Gondwana Research

Glimmerite: a product of melt-rock interaction within a crustal-scale high-strain zone
--Manuscript Draft--

Corresponding Author:	David Silva, Ph.D. Macquarie University Sydney, NSW AUSTRALIA
First Author:	David Silva, Ph.D.
Order of Authors:	David Silva, Ph.D.
	Nathan R. Daczko, Associate Professor
	Sandra Piazolo, Professor
	Tom Raimondo, Associate Professor

Abstract:	The paradigm for hydrous high-strain zones that cut dry host rocks is for fluid-rock interaction to have involved aqueous fluids. However, the role of silicate melt is increasingly recognised. This contribution examines the formation of glimmerite (biotitite) bands during melt migration in the Gough Dam shear zone, a high-strain zone in central Australia that was active during the Alice Springs Orogeny (c. 450–300 Ma). The glimmerite bands cut and replace a range of quartzo-feldspathic protoliths, including granitic gneiss and quartzite. Melt that migrated through the high-strain zone is interpreted to have penetrated relict layers along a network of fractures, enhancing dissolution of the precursor rock and causing replacement by glimmerite crystallisation. Microstructures indicative of the former presence of melt in the high-strain zone include: pseudomorphs of former melt pockets of granitic composition; small dihedral angles of interstitial phases; elongate grain boundary melt pseudomorphs; neighbourhoods of grains connected in three dimensions; and localised static grain growth and recovery. Other microstructures indicative of melt-present deformation include randomly oriented neosome grains, and evidence of activation of multiple slip systems during deformation. The degree of quartzite modification to glimmerite is recorded by an increase in biotite mode, and correlated with higher Ti concentrations in biotite (higher apparent temperature) and changes to trace element and REE compositions. Melt-assisted coupled dissolution-precipitation reactions during melt flux are interpreted to partially reset Proterozoic monazite U-Pb ages inherited from the protolith (> 1630 Ma) to younger Palaeozoic ages, with a complex age pattern partially congruent with the Alice Springs Orogeny (apparent ages range from c. 606–371 Ma, with a dominant age peak at c. 451 Ma). We propose that the glimmerite formed during dynamic melt migration of an externally-derived hydrous peraluminous melt, driving reaction replacement of various felsic protoliths during this orogenic event.

Response to Reviewers:	

- Reaction replacement glimmerite formed during dynamic hydrous melt migration

- Microstructures as evidence of former melt presence in high-strain zone

- Imperfect dissolution-precipitation reactions of monazite during melt flux

- Glimmerite layers in anhydrous terrains facilitates intracontinental orogenesis
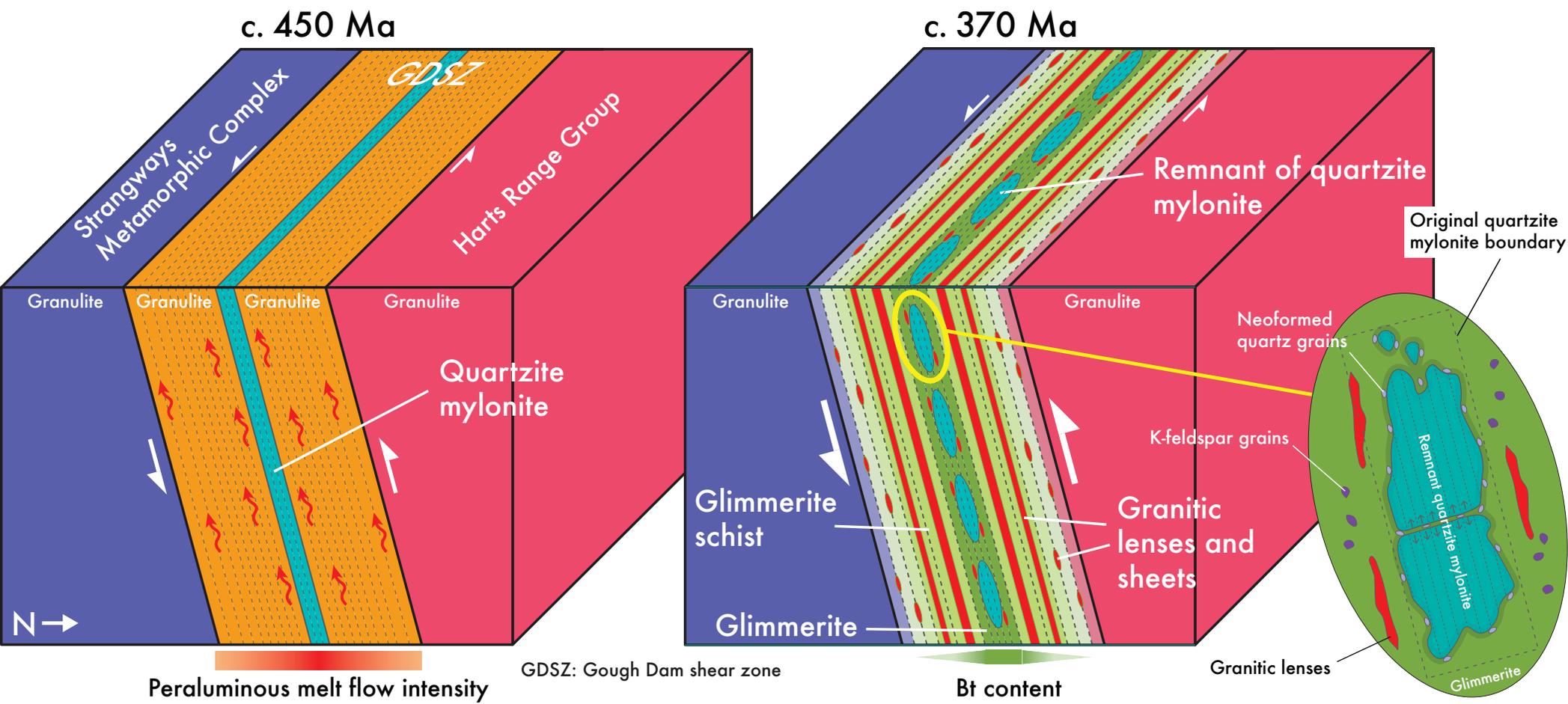
1 **Glimmerite: a product of melt-rock interaction within a crustal-scale high-**

2 **strain zone**

3

4 D. Silva[1*], N. R. Daczko[1], S. Piazolo[1,2] and T. Raimondo[3, 4]

5 [1] ARC Centre of Excellence for Core to Crust Fluid Systems and GEMOC, Department of

6 Earth and Environmental Sciences, Macquarie University, NSW 2109, Australia

7 [2] School of Earth and Environment, Institute of Geophysics and Tectonics, University of Leeds,

8 Leeds, LS2 9JT, United Kingdom

9 [3] UniSA STEM, University of South Australia, GPO Box 2471, Adelaide, SA 5001, Australia

10 [4] Future Industries Institute, University of South Australia, GPO Box 2471, Adelaide, SA 5001,

11 Australia

12

13 *Corresponding author. Email: david.barbosa-da-silva@students.mq.edu.au

14

15 **ABSTRACT**

16   The paradigm for hydrous high-strain zones that cut dry host rocks is for fluid-

17 rock interaction to have involved aqueous fluids. However, the role of silicate melt is

18 increasingly recognised. This contribution examines the formation of glimmerite (biotitite)

19 bands during melt migration in the Gough Dam shear zone, a high-strain zone in central

20 Australia that was active during the Alice Springs Orogeny (*c*. 450–300 Ma). The glimmerite

21 bands cut and replace a range of quartzo-feldspathic protoliths, including granitic gneiss and

22 quartzite. Melt that migrated through the high-strain zone is interpreted to have penetrated relict

23 layers along a network of fractures, enhancing dissolution of the precursor rock and causing

24 replacement by glimmerite crystallisation. Microstructures indicative of the former presence

25 of melt in the high-strain zone include: pseudomorphs of former melt pockets of granitic

26 composition; small dihedral angles of interstitial phases; elongate grain boundary melt

27 pseudomorphs; neighbourhoods of grains connected in three dimensions; and localised static

28 grain growth and recovery. Other microstructures indicative of melt-present deformation

29 include randomly oriented neosome grains, and evidence of activation of multiple slip systems

30 during deformation. The degree of quartzite modification to glimmerite is recorded by an

31 increase in biotite mode, and correlated with higher Ti concentrations in biotite (higher

32 apparent temperature) and changes to trace element and REE compositions. Melt-assisted

33 coupled dissolution-precipitation reactions during melt flux are interpreted to partially reset

34 Proterozoic monazite U-Pb ages inherited from the protolith (> 1630 Ma) to younger

35 Palaeozoic ages, with a complex age pattern partially congruent with the Alice Springs

36 Orogeny (apparent ages range from *c.* 606–371 Ma, with a dominant age peak at *c.* 451 Ma).

37 We propose that the glimmerite formed during dynamic melt migration of an externally-

38 derived hydrous peraluminous melt, driving reaction replacement of various felsic protoliths

39 during this orogenic event.

