# Geochemistry of Eocene-Early Oligocene low-temperature crustal melts from Greater Himalayan Sequence (Nepal): a nanogranitoid perspective

Omar Bartoli, Antonio Acosta-Vigil, Bernardo Cesare, Laurent Remusat, Adriana Gonzalez-Cano, Markus Wälle, Lucie Tajčmanová, Antonio Langone

# Geochemistry of Eocene-Early Oligocene low-temperature crustal melts from Greater Himalayan Sequence (Nepal): a nanogranitoid perspective

Omar Bartoli[1*], Antonio Acosta-Vigil[2], Bernardo Cesare[1], Laurent Remusat[3], Adriana Gonzalez-Cano[3], Markus Wälle[4], Lucie Tajčmanová [5], Antonio Langone[6]

[1] *Dipartimento di Geoscienze, Università degli Studi di Padova, Padova, Italy*
[2] *Instituto Andaluz de Ciencias de la Tierra, CSIC-Universidad de Granada, Spain*
[3] *Muséum National d'Histoire Naturelle, Sorbonne Université, UMR CNRS 7590, lRD, Institut de Minératogie, de Physique des Matériaux et de Cosmochimie, IMPMC, Paris, France*
[4] *REAIT, CRC and CFI Services (CCCS), Memorial University of Newfoundland, Canada*
[5] *Institute of Earth Sciences, Heidelberg University, Germany*
[6] *Istituto di Geoscienze e Georisorse, C.N.R.University of Pavia, Pavia, Italy*

\* corresponding author: omar.bartoli@unipd.it, +390498279832

*Abstract*

Despite melt inclusions in migmatites and granulites have provided a wealth of information on crustal anatexis in different geological scenarios, yet a complete compositional study (including trace elements and $H_2O$) is missing for the Himalayas.  In this contribution, we focus on nanogranitoids occurring in peritectic garnet of migmatites from Kali Gandaki valley in central Nepal (Greater Himalayan Sequence, GHS). Their microstructural position proves that these melts were produced at 650–720 °C and 1.0–1.1 GPa, during the Eohimalayan prograde metamorphism (41–36 Ma) associated to crustal thickening. Nanogranitoid compositions (mostly granodiorites, tonalities and trondhjemites) resemble those of experimental melts produced during $H_2O$-present melting of metasedimentary rocks. They have variable $H_2O$ concentrations (6.5-14.4 wt%), which approximate the expected minimum and maximum values for crustal melts produced at the inferred P–T conditions. These compositional signatures suggest that melt formation occurred in the proximity of the

H$_2$O-saturated solidus, in a rock-buffered system. The low to very low contents of Zr (3–8 ppm), Th (0.1–1.2 ppm) and LREE (4–11 ppm) along with the weak to moderate positive Eu anomalies (Eu/Eu* =1.2–3.3), the high B concentrations (200–3400 ppm) and the high U/Th ratio (up to 21) point to the lack of equilibration between melt and accessory minerals and are consistent with melting of plagioclase at low temperature. Kali Gandaki nanogranitoids have recorded the beginning of melting in a H$_2$O-present scenario that, in other GHS localities, may have produced voluminous crustal melts. We demonstrate that compositional comparison with nanogranitoids may be useful to reconstruct the petrogenesis of Eohimalayan granitoids.

# 1 Introduction

2 The Himalaya mountain belt is the largest active collisional orogen on our planet and is an

3 ideal natural laboratory for investigating orogenic processes (Kohn 2014). This mountain

4 chain has resulted from collision between India and Asia at c. 59 Ma (Hu et al. 2016). The

5 structural and metamorphic evolution of this orogenic belt is divided into two main phases

6 (Hodges 2000): an Eocene-Late Oligocene Eohimalayan phase during which crustal

7 thickening occurred and most metamorphic rocks reached peak conditions in the kyanite

8 stability field, and an Early Miocene to present Neohimalayan phase characterized by

9 formation of sillimanite-bearing rocks and leucogranites which intruded the higher portions

10 of the Himalayan belt – the so called "Himalayan leucogranites" – whose largest plutons

11 formed at c. 21 Ma (Harrison 2006).

12 Himalayan leucogranites and the associated Miocene migmatites have been

13 investigated for decades (Weinberg 2016 and references therein). A number of works have

14 established that Miocene anatexis and leucogranite formation occurred by fluid-absent

15 melting of muscovite during decompression (e.g., Inger and Harris 1992; Pognante and

16 Benna 1993; Harris et al. 1995; Harris and Massey 1994; Davidson et al. 1997). Recently,

17 some works have argued that decompression was a consequence rather than a cause of

18 melting (e.g., Rubatto et al. 2013) and the recent review paper by Weinberg (2016) has

19 demonstrated the complexity of Neohimalayan anatexis which records differences in P–T–

20 $X_{H2O}$ conditions, timing and duration across the Himalayas.

21 Unlike for the Neohimalayan phase, much fewer constraints are available on anatexis

22 occurred during crustal thickening (i.e., the Eohimalayan anatexis), partly because of

23 pervasive Miocene thermal overprint. Where evidence exists, the majority of works have

24 dealt with the age and P–T conditions of high-grade metamorphic processes (e.g. Hodges et

25 al. 1996; Godin et al. 2001; Groppo et al. 2010; Zhang et al. 2010, 2015, 2017; Guilmette et

26 al. 2011; Rubatto et al. 2013; Palin et al. 2014; Wang et al. 2013; 2015; Iaccarino et al. 2017),

27 without shedding light on the composition of crustal melts which, in turn, may provide

28 important, and otherwise impossible to retrieve, information on melting mechanisms (e.g.,

29 Acosta-Vigil et al. 2010, 2012, 2017; Sawyer 2010, Carvalho et al. 2016, 2017, 2018;

30 Weinberg and Hasalová 2015a). For instance, an unusual Ca-rich leucogranite of Late

31 Oligocene age (c. 27.5 Ma) was considered to provide a rare insight into the pre-Miocene

32 melting associated to crustal thickening (Zhang et al. 2004).

33    In this contribution, we present for the first time the trace element and $H_2O$ content of

34 melt inclusions (i.e., nanogranitoid inclusions or simply nanogranitoids; Cesare et al. 2015;

35 Bartoli et al. 2013a, 2016a; Ferrero and Angel 2018) from the Himalaya mountain belt.

36 Nanogranitoids come from the Greater Himalayan Sequence –which constitutes the high-

37 grade crystalline core of the orogen– and are present in a microstructural position that clearly

38 proves an Eohimalayan origin for these droplets of crustal melt (Carosi et al. 2015).

39

40


## Geological background and sampling

42    From south to north, the Himalayan belt is divided into four tectonic units –Sub-

43 Himalayas molasses, Lesser Himalayan Sequence, Greater Himalayan Sequence (GHS) and

44 Tethyan Himalayan Sequence (THS)– which are separated by important tectonic structures,

45 i.e. Main Boundary Thrust, Main Central Thrust (MCT) and South Tibetan Detachment

46 System (Hodges 2000 and references therein; Fig. 1a). The GHS is a 20-30 km thick

47 sequence of medium- to high-grade metamorphic rocks and represents the metamorphic core

48 of the Himalaya (Searle and Godin 2003). Several levels of tectono-metamorphic

49 discontinuities were identified within the GHS (Carosi et al. 2018).

50    The sample under investigation comes from the N–S trending Kali Gandaki valley of

51 central Nepal (Fig. 1b), one of the most classic study areas of the Himalaya (see Carosi et al.

52 2014 and references therein). Here, GHS reaches a structural thickness of 10–15 km (Godin

53 2003), shows a NE-dipping prominent high-grade foliation with SW-verging isoclinal folds

54 (Brown and Nazarchuk 1993; Godin 2003; Kellett and Godin 2009; Searle 2010) and can be

55 subdivided in three units (Searle and Godin 2003; Searle 2010): the lowest Unit 1 is

56 represented by kyanite-bearing gneisses and migmatites, the Unit 2 consists of

57 clinopyroxene-garnet-amphibole-bearing calcsilicate gneisses, whereas Unit 3 is made of

58 orthogneisses and minor metapelites.

59    The nanogranitoid-bearing garnet crystal selected for this study has been collected

60 from the debris at the base of an outcrop of kyanite-bearing migmatitic paragneiss (lower part

61 of Unit 1), located close to Titar village (North of Dana village, Fig. 1b). In this outcrop,

62 gneisses have a porphyroblastic texture and contain quartz + plagioclase + biotite + garnet +

63 kyanite + white mica (Fig. 2a). Monazite, xenotime, apatite, zircon, tourmaline, rutile,

64 ilmenite and pyrite are present as accessory minerals. Ilmenite is present in the matrix,

65 whereas rutile occurs only as relict cores in the former mineral and within garnet and kyanite

66 (Carosi et al. 2015; Iaccarino et al. 2015). Tonalitic leucosomes, mainly made of quartz +

67 plagioclase + kyanite, occur parallel to the main foliation defined by biotite and white mica

68 (Fig. 2a; Iaccarino et al. 2015). Garnet crystals both in the rock matrix and collected at the

69 base of the outcrop contain analogous mineral and melt inclusions and show exactly similar

70 compositional features (Carosi et al. 2015; Iaccarino et al. 2015), indicating that the dispersed

71 investigated garnet come from adjacent Ky-bearing migmatites.

72    Phase equilibria modeling of migmatitic paragneisses indicates peak conditions of

73 710–720 °C and 1.0-1.1 GPa, whereas the application of Zr-in-rutile thermometer yielded

74 temperatures of ≈650–720 °C (at P=1.1GPa), interpreted as prograde temperatures (Fig. 2b;

75 Iaccarino et al. 2015). These rocks underwent anatexis with the subsequent formation of

76 leucosomes close to peak conditions, and then decompression and cooling to 650–670 °C and

5

77  0.7-0.8 GPa (Iaccarino et al. 2015). The calculated amount of melt generated at peak

78  conditions is ≈7.5 vol% (Fig. 2b), approaching the critical threshold for melt interconnection

79  and migration (≈7 vol%; Rosenberg and Handy 2005). As a matter of fact, melt accumulation

80  is observed in leucosomes and in dilational structures, whose total volume is far more than

81  the predicted amount of melt at peak conditions (Fig. 2a) (see Iaccarino et al. 2015).

82      In situ U–Th–Pb dating of monazite has constrained the partial melting process at c.

83  41–28 Ma (Fig. 2b; Carosi et al. 2015; Iaccarino et al. 2015). On the basis of these and

84  previous geochronological constraints, crustal anatexis in the kyanite stability field in Kali

85  Gandaki valley, and in general in central and eastern Nepal, was ascribed to the

86  underthrusting of the GHS beneath THS following continental collision (Carosi et al. 2015).

87

## Nanogranitoid inclusions

89  The investigated garnet is 10 mm in diameter and contains hundreds of nanogranitoid

90  inclusions showing a systematic distribution in an annulus located in an intermediate position

91  between core and rim (Fig. 3a, b). Nanogranitoids are 10–20 μm in size, often show an

92  irregular shape and coexist with tiny single mineral inclusions of ilmenite, rutile, apatite,

93  zircon and monazite (Fig. 3c-f). These minerals also appear within melt inclusions (Fig. 3f)

94  and represent accidental trapped phases, as suggested by their relative large size compared

95  with the size of the inclusions, by their low solubility in crustal melts and by the presence of

96  indentations within the inclusion walls (e.g., Cesare and Maineri 1999; Barich et al. 2014).

