma (Fig. 7a, b). In sample NAM, the only concordant analysis of primary

635   monazite domain yields 106 ± 5 Ma (Fig. 7d). Both results are within error of the U−Pb

636   zircon ages of 105 ± 3 Ma and 105 ± 2 Ma derived from the same SHK and NAM samples,

637   respectively (Skrzypek et al. 2016). These age similarities, together with well-defined sector

638   and oscillatory zoning patterns, support a magmatic origin for both zircon and monazite, and

639   indicate that the age of magmatic crystallization is preserved in the primary monazite

640   domains. Previous works similarly reported that monazite may partly retain its magmatic age

641   in meta-igneous rocks metamorphosed up to ~700 °C, despite subsequent alteration (Finger et

642   al. 1998, 2016; Townsend et al. 2000). This view appears fully valid for sample SHK in which

643   the greenschist facies regional overprint did not lead to total monazite replacement, which

644   allows the magmatic age to be measured with a 10 μm laser spot. The magmatic age is less

645   well preserved in monazite from sample NAM, but is still accessible with the analytical

646   approach used in this study. The main observation is that the primary domain analyses are

647   commonly discordant (Fig. 7d). Given that such a discordance is not observed in SHK, it is

648   suspected that the upper-amphibolite facies metamorphic overprint was responsible for some

649   disturbance of the primary isotopic system in NAM. The BSE images of primary domains,

650   however, do not hint at a perturbation of the original, magmatic sector zoning (Figs 5 and 7c).

Identification of alteration age

652   Age dating of the secondary monazite domains in SHK was not attempted due to their small

653   size. In trying to analyze similar narrow veinlets by ion probe, Townsend et al. (2000) could

654   obtain nothing but mixed ages. Moreover, two spots located on micro-cracks show that severe

655   discordance is to be expected for the secondary domains (spot 16.6 in Fig. 7a). A few analyses

656   of the Th−U phases found next to altered monazite yield one Th−U−Pb date of 58 ± 5 and

657   three of 15−14 ± 2−3 Ma, with four nearly Pb-free spots (Fig. 4d). Because the Th−U phases

658   are tiny anhedral grains that are thought to host ~10 wt% H_2O, the probability of radiogenic

659   Pb loss is high and the Th−U−Pb dates should be treated with extreme caution. We can simply

660   remark that the oldest Th−U−Pb date (58 ± 5 Ma) agrees with the apatite fission track age of

661   57.4 ± 5 Ma that is reported for a granodiorite sample in the southern part of the Iwakuni-

662   Yanai area and ascribed to final exhumation of the Ryoke belt (Okudaira et al. 2001). It is

663   therefore possible that monazite replacement associated with chloritization actually occurred

664   during the Paleogene, when granitic rocks were ultimately brought closer to the surface.

665   The extensive recrystallization of secondary monazite in NAM gives the opportunity

666   to date the new domains *in situ*. However, the age of recrystallization is difficult to constrain

667   from the spread of concordant $^{206}Pb/^{238}U$ dates (102 ± 3 to 91 ± 2 Ma) that is obtained for

668   secondary domains (Fig. 7d). A first interpretation would be to consider all results as true

669   recrystallization ages, in which case a ca. 10 Ma duration could be ascribed to the whole

670   dissolution-reprecipitation process. A second interpretation is to consider that the results

671   reflect isotopic mixing between the primary and youngest secondary compositions. This view

672   is supported by several examples of Pb incorporation and incomplete isotopic resetting during

673   monazite recrystallization by dissolution-reprecipitation (Teufel and Heinrich 1997; Seydoux-

28   28

674   Guillaume et al. 2002, 2012; Grand'Homme et al. 2016). In such a case, the youngest

675   ^{206}Pb/^{238}U date of 91 ± 2 Ma can be considered as the minimum age for monazite

676   recrystallization.

Resolving the zircon-monazite age discrepancy and the age of high-*T* metamorphism

678   No discrepancy between U−Pb zircon (105 ± 3 Ma; Skrzypek et al. 2016) and monazite dates

679   (102 ± 2 Ma) is revealed for sample SHK, because monazite recrystallization remains limited.

680   The CHIME monazite dates of 87.3 ± 1.6 and 86.6 ± 2.1 Ma reported for the Shimokuhara

681   granite (Suzuki et al. 1996) were obtained from a sample collected at the northern edge of the

682   pluton, next to a fault zone (K. Suzuki, personal communication). The best way to explain

683   these younger CHIME results is to assume that monazite recrystallization was more

684   pronounced in that sample, so that the CHIME dates actually constrain monazite

685   recrystallization during regional metamorphism in the Ms-Crd zone (~500−550 °C; Ikeda,

686   2004). Conversely, dominant monazite replacement in SHK should be associated with lower

687   temperature chloritization (< 500 °C), and the Th−U−Pb dates obtained from Th−U phases

688   next to replaced monazite suggest that this alteration stage might be significantly younger

689   (Paleogene ?) and unrelated to regional metamorphism.

690   For sample NAM, the U−Pb age of secondary monazite domains (102−91 Ma) can be

691   up to ~10 Ma younger than that of magmatic monazite (106 ± 5 Ma) and zircon (105 ± 2 Ma;

692   Skrzypek et al. 2016). The comparison between our monazite analyses and those of Suzuki et

693   al. (1996) clearly shows that the CHIME dates (91−89.9 ± 3.2 Ma; Suzuki et al. 1996) were

694   calculated by grouping spots from primary and secondary domains. Because the secondary

695   domains are U-rich and thus have a lower analytical uncertainty, they presumably had more

696   weight in the calculation of the Th* vs. Pb isochron, which led to results of 91−90 Ma (Suzuki

697   et al. 1996). As confirmed by our combined petrological and geochronological data, the age of

29   29

698   ca. 91 Ma cannot be ascribed to magmatic crystallization but rather reflects monazite

699   recrystallization during a subsequent metamorphic overprint.

700   The present work proposes that the Kfs−Crd zone was still subjected to suprasolidus

701   conditions at ~ 91 Ma, i.e. about 10 Ma later than in the deeper Grt−Crd zone (U−Pb zircon

702   ages of 103−97 Ma; Nakajima et al. 2013; Skrzypek et al. 2018). It is another argument for

703   the diachronous origin of the metamorphic field gradient exposed in the Ryoke belt (Skrzypek

704   et al. 2016), and agrees with the persistence of medium-T conditions until 89 ± 5 Ma, as

705   suggested by U−Pb monazite ages obtained in a gneiss sample from the Kfs−Crd zone

706   (Skrzypek et al. 2018). The accumulation of petrochronological data reveals that in the Ryoke

707   belt − the type locality for low-P/high-T metamorphism − the age at which high-T conditions

708   ended seems to become younger with decreasing depth in the metamorphic section.

709   **CONCLUSIONS AND IMPLICATIONS**

710   We investigated two plutons for which the selected hand specimens appear as fresh, massive

711   or deformed biotite-granites. Under the microscope, the major minerals are not significantly

712   altered, except for plagioclase damouritization and variable biotite chloritization in the

713   massive granite, and the occurrence of Qtz−Ms veinlets and pinitization in the deformed

714   granite. However, the shape and/or composition of magmatic monazite is commonly altered in

715   both granites. The versatility of this accessory mineral offers a unique chance to constrain the

716   nature and timing of post-magmatic metasomatic events that are not easily detected otherwise.