40

41 **Key words:** Glimmerite; Metasomatism; High-strain zone; Monazite; Rheology

42

43 **1. INTRODUCTION**

44 Fluid (or the lack thereof) is key in crustal tectonometamorphic processes

45 because it plays a major role in the progression of reactions, changes to mineral assemblages,

46 metastability, activation of deformation processes and coupled rheological consequences (e.g.,

47 Yardley, 2009 and references therein). Both water or solute-rich fluids such as silicate melts

48 have a first-order effect on a wide range of geochemical and deformation processes, acting as

49 important agents at all scales. For example, fluids increase grain boundary mobility (e.g.,

50 Piazolo *et al.*, 2002; Mancktelow & Pennacchioni, 2004), cause crystal hydrolytic weakening

51 (e.g., Griggs & Blacic, 1965; Kronenberg *et al.*, 1990; Girard *et al.*, 2013) and rheological

52 weakening through fluid overpressure (e.g., Thompson & Connolly, 1992; Sibson, 1994),

53 induce water-fluxed melting (e.g., Weinberg & Hasalová, 2015 and references therein),

54 facilitate metasomatism (e.g., Etheridge *et al*., 1983; Ferry, 1994; Putnis & Austrheim, 2010),

55 enable fast intra- and inter-grain diffusion pathways (e.g., Watson & Baxter, 2007; Wheeler,

56 2014; Chapman *et al*., 2019; Gonçalves *et al*., 2021), and drive coupled dissolution-

57 precipitation replacement reactions (e.g., Putnis & Putnis, 2007; Putnis, 2009; Seydoux-

58 Guillaume *et al*., 2012). These fluid-aided processes can have major effects at the macroscale

59 for the exhumation of deep crust (e.g., Raimondo *et al*., 2014 and references therein), rates and

60 style of subduction (e.g., Pichon *et al*., 1993; Stern, 2002), arc volcanism (e.g., Morris *et al*.,

61 1990; Sobolev & Chaussidon, 1996; Schmidt & Poli, 1998), and progression and character of

62 orogenesis (e.g., Beaumont *et al*., 2001; Bercovici, 1998; Jamieson *et al*., 2011).

63 Externally sourced fluids are commonly in disequilibrium with the host rocks

64 through which they migrate and are therefore likely to trigger mineral reactions (Yardley, 2009;

65 Putnis, 2009). The most extreme examples of fluid-rock interaction and associated reaction are

66 usually observed in high-strain zones where time-integrated fluid fluxes and reactive surface

67 area at any one time are high and fluid migration is channelised (Etheridge *et al*., 1983; Oliver

68 *et al*., 1990; Selverstone *et al*., 1991; Cartwright & Barnicoat, 2003; Clarke *et al*., 2005; Daczko

69 *et al*., 2016; Stuart *et al*., 2016, 2017, 2018*a*). In these cases, the rheological consequences are

70 pronounced and fall into four general categories: (i) fluid-rock interaction results in the

71 production of minerals that have a different rheological behaviour to their precursors (e.g.,

72 Brodie & Rutter, 1985; Wintsch *et al*., 1995; Smith *et al*., 2015); (ii) fluids cause grain size

73 changes that may change the dominant deformation mechanism (Smith *et al*., 2015; Menegon

74 *et al*., 2015); (iii) fluids change the rate of the dominant deformation mechanism (e.g., Karato,

75 1986); and/or (iv) the actual physical presence of the fluid itself has a rheological effect (e.g.,

76 Faulkner & Rutter, 2001; Handy *et al*., 2001).

77   Until recently, the study of hydrated high-strain zones has been focused on those
78   that experienced interaction with water-rich fluids, most commonly during retrograde
79   metamorphic conditions (e.g., White & Knipe, 1978; McCaig *et al.*, 1990; Menegon *et al.*,
80   2008; Gonçalves *et al.*, 2012). Often, such fluid-fluxed areas exhibit extensive metasomatism,
81   changing not only the mineral assemblage but also the bulk chemistry of the fluid-fluxed rock
82   (e.g., Spruzeniece & Piazolo, 2015 and references therein). However, recently it has been
83   shown that hydration and strain localisation can also occur when a hydrous silicate melt fluxes
84   lower crustal rocks in the roots of magmatic arcs (Daczko *et al.*, 2016; Stuart *et al.*, 2016, 2017,
85   2018*a*; Meek *et al.*, 2019). In these cases, the main hydrous minerals produced are amphibole
86   and clinozoisite, controlled by the mafic to intermediate silicate melts inferred to have migrated
87   through the actively deforming high-strain zone. A similar process of melt-rock interaction as
88   melt migrates up high-strain zones has also recently been recognised in mid-crustal granulite
89   facies rocks of central Australia during the Alice Springs Orogeny (Ghatak, 2017; Piazolo *et*
90   *al.*, 2020). In this particular case, melt metasomatism is inferred to have involved hydrous
91   peraluminous melts and produced metasomatic rocks rich in biotite and garnet at pressures of
92   6–8 kbar. The enrichment in biotite and garnet, and depletion (i.e., replacement) of quartz and
93   feldspar in precursor rocks suggests that melt temperatures were close to the liquidus, where
94   fewer minerals are in equilibrium with the high-temperature melt (i.e., at temperatures above
95   the stability of quartz and feldspar such that these dissolve into the melt; Ghatak, 2017).

96   Due to the increased porosity and permeability that accompanies deformation,
97   high-strain zones are ideal pathways for fluid migration, either by aqueous fluids or melts (e.g.,
98   White & Knipe, 1978; Hutton, 1988; McCaig & Knipe, 1990; D'lemos *et al.*, 1992; Brown &
99   Solar, 1998; Rosenberg, 2004; Menegon *et al.*, 2008; Menegon *et al.*, 2015; Stuart *et al.*, 2018*a*;
100   Etheridge *et al.*, 2020; Gonçalves *et al.*, 2012, 2021). Both types of fluid migration in high-
101   strain zones are well understood and easily recognised if, in the case of melt, a certain threshold

102 of interconnected fluid (> 10%) is present during deformation (e.g., Brown & Solar, 1998;

103 Collins & Sawyer, 1996; Weinberg & Hasalová, 2015; Závada *et al.*, 2018). Such high

104 proportions are suggested in the field by the high percentage of leucosome (Sawyer, 2008). In

105 contrast, lower proportions of melt in a deforming rock (< 10%) will have limited rheological

106 effect (Rosenberg & Handy, 2005), although this is debatable.

107　　　　However, a lack of field evidence for high proportions of melt does not exclude

108 the possibility of melt-present rheological weakening, because during the last stages of

109 deformation the melt phase may have been largely evacuated (Schulmann *et al.*, 2008; Stuart

110 *et al.*, 2018*b*). Additionally, recent work has shown that high-strain rocks lacking typical sub-

111 solidus crystal-plastic deformation microstructures (e.g., crystallographic preferred

112 orientations, dynamic recrystallisation, undulose extinction) may be used as an indicator of

113 melt-present deformation and are suggestive of weakened rocks (Stuart *et al.*, 2018*b*). Key

114 microstructures used to recognise former melt flux through high-strain zones include: (1) small

115 dihedral angles of interstitial phases; (2) elongate interstitial grains; (3) neighbourhoods of

116 quartz or feldspar grains connected in three dimensions; (4) fine-grained multiphase aggregates

117 representing crystallised melt pockets or inclusions (e.g., Cesare *et al.*, 2015; Meek *et al.*,

118 2019); and (5) mm- to cm-scale felsic dykelets (Stuart *et al.*, 2018*b*). Other melt

119 microstructures such as those observed in migmatites (Sawyer, 1999; Vernon, 2011) may also

120 be useful indicators of melt-present deformation in high-strain zones.

121　　　　Glimmerite ("biotitite" in the IUGS classification) is defined as rock with a very

122 high proportion of mica (> 75%, mainly biotite or phlogopite) and has been reported globally

123 in diverse geological settings (e.g., Waters, 1987; Rakotondrazafy *et al.*, 1997; Becker *et al.*,

124 1999; Fuertes-Fuente *et al.*, 2000; Grégoire *et al.*, 2002; Rajesh *et al.*, 2004; Lin & Sawyer,

125 2019). Glimmerite is exceptionally hydrous, and its bulk composition does not resemble any

126 classic sedimentary or igneous rock type, apart from small-scale gradational concentrations of

127 mafic minerals in granites and migmatites, i.e., schlieren (Weinberg *et al*., 2001; Milord &

128 Sawyer, 2003; Žák *et al*., 2008). Hence, it is likely that glimmerites represent rocks that have

129 undergone extreme chemical change or metasomatism sensu lato (*s.l.*). The most commonly

130 reported occurrence of glimmerite is associated with highly metasomatised mantle rocks

131 caused by interaction with a percolating melt/fluid (e.g., Becker *et al*., 1999; Grégoire *et al*.,

132 2002; Smart *et al*., 2019). Glimmerites have also been reported to occur adjacent to granitic

133 intrusions in mantle rocks (e.g., Fuertes-Fuente *et al*., 2000), carbonatite intrusions and

134 associated fenitisation of surrounding rocks (e.g., Tappe *et al.,* 2006; Elliott *et al*., 2018) and

135 have been interpreted to represent progressive fractional crystallisation of mantle melts injected

136 into the lower crust (e.g., Rajesh *et al*., 2004).