97  Kyanite is not present as mineral inclusions within the investigated garnet, as already

98  observed for other garnet porphyroblasts occurring in Ky-bearing migmatites from this area

99  (Carosi et al. 2015). Cryptocrystalline aggregates of quartz + plagioclase + muscovite +

100  chlorite ± biotite commonly constitute nanogranitoids (Fig. 3c-f; see also Carosi et al. 2015),

101  suggesting the former presence of a Si- and Al-rich melt. Experimental re-homogenization

6

102 via piston cylinder apparatus at 800 °C (see details in Bartoli et al. 2013b) has confirmed the

103 granitoid and peraluminous composition of the trapped melt ($SiO_2$ >65 wt%, FeO ≈2–3 wt%,

104 CaO ≈1–4 wt%, $Na_2O$ ≈1–4 wt%, $K_2O$ ≈0.3–2, $P_2O_5$ ≈0.2 wt% ASI≥1.3; Carosi et al. 2015;

105 Cesare et al. 2015).

106

## Analytical methods

108 Trace elements in natural (experimentally untreated) and unexposed inclusions beneath

109 garnet surface (Table 1) have been analyzed at the Department of Earth Sciences, ETH

110 Zürich by laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS) using a

111 GeoLas system for ablation coupled to an Element XR (Thermo Fisher, Germany). The glass

112 NIST SRM 610 was employed as external standard. The acquired time-resolved signals were

113 processed using the software SILLS (Guillong et al. 2008). Owing to the small size of

114 nanogranitoids, also part of the host garnet has been ablated along with target inclusions.

115 Therefore, the deconvolution of the resulting mixed garnet-glass signal was applied following

116 the procedure of Halter et al. (2002). The $Na_2O$ and FeO contents determined by EMP were

117 used as reference standards for inclusion and garnet, respectively. For elements enriched in

118 the inclusion, the concentrations is accurate within 20 to 40%. Elements which are strongly

119 compatible in the host garnet (HREE, Y) are normally not detectable in the inclusions due to

120 the dominance of the host signal and were not collected. Gd, V, Zn and Sc show substantially

121 higher contents in the host garnet than in nanogranitoids (Table 1). Therefore, the amount of

122 these elements cannot be reliably estimated in the inclusions. Some inclusions show

123 anomalous very high concentrations of Zr, Th, U, Hf and Nb, suggesting the occurrence of

124 accidental trapped phases (e.g., zircon, monazite, rutile and ilmenite). For these inclusions,

125 only the content of Li, B, Rb, Sr, Ba was reported in Table 1 and discussed in the text.

126 The $H_2O$ measurements of re-homogenized melt (i.e., glassy) inclusions (Table 2) were

performed using the Cameca Nano Secondary Ion Mass Spectrometry 50 (NanoSIMS)

128 installed at Muséum National d'Histoire Naturelle (Paris). Polished experimental capsules

129 with glassy inclusions exposed on the garnet surface and standard glasses were mounted in

130 indium (Aubaud et al. 2007). Inclusions were identified by collecting secondary ion images

131 of Si, Al, K and Fe. For each analysis location, a pre-sputtering step on a 3x3 $\mu m^2$ surface

132 area was firstly done for 2 minutes with a 360 pA primary $Cs^+$ beam to remove the gold

133 coating, surface contamination and to reach a steady state sputtering regime (Thomen et al.

134 2014). Then a primary beam of 60 pA was used for data acquisition. Data were acquired by

135 rastering a 3x3 $\mu m^2$ surface area and collecting ions only from the inner 1x1 $\mu m^2$ (using the

136 beam blanking mode) to reduce surface and redeposition contamination. In order to

137 compensate charging effects, the electron gun was used while pre-sputtering and collecting

138 time. Each analysis is a stack of 200 cycles, a cycle being 1.024s long. $^{16}OH^-$ (used as a proxy

139 for $H_2O$), $^{28}Si^-$, $^{39}K^{16}O^-$ and $^{56}Fe^{16}O^-$ were recorded simultaneously in multicollection mode.

140 We checked that $^{16}OH^-/^{28}Si^-$ ratio was stable during analyses. Secondary ions were collected

141 by electron multipliers with a dead time of 44 ns. Mass resolution was set to 8000, enough to

142 resolve isobaric interference on $^{16}OH^-$. For NanoSIMS calibration we used granitic glasses

143 showing different $H_2O$ contents (for details on calibration procedure and correction curve see

144 Bartoli et al. 2014). Data corrections, using the aforementioned calibration, and error

145 calculations were performed using the R program. Errors combine counting statistic and

146 uncertainty of the calibration curve. However, the errors reported in Table 2 are dominated

147 by the uncertainty of the calibration curve, which corresponds to prediction interval at 68%.

148 During the session, the vacuum in the analysis chamber remained between 2.5 and 5 $x10^{-10}$

149 Torr.

150

151

## Results

152

153     Melt inclusions show highly variable concentrations of typical incompatible elements

154     such as Li (40–400 ppm), B (200–3400 ppm), Rb (40–400 ppm), Sr (5–100) and Ba (30–900

155     ppm) (Table 1). The contents of elements typically controlled by the dissolution of accessory

156     minerals are very low (Zr= 3–8 ppm; Th=0.1–1.2 ppm; LREE=4–11 ppm). U ranges from 1

157     to 8 ppm. The Nb/Ta and U/Th ratios are 3–15 and 3–21, respectively. Nanogranitoids also

158     contain appreciable amounts of Sc (3–76 ppm), V (0.4–20 ppm) and Zn (14–92 ppm), though

159     these concentrations might be affected by contamination from the host (see above and Fig.

160     5a). When compared with the composition of the upper continental crust, nanogranitoids

161     from Kali Gandaki are enriched in Pb, Li and B, and depleted in Ba, Th, Nb, LREE, MREE,

162     Sr, Zr and Hf (Fig. 4a). On the other hand, Rb, U and Ta show variable degrees of enrichment

163     and depletion. In the chondrite-normalized diagram, melt inclusions show weak to moderate

164     positive Eu anomalies (Eu/Eu* = 1.2-3.3) (Fig. 4b).

165     The $H_2O$ concentrations of the re-homogenized nanogranitoids determined by

166     NanoSIMS span a relatively wide range of values from 6.5 to 10.5 wt%, with one single

167     value up to 14.4 wt% (mean value of 8.9 ± 1.8 wt%; Fig. 5 and Table 2). Melt inclusions

168     were too small to perform replicated analyses within a single inclusion. However, the

169     distribution of $H_2O$ is expected to be quite homogeneous in glassy inclusions obtained from

170     remelting experiments (Bartoli et al. 2014; Acosta-Vigil et al. 2016). $H_2O$ estimated by

171     difference of electron microprobe totals from 100% yielded similar mean content (9.6 ± 1.4

172     wt%) compared to the NanoSIMS mean data (Fig. 5).

173     Figure 6 shows zircon and monazite saturation temperatures ($T_{Zrn}$ and $T_{Mnz}$,

174     respectively) for investigated melt inclusions, calculated following methods proposed by

175     Watson and Harrison (1984), Montel (1993) and Stepanov et al. (2012). $T_{Zrn}$ and $T_{Mnz}$ were

176     calculated from nanogranitoids not containing accidental trapped minerals; we carefully

9

177 avoided considering any data coming from analyses of nanogranitoids containing trapped

178 mineral inclusions such zircon or monazite (see above). $T_{Zrn}$ values vary from 541 to 594 °C.

179 The two different calibrations for the $T_{Mnz}$ thermometer provide different temperature ranges

180 such as 544–650 °C (Montel 1993) and 412–533 °C (Stepanov et al. 2012).

181

## Phase equilibria modeling

183 Migmatites from Kali Gandaki valley have been already investigated by means of phase

184 equilibria modeling (Iaccarino et al. 2015). In this contribution this approach was instead

185 applied considering only the average bulk composition of remelted nanogranitoids to

186 compare the predicted subsolidus mineral assemblages with those observed in the crystallized

187 melt inclusions. The average $H_2O$ content was estimated by EMP-difference and is consistent

188 with the NanoSIMS values (Fig. 5). Calculations were performed in the $Na_2O$–CaO–$K_2O$–

189 FeO–MgO–$Al_2O_3$–$SiO_2$–$H_2O$ (NCKFMASH) chemical system, using Perple_x software

190 version 6.8.3. Considering the pitfalls and uncertainties related to phase equilibria approach

191 (White et al. 2011; Bartoli et al. 2013c, 2016b; Forshaw et al. 2019) two *P–T* sections were

192 calculated for the same bulk composition, one using the internally consistent dataset hp04ver

193 (Holland and Powell 2003) and the older activity–composition (a–x) models, and another

194 calculated using dataset hp11ver (Holland and Powell 2011) and the re-parameterized a–x

195 models. The following a–x relations were selected. For hp11ver: melt, garnet, chlorite,

196 orthopyroxene, muscovite, biotite and cordierite from White et al. (2014), ilmenite from

197 White et al. (2000), plagioclase and K-feldspar from Holland & Powell (2003). For hp04ver:

198 melt from White et al. (2007), garnet, orthopyroxene and chlorite from Holland and Powell

199 (1998), biotite from Tajčmanová et al. (2009), white mica from Coggon and Holland (2002),

200 plagioclase from Newton et al. (1980) and K-feldspar from Thompson and Hovis (1979). An

201 ideal model was used for cordierite and ilmenite. In all calculations pure phases included

10

202    quartz, rutile, kyanite, sillimanite, andalusite and aqueous fluid.

203        The overall topology of the two $P$–$T$ sections appears somewhat different, but there

204    are not important variations in terms of phase assemblages (Fig. 7). Modeling of the bulk

205    composition for the investigated nanogranitoids shows the appearance of melt at T> 680-690

206    °C (Fig. 7). The subsolidus field is occupied by variable mineral assemblages which are

207    mainly characterized by the presence of chlorite, biotite, muscovite, paragonite, garnet,

208    plagioclase and quartz. At <500 °C and <0.4 GPa, the predicted stable mineral assemblages

209    are quartz + plagioclase + muscovite + chlorite + biotite and quartz + plagioclase +

210    muscovite + chlorite (Fig. 7). These assemblages resemble those observed in many

211    nanogranitoid inclusions (see above). Some pores observed during SEM analysis (Fig. 3)

212    likely reflect the former presence of volatiles (in particular $H_2O$) in excess to the amounts

213    used by phyllosilicates and therefore exsolved during melt crystallization (see Bartoli et al.

214    2013a), in agreement with free $H_2O$ predicted by the model at subsolidus conditions (Fig. 7).

215    In addition, the model indicates that chlorite is likely to be a phase formed in the subsolidus

216    during cooling, whose formation did not involve infiltration of external fluids (i.e., the

217    presence of chlorite in nanogranitoids does not imply that melt inclusions have been modify

218    by late stage fluid alteration). This finding is also consistent with the coexistence of chlorite

219    and unaltered biotite in nanogranitoids from Ivrea Zone (Carvalho et al., 2018).

220

## Discussion

**Nanogranitoids represent primary prograde Eohimalayan crustal melts**

223    Their microstructural distribution, together with the experimental re-homogenization, major

224    and trace element compositions and $H_2O$ contents, and results from phase equilibria modeling

225    (Carosi et al. 2015 and this work), confirm that these polycrystalline inclusions represent

226    former primary melt droplets that have crystallized into a cryptocrystalline aggregate; that is,

11

227  they are nanogranitoids (Cesare et al. 2015; Bartoli et al. 2016a). Their systematic

228  distribution in an annulus around the garnet core points to a primary origin (Roedder 1979),

229  implying that they were trapped by garnet during its growth. Therefore, the portion of garnet

230  containing melt inclusions certainly formed in the presence of an anatectic melt during

231  incongruent melting (i.e., this portion has certainly a peritectic origin: Cesare et al. 2015).

232  This is consistent with the increase of garnet mode (5 to 9 vol%; Iaccarino et al. 2015)

233  predicted by the modeling at the temperature of melting (Fig. 2b). The formation of peritectic

234  garnet at ≤700 °C might appear problematic, but it was already documented in Ronda

235  metatexites (S Spain), where it was interpreted as the consequence of a continuous melting

236  reaction consuming biotite (Bartoli et al. 2013c). Also, geochemical signatures typical of

237  muscovite melting were found in melt inclusions hosted in peritectic garnet from El Hoyazo

238  anatectic enclaves (SE Spain), indicating that this peritectic mineral likely grew (and trapped

239  melt droplets) through melting reactions consuming Fe-bearing muscovite at around 700-750

240  °C (Acosta-Vigil et al. 2010).