717   *Monazite alteration textures – clues to metasomatic conditions*

718   The two plutons arguably have a similar composition, but differences in their post-magmatic

719   histories led to contrasting monazite alteration processes. Depending on the response of the

720   granitic assemblage to post-emplacement $P−T$ and chemical conditions, the interaction

721   between the metamorphosed major and accessory minerals and a fluid medium will result in

722   monazite replacement or recrystallization. On the one hand, the hydration of a granitic

723 assemblage at low temperature (< 500 °C) will trigger the alteration of Ca-plagioclase and the

724 chloritization of biotite. This will enrich the externally-derived aqueous fluid in Ca+Si+Al±F

725 and favor *monazite replacement* by allanite and apatite. We introduce a new reaction that

726 accounts for this process and that is expected in most monazite-bearing granitoids affected by

727 greenschist facies hydrothermalism. On the other hand, a limited amount of hydrous melt,

728 probably of external origin and percolating through the granitic assemblage, can trigger

729 *monazite recrystallization* by dissolution-reprecipitation at upper-amphibolite facies

730 conditions (> 600 °C). The recrystallized monazite domains are expected to host inclusions,

731 and their new composition will be controlled by the dissolution of other phases in the

732 presence of melt, with for instance Y and U enrichment reflecting xenotime resorption.

733 *Monazite alteration textures – a chance for dating the thermal history*

734 Age dating of monazite alteration textures can yield information on both the magmatic and

735 subsequent alteration stages. Monazite *replacement* leaves primary domains that show no

736 detectable modification of the isotopic system and preserve the age of magmatic

737 crystallization. The minute Th−U-rich alteration products of monazite can additionally yield

738 precise constraints on the timing of low-*T* metasomatism. Partial monazite *recrystallization*

739 causes a disturbance of the isotopic system in primary domains and limits the access to the

740 magmatic crystallization age. It also produces secondary domains that are large enough to be

741 analyzed, but in which an incomplete resetting of the primary isotopic signature complicates

742 the interpretation of the alteration age. The recrystallized domains nevertheless provide a

743 minimum age for the end of high-*T* conditions. Regardless of the process involved, monazite

744 alteration textures offer the possibility to date precise steps of a high- to low-*T*, post-magmatic

745 thermal evolution.

746 **ACKOWLEDGMENTS**

755 **REFERENCES CITED**

756 Alderton, D.H.M., Pearce, J.A., and Potts, P.J. (1980) Rare earth element mobility during

757   granite alteration: Evidence from southwest England. Earth and Planetary Science Letters,

758   49, 149–165.

759 Aleinikoff, J.N., Schenck, W.S., Plank, M.O., Srogi, L., Fanning, C.M., Kamo, S.L., and

760   Bosbyshell, H. (2006) Deciphering igneous and metamorphic events in high-grade rocks

761   of the Wilmington Complex, Delaware: Morphology, cathodoluminescence and

762   backscattered electron zoning, and SHRIMP U-Pb geochronology of zircon and monazite.

763   Geological Society of America Bulletin, 118, 39–64.

764 Ayers, J.C., and Watson, E.B. (1991) Solubility of apatite, monazite, zircon, and rutile in

765   supercritical aqueous fluids with implications for subduction zone geochemistry.

766   Philosophical Transactions - Royal Society of London, A, 335, 365–375.

767 Be Mezeme, E.B., Cocherie, A., Faure, M., Legendre, O., and Rossi, P. (2006) Electron

768   microprobe monazite geochronology of magmatic events: examples from Variscan

769   migmatites and granitoids, Massif Central, France. Lithos, 87, 276–288.

770   Bea, F., and Montero, P. (1999) Behavior of accessory phases and redistribution of Zr, REE,

771   Y, Th, and U during metamorphism and partial melting of metapelites in the lower crust:

772   an example from the Kinzigite Formation of Ivrea-Verbano, NW Italy. Geochimica et

773   Cosmochimica Acta, 63, 1133–1153.

774   Bence, A.E., and Albee, A.L. (1968) Empirical correction factors for the electron

775   microanalysis of silicates and oxides. The Journal of Geology, 76, 382–403.

776   Broom-Fendley, S., Styles, M.T., Appleton, J.D., Gunn, G., and Wall, F. (2016) Evidence for

777   dissolution-reprecipitation of apatite and preferential LREE mobility in carbonatite-

778   derived late-stage hydrothermal processes. American Mineralogist, 101, 596–611.

779   Broska, I., and Siman, P. (1998) The breakdown of monazite in the West-Carpathian Veporic

780   orthogneisses and Tatric granites. Geologica Carpathica, 49, 161–167.

781   Broska, I., Williams, C.T., Janák, M., and Nagy, G. (2005) Alteration and breakdown of

782   xenotime-(Y) and monazite-(Ce) in granitic rocks of the Western Carpathians, Slovakia.

783   Lithos, 82, 71–83.

784   Brown, M. (1998) Unpairing metamorphic belts: P-T paths and a tectonic model for the

785   Ryoke Belt, southwest Japan. Journal of Metamorphic Geology, 16, 3–22.

786   Budzyn, B., Harlov, D.E., Williams, M.L., and Jercinovic, M.J. (2011) Experimental

787   determination of stability relations between monazite, fluorapatite, allanite, and REE-

788   epidote as a function of pressure, temperature, and fluid composition. American

789   Mineralogist, 96, 1547–1567.

790   Catlos, E.J. (2013) Versatile Monazite: resolving geological records and solving challenges in

791   materials science: Generalizations about monazite: Implications for geochronologic

792   studies. American Mineralogist, 98, 819–832.

793   Cherniak, D.J., Watson, E.B., Grove, M., and Harrison, T.M. (2004) Pb diffusion in monazite:

794   A combined RBS/SIMS study. Geochimica et Cosmochimica Acta, 68, 829–840.

795   Didier, A., Bosse, V., Boulvais, P., Bouloton, J., Paquette, J.-L., Montel, J.-M., and Devidal,

796   J.-L. (2013) Disturbance versus preservation of U-Th-Pb ages in monazite during fluid-

797   rock interaction: Textural, chemical and isotopic in situ study in microgranites (Velay

798   Dome, France). Contributions to Mineralogy and Petrology, 165, 1051–1072.

799   Dini, A., Rocchi, S., and Westerman, D.S. (2004) Reaction microtextures of REE–Y–Th–U

800   accessory minerals in the Monte Capanne pluton (Elba Island, Italy): a possible indicator

801   of hybridization processes. Lithos, 78, 101–118.

802   Engi, M. (2017) Petrochronology based on REE-minerals: monazite, allanite, xenotime,

803   apatite. Reviews in Mineralogy and Geochemistry, 83, 365–418.

804   Finger, F., Broska, I., Roberts, M.P., and Schermaier, A. (1998) Replacement of primary

805   monazite by apatite-allanite-epidote coronas in an amphibolite facies granite gneiss from

806   the eastern Alps. American Mineralogist, 83, 248–258.

807   Finger, F., Krenn, E., Schulz, B., Harlov, D., and Schiller, D. (2016) "Satellite monazites" in

808   polymetamorphic basement rocks of the Alps: Their origin and petrological significance.