137   Contrary to these latter examples, the glimmerite studied here is located in a

138 high-strain zone known as the Gough Dam shear zone, central Australia. They are spatially

139 adjacent to and/or surrounded by felsic mid-crustal rocks (granitic and felsic granulitic

140 gneisses). The chosen field example forms part of a regional-scale, 10 km long and 2–4 km-

141 thick belt of glimmerite and other biotite-rich schistose rocks of varying biotite mode. In this

142 contribution, we explore the mechanism of formation of glimmerite by examining well-

143 exposed continuous outcrops that allow detailed field and multi-technique analysis. Based on

144 the glimmerite assemblage, geochemistry, deformation, reaction textures and geochronology,

145 we suggest that this intriguing rock type formed as a consequence of extreme melt-rock

146 interaction during peraluminous melt migration through high-strain zones during orogenesis.

147

2. GENERAL GEOLOGICAL BACKGROUND

149   The area of interest lies in the Strangways Metamorphic Complex (SMC),

150 Arunta Region, central Australia. This basement terrane is a broad belt of dominantly

151 anhydrous upper amphibolite and granulite facies rocks, up to ~125 km wide, metamorphosed

152 during the Strangways Event at *c*. 1735–1690 Ma (Fig. 1) (Shaw *et al*., 1984; Collins &

153 Teyssier, 1989; Claoué-Long *et al*., 2008; Scrimgeour, 2003, 2013). Subsequently, it was

154 extensively reworked during the Alice Springs Orogeny (ASO), an intracontinental orogeny

155 active during the middle to late Palaeozoic (450–300 Ma) (Collins & Teyssier, 1989; Raimondo

156 *et al*., 2014). This event caused crustal thickening and deep exhumation of ensialic granulitic

157 basement in the Arunta Region, resulting in an orogen with an estimated horizontal shortening

158 of up to 100 km at its eastern margin (Shaw *et al*., 1984; Teyssier, 1985; Collins & Teyssier,

159 1989; Fig. 1). An extensive regional network of anastomosing, km-wide high-strain zones

160 bounding the Strangways Metamorphic Complex were reactivated during the ASO, facilitating

161 much of the observed exhumation (e.g., Cartwright *et al*., 1999; Hand & Sandiford, 1999;

162 Mawby *et al*., 1999; Ballèvre *et al*., 2000; Scrimgeour, 2013; Raimondo *et al*., 2011, 2014;

163 Prent *et al*., 2020; Fig. 1). These schistose high-strain zones include the West Bore shear zone

164 and Gough Dam shear zone at the northern boundary of the SMC, and the Harry Creek shear

165 zone toward the south.

166   Shear zones in the Arunta Region display hydrous mineral assemblages with

167 high modal proportions of biotite and muscovite (e.g., SMC and Reynolds-Anmatjira Ranges

168 (Fig. 1; Raimondo *et al*., 2011, 2014). In the Reynolds-Anmatjira Ranges, hydrous shear zones

169 are linked to metasomatic processes and retrogression of amphibolite facies rocks and gneisses,

170 instead of common alternatives such as stress-induced mineral segregation or highly evolved

171 schlieren bands (Cartwright & Buick, 1999; Raimondo *et al*., 2011, 2012; Prent *et al*., 2020).

172 Similar to the Reynolds-Anmatjira Ranges, the SMC is cut by shear zones forming broad

173 hydrous schist belts containing a high proportion of micas. These shear zones and schist belts

174 are interpreted as strongly deformed and hydrated versions of granulite precursor rocks, which

175 formed at deeper crustal levels when compared to those of the Reynolds-Anmatjira Ranges

176 (Ballèvre *et al*., 2000; Raimondo *et al*., 2011, 2014).

177   The studied glimmerite occurs in the Gough Dam shear zone (GDSZ; labelled

178   #1 in Fig. 1), located at the northern boundary of the Strangways Metamorphic Complex. The

179   GDSZ trends roughly E-W to NNW-SSE, is steeply dipping (60–90°) to the north and presents

180   a typical width of 1–2 km but can reach up to 4 km wide. The GDSZ is interpreted to have

181   deformed in the presence of externally-derived melt that induced reaction softening and

182   rheological weakening (Piazolo *et al.*, 2020). The high-strain zone is described as a significant

183   tectonic discontinuity in the Arunta Region, controlling crustal block separation and reverse

184   south-directed thrusting of the Harts Range Group relative to the Strangways Metamorphic

185   Complex, kinematically informed by shear bands and sigma tails around porphyroclasts

186   (Collins & Shaw, 1995; Bendall, 2000; Fig. 1). The GDSZ high-strain zone is characterised by

187   hydration of the bimodal interlayered mafic-felsic granulites, quartzo-feldspathic and quartzite

188   gneisses, and minor calcsilicate and amphibolite rocks constituting the Paleoproterozoic

189   basement of the Arunta Region.

190

2. 1. Gough Dam shear zone field relationships

192   The GDSZ features several different rock types throughout the schist belt, the

193   most recurrent being phyllonitic biotite-rich quartzo-feldspathic gneiss, felsic and mafic

194   granulite pods of comparatively lower strain, and quartz-rich rafts (Norman, 1991; Ballèvre *et*

195   *al.*, 1997). Some low-strain areas of the GDSZ include garnet-bearing granulites with garnet

196   crystals up to 8 cm in diameter. A steeply northeast-dipping/plunging foliation/mineral

197   lineation is defined by biotite and sillimanite, and locally kyanite. Highly strained rock types

198   have variable bulk rock compositions and include granitic mylonite closer to the shear zone's

199   southern boundary (Fig. 2a), and feldspar-biotite-quartz schist (identified hereafter in the text

200   as glimmerite schist) with elongated granitic lenses located towards and in the centre of GDSZ

201   (Fig. 2b, c). The width of granitic lenses increases towards the centre of GDSZ (Fig. 2b to 2c)

202 and these have been shown to have igneous microstructure consistent with syntectonic melt

203 migration through the GDSZ (Piazolo *et al.*, 2020).

204   The selected study area is localised in the central section of the GDSZ and

205 comprises an outcrop of approximately 10 m width by 15 m length of glimmerite schist, which

206 contains cm to dm scale lenses and layers of strained quartzite (up to 50 cm thick and 10 m in

207 length; Fig. 2d, g–i), and felsic granulite and granitic gneiss that are elongate parallel to the

208 schistose foliation (Fig. 2d–f). The matrix of the glimmerite schist is dominated by biotite

209 (~50%) and subordinate quartz, muscovite and sillimanite, with mm to cm-wide feldspar grains

210 (Fig. 2d–i). Within 200 m of the study area, kyanite and staurolite are also observed.

211 Centimetre-wide glimmerite sleeves (identified hereafter as glimmerite) with high proportions

212 of biotite (> 75%) are present adjacent to, cross-cutting, or 'invading' felsic bodies (i.e.,

213 quartzite mylonite, granitic gneiss and pegmatite), and dispersed as elongated lenses within

214 glimmerite schist (Fig. 2f–i). The relationships detailed at our study site are representative of

215 the most biotite-rich (i.e., most extensively hydrated) rocks within the hydrous shear zones of

216 the Alice Springs Orogen.

217   In the field, quartzite lenses exhibit classic mylonite features including a strong

218 alignment of biotite and quartz to form a steeply plunging lineation. Mylonitic foliations in

219 quartzite and foliations in granitic gneiss are parallel to the foliation in glimmerite schist (Fig.

220 2e, f). Quartzite mylonite and granitic gneiss are partially dismembered by perpendicular

221 fractures of mm to cm size, exhibiting a range of isolated lenses and blocks ranging from mm

222 to dm size (Fig. 2d–i). Glimmerite is seen in brittle fractures oriented perpendicular to the

223 foliation, and in embayments and irregular partings that define the lenses and blocks (Fig. 2d–

224 i), where it contains the same mineral assemblage as observed in the main glimmerite schist

225 unit. Field evidence suggests that there is a near continuum between quartzite mylonite,

226 glimmerite schist and glimmerite, with an increasing amount of biotite content from < 10%

227 biotite (Bt) representing the original quartzite mylonite, to > 80% Bt in glimmerite (Fig. 2h, i).

228 This progressive increase in Bt content, commonly associated with structural features such as

229 fractures, suggests that there is a genetic relationship between Bt content and dismemberment.

230 Hence, field observations suggest an association between chemical change (i.e., reduced quartz

231 and feldspar content and increased Bt content), reaction and dismemberment commonly

232 observed in metasomatic zones (Ferry, 1994).

233

3. SAMPLE SELECTION AND ANALYSIS

3. 1. Sample selection

236 The majority of samples for this study (Fig. 2d–i and Fig. 4 onward) were

237 collected along a transect in the central section of the Gough Dam shear zone (GPS coordinates

238 23.1469°S, 134.56517°E, WGS84) and include components of quartz-rich mylonite and

239 glimmerite, along with the contacts between them. The remaining samples (Fig. 2a–c and Fig.