241  The investigated garnet has been collected at the base of a rock wall made of Ky-

242  bearing, partially- melted gneiss. Rocks from this outcrop have been intensively studied in

243  the recent years. On the basis of thermodynamic modeling, peritectic garnet likely grew in

244  the presence of a melt phase from 650 to 720 °C, at 1.0–1.1 GPa (Fig. 2b; Iaccarino et al.

245  2015). In situ U–Th–Pb dating of monazite grains from rock matrix and included in melt

246  inclusion-bearing garnets indicates the beginning of anatexis at c. 41–36 Ma (Fig. 2a; see also

247  Carosi et al. 2015). Near peak-pressure melting has been dated at c. 36–28 Ma by Iaccarino et

248  al., (2015). These geochronological constraints are in agreement with the age of c. 35 Ma

249  obtained by U–Pb monazite and zircon dating of a Ky-bearing leucosome collected close to

250  Taglung, 10-12 km from the Titar village (Godin et al. 2001).

251  Considering their primary origin and the geochronological constraints reported above,

252 it is reasonable to interpret that the investigated nanogranitoids reflect primary anatectic

253 melts produced during the Eocene-Early Oligocene (Eohimalayan) prograde metamorphism

254 associated to crustal thickening during the early stages of the India–Asia collision.

255

**Conditions and mechanisms of Eohimalayan anatexis in Kali Gandaki**

257    Most re-homogenized nanogranitoids from Kali Gandaki valley are granodiorites,

258 tonalites and trondhjemites, and minor granites, showing low $K_2O$ contents ($< 2.5$ wt%),

259 moderate to very high $Na_2O/K_2O$ ratio (1.1-14.1) and CaO up to 4 wt% (Carosi et al., 2015;

260 Bartoli et al., 2016a; Fig. 8). Notably, leucosomes in Kali Gandaki gneisses exhibit a tonalitic

261 composition (Iaccarino et al. 2015). These peculiar compositions resemble those of

262 experimental melts produced during $H_2O$-present melting of metasedimentary rocks (Fig. 8).

263 For instance, Patiño Douce and Harris (1998) experimentally remelted two GHS schists from

264 the hanging wall of the Main Central Thrust. Melts produced by dehydration melting of mica

265 at >750 °C (0.6-1.0 GPa) are granitic in compositions ($Na_2O/K_2O = 0.7–0.9$, one value up to

266 1.4). Instead, near-solidus melts formed at <750 °C by $H_2O$-present melting of muscovite

267 correspond to trondhjemites ($Na_2O/K_2O =1.4–3.9$, one value of 0.9). The $Na_2O/K_2O > 1$ of

268 melt has been ascribed to the decreasing temperature of quartz + plagioclase solidus with

269 increasing $H_2O$ amount (Patiño Douce and Harris 1998; García-Casco et al. 2003). The weak

270 to moderate positive Eu anomaly of Kali Gandaki nanogranitoids (Fig. 4b) would support the

271 role of plagioclase in melting reaction (Gao et al. 2017). Therefore, based on the major

272 element compositions of rehomogenized nanogranitoids, the presence of an aqueous fluid

273 phase during the associated anatexis could be proposed.

274    The peculiar Na-rich composition of Kali Gandaki nanogranitoids with respect to

275 classic granitic s.s. anatectic melts cannot be the result of melt-garnet interaction during

276 experimental remelting at temperature (800 °C) higher than that of entrapment (650-720 °C),

13

277 because the host mineral is virtually Na-free. Rather, this discrepancy is likely to be related to

278 kinetics (see below).

279      Rehomogenized nanogranitoids have high $H_2O$ concentrations (6.5-10.5, one value up

280 to 14.4 wt%; Fig. 5). This range approximates the expected minimum and maximum $H_2O$

281 contents for anatectic melts produced at these P–T conditions ($\approx$6.5-15 wt%; Holtz et al.,

282 2001). The $H_2O$ content approaching the maximum solubility value ($\approx$15 wt.% at 1.1 GPa)

283 was found in a single nanogranitoid inclusion and it would be consistent with partial melting

284 under fluid-present conditions at the $H_2O$-saturated solidus. The lower $H_2O$ concentrations

285 may reflect different processes. 1) Weinberg and Hasalová (2015b) argue that in the case of a

286 rock-buffered system (i.e., a system characterized by a high rock/fluid ratio) the $H_2O$-fluxed

287 melting would produce $H_2O$-undersaturated melts and that the $H_2O$ contents would range

288 between the minimum and maximum values. 2) Alternatively, different buffering

289 assemblages at the micro-scale may be responsible for variable melt $H_2O$ content. Bartoli et

290 al. (2014) documented that melt inclusions trapped in different garnet crystals in Ronda

291 migmatites (Spain) showed variable $H_2O$ content and interpreted these heterogeneities as the

292 result of melt formation under conditions of "mosaic equilibrium" affecting the activity of

293 $H_2O$ in the coexisting fluid phase and/or the solid assemblages. This explanation seems to be

294 highly improbable for Kali Gandaki sample where the distance among analyzed

295 nanogranitoids is in the order of a few hundreds of μm. 3) The loss of $H_2O$ after melt

296 inclusion entrapment and/or during remelting experiments cannot be excluded *a priori*,

297 considering the presence of cracks in some inclusions (Fig. 3c). However, a systematic loss

298 of $H_2O$ from nanogranitoids during subsolidus retrograde history is expected to result in a

299 larger variability of the melt $H_2O$ contents, with compositions approaching dry values. This is

300 not the case here, where the majority of melt $H_2O$ contents define a relatively narrow range,

301 never below 6.5 wt% (Fig. 5).

302    One could argue that $H_2O$ loss after melt inclusion entrapment could partly explain

303    the discrepancy between the trapping temperature and the temperature at which

304    nanogranitoids were experimentally remelted (800 °C). However, again, this discrepancy

305    may be explained without taking into account the loss of $H_2O$. Indeed, higher remelting

306    temperatures could be needed because melting kinetics play a fundamental role during

307    reheating experiments. Smaller inclusions usually remelt more rapidly than larger ones

308    (Thomas and Klemm 1997) and nanogranitoids from Kali Gandaki show larger size than melt

309    inclusions for which a run duration of 24 h was tested (Bartoli et al., 2013c and discussion

310    therein). Therefore, a process of $H_2O$ loss affecting the majority of analyzed inclusions does

311    not seem to be supported by convincing evidence.

312    To summarize, the occurrence of near $H_2O$-saturated melts would support a $H_2O$-

313    present melting scenario, confirming the inferences based on major element compositions of

314    nanogranitoids and temperature estimates of Kali Gandaki migmatites (see above). The

315    observed $H_2O$ variability, which is likely to be primary and ranges between the expected

316    minimum and maximum values, would suggest the existence of a $H_2O$-present, rock-buffered

317    system during Eohimalayan anatexis. These observations are consistent with anatexis and

318    melt formation in the proximity of the $H_2O$-saturated solidus (Fig. 2).

319    $T_{Zrn}$ and $T_{Mnz}$ are ≈100-250 °C lower than suprasolidus temperatures obtained for

320    migmatites collected in the same area (Fig. 6). $T_{Zrn}$ and $T_{Mnz}$ values are also notably lower

321    than the minimum temperature at which crustal protoliths melt in the presence of an aqueous

322    fluid (≈650 °C). Migmatites from Kali Gandaki valley contain abundant grains of zircon and

323    monazite occurring in matrix as well as in garnet, kyanite, biotite and quartz (Carosi et al.

324    2015; Iaccarino et al. 2015). In addition, these accessory minerals are present as accidentally

325    trapped phases within some nanogranitoids (Fig. 3d, f). One, therefore, could assume that

326    melt pockets were likely in contact with zircon and monazite during the early stages of

327 melting. Because zircon and monazite are expected to break down and dissolve at

328 suprasolidus conditions during prograde metamorphism to maintain saturation of the melt

329 (Yakymchuk and Brown 2014; Yakymchuk 2017), the undersaturated composition of melt

330 inclusions with respect to the accessory minerals would suggest a limited time of interaction

331 between zircon/monazite and anatectic melt. On the other hand, the amount of melt was

332 likely low (around the percolation threshold; Fig. 2) and, therefore, the lack of contact with

333 accessory phases cannot be excluded. In any case, the very low $T_{Zrn}$ and $T_{Mnz}$ values have to

334 be ascribed to the lack of equilibration between melt pockets and accessory minerals present

335 in the rock. Some degrees of disequilibrium involving accessory minerals and melt inclusions

336 in anatectic rocks have been already documented during the entrapment of low-temperature

337 (<750 °C) melt droplets in El Hoyazo metapelitic enclaves (Acosta-Vigil et al. 2010). In that

338 case, a much smaller difference of 25-50 °C between calculated and expected temperatures

339 has been reported (see discussion in Acosta-Vigil et al. 2012).

340       To date, the most complete dataset of trace element compositions of melt inclusions

341 from metapelitic anatectic rocks is that presented by Acosta-Vigil et al. (2010, 2012). In

342 residual anatectic enclaves within peraluminous dacites of the Neogene Volcanic Province

343 (NVP) of SE Spain, a systematic evolution of trace element content of anatectic melt

344 (recorded as melt inclusions in several minerals and as matrix glass) is observed. Combining

345 microstructural and geochemical data, these authors interpreted that concentrations in melt of

346 Li, Cs and B decrease whereas those of Zr, Th, REE, V, Cr, Co, Zn increase with increasing

347 melting temperature, due to the consumption of muscovite (main carrier of Li, B and Cs) at

348 low temperature and biotite (main reservoir of Rb) and accessory phases (LREE, Zr, Hf, U,

349 Th) at higher temperature (Acosta-Vigil et al. 2010). These authors interpreted the melt

350 inclusions in plagioclase and garnet to represent crustal melts formed by the $H_2O$-present to

351 $H_2O$-absent muscovite melting reactions, whereas the matrix glass formed by biotite fluid-

16

352 absent melting (see also Acosta-Vigil et al. 2017).

353 The majority of investigated nanogranitoids from Kali Gandaki valley show higher B

354 content, similar Li concentration and Rb/Sr ratio, and lower Ba, Ce, Th, Zr, Hf and Rb/B

355 values with respect to NVP melts produced at ≤750 °C (Fig. 9). The U/Th ratio is variable,

356 from values approaching those observed in El Hoyazo low-temperature melts up to much

357 greater contents (> 10; Fig. 9g), suggesting that Kali Gandaki nanogranitoids could reflect the

358 composition of the first melt produced. All these compositional features are consistent with i)

359 melting at low temperature and ii) the undersaturation of melt with respect to the accessory

360 minerals (see above). Two melt inclusions from the investigated garnet display very high Ba

361 amount (> 600 ppm; Fig. 9c). The high B contents (Fig. 9a, b) may point to the presence and

362 dissolution of tourmaline during the melt-producing reaction. Indeed, this mineral is a

363 common accessory phase in Ky-bearing gneisses from this area (Iaccarino et al. 2015). Wolf

364 and London (1997) experimentally demonstrated that at the onset of crustal melting

365 tourmaline breaks down incongruently to assemblages containing Al-, Fe- and/or Mg-bearing

366 solid phases (i.e., garnet, biotite, sillimanite) and a B-enriched melt. Alternatively, the

367 infiltration of B-rich fluids into the anatectic zone and formation of B-enriched melts cannot

368 be ruled out (Acosta et al. 2001; and references therein) and it would be consistent with a

369 melting scenario characterized by the presence of an aqueous fluid (see above).

370 Acosta-Vigil et al. (2012) modeled REE abundance in crustal melts of El Hoyazo

371 considering different scenarios (e.g., disequilibrium melting of feldspars, equilibrium

372 dissolution of accessory phases and equilibrium with major and accessory minerals).