809   American Mineralogist, 101, 1094–1103.

810   Förster, H.-J. (2001) Synchysite-(Y)–synchysite-(Ce) solid solutions from Markersbach,

811   Erzgebirge, Germany: REE and Th mobility during high-T alteration of highly

812   fractionated aluminous A-type granites. Mineralogy and Petrology, 72, 259–280.

813   Foster, G., Parrish, R.R., Horstwood, M.S., Chenery, S., Pyle, J., and Gibson, H.D. (2004)

814   The generation of prograde P–T–t points and paths; a textural, compositional, and

815   chronological study of metamorphic monazite. Earth and Planetary Science Letters, 228,

816   125–142.

817   Gardés, E., Jaoul, O., Montel, J.-M., Seydoux-Guillaume, A.-M., and Wirth, R. (2006) Pb

818   diffusion in monazite: An experimental study of $Pb^{2+} + Th^{4+} \Leftrightarrow 2\ Nd^{3+}$ interdiffusion.

819   Geochimica et Cosmochimica Acta, 70, 2325–2336.

820   Geological Survey of Japan (2012) Seamless digital geological map of Japan 1: 200,000. July

821   3, 2012 version. Research Information Database DB084. Geological Survey of Japan,

822   National Institute of Advanced Industrial Science and Technology (ed.).

823   Gonçalves, G.O., Lana, C., Scholz, R., Buick, I.S., Gerdes, A., Kamo, S.L., Corfu, F.,

824   Marinho, M.M., Chaves, A.O., and Valeriano, C. (2016) An assessment of monazite from

825   the Itambé pegmatite district for use as U–Pb isotope reference material for microanalysis

826   and implications for the origin of the "Moacyr" monazite. Chemical Geology, 424, 30–50.

827   Grand'Homme, A., Janots, E., Seydoux-Guillaume, A.-M., Guillaume, D., Bosse, V., and

828   Magnin, V. (2016) Partial resetting of the U-Th-Pb systems in experimentally altered

829   monazite: Nanoscale evidence of incomplete replacement. Geology, 44, 431–434.

830   Grand'Homme, A., Janots, E., Seydoux-Guillaume, A.M., Guillaume, D., Magnin, V.,

831   Hövelmann, J., Höschen, C., and Boiron, M.C. (2018) Mass transport and fractionation

832   during monazite alteration by anisotropic replacement. Chemical Geology, 484, 51–68.

833   Harlov, D.E. (2011) Formation of monazite and xenotime inclusions in fluorapatite

834   megacrysts, Gloserheia Granite Pegmatite, Froland, Bamble Sector, southern Norway.

835   Mineralogy and Petrology, 102, 77.

836   Harlov, D.E., Wirth, R., and Hetherington, C.J. (2011) Fluid-mediated partial alteration in

837   monazite: The role of coupled dissolution-reprecipitation in element redistribution and

838   mass transfer. Contributions to Mineralogy and Petrology, 162, 329–348.

839   Harrison, T.M., McKeegan, K.D., and LeFort, P. (1995) Detection of inherited monazite in

840   the Manaslu leucogranite by 208Pb232Th ion microprobe dating: crystallization age and

841   tectonic implications. Earth and Planetary Science Letters, 133, 271–282.

842   Hawkins, D.P., and Bowring, S.A. (1997) U-Pb systematics of monazite and xenotime: case

843   studies from the Paleoproterozoic of the Grand Canyon, Arizona. Contributions to

844   Mineralogy and Petrology, 127, 87–103.

845   Heinrich, W., Andrehs, G., and Franz, G. (1997) Monazite–xenotime miscibility gap

846   thermometry. I. An empirical calibration. Journal of Metamorphic Geology, 15, 3–16.

847   Herzig, C.T., Kimbrough, D.L., Tainosho, Y., Kagami, H., Iizumi, S., and Hayasaka, Y.

848   (1998) Late Cretaceous U/Pb zircon ages and Precambrian crustal inheritance in Ryoke

849   granitoids, Kinki and Yanai districts, Japan. Geochemical journal, 32, 21–31.

850   Hetherington, C.J., Harlov, D.E., and Budzyń, B. (2010) Experimental metasomatism of

851   monazite and xenotime: Mineral stability, REE mobility and fluid composition.

852   Mineralogy and Petrology, 99, 165–184.

853   Higashimoto, S., Nureki, T., Hara, I., Tsukuda, E., and Nakajima, T. (1983) Geology of the

854   Iwakuni district. Quadrangle Series, Scale 1:50,000, Geological Survey of Japan, 78, 79p.

855   Ikeda, T. (1993) Compositional zoning patterns of garnet during prograde metamorphism

856   from the Yanai district, Ryoke metamorphic belt, southwest Japan. Lithos, 30, 109–121.

857   ——— (1998) Progressive sequence of reactions of the Ryoke metamorphism in the Yanai

858   district, southwest Japan: The formation of cordierite. Journal of Metamorphic Geology,

859   16, 39–52.

860   ——— (2004) Pressure-temperature conditions of the Ryoke metamorphic rocks in Yanai

861   district, SW Japan. Contributions to Mineralogy and Petrology, 146, 577–589.

862   Ishihara, S. (1977) The Magnetite-series and Ilmenite-series Granitic Rocks. Mining Geology,

863   27, 293–305.

864   Janoušek, V., Farrow, C.M., and Erban, V. (2006) Interpretation of whole-rock geochemical

865   data in igneous geochemistry: introducing Geochemical Data Toolkit (GCDkit). Journal of

866   Petrology, 47, 1255–1259.

867   Jefferies, N.L. (1985) The distribution of the rare earth elements within the Carnmenellis

868   pluton, Cornwall. Mineralogical Magazine, 49, 495–504.

869   Kato, T. (2005) New Accurate Bence-Albee α-Factors for Oxides and Silicates Calculated

870   from the PAP Correction Procedure. Geostandards and Geoanalytical Research, 29, 83–

871   94.

872   Kawakami, T., and Ikeda, T. (2003) Boron in metapelites controlled by the breakdown of

873   tourmaline and retrograde formation of borosilicates in the Yanai area, Ryoke

874   metamorphic belt, SW Japan. Contributions to Mineralogy and Petrology, 145, 131–150.

875   Kawano, Y., and Ueda, Y. (1966) K-A dating on the igneous rocks in Japan (V). The Journal

876   of the Japanese Association of Mineralogists, Petrologists and Economic Geologists, 56,

877   191–211.

878   Kelly, N.M., Harley, S.L., and Möller, A. (2012) Complexity in the behavior and

879   recrystallization of monazite during high-T metamorphism and fluid infiltration. Chemical

880   Geology, 322–323, 192–208.

881   Kretz, R. (1983) Symbols of rock-forming minerals. American Mineralogist, 68, 277–279.

882   Kusiak, M.A., Williams, I.S., Dunkley, D.J., Konečny, P., S\laby, E., and Martin, H. (2014)

883   Monazite to the rescue: U–Th–Pb dating of the intrusive history of the composite

884   Karkonosze pluton, Bohemian Massif. Chemical Geology, 364, 76–92.

885   Ludwig, K.R. (2004) Users manual for ISOPLOT/EX, version3.1. A geochronological toolkit

886   for Microsoft Excel. Berkeley Geochronology Center, Special Publication, 4.