240 3) were collected close to the southern GDSZ boundary (Fig. 2a; GPS coordinates 23.1519°S,

241 134.56388°E, WGS84) and approximately halfway towards the centre of GDSZ (Fig. 2b, c;

242 GPS coordinates 23.1453°S, 134.56722°E, WGS84). The transect samples from the central

243 GDSZ were selected to represent the observed change in Bt content from a highly quartz-rich

244 mylonite (< 10 vol.% biotite) to glimmerite sleeves (> 80 vol.% biotite). We therefore selected

245 a sequence of four domains represented by three hand specimens and thin sections: (i) low,

246 < 10% Bt content, least-reacted quartzite (Sample L-Qtz [field ID number GD1609A]; Fig. 2h

247 and 4a), (ii) medium, ~10% Bt content, medium-reacted quartzite (sample M-Qtz [field ID

248 number GD1609C]; Fig. 2i and lower part of Fig. 4c), (iii) high, ~15% Bt content, highly-

249 reacted quartzite (sample H-Qtz [field ID number GD1609B]; Fig. 2i and 4b) and (iv)

250 glimmerite with > 80% Bt (Sample GL [field ID number GD1609C]; Fig. 2g–i and upper part

251 of Fig. 4c). It should be noted that within one hand specimen [field ID number GD1609C], the

252 boundary between M-Qtz and GL is contained. Unless otherwise stated, the two parts of the

253 same sample will be treated separately.

254

3. 2. Methods of analysis

256 *3. 2. 1. Petrography and quantitative orientation analysis*

257 Polished thin sections were made from blocks cut perpendicular to foliation and

258 parallel to lineation in both the quartz mylonite and glimmerite samples. Petrographic

259 observations of the mineral assemblages were made using a petrographic microscope, the

260 Virtual Petrographic Microscope (Tetley & Daczko, 2014) and ImageJ 1.47v (Rasband, 1997-

261 2018). Microstructural/crystallographic characterisation of thin sections was performed using

262 a FEI Quanta 650 FEG-ESEM with AZtec software and an Oxford/HKL Nordlys S EBSD

263 system at the University of Leeds, UK. EBSD mapping was performed covering large areas of

264 the thin section. Working conditions were 20 kV accelerating voltage, 20–26 mm working

265 distance, 70° specimen tilt and a step size of 8 μm. Automatic indexation was performed using

266 AZtec software (Oxford Instruments). HKL software was used to execute standard noise

267 reduction and to extrapolate missing data using at least and in succession 8, 7, 6 and finally 5

268 identical neighbours with similar orientation. Grain orientation maps using Euler angles and

269 inverse pole figures (IPF) were generated using the MTEX package (Bachmann *et al*., 2010;

270 Henry *et al*., 2017; Henry, 2018). Grains were calculated using a boundary angle > 10°

271 misorientation. Dauphine twin boundaries for quartz were defined as quartz-quartz grain

272 boundaries with a 60° rotation around the c-axis. Crystallographic preferred orientation (CPO)

273 of quartz was assessed using pole figures plotted on lower hemisphere, equal area projections

274 with one point per grain. J- and M-indices (Bunge, 2013; Skemer *et al.,* 2005) were computed

275 to quantify the intensity of the CPO. Average grain size of minerals was determined using the

276 EBSD data. We present grain size data where Dauphine twins were disregarded.

277

3. 2. 2. *Imaging and geochemical analysis*

278 *3. 2. 2. Imaging and geochemical analysis*

279   Micro X-ray Fluorescence (μ-XRF) analysis of the polished thin sections was

280 used for mineral identification and spatial distribution mapping, quantification of modal

281 proportions, and analysis of thin section-scale "whole-rock" major oxides. μ-XRF analyses

282 were performed using the Bruker M4 Tornado spectrometer at Macquarie University

283 Geoanalytical (MQGA), Sydney, Australia. The μ-XRF analyses were run with a tube voltage

284 of 50 kV, a beam current of 200 μA, a chamber pressure of 20 mbar, an acquisition time of 15

285 ms/pixel and a step size of 25 μm. AMICS (Advanced Mineral Identification and

286 Characterization System) was used to convert the X-ray fluorescence spectra to produce

287 detailed mineral maps.

288   High-resolution, high-sensitivity element mapping of polished thin sections on

289 pure silica glass slides was performed to assess the trace element distribution within the reacted

290 quartzite. Mapping was completed using the X-Ray Fluorescence Microscopy (XFM) beamline

291 and the Maia-384 detector on the Kirkpatrick-Baez mirror microprobe at the Australian

292 Synchrotron, Melbourne (Ryan *et al*., 2010; Paterson *et al*., 2011). A beam energy of 18.5 keV

293 focussed to a 2 μm spot size was used. Samples were scanned using a 4 mm step size in both

294 the x and y directions, at a speed of 4 mm/s and a dwell time of 1 ms/pixel. A set of standard

295 foils (Pt, Mn, Fe, YF3) were periodically analysed for calibration. Data reduction was

296 performed using GeoPIXE (Ryan *et al*., 1990), which uses the fundamental parameter model

297 for the layered sample, the Maia detector efficiency model, and the DA matrix method to

298 deconvolute spectra (Ryan *et al*., 1995, 2015).

299   Backscatter Electron (BSE) images and associated Energy-dispersive X-ray

300 spectroscopy (EDS) point analysis were used for both mineral identification and imaging of

301 microstructures. Polished thin sections were carbon coated and imaged in a Hitachi Desktop

302 Scanning Electron Microscope (SEM) at the OptoFab node of the Australian National
303 Fabrication Facility, Macquarie University, Sydney, Australia. The operating conditions of the
304 SEM were low vacuum and 15 kV accelerating voltage. For rapid identification of monazite
305 grains the BSE image based Mineral Liberation Analysis on the FEI 650 ESEM at the
306 University of Tasmania, Australia, was used at low vacuum and 20 kV accelerating voltage.

307 Cathodoluminescence (CL) images were used to identify alteration of elemental
308 concentration and/or crystallographic defects within quartz grains by migrating fluids
309 interacting with the quartzite samples from the study area. The CL images were acquired using
310 a Tescan VEGA3 XM SEM (University of Leeds, UK), operating at 20 kV and a beam current
311 of 20 nA.

312 Electron microprobe (EMP) analysis was used to acquire compositional data of
313 silicates using a JEOL JXA 8530F Plus field emission electron microprobe at the Central
314 Science Laboratory, University of Tasmania. The instrument is equipped with a field emission
315 source, running an accelerating voltage of 15 kV, a beam current of 15 nA and a beam size of
316 10 μm. The instrument has 5 wavelength dispersive spectrometers and is operated using the
317 Probe Software Inc. "Probe For EPMA" software package. Plagioclase Lake County,
318 Hornblende Kakanui, Augite Kakanui, Pyrope Kakanui, Olivine Springwater, Garnet Roberts
319 Victor Mine (all Smithsonian; Jarosewich *et al*., 1980) and Orthoclase from P&H
320 Developments UK were analysed as secondary standards to confirm the quality of the analysis
321 of the unknown material. A time dependent intensity correction was applied on Na and K if
322 applicable. Oxygen was calculated by cation stoichiometry and included in the matrix
323 correction. Hydrogen was calculated based on the mineral formula and included in the matrix
324 correction as well. The matrix correction algorithm utilised was Armstrong/Love Scott
325 (Armstrong, 1988) and the mass absorption coefficients dataset was LINEMU < 10 keV
326 (Henke, 1985) and CITZMU > 10 keV (Heinrich, 1966).

327   Laser Ablation Inductively Coupled Plasma Mass Spectrometry (LA-ICP-MS)

328 was used to collect monazite U-Pb data for age determination using a Resonetics M-50-LR 193

329 nm excimer laser coupled to an Agilent 7700x Quadrupole ICP-MS housed at Adelaide

330 Microscopy, University of Adelaide. Data were collected from thin sections using 5 pre-

331 ablation shots, 30 seconds of background measurement and 30 seconds sample ablation at 5

332 Hz, 2 J/cm^2 fluence and a spot size of 19 or 13 µm. The raw data signal was reduced using

333 Iolite software (Paton *et al.*, 2011). U-Pb fractionation was corrected using the MAdel monazite

334 standard (TIMS normalisation data: $^{207}Pb/^{206}Pb$ age = 492.01 ± 0.77 Ma; $^{206}Pb/^{238}U$ age = 517.9

335 ± 2.6 Ma; $^{207}Pb/^{235}U$ age = 513.13 ± 0.20 Ma; updated from Payne *et al.* (2008) with additional

336 TIMS analyses). Accuracy was also monitored by repeat analyses of the in-house internal

337 standard 94-222/Bruna-NW ($^{206}Pb/^{238}U$ = 450.2 ± 3.4 Ma; Maidment *et al.*, 2005).