373 Similarly, Prince et al. (2001) calculated REE profiles of crustal melts derived from GHS

374 metapelites assuming that REE abundances are controlled by accessory phase dissolution and

375 by disequilibrium melting of plagioclase. Collectively these calculations indicate that REE-

376 poor melts with a strong positive Eu anomaly can be due to the disequilibrium melting of

17

377 feldspars, whereas dissolution of accessory phases results in higher REE contents and

378 negative Eu anomalies (Fig. 10). The REE pattern and positive Eu anomalies of Kali Gandaki

379 nanogranioids resemble those of El Hoyazo melt inclusions hosted in garnet and models

380 based on the dissolution of feldpars, and differ significantly from models associated with the

381 dissolution of accessory minerals (Fig. 10). Therefore, it follows that the latter process is not

382 likely responsible for the REE abundances of nanogranitoids. Also, the extremely variable,

383 LREE concentrations in melt suggest the lack of equilibration between melt and residue.

384

**Eohimalayan anatexis and granitoid formation**

386 Kyanite-bearing migmatitic gneisses of the lower GHS in the Kali Gandaki valley underwent

387 low temperature (up to 720 °C) Eocene to Early Oligocene anatexis, with the formation of

388 low amounts of melt (≈7.5 vol.%; Fig. 2b) at peak conditions (Carosi et al. 2015; Iaccarino et

389 al. 2015). However, similar rocks, which represent significant portions of the metamorphic

390 core of the Himalayas, experienced higher temperatures and/or higher melting degrees in

391 other localities of GHS, between Eocene and Late Oligocene.

392 　　　　For example, the Ky-bearing migmatites in the Main Central Thrust Zone (Arun

393 valley, eastern Nepal Himalaya) reached peak conditions at 820 °C and 1.3 GPa, during Early

394 Oligocene (≈ 31 Ma), producing ≈15 vol.% melt (Groppo et al. 2010). Guilmette et al. (2011)

395 reported peak conditions of 820 °C and > 1.4 GPa for Ky-bearing anatectic paragneisses from

396 the Namche Barwa (Eastern Himalayan Syntaxis). Here partial melting is considered to have

397 generated 20–23 vol.% total melt. The HP metamorphism was dated at 37–32 Ma (Zhang et

398 al. 2010). Zhang et al. (2015) documented melting of pelitic granulites from the same area

399 (Namche Barwa Complex) at peak conditions of 840–880 °C and 1.3–1.6 GPa; anatexis of

400 these rocks produced up to ≈30 vol.% melt from c. 40 to 30 Ma. Here metagreywackes

401 reached peak consitions of 820–845 °C and 1.5–1.6 GPa (Tian et al. 2016). Palin et al. (2014)

402 dated the melting at ≈700 °C and ≈1.0 GPa in the northwestern flank of the Eastern

403 Himalayan Syntaxis between c. 44 and 33 Ma. Leucosomes in Ky-bearing migmatites

404 comprise up to 20% of the outcrop and formed as a result of Kfs-absent, fluid present melting

405 reactions. In the Kharta Area (Southern Tibet) metasedimentary migmatites experienced

406 anatectic conditions of 750–800 °C and 1.4 GPa, at ≈33 Ma (Liu et al. 2007). Phase equilibria

407 modeling demonstrated that Barun gneisses (Eastern Nepal) reached peak conditions of 800–

408 850 °C in the kyanite field, with the formation of 15-20 vol.% melt (Groppo et al. 2012).

409 Melting at the bottom of the GHS (e.g., Barun gneisses) was dated at 33–28 Ma (Carosi et al.

410 2015 and references therein).

411 Considering their low melting temperature and degree, Kali Gandaki anatectic

412 gneisses probably did not act as source region of voluminous anatectic melts. However, from

413 the above, it is clear that Ky-bearing migmatites have played an important role during

414 Eohimalayan anatexis. As a matter of fact, melting of metasedimentary lithologies from GHS

415 is considered to have contributed to the granitoid magmatism occurred during crustal

416 thickening at variable temperatures and under variable fluid regimes (Prince et al. 2001; Zeng

417 et al. 2009; 2011; Aikman et al. 2012). Therefore, nanogranitoids from Kali Gandaki gneisses

418 may provide us with a snapshot of anatectic melts whose extraction from similar lithologies

419 and emplacement at shallower levels have resulted in the formation of Eohimalayan

420 granitoids.

421 Granitic to trondhjemitic bodies from the upper GHS of Garhwal Himalaya are

422 considered to be derived from $H_2O$-present melting of Ky-bearing GHS metasedimentary

423 rocks at 650 °C during Eohimalayan prograde metamorphism (39 ± 3 Ma) (Prince et al.

424 2001). These Eohimalayan granitoids, which differ from the classic Miocene leucogranites,

425 share some key features with the Kali Gandaki melt inclusions, such as Ca- and Na-

426 enrichment (up to 2.0 and 3.9 wt%, respectively), very low melting temperatures and a

427 positive Eu anomaly (Prince et al. 2001). In the haplogranitic Qtz-Ab-Or diagram the

428 Garhwal rocks plot in the proximity of Kali Gandaki nanogranitoids and experimental glasses

429 produced via $H_2O$-present melting (Fig. 8). Anatectic gneisses of Kali Gandaki valley may

430 reasonably be considered the analog source rocks of Eohimalayan Garhwal granitoids. Active

431 shear zones in this area could have facilitated the ingress of external fluids, enhancing the

432 amount of melt produced (Prince et al. 2001). Instead, a large supply of external aqueous

433 fluids did not occur in Kali Gandaki area where melting stopped at low degrees.

434     Granite compositions have been often interpreted in the light of "restite unmixing"

435 model (Chappell et al. 1987), which states that the majority of granites represent mixtures

436 between a minimum or near-minimum temperature melt and the restite (residuum), and that

437 the compositional variability in granites reflects varying degrees of restite unmixing.

438     The Ba and Sr content of three Garhwal granites can be modeled as the variable

439 mixture of a starting melt (represented by nanogranitoids) and a residuum composed of

440 variable amounts of biotite, plagioclase and K-feldspar (Fig. 9d). One rock needs solely the

441 addition of 5 vol% K-feldspar. However, the same mixtures with a biotite-bearing residuum

442 would result in Rb/Sr ratios much higher (≈12) than those measured (<1; Fig. 9c). The very

443 high Ba content and very low Rb/Sr of Garhwal rocks could, instead, be explained by

444 addition of solely K-feldspar, up to 20 vol% (Fig. 9c). These mismatches clearly indicate that

445 entrapment of residual minerals does not explain the trace element concentrations controlled

446 by major minerals. Although the addition of very tiny amounts of zircon and monazite could

447 explain Ce, Th, U and Zr variability in Garhwal rocks (Fig. 9e, f), this is unlikely. Firstly, the

448 entrainment of phases from the source region is not a selective process (Sawyer 2014).

449 Rather, the inheritance of accessory minerals should be coupled with that of major phases in

450 which they are usually included (micas and feldspars). Again, the mixing with a biotite-,

451 feldspar-bearing residuum is highly improbable (see above). Secondly, the higher Zr, Th and

20

452 Ce contents are perfectly consistent with higher degrees of melting compared to those in Kali

453 Gandaki migmatites. Accessory phase dissolution and disequilibrium melting of plagioclase

454 are considered to have shaped the LREE patterns and positive Eu anomaly, respectively, of

455 Garhwal rocks (Fig. 10; Prince et al. 2001). This is mostly in agreement with the diverse La,

456 Ce and Eu contents of Kali Gandaki nanogranitoids whose composition seems to be

457 controlled by the dissolution of plagioclase and lack of equilibrium with accessory minerals

458 (Fig. 10).

459 Prince et al. (2001) ascribed the very high Ba content of these rocks (> 600 ppm; Fig.

460 9c) to fluid transport of Ba, which "is a well-documented phenomenon during crustal melting

461 (Weinberg and Searle 1999)". Surprisingly, Ba is highly variable in Kali Gandaki

462 nanogranitoids, formed in a $H_2O$-present scenario (see above), with the highest contents

463 resembling those observed in some Garhwal rocks (Fig. 9c).

464 Summarizing, the comparison between nanogranitoids and granites indicates i) that

465 the restite unmixing process did not shape the composition of Garhwal rocks and that an

466 aqueous fluid was involved in melting process, reinforcing the inferences of Prince et al.

467 (2001), and ii) that Kali Gandaki melt inclusions are recording the very beginning of melting

468 in a scenario that, in other GHS localities, has produced Eohimalayan granitoids.

469

**Conclusions**

471    1. Kali Gandaki nanogranitoids occur within peritectic garnet whose formation has been

472        ascribed to the Eohimalayan prograde metamorphism at conditions of 650–720 °C

473        and 1.0–1.1 GPa and has been dated at c. 41–36 Ma (see Carosi et al. 2015; Iaccarino

474        et al. 2015).

475    2. During their crystallization, nanogranitoids developed a quartz + plagioclase +

476        muscovite + chlorite ± biotite assemblage.

21

3. Major element compositions and $H_2O$ contents of rehomogenized nanogranitoids suggest the presence of an aqueous fluid phase during melting, in agreement with the inferred melting conditions.

4. The low Zr, Th and LREE concentrations along with the very low zircon and monazite saturation temperatures indicate some degrees of disequilibrium between accessory minerals and coexisting melt.

5. Investigated nanogranitoids are incipient melts, which did not have time to equilibrate with their source rock.

6. Kali Gandaki nanogranitoids share some key features with Eohimalayan granitoids from Garhwal Himalaya (i.e., major elements, very low melting temperature, Ba enrichment). A comparative study supports previous inferences of Prince et al. (2001): Garhwal rocks likely represent pristine crustal melts formed under $H_2O$-present conditions, whose composition was not modified by entrainment of residuum.

500　　FIGURE and TABLE CAPTIONS

501

502　**Fig. 1** Geological sketch map of the Kali Gandaki valley (redrawn from Carosi et al. 2015)

503

504　**Fig. 2** a) Representative hand specimen of the anatectic paragneisses. b) Phase equilibria

505　modeling and P–T path of Ky-bearing gneisses from Kali Gandaki valley (redrawn from

506　Iaccarino et al., 2015). Green rectangle: temperature range from the thermometer based on Zr

507　content in rutile. C2015 and I2015: geochronological constraints from Carosi et al. (2015)

508　and Iaccarino et al. (2015), respectively

509

510　**Fig. 3** Melt inclusions in garnet from Kali Gandaki. a) plane-polarized photomicrograph of

511　the equatorial section of the investigated garnet. Dashed white lines define the annulus

512　containing nanogranitoid inclusions (redrawn after Cesare et al. 2015). b) plane-polarized

513　photomicrograph showing a cluster of polycrystalline inclusions. c-f) SEM-BSE images of

514　nanogranitoids. Red arrows: decrepitation cracks. Yellow arrows: micro- to nano-porosity.

515　Scale bar: 5 μm

516

517　**Fig. 4** Trace element and REE composition of nanogranitoid inclusions (NG) and host garnet.

518　a) Values normalized to the average upper continental crust (UCC) composition (data from

519　Rudnick and Gao 2014). b) REE patterns. Values normalized to the chondrite C1 (Sun and

520　McDonough 1989)

521

522　**Fig. 5** $H_2O$ concentrations of nanogranitoid inclusions determined by NanoSIMS. The error

523　associated to these values is smaller than the symbol (see Table 2). White symbols: average

524　contents. The average $H_2O$ content estimated by EMP-difference is from Cesare et al. (2015).