887   Mielke, P., and Winkler, H.G.F. (1979) Eine bessere Berechnung der Mesonorm für

888   granitische Gesteine. Neues Jahrbuch für Mineralogie, Monatshefte, 10, 471–480.

889   Montel, J.-M. (1986) Experimental determination of the solubility of Ce-monazite in SiO2-

890   Al2O3-K2O-Na2O melts at 800 C, 2 kbar, under H2O-saturated conditions. Geology, 14,

891   659–662.

892   Moutte, J., and Iiyama, J.T. (1984) The Ryoke-Sanyo granite series in the Iwakuni-Yanai

893   district, Southwest Honshu, Japan. Mining Geology, 34, 425–436.

894   Murakami, N. (1971) An example of the emplacement of the Chugoku batholith-the Kuga

895   granites, SW Japan. Pacific Geology, 45−56.

896   Murata, K.J., Rose Jr, H.J., Carron, M.K., and Glass, J.J. (1957) Systematic variation of rare-

897   earth elements in cerium-earth minerals. Geochimica et Cosmochimica Acta, 11, 141−161.

898   Nakajima, T. (1994) The Ryoke plutonometamorphic belt: crustal section of the Cretaceous

899   Eurasian continental margin. LITHOS, 33, 51−66.

900   Nakajima, T., Horie, K., Adachi, T., Miyazaki, K., Dunkley, D.J., and Hokada, T. (2013)

901   SHRIMP U-Pb ages of zircons from Ryoke metamorphic rocks. Geological Society of

902   Japan Annual Meeting, Abstract Volume, 51.

903   Nasdala, L. Corfu, F., Blaimauer, D., Chanmuang, C., Ruschel, K., Škoda, R., Wildner, M.,

904   Wirth, R., Zeug, M., and Gamini Zoysa, E. (2017). Neoproterozoic amorphous "ekanite"

905   ($Ca_2Th_{0.9}U_{0.1}Si_8O_{20}$) from Okkampitiya, Sri Lanka: A metamict gemstone with excellent

906   lead-retention performance. Geology, 45, 919−922.

907   Ni, Y., Hughes, J.M., and Mariano, A.N. (1995) Crystal chemistry of the monazite and

908   xenotime structures. American Mineralogist, 80, 21−26.

909   Nureki, T. (1974) Contact metamorphism in the So-o district, Yamaguchi Prefecture, Japan -

910   With special reference to the occurrence of sillimanite. Memoirs of the Geological Society

911   of Japan, 11, 251−281.

912   Oelkers, E.H., and Poitrasson, F. (2002) An experimental study of the dissolution

913   stoichiometry and rates of a natural monazite as a function of temperature from 50 to 230

914   °C and pH from 1.5 to 10. Chemical Geology, 191, 73−87.

915   Okudaira, T., Hayasaka, Y., Himeno, O., Watanabe, K., Sakurai, Y., and Ohtomo, Y. (2001)

916   Cooling and inferred exhumation history on the Ryoke metamorphic belt in the Yanai

917   district, South-west Japan: Constraints from Rb-Sr and fission-track ages of gneissose

918   granitoid and numerical modeling. Island Arc, 10, 98−115.

919   Ondrejka, M., Uher, P., Putiš, M., Broska, I., Bačik, P., Konečný, P. and Schmiedt, I. (2012).

920   Two-stage breakdown of monazite by post-magmatic and metamorphic fluids: An

921   example from the Veporic orthogneiss, Western Carpathians, Slovakia. Lithos, 142–143,

922   245–255.

923   Pearce, N.J.G., Perkins, W.T., Westgate, J.A., Gorton, M.P., Jackson, S.E., Neal, C.R., and

924   Chenery, S.P. (1997) A Compilation of New and Published Major and Trace Element

925   Data for NIST SRM 610 and NIST SRM 612 Glass Reference Materials. Geostandards

926   and Geoanalytical Research, 21, 115–144.

927   Piechocka, A.M., Gregory, C.J., Zi, J.-W., Sheppard, S., Wingate, M.T.D., and Rasmussen, B.

928   (2017) Monazite trumps zircon: applying SHRIMP U–Pb geochronology to systematically

929   evaluate emplacement ages of leucocratic, low-temperature granites in a complex

930   Precambrian orogen. Contributions to Mineralogy and Petrology, 172, 63.

931   Poitrasson, F., Chenery, S., and Bland, D.J. (1996) Contrasted monazite hydrothermal

932   alteration mechanisms and their geochemical implications. Earth and Planetary Science

933   Letters, 145, 79–96.

934   Poitrasson, F., Chenery, S., and Shepherd, T.J. (2000) Electron microprobe and LA-ICP-MS

935   study of monazite hydrothermal alteration: Implications for U-Th-Pb geochronology and

936   nuclear ceramics. Geochimica et Cosmochimica Acta, 64, 3283–3297.

937   Putnis, A. (2002) Mineral replacement reactions: From macroscopic observations to

938   microscopic mechanisms. Mineralogical Magazine, 66, 689–708.

939   Putnis, A. (2009) Mineral Replacement Reactions. Reviews in Mineralogy and Geochemistry,

940   70, 87–124.

941   Putnis, A., and Putnis, C.V. (2007) The mechanism of reequilibration of solids in the presence

942   of a fluid phase. Journal of Solid State Chemistry, 180, 1783–1786.

39   39

943 Pyle, J.M., Spear, F.S., Rudnick, R.L., and McDonough, W.F. (2001) Monazite-xenotime-

944 garnet equilibrium in metapelites and a new monazite-garnet thermometer. Journal of

945 Petrology, 42, 2083–2107.

946 Pyle, J.M., Spear, F.S., Wark, D.A., Daniel, C.G., and Storm, L.C. (2005) Contributions to

947 precision and accuracy of monazite microprobe ages. American Mineralogist, 90, 547–

948 577.

949 Rubatto, D., Müntener, O., Barnhoorn, A., and Gregory, C. (2008) Dissolution-reprecipitation

950 of zircon at low-temperature, high-pressure conditions (Lanzo Massif, Italy). American

951 Mineralogist, 93, 1519–1529.

952 Sato, H. (1933) Yanaizu Geological Map, 1:75 000. Imperial Geological Survey of Japan.

953 Scherrer, N.C., Engi, M., Gnos, E., Jakob, V., and Liechti, A. (2000) Monazite analysis; from

954 sample preparation to microprobe age dating and REE quantification. Schweizerische

955 Mineralogische und Petrographische Mitteilungen, 80, 93–105.

956 Schmidt, C., Rickers, K., Bilderback, D.H., and Huang, R. (2007) In situ synchrotron-

957 radiation XRF study of REE phosphate dissolution in aqueous fluids to 800 °C. Lithos, 95,

958 87–102.

959 Seydoux-Guillaume, A.-M., Paquette, J.-L., Wiedenbeck, M., Montel, J.-M., and Heinrich, W.

960 (2002) Experimental resetting of the U-Th-Pb systems in monazite. Chemical Geology,

961 191, 165–181.