338   The concentration of rare earth elements in biotite was determined in thin

339 section using a Teledyne Analyte Excite 193 nm excimer laser coupled to an 7700x ICP-MS

340 housed at Macquarie University Geoanalytical (MQGA), Sydney, Australia. Data were

341 collected from thin sections using 30 seconds of background measurement and 30 seconds

342 sample ablation at 10 Hz, 5 J/cm^2 fluence and spot size of 150 µm. Silicon (measured by EMP)

343 was used as an internal standard for all minerals, and NIST 610 and 612, and basalt from the

344 Columbia River (BCR-2) were used as external standards. The raw data signal was reduced

345 using GLITTER software (Griffin, 2008).

346

347 *3.2.3. Biotite and phengite geothermobarometers*

348   The *P-T* conditions for each sample are estimated using the titanium-in-biotite

349 thermometer of Henry *et al.* (2005) and the silicon-in-phengite geobarometer of Massonne &

350 Schreyer (1987). The Henry *et al.* (2005) geothermometer is based on the titanium (Ti a.p.f.u.)

351 composition of biotite in a peraluminous metapelite with Ti-bearing minerals (ilmenite or

352 rutile) and graphite in the mineral assemblage, equilibrated at 4–6 kbar. Temperature estimates

353 are made by using the concentration of Ti, Fe and Mg in the biotite present in all samples (Table

354 1 and Table S1). Compared to the Henry *et al*. (2005) experimental conditions for the

355 geothermometer, the expected accuracy of the temperature estimates is approximately ± 50°C.

356 This large error is due to the absence of graphite and accessory Ti-bearing minerals in the

357 samples from this study (Fig. 4). Henry *et al*. (2005) describes the Ti-in-biotite temperature

358 estimate calculations in the absence of Ti-saturating phases as being minimum temperatures

359 estimates.

360   The Massonne & Schreyer (1987) geobarometer is based on the silicon (Si)

361 composition in magnesium muscovite (phengite) coexisting with alkali feldspar,

362 phlogopite/biotite and quartz. The maximum Si composition in phengite can only be attained

363 using the geobarometer if the phengite coexists with the aforementioned mineral assemblage.

364 For phengite present in glimmerite schist and glimmerite, the pressure estimates are expected

365 to be precise since all the necessary minerals for the best pressure calculations are present (Fig.

366 4). Phengite in quartzite does not coexist with alkali feldspar, limiting the estimates to a

367 minimum pressure of formation (Fig. 4).

368

4. RESULTS

4. 1. Petrography and microstructures

4. 1. 1. Least Reacted Quartzite (sample L-Qtz; < 10 vol.% biotite)

372   Sample L-Qtz (field ID number GD1609A) contains the mineral assemblage

373 quartz (Qz; ~90 vol.%), biotite (Bt; 8 vol.%) and magnetite (Mt; < 1 vol.%) (Fig. 4a), along

374 with accessory zircon and monazite (see Section 4.4 for geochronology). The quartz grains

375 show variable morphology from equant, mostly visible in the smaller grains (100 μm to 1.5

376 mm), to elongate and ribbon-like in the coarser grains (2 to 3.5 mm) that define small shear

377 bands (Fig. 4a). Most quartz grains show signs of high mobility of the grain boundaries as

378 deeply sutured, curved and interlobate quartz-quartz boundaries (Fig. 4a). In areas of interstitial

379 biotite between quartz grains, the mobility of the quartz grain boundaries is retarded, from

380 curved or lobate quartz-quartz boundaries to more rectilinear quartz-biotite boundaries (Fig.

381 4a).

382   Biotite grains in quartzite are light to dark green, usually of rectangular shape

383 and show variable grain dimensions from 100 μm to 5 mm in length and up to 1 mm in width

384 (Fig. 4a). Magnetite and biotite are only locally associated and share straight grain boundaries.

385 However, magnetite has irregular boundaries with quartz. Biotite grains have a shape preferred

386 orientation consistent with the tectonic reverse sense of shear (i.e., ASO mid-crustal terrane

387 exhumation) for the high-strain zone. Biotite is concentrated between quartz grains or in

388 fractures present in quartzite (Fig. 4a).

389

4.1.2. Highly Reacted Quartzite (sample H-Qtz; < 15 vol.% biotite)

391   In comparison to sample L-Qtz, sample H-Qtz (field ID number GD1609B) is

392 similarly dominated by quartz (> 80 vol.%) and biotite (~15 vol.%), but has lower magnetite

393 (a few isolated grains) and contains sillimanite (Sil; < 5 vol.%) (Fig. 4b, d) and accessory zircon

394 and monazite. Quartz grains generally display an elongate or ribbon-shaped morphology and

395 are coarser than sample L-Qtz (Fig. 4b). The shape preferred orientation of the quartz

396 elongation is again consistent with the tectonic reverse sense of shear observed in the high-

397 strain zone. Sample H-Qtz shares with sample L-Qtz a high degree of grain boundary mobility

398 in the quartz grains, recorded as deeply sutured, curved and interlobate quartz-quartz

399 boundaries (Fig. 4b). However, in contrast to sample L-Qtz, sillimanite-rich bands are observed

400 where quartz-quartz boundaries are not as lobate (i.e., lower half of sample H-Qtz

401 photomicrograph in Fig. 4b), inverting the overall boundary migration to more rectilinear

402 quartz-quartz boundaries and to more common ribbon quartz morphology.

403 Biotite grains in sample H-Qtz are light to dark green, similar to the colouration

404 observed in sample L-Qtz (Fig. 4d). In comparison with sample L-Qtz, there is a two-fold

405 increase in biotite abundance, accompanied by an overall increase in grain size and the presence

406 of contiguous bands (Fig. 4b). Biotite grains present rare finger-like protrusions into adjacent

407 quartz grains and quartz-quartz boundaries (Fig. 5a–c). The protrusions terminate with low

408 apparent dihedral angles between the quartz-quartz grains (Fig. 5b, c). Biotite and muscovite

409 are locally intergrown, or muscovite truncates biotite (Fig. 5b, c). A sillimanite-biotite-rich

410 band is observed in the ribbon quartz domains and is highlighted by blue arrows in Fig. 4. In

411 this band (at left), a circular-shaped lens contains equigranular, euhedral to subhedral biotite,

412 plagioclase, muscovite and quartz, with the biotite interlocked with the host quartz (Figs. 4b, d

413 and 5a). A very fine-grained Al-Si-rich alteration product after plagioclase and sillimanite,

414 inferred to be kaolinite or pyrophyllite (labelled on figures as Alt.), occurs in grain boundaries

415 and micro-fractures (Figs. 4b, d, e and 5a).

416

417 *4. 1. 3. Medium Reacted Quartzite (M-Qtz) and Glimmerite (GL) contact (composite sample*

418 *M-Qtz/GL; ~10 vol.% and ~80 vol.% biotite)*

419 The M-Qtz and GL composite sample (field ID number GD1609C) includes the

420 contact between a quartzite sample (labelled M-Qtz) similar in mode to H-Qtz described above

421 and a glimmerite seam (labelled GL; both in Fig. 4c). The GL domain has 80 vol.% biotite, 8

422 vol.% quartz, 6 vol.% sillimanite and 6 vol.% muscovite (Fig. 4c), along with accessory zircon

423 and monazite (see Section 4.4 for geochronology).

424 Quartz grains in sample M-Qtz have strong similarities in their shape and

425 boundary morphology to those previously described for sample H-Qtz. Dissimilar to H-Qtz,

426 unusually shaped quartz grains with complex embayment-like grain concavities and

427 protrusions are observed in proximity to the large sillimanite-rich band present in the quartzite

428 (Figs. 4c and 5d, e). These unusually shaped quartz grains present a variation in

429 crystallographic grain orientation less than or equal to 10° within the respective selected group

430 (Fig. 6c, d). Multiple quartz grains near the quartzite-glimmerite contact are surrounded by

431 biotite, where biotite-filled embayments or fractures segment quartz grains with similar

432 crystallographic orientation (highlighted by white arrows in Fig. 6e).

433　　　　Biotite in sample M-Qtz displays generally similar grain shape and

434 microstructure to that described for sample H-Qtz, with the presence of finger-like protrusions

435 into adjacent quartz-quartz boundaries (Fig. 5d). Sillimanite occurs in bands within the

436 quartzite in association with biotite (Figs. 4c and 6a), forming elongate prisms that contain

437 extension fractures perpendicular to the sense of elongation. An alteration product inferred to

438 be kaolinite or pyrophyllite is observed in the fractures (Fig. 5e).

439　　　　All the major minerals show a strong preferred orientation in the GL domain

440 (Fig. 5f). Similar to sample H-Qtz, biotite and muscovite are sometimes intergrown or truncate

441 each other (Fig. 5f), with muscovite truncating biotite more commonly. Sillimanite has an

442 elongate morphology similar to that observed in M-Qtz and H-Qtz samples, with fractures

443 perpendicular to their long axes (Fig. 5e), and is likewise concentrated into bands (Fig. 4c).