23

525 Horizontal bars are two standard deviations on average values. The number of analyses is

526 indicated next to horizontal bars

527

528 **Fig. 6** Monazite and zircon saturation temperatures calculated for the investigated melt

529 inclusions. Black dots: $T_{Mnz}$ calibration of Montel (1993). White dots: $T_{Mnz}$ calibration of

530 Stepanov et al (2012). $T_{Zrn}$ from Watson and Harrison (1984). Grey field: anatectic

531 temperatures estimated for Kali Gandaki gneisses by Iaccarino et al. (2015)

532

533 **Fig. 7** Comparative P–T sections for nanogranitoid inclusions calculated in the NCKFMASH

534 system with two internally consistent datasets: hp04ver for (a) and hp11ver for (b). The bulk

535 composition (mol%) used for calculation is indicated in the upper right inset and represents

536 the average composition obtained by EMP. The average $H_2O$ content was estimated by EMP-

537 difference and is consistent with the NanoSIMS values

538

539 **Fig. 8** Qtz-Ab-Or diagram showing the normative compositions of Kali Gandaki

540 nanogranitoids (square represents the average value). Compositions of experimental glasses

541 (PDH98: Patiño Douce and Harris 1998; GC03: García-Casco et al. 2003) and Eohimalayan

542 granites (P2001: Prince et al. 2001) are plotted for comparison

543

544 **Fig. 9** Comparison among the trace element contents (ppm) of Kali Gandaki nanogranitoids,

545 El Hoyazo melts and Eohimalayan granitoids. Ba and Sr content of melt inclusions in

546 plagioclase were not reported because these elements are depleted being compatible with

547 respect the host phase; Hf concentrations were not reported for inclusions in plagioclase in

548 the original database (see Acosta-Vigil et al. 2012). Temperature estimates correspond to

549 zircon and monazite saturation temperatures reported by Acosta-Vigil et al. (2010). Lines

24

550 track the result of melt–residuum mixing models. Tick marks on the lines reflect increments

551 of the residuum. The average composition of nanogranitoids was considered as representative

552 of the starting melt composition, except for Ba–Rb/Sr and Ba–Sr diagrams where the two Ba-

553 rich melt inclusions were not considered for the average value. This assumption does not

554 change the inferences obtained from the geochemical modeling. Mineral compositions are

555 from Acosta-Vigil et al. (2010). See text for details

556

557 **Fig. 10** Comparison of REE abundances of Kali Gandaki nanogranitoids, Garwhal granites

558 and low-temperature anatectic melts modelled by Prince et al. (2001) [P2001] and Acosta-

559 Vigil et al. (2012) [AAV2012]. See text for details.

560

561 **Table 1** LA-ICP-MS analyses (ppm) of melt inclusions from Kali Gandaki garnet

562

563 **Table 2** $H_2O$ concentrations measured in nanogranitoids by NanoSIMS

564

**References**

566 Acosta A, Pereira MD, Shaw DM, London D (2001) Contrasting behaviour of B during
567     crustal anatexis. Lithos 56:15–31
568 Acosta-Vigil A, Barich A, Bartoli O, Garrido C, Cesare B, Remusat L, Poli S, Raepsaet C
569     (2016) The composition of nanogranitoids in migmatites overlying the Ronda peridotites
570     (Betic Cordillera, S Spain): the anatectic history of a polymetamorphic basement. Contrib
571     Mineral Pet 171:24
572 Acosta-Vigil A, Buick I, Cesare B, London D, Morgan GB VI (2012b) The extent of
573     equilibration between melt and residuum during regional anatexis and its implications for
574     differentiation of the continental crust: a study of partially melted metapelitic enclaves. J
575     Petrol 53:1319–1356
576 Acosta-Vigil A, Buick I, Hermann J, Cesare B, Rubatto D, London D, Morgan GB VI (2010)
577     Mechanisms of crustal anatexis: a geochemical study of partially melted metapelitic
578     enclaves and host dacite, SE Spain. J Petrol 51:785–821
579 Acosta-Vigil A, London D, Morgan VI GB, Cesare B, Buick I, Hermann J, Bartoli O (2017)
580     Primary crustal melt compositions: Insights into the controls, mechanisms and timing of
581     generation from kinetics experiments and melt inclusions. Lithos 286–287:454–479
582 Aikman AB, Harrison TM, Hermann J (2012) The origin of Eo- and Neo-himalayan
583     granitoids, Eastern Tibet. J Asian Earth Sc 58:143–157

25

Aubaud C, Withers AC, Hirschmann MM, Guan Y, Leshin LA, Mackwell SJ, Bell DR (2007) Intercalibration of FTIR and SIMS for hydrogen measurements in glasses and nominally anhydrous minerals. Am Mineral 92:811–828

Barich A, Acosta-Vigil A, Garrido CJ, Cesare B, Tajčmanová L, Bartoli O (2014) Microstructures and petrology of melt inclusions in the anatectic sequence of Jubrique (Betic Cordillera, S Spain): implications for crustal anatexis. Lithos 206–207:303–320

Bartoli O, Acosta-Vigil A, Ferrero S, Cesare B (2016a) Granitoid magmas preserved as melt inclusions in high-grade metamorphic rock. Am Mineral 101:1543–1559

Bartoli O, Acosta-Vigil A, Tajčmanová L, Cesare B, Bodnar RJ (2016b) Using nanogranitoids and phase equilibria modeling to unravel anatexis in the crustal footwall of the Ronda peridotites (Betic Cordillera, S Spain). Lithos 256–257: 282–299

Bartoli O, Cesare B, Poli S, Acosta-Vigil A, Esposito R, Turina A, Bodnar RJ, Angel RJ, Hunter J (2013b) Nanogranite inclusions in migmatitic garnet: behavior during piston cylinder re-melting experiments. Geofluids 13:405–420

Bartoli O, Cesare B, Poli S, Bodnar RJ, Acosta-Vigil A, Frezzotti ML, Meli S (2013a) Recovering the composition of melt and the fluid regime at the onset of crustal anatexis and S-type granite formation. Geology 41:115–118

Bartoli O, Tajčmanová L, Cesare B, Acosta-Vigil A (2013c) Phase equilibria constraints on melting of stromatic migmatites from Ronda (S. Spain): insights on the formation of peritectic garnet. J Metamorphic Geol 31:775–789

Bartoli O, Cesare B, Remusat L, Acosta-Vigil A, Poli S (2014) The $H_2O$ content of granite embryos. Earth Planet Sci Lett 395:281–290

Brown RL, Nazarchuk JH (1993) Annapurna detachment fault in the Greater Himalaya of Central Nepal. Geol Soc London Sp Publ 74:461–473

Carosi R, Gemignani L, Godin L, Iaccarino S, Larson KP, Montomoli C, Rai SM (2014) A geological journey through the deepest gorge on Earth: the Kali Gandaki valley section, west-central Nepal. Journal of Virtual Explorer 47 (paper 7)

Carosi R, Montomoli C, Iaccarino S (2018) 20 years of geological mapping of the metamorphic core across Central and Eastern Himalayas. Earth-Science Rev 177:124–138

Carosi R, Montomoli C, Langone A, Turina A, Cesare B, Iaccarino S, Fascioli L, Visonà D, Ronchi A, Rai SM (2015) Eocene partial melting recorded in peritectic garnets from kyanite-gneiss, Greater Himalayan Sequence, central Nepal. Geol Soc London Sp Publ 412:111

Carvalho BB, Bartoli O, Ferri F, Cesare B, Ferrero F, Remusat L, Capizzi L, Poli S (2018) Anatexis and fluid regime of the deep continental crust: new clues from melt and fluid inclusions in metapelitic migmatites from Ivrea Zone (NW Italy). J Metamorphic Geol, DOI: 10.1111/jmg.12463

Carvalho BB, Sawyer EW, Janasi VA (2016) Crustal reworking in a shear zone: transformation of metagranite to migmatite. J Metamorphic Geol 34:237–264

Carvalho BB, Sawyer EW, Janasi VA (2017) Enhancing maficity of granitic magma during anatexis: entrainment of infertile mafic lithologies. J Petrol 58:1333–1362

Cesare B, Maineri C (1999) Fluid-present anatexis of metapelites at El Joyazo (SE Spain): constraints from Raman spectroscopy of graphite. Contrib Mineral Pet 135:41–52

Cesare B, Acosta-Vigil A, Bartoli O, Ferrero S (2015) What can we learn from melt inclusions in migmatites and granulites? Lithos 239:186–216

Chappell BW, White AJR, Wyborn D (1987) The importance of residual source material (Restite) in granite petrogenesis. J Petrol 28:1111–1138

Coggon R, Holland TJB (2002) Mixing properties of phengitic micas and revised garnet–phengite thermobarometers. J Metamorphic Geol 20:683–696

Davidson C, Grujic DE, Hollister LS, Schmid SM (1997) Metamorphic reactions related to

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26
27
28
29
30
31
32
33
34
35
36
37
38
39
40
41
42
43
44
45
46
47
48
49
50
51
52
53
54
55
56
57
58
59
60
61
62
63
64
65

634 decompression and synkinematic intrusion of leucogranite, High Himalayan Crystallines,
635 Bhutan. J Metamorphic Geol 15:593–612

636 Ferrero F, Angel (2018) Micropetrology: Are inclusions grains of truth? J Petrol 59:1671–
637 1700

638 Forshaw JB, Waters DJ, Pattison DRM, Palin RM, Gopon PA (2019) Comparison of
639 observed and thermodynamically predicted phase equilibria and mineral compositions in
640 mafic granulites. J Metamorphic Geol 37:153–179

641 Gao Li-E, Zeng L, Asimow PD (2017) Contrasting geochemical signatures of fluid-absent
642 versus fluid-fluxed melting of muscovite in metasedimentary sources: The Himalayan
643 leucogranites. Geology 45:39–42

644 García-Casco A, Haissen F, Castro A, El-Hmidi H, Torres-Roldán RL, Millán G (2003)
645 Synthesis of staurolite in melting experiments of a natural metapelite: consequences for
646 the phase relations in low-temperature pelitic migmatites. J Petrol 44:1727–1757

647 Godin L (2003) Structural evolution of the Tethyan sedimentary sequence in the Annapurna
648 area, central Nepal Himalaya. J As Earth Sc 22:307–328

649 Godin L, Parish RR, Brown L, Hodges KV (2001) Crustal thickening leading to exhumation
650 of the Himalayan Metamorphic core of central Nepal: insight from U–Pb geochronology
651 and $^{40}Ar/^{39}Ar$ thermochronology. Tectonics 20:729–747

652 Groppo C, Rolfo F, Indares A (2012) Partial melting in the Higher Himalayan Crystallines of
653 Eastern Nepal: the effect of decompression and implications for the 'Channel Flow'
654 model. J Petrol 53:1057–1088

655 Groppo C, Rubatto D, Rolfo F, Lombardo B (2010) Early Oligocene partial melting in Main
656 Central Thrust Zone (Arun valley, eastern Nepal Himalaya). Lithos 118:287–301.

657 Guillong M, Meier DL, Allan MM, Heinrich CA, Yardley BWD (2008) SILLS: A
658 MATLAB-based program for the reduction of laser ablation ICP-MS data of
659 homogeneous materials and inclusions. Mineralogical Association of Canada Short Course
660 40, Vancouver, B.C., p. 328–333.

661 Guilmette C, Indares A, Hébert R (2011) High-pressure anatectic paragneisses from the
662 Namche Barwa, Eastern Himalayan Syntaxis: textural evidence for partial melting, phase
663 equilibria modeling and tectonic implications. Lithos 124:66–81

664 Halter WE, Pettke T, Heinrich CA, Rothen-Rutishauser B (2002) Major to trace element
665 analysis of melt inclusions by laser-ablation ICP–MS: methods of quantification. Chem
666 Geol 183:63–86

667 Harris N, Massey (1994) Decompression and anatexis of Himalayan metapelites. Tectonics
668 13:1537–1546

669 Harris N, Ayres M, Massey J (1995) Geochemistry of granitic melts produced during the
670 incongruent melting of muscovite: implication for the extraction of Himalayan
671 leucogranite magmas: J Geoph Res 100:15,767–15,777

672 Harrison TM (2006) Did the Himalayan crystallines extrude partially molten from beneath
673 the Tibetan Plateau? In: Geological Constraints on Channel Flow and Ductile Extrusion as
674 an Important Orogenic Process – Himalaya-Tibetan Plateau (eds Law, R.D., Searle, M.P.
675 & Godin, L.), Geological Society, London, Special Publication, 268:237–254

676 Hodges KV (2000) Tectonics of the Himalaya and southern Tibet from two perspectives.
677 Geol Soc Am Bull 112:324–350

678 Hodges KV, Parrish RR, Searle MP (1996) Tectonic evolution of the central Annapurna
679 Range, Nepalese Himalayas. Tectonics 15:1264–1291.