962 Seydoux-Guillaume, A.-M., Montel, J.-M., Bingen, B., Bosse, V., De Parseval, P., Paquette,

963 J.-L., Janots, E., and Wirth, R. (2012) Low-temperature alteration of monazite: Fluid

964 mediated coupled dissolution–precipitation, irradiation damage, and disturbance of the U–

965 Pb and Th–Pb chronometers. Chemical Geology, 330, 140–158.

966 Skrzypek, E., Kawakami, T., Hirajima, T., Sakata, S., Hirata, T., and Ikeda, T. (2016)

967   Revisiting the high temperature metamorphic field gradient of the Ryoke Belt (SW Japan):

968   New constraints from the Iwakuni-Yanai area. Lithos, 260, 9–27.

969 Skrzypek, E., Kato, T., Kawakami, T., Sakata, S., Hattori, K., Hirata, T., and Ikeda, T. (2018)

970   Monazite Behaviour and Time-scale of Metamorphic Processes along a Low-

971   pressure/High-temperature Field Gradient (Ryoke Belt, SW Japan). Journal of Petrology,

972   59, 1109–1144.

973 Spear, F.S. (2010) Monazite–allanite phase relations in metapelites. Chemical Geology, 279,

974   55–62.

975 Suzuki, K., and Kato, T. (2008) CHIME dating of monazite, xenotime, zircon and polycrase:

976   protocol, pitfalls and chemical criterion of possibly discordant age data. Gondwana

977   Research, 14, 569–586.

978 Suzuki, K., Adachi, M., and Nureki, T. (1996) CHIME age dating of monazites from

979   metamorphic rocks and granitic rocks of the Ryoke belt in the Iwakuni area, Southwest

980   Japan. Island Arc, 5, 43–55.

981 Takatsuka, K., Kawakami, T., Skrzypek, E., Sakata, S., Obayashi, H., and Hirata, T. (2018a)

982   Age gap between the intrusion of gneissose granitoids and regional high-temperature

983   metamorphism in the Ryoke belt (Mikawa area), central Japan. Island Arc, 27, e12224.

984 ——— (2018b) Spatiotemporal evolution of magmatic pulses and regional metamorphism

985   during a Cretaceous flare-up event: Constraints from the Ryoke belt (Mikawa area, central

986   Japan). Lithos, 308, 428–445.

987 Taylor, M., and Ewing, R.C. (1978) The crystal structures of the $ThSiO_4$ polymorphs:

988   huttonite and thorite. Acta Crystallographica Section B, 34, 1074–1079.

989 Taylor, R.J.M., Clark, C., Fitzsimons, I.C.W., Santosh, M., Hand, M., Evans, N., and

990   McDonald, B. (2014) Post-peak, fluid-mediated modification of granulite facies zircon

41   41

991   and monazite in the Trivandrum Block, southern India. Contributions to Mineralogy and

992   Petrology, 168, 1–17.

993   Terra, O., Dacheux, N., Clavier, N., Podor, R., and Audubert, F. (2008) Preparation of

994   Optimized Uranium and Thorium Bearing Brabantite or Monazite/Brabantite Solid

995   Solutions. Journal of the American Ceramic Society, 91, 3673–3682.

996   Teufel, S., and Heinrich, W. (1997) Partial resetting of the UPb isotope system in monazite

997   through hydrothermal experiments: An SEM and UPb isotope study. Chemical Geology,

998   137, 273–281.

999   Tomaschek, F., Kennedy, A.K., Villa, I.M., Lagos, M., and Ballhaus, C. (2003) Zircons from

1000   Syros, Cyclades, Greece—Recrystallization and Mobilization of Zircon During High-

1001   Pressure Metamorphism. Journal of Petrology, 44, 1977–2002.

1002   Townsend, K.J., Miller, C.F., D'Andrea, J.L., Ayers, J.C., Harrison, T.M., and Coath, C.D.

1003   (2001) Low temperature replacement of monazite in the Ireteba granite, Southern Nevada:

1004   Geochronological implications. Chemical Geology, 172, 95–112.

1005   Trail, D. (2018) Redox-controlled dissolution of monazite in fluids and implications for phase

1006   stability in the lithosphere. American Mineralogist, 103, 453–461.

1007   Villa, I.M., and Williams, M.L. (2013) Geochronology of Metasomatic Events. In

1008   Metasomatism and the Chemical Transformation of Rock pp. 171–202. Springer Berlin

1009   Heidelberg.

1010   Ward, C.D., McArthur, J.M., and Walsh, J.N. (1992) Rare earth element behaviour during

1011   evolution and alteration of the Dartmoor granite, SW England. Journal of Petrology, 33,

1012   785–815.

1013   Watson, E.B. (1979) Zircon saturation in felsic liquids: experimental results and applications

1014   to trace element geochemistry. Contributions to Mineralogy and Petrology, 70, 407–419.

1015   Weinberg, R.F., and Hasalová, P. (2015) Water-fluxed melting of the continental crust: A

1016   review. Lithos, 212–215, 158–188.

1017   Williams, M.L., Jercinovic, M.J., and Terry, M.P. (1999) Age mapping and dating of

1018   monazite on the electron microprobe: deconvoluting multistage tectonic histories.

1019   Geology, 27, 1023–1026.

1020   Williams, M.L., Jercinovic, M.J., Harlov, D.E., Budzyń, B., and Hetherington, C.J. (2011)

1021   Resetting monazite ages during fluid-related alteration. Chemical Geology, 283, 218–225.

1022   Yuguchi, T., Sasao, E., Ishibashi, M., and Nishiyama, T. (2015) Hydrothermal chloritization

1023   processes from biotite in the Toki granite, Central Japan: Temporal variations of of the

1024   compositions of hydrothermal fluids associated with chloritization. American

1025   Mineralogist, 100, 1134–1152.

1026   **FIGURE CAPTIONS**

1027   **FIGURE 1.** Geological map of the Ryoke belt in the Iwakuni-Yanai area (SW Japan), with

1028   sampling localities. Inset shows the area within the Inner Zone of SW Japan (MTL=Median

1029   Tectonic Line). Granitoids are classified according to their emplacement timing relative to the

1030   main, regional tectono-metamorphic event (D1). Lithologies are after Sato (1933),

1031   Higashimoto et al. (1983) and the Geological Survey of Japan (2012). Metamorphic zone

1032   boundaries are after Ikeda (1993). Schist/gneiss boundary modified after Higashimoto et al.

1033   (1983). Mineral abbreviations follow IUGS recommendations after Kretz (1983).

1034   **FIGURE 2.** Microscopic texture of SHK and NAM samples on (**a, b**) EMP scans of thin

1035   sections and (**c, d**) optical microscope photographs. (**a**) Sample SHK showing rounded Qtz

1036   aggregates, Kfs−Pl with straight margins and variably chloritized Bt. (**b**) Sample NAM

1037   showing elongated Qtz layers, aligned Bt and Kfs−Pl with irregular margins. The trace of the

1038   gneissosity is subparallel to the long axis of the thin section. (**c**) Occurrence of Mnz and Aln

1039   in relationship with variably chloritized Bt. SHK, right-hand side of the thin section, plane-

1040   polarized light. (**d**) Ms−Qtz network inside fragmented Kfs, and Mnz grains inside Bt. NAM,

1041   bottom left-hand side of the thin section, crossed polarizers.