444 Some of the coarser muscovite grains show a sigmoidal shape with biotite wrapping around

445 the larger grains (Fig. 5f). A notable higher proportion of monazite and zircon grains is

446 observed in the GL component compared to the quartzite domain (i.e., M-Qtz) and the other

447 L-Qtz and H-Qtz samples (see Section 4.4 for geochronology).

448　　　　SEM cathodoluminescence (CL) images (Fig. 5g–j) show darker CL bands that

449 transgress single grains of quartz. The darker luminescence intensity is also observed at the

450 majority of quartz grain boundaries. These microstructures observed in the CL images are more

451 common in domains where biotite mode is higher (within or close to sample GL) (Fig. 5g, h),

452 and are less common in domains with lower biotite mode (sample M-Qtz) (Fig. 5i, j).

453

4. 2. Quantitative quartz orientation analysis

455 Inverse pole figure (IPF) coloured maps for both the M-Qtz and H-Qtz samples

456 show a crystallographic preferred orientation (CPO) of quartz grains. The majority of grains

457 contain c-axes oriented perpendicular to the stretching lineation of the sample (Fig. 7a, d).

458 Crystallographic orientation data for quartz grains in samples M-Qtz, GL and

459 H-Qtz are plotted as pole figures (Fig. 7b, e). The J-index for all quartz grains in the mapping

460 area is 2.86 and 3.35 for the composite M-Qtz/GL sample and H-Qtz sample, respectively (Fig.

461 7b, e). Sample M-Qtz/GL quartz CPO features a [c]-axis maximum parallel to the Z-axis and

462 spreading 45° in the NE sector of the pole figure. <a> axes define a distribution close to the

463 primitive circle with a main cluster for the <a> axis at ~30–45° from X-axis (Fig. 7b).

464 Results from the three regions delineated for the composite M-Qtz and GL

465 sample map show a discrepancy in the quartz CPO, J- and M-index between the different

466 regions (region 1 vs regions 2 and 3), or put simply, sample GL vs sample M-Qtz (Fig. 7a, b).

467 Region 1 quartz CPO presents a [c]-axis maximum close to the primitive circle at ~45° from X

468 and Y in the SE sector. This distribution of the <c> axes in region 1 contrasts from the observed

469 <c> axes distribution close to Z for both regions 2 and 3 in sample M-Qtz. Overall, axes are

470 more randomly distributed in region 1, i.e., sample GL, when compared to regions 2 and 3, i.e.,

471 sample M-Qtz (Fig. 7b). In regions 2 and 3, a clustering of <a> axes close to the primitive

472 circle is distinguishable (Fig. 7b).

473 For quartz in sample H-Qtz, variations between regions are not as extreme as

474 those observed for the composite M-Qtz/GL sample. The <c> axes in regions 1 and 2 are

475 similar to the entirety of quartz in the H-Qtz sample – a NE-SW oriented girdle. However, a

476 [c]-axis maximum is observed in region 1 close to the primitive circle, whereas for region 2,

477 the two maxima are close to Z. The <a> axes show similar orientation and distribution for both

478 regions, with a slightly greater girdle like distribution in region 1 (Fig. 7e).

479 The distribution of misorientation between pairs of quartz grains shows a higher

480 proportion of angles below 60° than that expected for a random quartz distribution for the

481 composite sample M-Qtz/GL (Fig. 7c), when compared to sample H-Qtz data (Fig. 7f). In

482 sample M-Qtz/GL, quartz misorientation contrasts between region 1 and the full sample quartz

483 misorientation data set (Fig. 7c). Region 1 shows a nearly perfect random distribution, whereas

484 sample M-Qtz/GL returns a stronger correlated misorientation distribution. H-Qtz quartz

485 misorientations are similar to sample M-Qtz/GL, a slightly correlated distribution below 60°

486 (Fig. 7f).

487 Mean area weighted (MAW) quartz grain size analysis was performed for the

488 regions of samples M-Qtz/GL and H-Qtz indicated in Fig. 7. A higher MAW quartz grain size

489 is observed for region 2 in sample M-Qtz (1752.63 mm^2) and region 1 in sample H-Qtz

490 (1638.63 mm^2), compared to region 3 (1258.39 mm^2) and region 2 (1067.04 mm^2) in samples

491 M-Qtz and H-Qtz, respectively. Quartz grains in sample GL have an order of magnitude lower

492 MAW quartz grain size (130.73 mm^2) compared to the remaining regions described above (Fig.

493 8).

494

4. 3. Geochemistry

496 *4. 3. 1. Glimmerite whole-rock major oxides*

497 Harker diagrams for whole rock compositions of representative samples of the

498 Strangways Metamorphic Complex (SMC; Fig. 1), including ferromagnesian granulites,

499 quartzo-feldspathic granulites, metasedimentary rocks and the schistose high-strain zones that

500 cut them (Table S2), are compared with data from sample GL in Fig. 9. The felsic and mafic

501 granulites generally form a distribution similar to fractionation trends in igneous suites (e.g.,

502 Winter, 2013; Frost & Frost, 2013). Data for granulite facies metasedimentary rocks of the

503 SMC and the schistose high-strain zones largely follow the same trends, forming a distribution

504 from lower SiO_2 metapelitic rocks to higher SiO_2 metapsammitic rocks. Quartzo-feldspathic

505 granulites are distinct in their high SiO_2 content and lower concentrations of Al_2O_3, TiO_2, FeO,

506 and MgO. Data for schistose high-strain zones is difficult to distinguish from the geochemical

507 pattern of metasedimentary rocks, having generally lower K_2O concentrations. Sample GL has

508 lower Al_2O_3, higher K_2O and higher TiO_2 compared to the majority of the other rocks from

509 schistose high-strain zones in the SMC, and lower concentration of TiO_2 and FeO compared to

510 average biotite in peraluminous granite (Neiva *et al*., 2002).

511

512 *4.3.2. Mineral major element chemistry and geothermobarometry*

513   The totals for added H_2O in biotite from all samples averages 3.97 wt% (Table

514 1 and Table S1). Biotite in sample L-Qtz has variable composition, whereas biotite in the

515 remaining samples is relatively homogeneous (Table 1 and Table S1; Fig. 10a–c). Comparison

516 between samples reveals a higher biotite TiO_2 content (ranging from 0.29 to 2.7 wt%) and

517 Al_2O_3 content (16.64 to 22.03 wt%) from sample L-Qtz to samples H-Qtz/GL (Table 1 and

518 Table S1). Conversely, X_{Mg} [= Mg/(Fe+Mg)] is higher in sample L-Qtz relative to sample H-

519 Qtz (Fig. 10a). Biotite compositions in all samples have moderate concentrations of F (average

520 0.71 wt%) and Cl (average 0.26 wt%; Table 1 and Table S1) compared to published igneous

521 biotite data (Neiva *et al*., 2002; Cesare *et al*., 2008).

522   Fig. 10b employs the Nachit *et al*. (1985) biotite classification and the vast

523 majority of biotite analyses (~90%) plot in the combined peraluminous fluid domain and

524 overlap with igneous biotite data from Stussi & Cuney (1996). Samples H-Qtz, M-Qtz and GL

525 plot tightly in the same region of the diagram, each cluster showing a minor range in Al and

526 Mg. In contrast, biotite from sample L-Qtz contains noticeably higher Mg and a broader range

527 of values (Fig. 10b). Using the Nachit *et al.* (2005) biotite classification diagram (Fig. 10c),

528 based on the assumption that the magmatic biotite composition (primary biotite) is altered with

529 hydrothermal interaction (reequilibrated biotite) or neocrystallisation at equilibrium with that

530 fluid (neoformed biotite), analyses from samples M-Qtz, H-Qtz and GL cluster within the

531 reequilibrated biotite domain and adjacent to the primary biotite domain, whereas sample L-

532 Qtz is scattered between the neoformed and reequilibrated domains (Fig. 10c).

533 A plot of Ti a.p.f.u. vs X_{Mg} shows a trend of increasing Ti with decreasing X_{Mg}

534 in biotite from samples L-Qtz to M-Qtz/GL (Fig. 10a). The application of the Ti-in-biotite

535 thermometer of Henry *et al.* (2005) indicates highly variable minimum formation temperatures,

536 300–570 ± 50°C for sample L-Qtz, ~630 ± 50°C for biotite in the quartzite and glimmerite

537 domains of sample M-Qtz/GL, and ~660 ± 50°C for sample H-Qtz (Fig. 10d). In samples where

538 phengite is present (H-Qtz and M-Qtz/GL), the Si-in-phengite geobarometer of Massonne &

539 Schreyer (1987) was applied. Phengite from these samples has Al slightly lower than 3.1

540 a.p.f.u., which in combination with the Ti-in-biotite thermometer (Henry *et al.*, 2005) allows

541 us to discriminate their pressure of formation at ~5–6 kbar (Fig. 10d).