680 Holland TJB, Powell R (1998) An internally consistent thermodynamic data set for phases of
681 petrological interest. J Metamorphic Geol 16:309–343

682 Holland TJB, Powell R (2003) Activity-composition relations for phases in petrological
683 calculations: An asymmetric multicomponent formulation. Contrib Mineral Pet 145:492–

684    501
685    Holland TJB, Powell R (2011) An improved and extended internally consistent
686        thermodynamic dataset for phases of petrological interest, involving a new equation of
687        state for solids. J Metamorphic Geol 29(3):333–383
688    Holtz F, Johannes W, Tamic N, Behrens H (2001) Maximum and minimum water contents of
689        granitic melts generated in the crust: a reevaluation and implications. Lithos 56:1–14
690    Hu X, Garzanti E, Wang J, Huang W, An W, Webb A (2016) The timing of India–Asia
691        collision onset – facts, theories, controversies. Earth-Sci Rev160: 264–299
692    Iaccarino S, Montomoli C, Carosi R, Massonne H-Jì, Langone A, Visonà D (2015) Pressure–
693        temperature–time–deformation path of kyanite-bearing migmatitic paragneiss in the Kali
694        Gandaki valley (Central Nepal): investigation of Late Eocene-Early Oligocene melting
695        processes. Lithos 231:103–121
696    Iaccarino S, Montomoli C, Carosi R, Massonne H-Jì, Langone A, Visonà D (2017) Geology
697        and tectono-metamorphic evolution of the Himalayan metamorphic core: insights from the
698        Mugu Karnali transect, Western Nepal (Central Himalaya). J Metamorphic Geol 35:301–
699        325
700    Inger S, Harris NBW (1992) Tectonothermal evolution of the High Himalayan crystalline
701        sequence, Langtang Valley, northern Nepal. J Metamorphic Geol 10:439–452
702    Kellett DA, Godin L (2009) Pre-Miocene deformation of the Himalayan superstructure,
703        Hidden valley, central Nepal. J Geol Soc London 166:261–275
704    Kohn, M.J., 2014. Himalayan metamorphism and its tectonic implications. An. Rev Earth
705        Planet Sc 42:381–419
706    Liu Y, Siebel W, Massonne H-J, Xiao X (2007) Geochronological and petrological
707        constraints for the tectonic evolution of the central Greater Himalayan Sequence in the
708        Kharta area, southern Tibet. J Geol 115:215–230
709    Montel JM (1993) A model for monazite/melt equilibrium and applications to the generation
710        of granitic magmas. Chem Geol 110:127–146
711    Newton RC, Charlu TV, Kleppa OJ (1980) Thermochemistry of high structural state
712        plagioclases. Geochim Cosmochim Acta 44:933–941
713    Palin RM, Searle MP, St-Onge MR, Waters DJ, Roberts NMW, Horstwood MSA, Parrish
714        RR, Weller OM, Chen S, Yang J (2014) Monazite geochronology and petrology of
715        kyanite- and sillimanite-grade migmatites from the northwestern flank of the eastern
716        Himalayan syntaxis. Gondwana Res 26:323–347
717    Patiño Douce AE, Harris N (1998) Experimental constraints on Himalayan anatexis. J Petrol
718        39:689–710
719    PognanteU, Benna P (1993) Metamorphic zonation, migmatization and leucogranites along
720        the Everest transect of Eastern Nepal and Tibet: record of an exhumation history. Geol
721        Soc London Sp Publ  74:323–340
722    Prince C, Harris N, Vance D (2001) Fluid-enhanced melting during prograde metamorphism.
723        J Geol Soc 158:233–242
724    Roedder E (1979) Fluid inclusions as samples of ore fluids. In H. L. Barnes (Ed.),
725        Geochemistry of hydrothermal ore deposits (2nd ed., pp. 684–737). New York: Wiley.
726    Rosenberg CL, Handy MR (2005) Experimental deformation of partially melted granite
727        revisited: Implications for the continental crust. J Metamorphic Geol 23:19–28
728    Rubatto D, Chakraborty S, Dasgupta S (2013) Timescales of crustal melting in the Higher
729        Himalayan crystallines (Sikkim, Eastern Himalaya) inferred from trace element-
730        constrained monazite and zircon chronology. Contrib Mineral Pet 165:349–372
731    Rudnick RL, Gao S (2014) Composition of the continental crust. In RL Rudnick Ed, Treatise
732        on Geochemistry, 2nd, 1–51
733    Sawyer EW (2010) Migmatites formed by water-fluxed partial melting of a leucogranodiorite

734       protolith: microstructures in the residual rocks and source of the fluid. Lithos 116:273–286

735 Sawyer EW (2014) The inception and growth of leucosomes: microstructure at the start of
736       melt segregation in migmatites. J Metamorphic Geol 32:695–712

737 Searle MP (2010) Low-angle normal faults in the compressional Himalayan orogen; evidence
738       from the Annapurna–Dhaulagiri Himalaya, Nepal. Geosph 6:296–315

739 Searle MP, Godin L (2003) The South Tibetan Detachment System and the Manaslu
740       Leucogranite: a structural reinterpretation and restoration of the Annapurna-Manaslu
741       Himalaya, Nepal. J Geol 111:505–523

742 Stepanov AS, Hermann J, Rubatto D, Rapp RP (2012) Experimental study of monazite/melt
743       partitioning with implications for the REE, Th and U geochemistry of crustal rocks. Chem
744       Geol 300–301:200–220

745 Sun SS, McDonough WF (1989) Chemical and Isotopic Systematics of Oceanic Basalts:
746       Implications for Mantle Composition and Processes. In: Sanders AD, Norry MJ (Eds),
747       Magmatism in the Ocean Basins. Geological Society, London, Special Publications 42, pp.
748       313–345

749 Tajčmanová L, Conolly JAD, Cesare B (2009) A thermodynamicmodel for titanium and
750       ferric iron solution in biotite. J Metamorphic Geol 27:153–165

751 Tian Z, Zhang Z, Dong X (2016) Metamorphism of high-P metagreywacke from the Eastern
752       Himalayan syntaxis: phase equilibria and P–Tpath. J Metamorphic Geol 34:697–718

753 Thomas R, Klemm W (1997) Microthermometric study of silicate melt inclusions in Variscan
754       granites from SE Germany: volatile contents and entrapment conditions. J Metamorphic
755       Geol 38:1753–1765

756 Thomen A, Robert F, Remusat L (2014) Determination of the nitrogen abundance in organic
757       materials by NanoSIMS quantitative imaging. J Anal At Spectr 29:512–519

758 Thompson JB, Hovis GL (1979) Entropy of mixing in sanidine. Am Mineral 64:57–65

759 Wang JM, Zhang JJ, Wang XX (2013) Structural kinematics, metamorphic P–T profiles and
760       zircon geochronology across the Greater Himalayan Crystalline Complex in south–central
761       Tibet: implication for a revised channel flow. J Metamorphic Geol 31:607–628

762 Wang JM, Rubatto D, Zhang JJ (2015) Timing of partial melting and cooling across the
763       Greater Himalayan Crystalline Complex (Nyalam, Central Himalaya): in-sequence
764       Thrusting and its Implications. J Petrol 56/9, 1677–1702

765 Watson EB, Harrison TM (1984). Accessory minerals and the geochemical evolution of
766       crustal magmatic systems: a summary and prospectus of experimental approaches. Phys
767       Earth Planet Inter 35:19–30.

768 Weinberg (2016) Geology and tectono-metamorphic evolution of the Himalayan
769       metamorphic core: insights from the Mugu Karnali transect, Western Nepal (Central
770       Himalaya). J Metamorphic Geol 35:301–325

771 Weinberg R, Hasalová P (2015a) Water-fluxed melting of the continental crust: a review.
772       Lithos 212–215:158–188.

773 Weinberg R, Hasalová P (2015b) Reply to comment by JD Clemens and G. Stevens on
774       "Water-fluxed melting of the continental crust: A review". Lithos 234:102–103

775 Weinberg R, Searle MP (1999) Volatile-assisted intrusion and autometasomatism of
776       leucogranites in the Khumbu Himalaya, Nepal. J Geol 107:27–48

777 White RW, Powell R, Holland TJB (2007) Progress relating to calculation of partial melting
778       equilibria for metapelites. J Metamorphic Geol 25, 511–527

779 White RW, Powell R, Holland TJB, Worley BA (2000) The effect of $TiO_2$ and $Fe_2O_3$ on
780       metapelitic assemblages at greenschist and amphibolite facies conditions: mineral
781       equilibria calculations in the system $K_2O–FeO–MgO–Al_2O_3–SiO_2–H_2O–TiO_2–Fe_2O_3$. J
782       Metamorphic Geol 18:497–511

783 White RW, Stevens G, Johnson TE (2011) Is the crucible reproducible? Reconciling melting

784     experiments with thermodynamic calculations. Elements 7:241–246

785 White RW, Powell R, Holland TJB, Johnson TE, Green ECR (2014a) New mineral activity–
786     composition relations for thermodynamic calculations in metapelitic systems. J
787     Metamorphic Geol 32:261–286

788 Wolf MB, London D (1997) Boron in granitic magmas: stability of tourmaline in equilibrium
789     with biotite and cordierite. Contrib Mineral Petrol 130:12–30

790 Yakymchuk C (2017) Behaviour of apatite during partial melting of metapelites and
791     consequences for prograde suprasolidus monazite growth. Lithos 274–275:412–426

792 Yakymchuk C, Brown M (2014) Behaviour of zircon and monazite during crustal melting. J
793     Geol Soc 171:465–479

794 Zeng LS, Gao LE, Xie KJ, Liu-Zeng J (2011) Mid-Eocene high Sr/Y granites in the Northern
795     Himalayan Gneiss Domes: melting thickened lower continental crust. Earth Planet Sci Lett
796     303:251–266

797 Zeng LS, Liu J, Gao LE, Xie KJ, Wen L (2009) Early Oligocene crustal anatexis in the
798     Yardoi gneiss dome, southern Tibet and geological implications. Chin Sci Bull 54:104–
799     112

800 Zhang H, Harris N, Parrish R, Kelley S, Zhang L, Rogers N, Argles T, King J (2004) Causes
801     and consequences of protracted melting of the mid-crust exposed in the North Himalayan
802     antiform. Earth Planet Sci Lett 228:195–212

803 Zhang ZM, Zhao GC, Santosh M, Wang JL, Dong X, Liou JG (2010) Two stages of granulite
804     facies metamorphism in the eastern Himalayan syntaxis, south Tibet: petrology, zircon
805     geochronology and implications for the subduction of Neo-Tethys and the Indian
806     continent beneath Asia. J Metamorphic Geol 28:719–733

807 Zhang Z, Xiang H, Dong X, Ding H, He Z (2015) Long-lived high-temperature
808     granulite-facies metamorphism in the Eastern Himalayan orogen, south Tibet. Lithos
809     212-215:1–15

810 Zhang Z, Xiang H, Dong X, Li W, Ding H, Gou Z, Tian Z (2017) Oligocene HP
811     metamorphism and anatexis of the Higher Himalayan Crystalline Sequence in Yadong
812     region, east-central Himalaya. Gondwana Res 41:173–187

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

Figure 2                                                                                           Click here to access/download;Figure;Fig. 2.tif ⬇

Figure 3

Click here to access/download;Figure;Fig. 3.tif

Figure 4

Figure 4

Figure 5

Figure 5

Figure 6

Figure 7

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



Figure 7

(a) and (b): Pressure-temperature pseudosection diagrams with P (GPa) on the vertical axis (0.2 to 1.2) and T (°C) on the horizontal axis (400 to 700).