1042   **FIGURE 3.** Monazite textures in sample SHK. (**a**) Optical microscope photographs of variably

1043   altered, separated monazite grains. White crosses indicate the location of laser pits. (**b-f**) BSE

1044   images of *in situ* textures and separated grains. (**b**) Euhedral monazite cross-cut by cracks and

1045   surrounded by Ap, Aln, Xtm and a Th−U phase next to chloritized Bt. (**c**) Anhedral monazite

1046   surrounded by zoned allanite, Th−U phase, Xtm and chloritized Bt. (**d**) Oscillatory zoned

1047   allanite protruding into the cleavages of chloritized Bt. Allanite is separated from chlorite by

1048   Al-rich titanite. (**e**) Virtually unaltered monazite with a sector-zoned core and a darker

1049   outermost rim. Limited alteration (darkening) is visible in the bottom right hand part. (**f**)

1050   Altered monazite with Ap, Aln, Xtm, Th−U phase, Chl and Bt present along micro-cracks.

1051   Recrystallization of secondary domains occurred around cracks (darkened parts) and veinlets

1052   (brightened parts). Contrast is adjusted to see the secondary minerals (left) and internal zoning

1053   (right). Same grain as on the right hand photograph in (a).

1054   **FIGURE 4**. Monazite composition in samples from the Shimokuhara (SHK, SHK2) and

1055   Namera (NAM, NAM2) granites. Data are from *in situ* and separated grains. (**a**) REE+Y+P

1056   vs. Th+U+Si diagram (cations for 16 oxygens) highlighting the dominant huttonic substitution

1057   for primary domain analyses in both samples, and the dominant cheralitic substitution for

1058   secondary domain analyses in NAM. (**b**) Y vs. U diagram (cation fraction on A site) showing

1059   Y−U depletion for some secondary domain analyses in SHK, and Y−U enrichment for

1060   secondary domain analyses in NAM. (**c**) Ternary diagram showing proportions of the

1061   monazite [REEPO$_4$], "cheralite" [Ca$_{0.5}$(Th,U)$_{0.5}$PO$_4$] and "huttonite" [(Th,U)SiO$_4$] end-

1062   members for analyses of monazite (all samples) and Th−U phases (SHK & SHK2).

1063   **FIGURE 5.** Monazite textures in sample NAM. (**a**) Optical microscope photographs of

1064   variably altered, separated monazite grains. White crosses indicate the location of laser pits.

44   44

1065   (**b-f**) BSE images of *in situ* textures and separated grains. (**b**) Typical mode of occurrence of

1066   monazite inside matrix Bt, with (**c**) an internal structure showing relicts of a sector- or

1067   oscillatory-zoned primary domain cross-cut by a darker secondary domain. (**d**) Virtually

1068   unaltered monazite with a dominant oscillatory-zoned primary domain. A convoluted

1069   secondary domain is slightly developed in the bottom part. (**e**) Altered monazite with a dark

1070   and inclusion-rich secondary domain preferentially developed in the core (included minerals

1071   indicated). Isolated "islands" of primary monazite are left inside the secondary domain. (**f**)

1072   Altered monazite with a dark and inclusion-rich secondary domain developed at the grain

1073   margin (included minerals indicated). The right hand part preserves primary oscillatory

1074   zoning. (**g-h**) Close-up view of rounded and tubular, multi-phase inclusions in monazite.

1075   **FIGURE 6.** Characterization of selected inclusions in secondary monazite domains from

1076   sample NAM. (**a**) Energy dispersion spectra with the most intense element peaks labelled.

1077   The analyzed inclusions are shown in Fig. 5g, h. (**b**) Raman spectra with the shift of the main

1078   vibration bands labelled. The analyzed inclusions are different from those in (**a**). All spectra

1079   are background-corrected and normalized to the maximum intensity of the most intense

1080   spectrum in each set (vertical intensity scale is arbitrary). The spectra of host monazite are

1081   shown for comparison.

1082   **FIGURE 7.** Results of LA-ICP-MS monazite dating in (**a,b**) sample SHK and (**c,d**) sample

1083   NAM. (**a**, **c**) BSE images of analyzed monazite grains with representative results. Each spot

1084   location is labelled with spot name, $^{206}Pb/^{238}U$ date $\pm$ 2σ error (Ma) and concordance (%).

1085   Scale bar is always 20 μm. (**b**, **d**) Concordia diagrams showing concordant (cc=97−103%)

1086   and discordant (cc < 97% or cc > 103%) analyses for the different monazite domains. Gray-

1087   coded symbols according to U/Th ratio are added in the centre of error ellipses for sample

1088   NAM.

Table 1. Results of mineral mode calculations

	Or	Ab	An	Pl	Qtz	Bt	Chl	Ms	Grt	Ap	Mnz	Ttn*	Hem	Ilm	Crn	Sum	Rest
Mesonorm																	
SHK	17.4	29.5	10.2	39.7	31.8	7.1	-	-	-	0.17	-	-	-	0.21	2.11	98.15	-0.26
NAM	15.3	30.7	8.8	39.6	33.9	6.9	-	-	-	0.24	-	-	-	0.19	2.31	98.22	-0.26
EMP mapping																	ND
SHK	24.7	-	-	30.2	32.7	3.6	4.1	-	0.07	0.05	0.01	0.36	-	-	-	95.71	4.29
NAM	19.1	-	-	38.6	31.2	7.1	-	1.4	0.01	0.05	0.01	-	*low*	-	-	97.48	2.52

Note: * Titanite mode includes allanite in the case of EMP mapping; *low* =<0.01; ND = no data.