542

543 *4.3.3. Biotite trace element chemistry*

544 For all samples, the biotite concentrations for Zr, Hf, Nb and Ta (Table 2 and

545 Table S3) normalised to chondrite (McDonough & Sun, 1995) follow similar patterns. Zr

546 values are consistently below chondrite (Fig. 11a–c), except for a single biotite analysis from

547 the M-Qtz/GL sample (Fig. 11c). Hf values are likewise depleted but to a lesser extent (Fig.

548 11a–c), except the distinctive Zr-enriched biotite grain conjointly enriched in Hf (Fig. 11c). In

549 contrast, Nb and Ta show consistently enriched values (Fig. 11a–c). A distinction in Nb and

550 Ta concentration is observed between the values for sample L-Qtz and the other samples (H-

551 Qtz and M-Qtz/GL), with the former having lower Nb concentrations (Figs. 11a–c), but with

552 equivalent values of Ta (L-Qtz ~167–232; pair H-Qtz ~143–225 times chondrite; Table 2 and

553 Table S3; Figs. 11a–c and 12).

554 Biotite compositions in all samples have a consistent depletion in rare-earth

555 elements (REE) compared to the chondrite values (Table 2 and Table S3; Fig. 11a–c). Similar

556 relations are observed for all chondrite-normalised REE patterns, with a small decrease in

557 LREE trending down from La to Sm in comparison to the HREE (Fig. 11a–c). Compared to

558 biotite in samples H-Qtz and M-Qtz/GL, sample L-Qtz shows less variability and a general

559 increase in normalised LREE concentrations (Fig. 11a–c). Overall, biotite HREE patterns are

560 concave-up due to higher Lu normalised concentrations relative to the remaining HREE.

561 Pronounced positive Eu anomalies are evident for sample H-Qtz and the combined M-Qtz and

562 GL biotite analyses (Fig. 11a–c), with a slightly smaller anomaly for sample L-Qtz.

563 Fig. 11d and Table S4 contain compilations of published values and sample

564 descriptions for natural and experimental igneous and metamorphic biotite normalised to

565 chondrite (McDonough & Sun, 1995). The Gough Dam samples have lower normalised

566 concentrations of Zr, Hf and REE compared to the compiled data, but have similar values of

567 Nb and Ta, especially in relation to the metapelitic peraluminous El Hoyazo enclaves (Fig. 11;

568 Cesare *et al*., 1997; Acosta-Vigil *et al*., 2010) and the peraluminous migmatite biotite of Bea

569 *et al*. (1994*b*).

570 Two groups of biotite composition are distinguishable on bivariate plots of Ba,

571 Sc, Nb/Ta and Li versus Ti (Fig. 12). Sample L-Qtz is depleted compared to H-Qtz and the

572 combined M-Qtz and GL biotite analyses, with the pooled data forming a positive correlation

573 trend. The enrichment trend is also visible if compared to El Hoyazo biotite (Acosta-Vigil *et*

574 *al*., 2010), which plots as either the most Ba and Li enriched analyses or at equivalent Nb/Ta

575 and Sc concentrations with samples H-Qtz, M-Qtz and GL (Fig. 12). Within the group of Nb-

576 enriched biotite (samples H-Qtz, M-Qtz and GL), Nb/Ta values above the chondritic

577 concentrations of McDonough & Sun (1995) is observed for H-Qtz and the combined M-

578 Qtz/GL samples. In general, the order of Ba, Li and Sc in biotite concentration is observed as

579 L-Qtz < M-Qtz/GL ≤ H-Qtz.

580

4. 4. Geochronology (U-Pb monazite)

582 Monazite grains from samples GL (higher biotite vol%; Fig. 13a–b) and M-Qtz

583 (lower biotite vol%; Fig. 13c–d) are generally located in contact with biotite and sillimanite,

584 with few grains adjacent to or in contact with quartz grain boundaries or quartz fractures.

585 Monazite grains are rounded to elongate and commonly 50–200 μm in the longest dimension

586 (Fig. 14a), but with size range of ~30–200 μm. Several grains show zonation in BSE images,

587 usually with a higher BSE response in core domains or multiple zones of variable BSE

588 responses (Fig. 14a). Curved, inward-penetrating boundaries and cross-cutting textures are

589 visible in some grains between core and rim domains. Thorite inclusions (after monazite) are

590 mostly contained in brighter zones near grain boundaries or in micro-fractures (Fig. 14a). The

591 remaining monazite grains are mostly smaller (< 100 μm) and do not display any observable

592 BSE zonation.

593 In all, 56 U-Pb spot analyses on 39 monazite grains from sample M-Qtz/GL

594 were obtained using LA-ICP-MS. Data with < 2% discordance show two discernible age

595 groups (Fig. 14b and Table S5): (i) Palaeoproterozoic ages ranging from 1740 ± 16 Ma to 1630

596 ± 19 Ma with a density age estimation at c. 1669 Ma (all located in sample M-Qtz); and (ii)

597 Neoproterozoic to Palaeozoic ages ranging from 606.1 ± 5.8 Ma to 370.9 ± 3.6 Ma with a

598 density age estimation at c. 451 Ma. The remaining discordant analyses from sample M-Qtz fit

599 a linear regression with an upper intercept age of 1669 ± 4.6 Ma and a lower intercept age of

600 482 ± 2.5 Ma, resulting in a high MSWD of 11 due to significant dispersion (Fig. 14b). The

601 majority of these analyses are spread along the discordia line towards its upper intercept, with

602 the remaining analyses together with those of sample GL forming a concordant but widely

603 spread Palaeozoic age group, creating a complex age distribution at the lower intercept (Fig.

604 14b).

605 The obtained monazite ages are not homogenously distributed throughout

606 samples M-Qtz and GL (Fig. 14 and Fig. S1). All the Early Palaeoproterozoic ages are derived

607 from sample M-Qtz, which also contains isolated grain clusters that record a spread of

608 Neoproterozoic-Palaeozoic ages (Fig. S1). In contrast, the adjacent glimmerite component

609 (sample GL) hosts monazite with a much narrower age range, with most monazite returning

610 Palaeozoic ages and a greater proportion falling within the age bounds of the ASO (*c*. 450–

611 300 Ma; Fig. 14 and Fig. S1). Discordant ages from both samples are derived from rim domains

612 and/or monazite grains with lower BSE response zones (Fig. 14a).

613

5. DISCUSSION

5. 1. Conditions and timing of glimmerite formation

616 Combined petrography, mineral geochemistry and monazite geochronology

617 across the cm-scale transect of variably modified quartzites from the GDSZ provides insight

618 into the geological environment and *P-T-t* evolution responsible for their conversion to

619 glimmerite. The *P-T* estimation using mica in the studied samples suggests mid-crustal

620 conditions of $P > 4$–5 kbar and $T > 570$–660°C (Fig. 10), similar to previously published

621 conditions from the GDSZ and other Palaeozoic shear zones in the Strangways region (Ballèvre

622 *et al*., 2000; Bendall, 2000). Due to the lack of graphite and Ti-bearing accessory minerals in

623 all samples, temperature estimates are minimum values. Note that this also influences the

624 pressure estimates, suggesting they are likewise minimum values. The upper end of the

625 calculated temperature range lies above the hydrous granite solidus and is consistent with field

626 observations of syn-deformational granitic lenses in the GDSZ (Fig. 2e), along with

627 microstructural evidence for the former presence of melt (Figs. 3 and 5; see Section 5.2).

628 The relationship between age and textural position of monazite grains in the

629 studied samples suggests that the precursor quartzite initially experienced metamorphism

630 during the Strangways Event (*c*. 1735–1690 Ma), followed by a subsequent metamorphic

631 overprint during the Alice Springs Orogeny (*c*. 450–300 Ma) that is largely recorded in the

632 cross-cutting glimmerite (Fig. 14). The morphology and zonation of the monazite grains (e.g.,

633 inward-penetrating boundary interfaces, radial or spotted domains of lower BSE intensity,

634 porous textures usually present near monazite rims and micro-fractures, and thorite inclusions

635 near micro-fractures) suggests that they were modified by fluid-assisted (melt-assisted)

636 coupled dissolution-precipitation reactions (Fig. 14a; Ruiz-Agudo *et al.*, 2014; Varga *et al.*,

637 2020). The lower intercept age of *c*. 482 Ma intersects a significant spread of Neoproterozoic

638 to Palaeozoic apparent ages (*c*. 606–371 Ma; Fig. 14). These complex age data can be

639 interpreted in two ways: (i) the individual analyses are geologically meaningful and reflect

640 multiple reactivation and fluid flow events in the GDSZ; or (ii) the data reflect partial age-

641 resetting of inherited monazite grains during one or more Palaeozoic reactivation and fluid

642 flow events in the GDSZ.