Panel a mineral assemblages:
- Chl Pl Ms Pg Grt Qtz H₂O
- Pl Ms Pg Grt Qtz H₂O
- Pl Ms Grt Qtz Liq H₂O
- Bt Pl Ms Pg Grt Qtz H₂O
- Bt Chl Pl Ms Pg Grt Qtz H₂O
- Bt Pl Ms Grt Qtz Liq H₂O
- Bt Pl Ms Grt Qtz H₂O
- Bt Chl Pl Ms Pg Qtz H₂O
- Bt Pl Grt Sil Qtz Liq H₂O
- Chl Pl Ms Grt Qtz H₂O
- Chl Pl Ms Qtz H₂O
- Bt Chl Pl Ms Qtz H₂O
- Bt Pl Ms And Qtz H₂O
- Liq-in

Panel b mineral assemblages:
- Bt Ms Pg Pl Grt Qtz H₂O
- Ms Pl Grt Qtz H₂O
- Pl Ms Grt Qtz Liq H₂O
- Bt Ms Pl Grt Qtz H₂O
- Chl Bt Pl Ms Pg Grt Qtz H₂O
- Chl Pl Bt Ms Grt Qtz H₂O
- Chl Pl Ms Pg Grt Qtz H₂O
- Pl Bt Grt Sil Qtz Liq H₂O
- Bt Pl Ms Sil Qtz H₂O
- Bt Chl Pl Ms Qtz H₂O
- Bt Pl Ms And Qtz H₂O
- Liq-in

Composition (mol.%): Na₂O 2.17, CaO 2.44, K₂O 0.69, FeO 1.94, MgO 0.58, Al₂O₃ 7.10, SiO₂ 57.45, H₂O 27.64

Figure 8

Click here to access/download;Figure;Fig. 8.tif ±

Figure 9

Figure 10

Click here to access/download;Figure;Fig. 10.tif ⬇



Figure 10

Table 1

Table 1. LA-ICP-MS analyses (ppm) of melt inclusions and garnet from Kali Gandaki

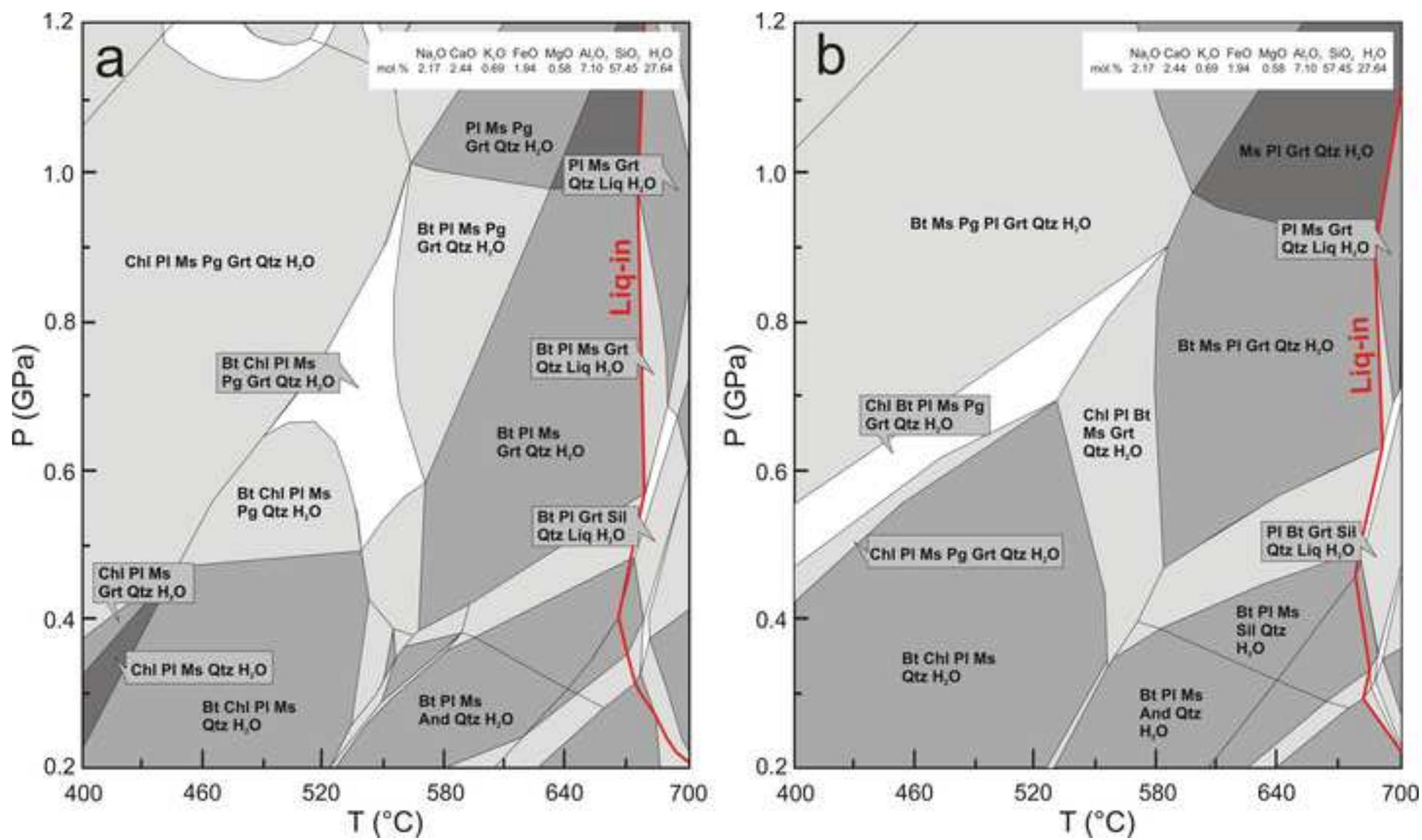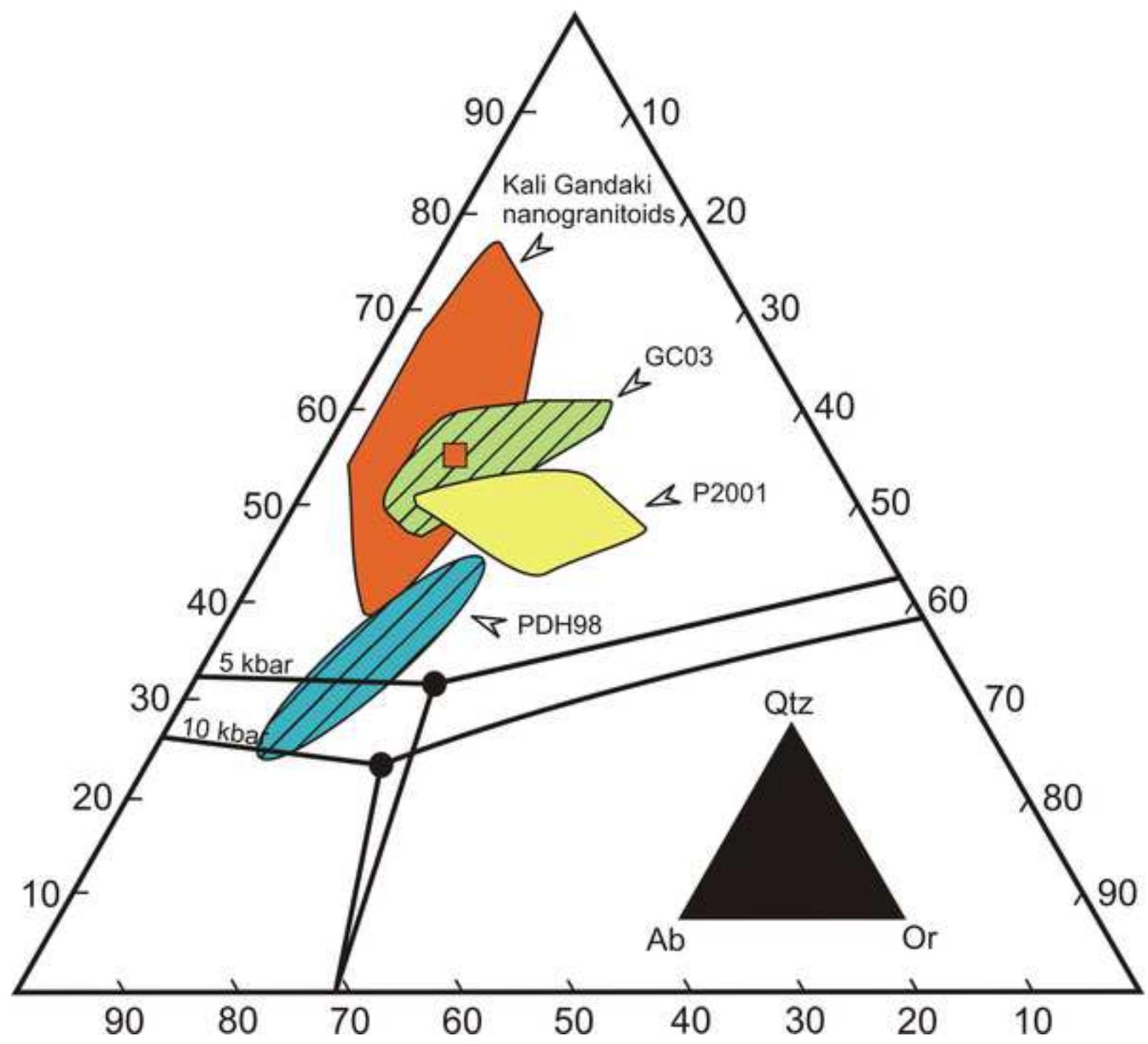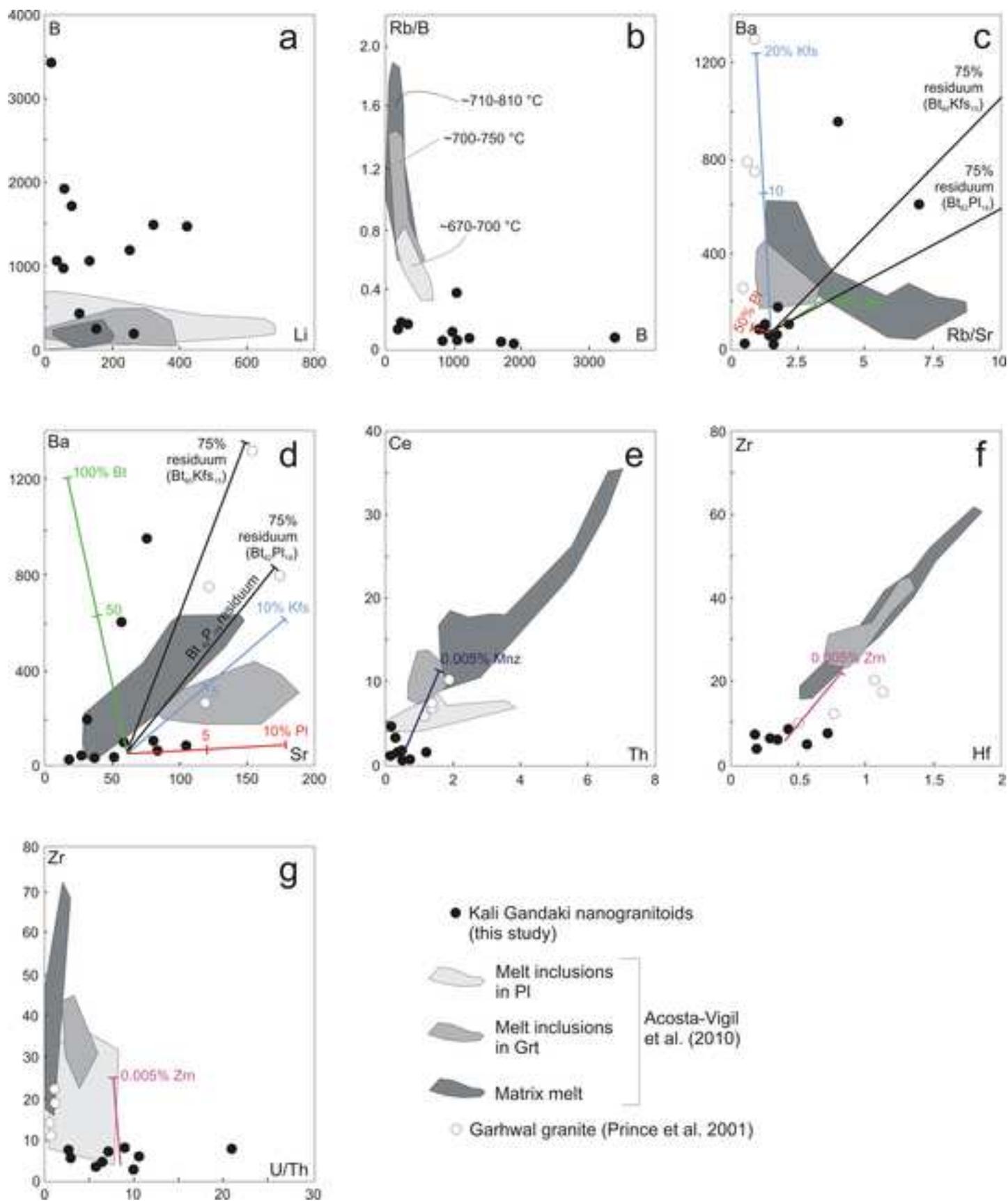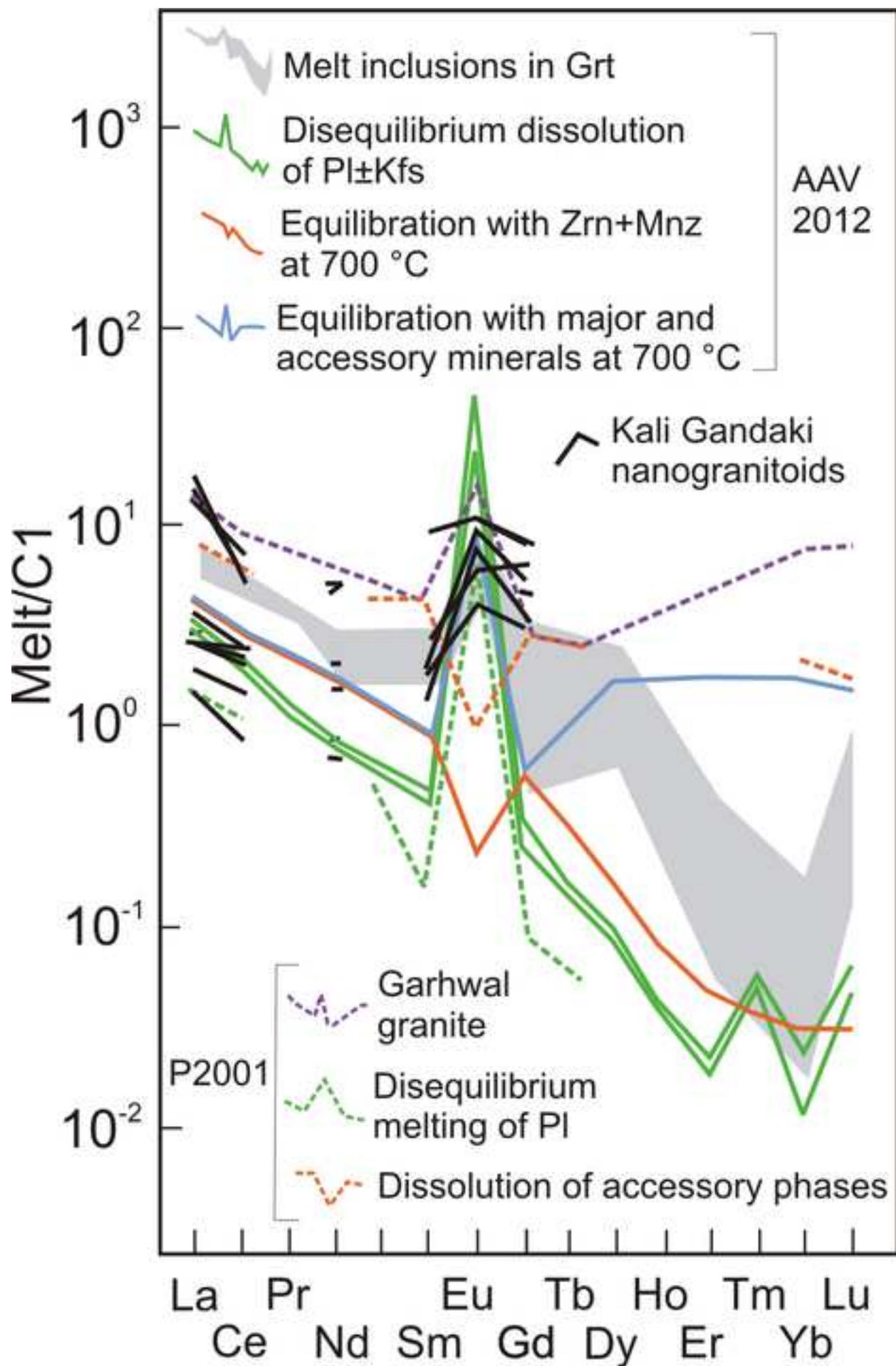|        | KGD_27 | KGD_29 | KGD_32 | KGD_38 | KGD_39 | KGD_40 | KGD_45 | KGD_46 |
|--------|--------|--------|--------|--------|--------|--------|--------|--------|
| Rb     | 103    | 59     | 81     | 127    | 60     | 28     | 43     | 46     |
| Ba     | 77     | 32     | 42     | 101    | 185    | 44     | 50     | 50     |
| Pb     | 60     | 36     | 47     | 88     | 20     | 20     | 37     | 37     |
| Li     | 79     | 58     | 41     | 50     | 329    | 262    | 153    | 153    |
| B      | 1731   | 847    | 1068   | 1005   | 325    | 191    | 244    | 242    |
| Th     | 1.2    | 0.48   | 0.68   | 0.48   | 0.43   | 0.32   | 0.20   | 0.17   |
| U      | 8.5    | 5.0    | 6.1    | 4.8    | 3.4    | 1.9    | 0.56   | 0.50   |
| Nb     | b.d.l. | 18.28  | 8.9    | 1.1    | 7.3    | 4.8    | 8.5    | 8.3    |
| Ta     | b.d.l. | 1.2    | 0.86   | 0.33   | 2.6    | 0.31   | 1.2    | 1.1    |
| La     | 0.91   | 0.61   | 0.70   | 0.34   | 0.57   | 0.59   | 3.4    | 3.3    |
| Ce     | 1.7    | 1.3    | 0.54   | 0.53   | 1.5    | 1.4    | 4.0    | 4.8    |
| Sr     | 84     | 37     | 53     | 59     | 34     | 18     | 29     | 27     |
| Nd     | b.d.l. | 0.41   | b.d.l. | b.d.l. | b.d.l. | 0.33   | 2.50   | 2.48   |
| Sm     | 1.7    | 0.29   | b.d.l. | b.d.l. | b.d.l. | b.d.l. | 0.32   | 0.22   |
| Zr     | 7.3    | 6.2    | 8.2    | 3.2    | b.d.l. | 3.9    | 7.5    | 6.2    |
| Hf     | b.d.l. | 0.34   | 0.42   | b.d.l. | b.d.l. | 0.19   | 0.18   | 0.29   |
| Eu     | 0.72   | 0.23   | 0.67   | b.d.l. | b.d.l. | b.d.l. | 0.51   | 0.45   |
| Gd     | 1.9    | 0.68   | 1.9    | 0.96   | b.d.l. | 0.96   | 1.3    | 0.77   |
| V      | b.d.l. | b.d.l. | b.d.l. | 13     | 20     | 2.7    | 0.43   | b.d.l. |
| Zn     | 61     | 30     | 92     | 62     | b.d.l. | 14     | 26     | 23     |
| Sc     | 66     | 8.0    | 57     | 44     | 76     | 13     | 3.1    | 4.5    |
| U/Th   | 7      | 11     | 9      | 10     | 8      | 6      | 3      | 3      |
| Rb/Sr  | 1.2    | 1.6    | 1.5    | 2.2    | 1.8    | 1.6    | 1.5    | 1.7    |
| Nb/Ta  |        | 15.2   | 10.3   | 3.2    | 2.8    | 15.5   | 6.8    | 7.9    |
| Eu/Eu* | 1.2    | 1.6    |        |        |        |        | 2.5    | 3.3    |