Table 2. Representative analyses of monazite domains and Th-U phase

Sample	SHK				NAM	
Spot	99a-m41	99a-m34	99a-Mnz9	99a-Ac2	117a-m35	117a-Mnz17
Domain	primary core	primary rim	secondary	Th-U phase	primary	secondary
Wt% oxides						
SiO_2	2.30	0.60	0.63	18.16	0.70	0.44
P_2O_5	27.31	29.70	29.39	0.78	29.63	29.29
CaO	1.08	0.83	0.11	1.26	0.86	1.30
ThO_2	14.62	5.89	2.69	62.77	6.88	6.84
UO_2	0.73	0.45	d.l.	5.64	0.15	0.45
Y_2O_3	2.53	3.22	0.74	1.10	0.65	1.46
La_2O_3	11.35	13.31	12.80	0.195	16.90	11.50
Ce_2O_3	22.28	26.60	30.06	0.32	29.77	26.58
Pr_2O_3	2.53	2.82	3.72	d.l.	2.92	3.14
Nd_2O_3	10.88	11.08	13.41	0.34	10.56	11.49
Sm_2O_3	2.28	2.51	3.45	d.l.	1.27	3.19
Gd_2O_3	2.09	2.30	2.43	0.35	0.70	2.01
Tb_2O_3	n.a.	n.a.	0.19	n.a.	n.a.	0.19
Dy_2O_3	n.a.	n.a.	0.57	n.a.	n.a.	0.61
Ho_2O_3	n.a.	n.a.	d.l.	n.a.	n.a.	d.l.
Er_2O_3	n.a.	n.a.	d.l.	n.a.	n.a.	0.14
Tm_2O_3	n.a.	n.a.	0.34	n.a.	n.a.	0.29
Yb_2O_3	n.a.	n.a.	d.l.	n.a.	n.a.	d.l.
PbO	0.04	0.02	d.l.	0.198	d.l.	0.02
Total	99.98	99.34	100.53	91.12	100.97	98.92
Cations						
Si	0.366	0.094	0.099	0.988	0.109	0.069
P	3.681	3.932	3.912	0.036	3.908	3.934
Ca	0.184	0.138	0.018	0.073	0.143	0.221
Th	0.530	0.210	0.096	0.777	0.244	0.247
U	0.026	0.016	0.000	0.068	0.005	0.016
Y	0.214	0.268	0.062	0.032	0.053	0.123
La	0.667	0.768	0.742	0.004	0.971	0.673
Ce	1.299	1.523	1.730	0.006	1.698	1.544
Pr	0.147	0.161	0.213	-	0.166	0.182
Nd	0.619	0.619	0.753	0.007	0.588	0.651
Sm	0.125	0.135	0.187	-	0.068	0.174
Gd	0.110	0.119	0.127	0.006	0.036	0.106
Tb	-	-	0.010	-	-	0.010
Dy	-	-	0.029	-	-	0.031
Ho	-	-	-	-	-	-
Er	-	-	-	-	-	0.007
Tm	-	-	0.017	-	-	0.014
Yb	-	-	-	-	-	-
Pb	0.001	0.001	-	0.003	-	0.001
Total	7.968	7.985	7.995	2.000	7.990	8.003
Xmnz	0.811	0.908	0.971	0.059	0.901	0.879
Xcher	0.094	0.070	0.009	0.150	0.072	0.110
Xhut	0.095	0.022	0.020	0.792	0.027	0.011
T (°C)[a]	-	-	482	-	-	533
T (°C)[b]	-	-	572	-	-	618