b.d.l.: below detection limit. n.c.: not considered. *) nanogranitoid containing accidentally tr

| KGD_58 | KGD_77 | KGD_03* | KGD_16* | KGD_08* | KGD_11* | KGD_12* | KGD_34* | KGD_37* |
|---|---|---|---|---|---|---|---|---|
| 400 | 305 | b.d.l. | b.d.l. | b.d.l. | b.d.l. | b.d.l. | 104 | 104 |
| 606 | 956 | b.d.l. | b.d.l. | b.d.l. | b.d.l. | b.d.l. | 85 | 105 |
| 9.6 | 49 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 135 | 18 | 424 | 324 | b.d.l. | 102 | 256 | 58 | b.d.l. |
| 1054 | 3412 | 1485 | 1500 | 1604 | 413 | 1195 | 1914 | 1245 |
| 0.27 | 0.12 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 1.8 | 2.5 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 10.7 | 5.5 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 2.1 | 0.64 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 4.0 | 0.43 | b.d.l. | b.d.l. | n.c. | n.c. | n.c. | n.c. | b.d.l. |
| 3.3 | 0.88 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 57 | 76 | b.d.l. | b.d.l. | 15 | 4.4 | 16 | 106 | 81 |
| 0.95 | 0.69 | b.d.l. | b.d.l. | n.c. | n.c. | n.c. | n.c. | b.d.l. |
| b.d.l. | 0.41 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 4.9 | 7.8 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 0.56 | 0.71 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| b.d.l. | 0.33 | b.d.l. | b.d.l. | n.c. | n.c. | n.c. | n.c. | b.d.l. |
| b.d.l. | 1.3 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 10.2 | b.d.l. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 54 | 48 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 16 | 34 | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. | n.c. |
| 6 | 21 | | | | | | | |
| 7.0 | 4.0 | | | | | | 1.0 | 1.3 |
| 5.0 | 8.5 | | | | | | | |
| | 1.4 | | | | | | | |

rapped minerals (e.g., zircon, monazite, rutile)

| garnet |
| --- |
| 1.98 |
| 3.48 |
| 0.39 |
| 7.06 |
| 12.7 |
| 0.02 |
| 0.10 |
| 1.89 |
| 0.43 |
| 0.19 |
| 0.40 |
| 0.72 |
| 0.40 |
| 0.45 |
| 3.66 |
| 0.35 |
| 0.12 |
| 4.31 |
| 44.2 |
| 77.4 |
| 87.1 |

Table 2

Table 2. H2O concentrations measured in nanogranitoids by NanoSIMS

| Sample | Water content (wt %) | $1\sigma$ error |
|--------|----------------------|-----------------|
| KGD_1  | 14.4 | 0.1 |
| KGD_2  | 10.5 | 0.1 |
| KGD_3  | 10.1 | 0.1 |
| KGD_4  | 10.0 | 0.1 |
| KGD_5  | 7.7  | 0.1 |
| KGD_6  | 8.1  | 0.1 |
| KGD_7  | 8.1  | 0.1 |
| KGD_8  | 6.9  | 0.1 |
| KGD_9  | 9.5  | 0.1 |
| KGD_10 | 6.5  | 0.1 |
| KGD_11 | 8.8  | 0.1 |
| KGD_12 | 8.9  | 0.1 |
| KGD_13 | 8.4  | 0.1 |
| KGD_14 | 7.3  | 0.1 |
| KGD_15 | 8.7  | 0.1 |
| KGD_16 | 8.0  | 0.1 |

[c] Average value and $1\sigma$ standard deviation (in parentheses) regarding all the measurments

[d] Water contents determined by difference of EMP totals from 100 %