Notes: n.a. = not analyzed; d.l. = below detection limit
Monazite: cations normalized to 32 charges; Th-U phase: normalization to 2 cations
[a]Calibration after Pyle et al. (2001)
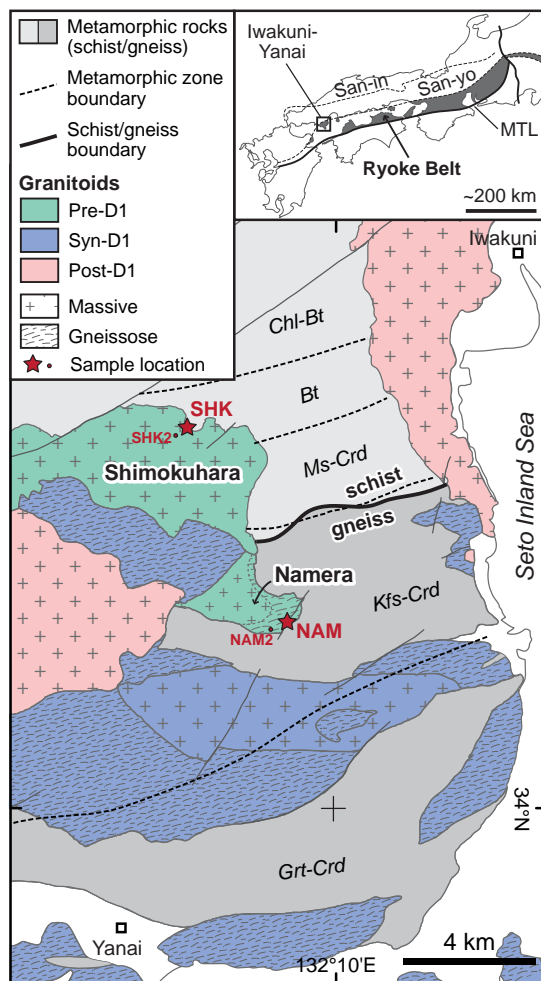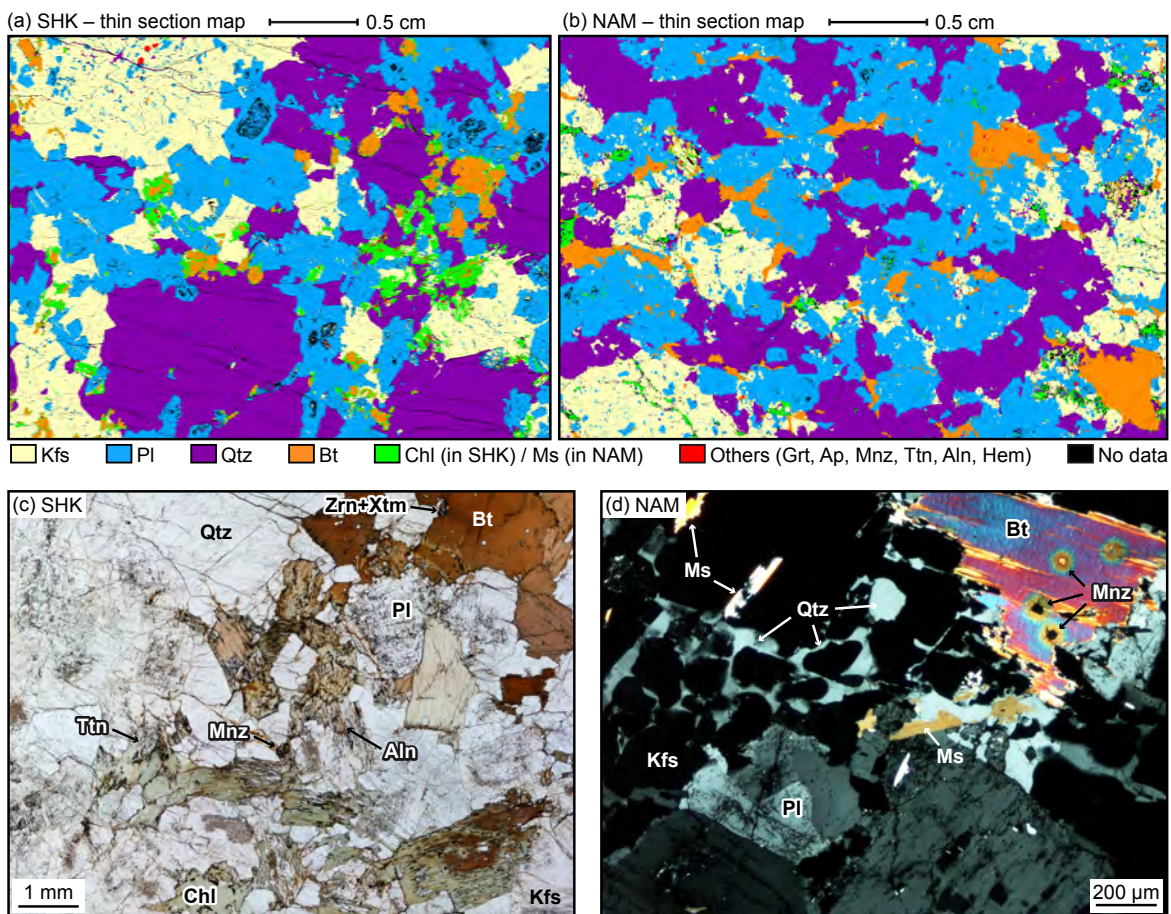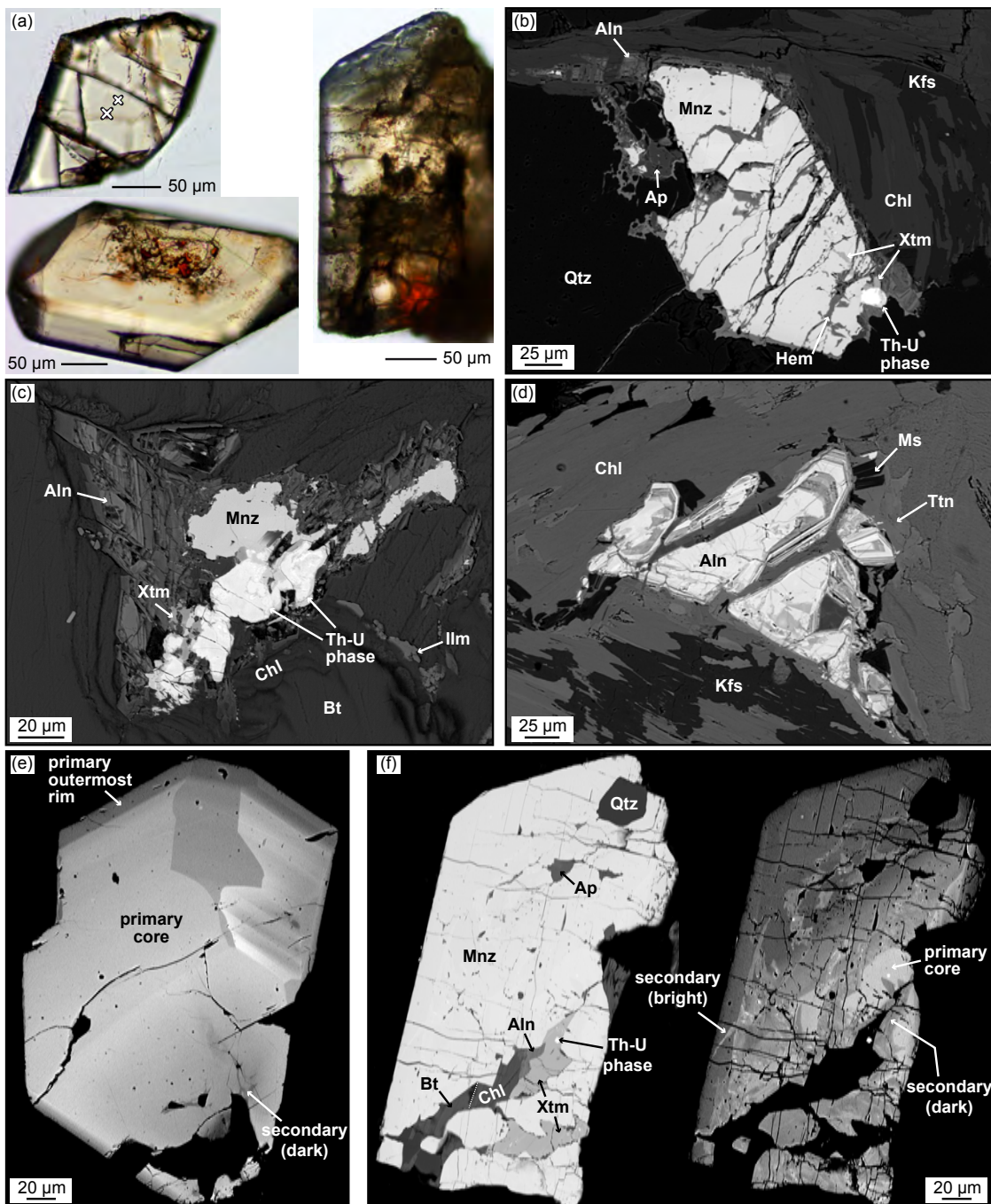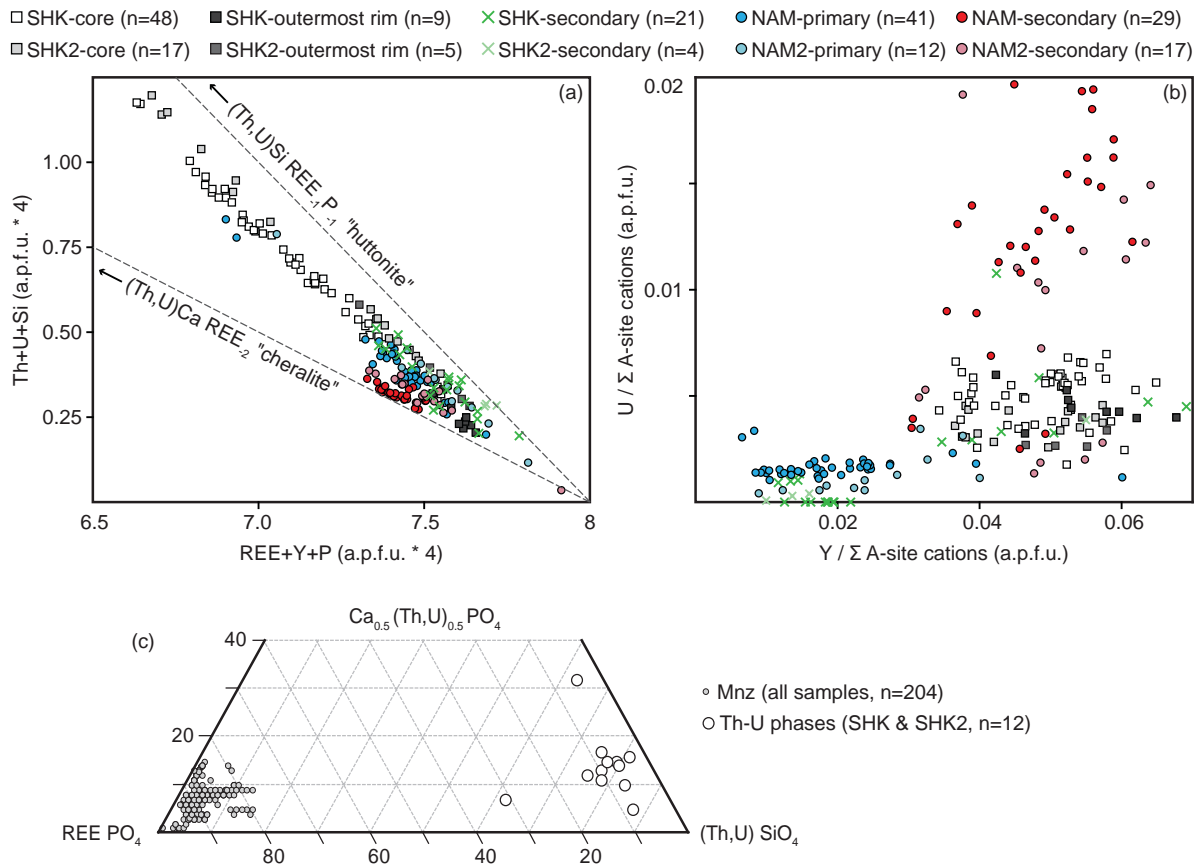[b]Calibration after Heinrich et al. (1997)

Figure 1

Figure 2

Figure 3

Figure 4

Figure 6

Figure 7