643 Coupled dissolution-precipitation experiments have demonstrated that the

644 replacement of precursor monazite can be incomplete, either (i) forming nanomixtures of old

645 and new monazite (Grand'Homme *et al.*, 2016, 2018); or (ii) imperfectly resetting inherited

646 grains (Varga *et al.*, 2020). Temperature and other physicochemical parameters (i.e., fluid

647 composition, mineral composition, solubility, primary zoning, stress, reaction kinetics) greatly

648 affect the replacement mechanism that redistributes U-Pb cargo in the precipitated monazite

649 (Ruiz-Agudo *et al.*, 2014; Grand'Homme *et al.*, 2016, 2018; Varga *et al.*, 2020). Additional

650 detailed nano-scale studies of the monazite reaction front microstructures (i.e., nanocracks,

651 dislocations, pores/inclusions) are therefore necessary to better understand the exact extent of

652 monazite partial age resetting and significance of the age spread observed in Fig. 14b. As such,

653 we prefer the interpretation that the Neoproterozoic to Palaeozoic dates from individual

654 monazite grains in this study are unlikely to be geologically meaningful. The younger age

655 cluster ending at c. 370 Ma cannot be used to precisely date the timing of glimmerite formation,

656 beyond stating that it aligns with the Alice Springs Orogeny.

657 Despite this, the spread of concordant ages across this interval may hold some

658 utility in establishing the broad context for melt-rock interaction. Notably, the most significant

659 age peak from the kernel density estimation occurs at c. 451 Ma (Fig. 14b), consistent with U-

660 Pb monazite ages from similar shear zones of the Strangways Range (Möller *et al*., 1999;

661 Howlett *et al*., 2014; Fournier *et al*., 2016). This timing reflects the earliest onset of the Alice

662 Springs Orogeny recorded regionally (Mawby *et al*., 1999; Raimondo *et al*., 2014), perhaps

663 suggesting a peak in melt-rock interaction and the replacement or resetting of inherited

664 monazite driven by large-scale changes to deep crustal architecture and basin dynamics that

665 initiated at this time. Similarly, the youngest monazite age of c. 370 Ma most probably

666 represents the last recorded stage of overprinting metamorphism and/or monazite resetting,

667 perhaps due to either (i) the cessation of melt migration in the shear zone; or (ii) shear zone

668 armouring physically buffering melt interaction with the monazite grains incompletely reset by

669 coupled dissolution-precipitation processes (Daczko *et al.,* 2016; Varga *et al*., 2020).

670

5. 2. Microstructural evidence for melt-present deformation

672 Of the five key microstructures indicative of the former presence of melt in

673 high-strain zones outlined by Stuart *et al*. (2018*b*), four were observed in the GDSZ samples:

674 (1) small dihedral angles between biotite, muscovite, plagioclase, magnetite and quartz (Figs.

675 3 and 5b, c); (2) elongated grain boundary films pseudomorphed by biotite (Figs. 3 and 5a–e);

676 (3) neighbourhoods of small quartz grains connected in three dimensions (Figs. 4c and 6c, d),

677 presenting less than ten degrees of crystallographic misorientation between grains; (4)

678 pseudomorphs of inferred cm-scale granitic melt pockets (Fig. 4b, d). Other microstructures

679 that support the interpretation of melt-present deformation in the GDSZ include myrmekitic

680 intergrowth of quartz and plagioclase (Fig. 3b, f and h), biotite finger-like protrusions into

681 adjacent quartz grains (Fig. 5a–d) and CL images highlighting bands of darker CL response in

682 quartz near biotite (Fig. 5g–j), possibly reflecting microscale fracturing, melt migration

683 pathways or a CL signature of quartz grain boundary migration (Holness & Watt, 2001;

684 Bergman & Piazolo, 2012).

685 Quartz crystallographic orientation data also supports the former presence of

686 melt in the GDSZ. Quartz grains in the H-Qtz and M-Qtz samples have a strong CPO, whereas

687 quartz grains present in sample GL have a weak CPO and apparent activation of multiple slip

688 systems (Fig. 7). The strong CPO for the highly strained quartzite represented by H-Qtz and

689 M-Qtz samples is not accompanied by intense recrystallisation of the quartz grains. Rather,

690 they record widespread grain boundary migration, implying deformation temperatures > 500°C

691 (Stipp *et al*., 2002*a, b*). The observed CPO in quartzite is most probably an inherited structure

692 formed at granulite facies, suggesting that the prism <a> slip system was activated during the

693 Strangways Event rather than the ASO. This interpretation is informed by the lack of plastic

694 shear band microstructures in the quartzite samples and quartz grains (Gapais & White, 1982;

695 Menegon *et al*., 2011) or quartz grain size reduction by recrystallisation (Hobbs, 1968; Lloyd

696 & Freeman, 1994; Lloyd, 2004; Stipp *et al*., 2010) (Figs. 4b, c, 7 and 8). The absence of these

697 features supports that melt flux pathways initiated within the quartzite mylonite, subsequently

698 crystallising biotite-sillimanite-muscovite and resulted from shear or dilation fractures in a

699 brittle regime (Lloyd & Knipe, 1992).

700 Alongside the deformation processes documented above, diffusion or

701 dissolution-precipitation creep likely played a role in quartz deformation within the proposed

702 fluid saturated high-strain zone (Farver & Yund, 2000; Bestmann *et al*., 2004). However, the

703 presence of fluid could imply reaction with quartz grains and silica loss to the local open

704 system, undercutting these creep processes (Farver & Yund, 2000). Small quartz grains in

705 sample GL (Fig. 7a) have a weak CPO and apparent activation of multiple slip systems (Fig.

706 7b, c), suggesting that they are not relict grains from the main quartzite present in sample M-

707 Qtz, but instead are the product of neoformed quartz grains that may reflect melt

708 pseudomorphs. These results are consistent with the concomitant deformation of quartz grains

709 in the presence of melt, causing strain to be accommodated into the weaker surrounding melt

710 and preventing the development of a strong CPO similar to that observed for quartz grains in

711 samples M- and H-Qtz (Fig. 7; Lee *et al*., 2018; Stuart *et al*., 2018*b*; Prakash *et al*., 2018).

712 A larger mean quartz grain size is observed in domains of the quartzite samples

713 close or in contact with the inferred melt pathways (Fig. 8). This may be explained by a local

714 increase in temperature due to melt flux (Fig. 7a, d), leading to static grain growth and recovery

715 of quartz grains in samples M-Qtz (close to glimmerite) and H-Qtz (distant from glimmerite).

716 An enhanced local effect suggests that melt flux was short-lived and probably involved several

717 pulses, rather than a continuous and long-lived event that would anneal a larger area of the

718 sample (Heilbronner & Tullis, 2002; Piazolo *et al*., 2006). During annealing, grain boundaries

719 migrate to cross biotite boundaries (e.g., Piazolo *et al*., 2006), matching the increase in biotite

720 enclosure with proximity to the inferred melt pathways (Figs. 6e and 7). Furthermore, subgrain

721 boundaries are rearranged and decrease in abundance (e.g., Borthwick & Piazolo, 2010), while

722 the intensity of CPO simultaneously increases (Piazolo *et al*., 2010). The much smaller mean

723 size of quartz present in sample GL can be related to its neoformation by melt pseudomorphism

724 as previously described.

725 Taken individually, the microstructures described above are not sufficient

726 evidence to argue convincingly for the former presence of quartzo-feldspathic melt in high-

727 strain zones. However, it is the combination and consistency of several key microstructural

728 features that lends weight to our interpretation. It should also be noted that preservation of such,

729 often delicate, microstructures is highly dependent on subsequent metamorphic and tectonic

730 overprinting during the retrograde path, which can potentially obliterate indicators of former

731 melt presence (Holness *et al.,* 2011). This suggests that once the melt had crystallised, the

732 quartzites were rheologically strong, enabling their microstructural features to be preserved.

733

5. 3. Glimmerite geochemistry and links to melt-rock interaction processes

735 Comparing the whole rock major oxide geochemistry of sample GL against the

736 range of precursor rock types present in the SMC is necessary to understand the chemical

737 evolution between the different glimmerite rocks present in the study area and their possible

738 protoliths. This analysis reveals that the glimmerite composition generally plots between the

739 major oxide values for ferromagnesian granulites, schistose high-strain zone metapelites and

740 average peraluminous granitic biotite (Fig. 9). It does not demonstrate a strong relationship

741 between the glimmerite rock observed in the GDSZ and the surrounding rocks from the SMC,

742 suggesting that glimmerite formation involved a high degree of geochemical modification (i.e.,

743 metasomatism *s.l.*). We infer that reactive flow of melt through high-strain zones led to the

744 near-complete dissolution of a SMC protolith, likely one of the quartzo-feldspathic rocks based

745 on the scarcely preserved layers and lenses within the glimmerite outcrop. Coupled

746 precipitation of abundant biotite pushed the whole rock composition of the resulting glimmerite

747 rock to values similar to the average composition of peraluminous granitic biotite (Fig. 9).

748 Biotite geochemistry affords more detailed insights into the mechanism of melt-

749 present deformation driving glimmerite formation in the GDSZ. Micas (particularly biotite)

750 dominate both the glimmerite mineral assemblage and the inferred reactive melt pathways

751 within the partially modified quartzite samples (Figs. 2d–i, 4–7). Low to moderate F and Cl

752 concentrations (< 1 wt%; Table 1 and Table S1) suggest their formation from a halogen-

753 depleted melt in an open system. This interpretation is based on the recognition of F- and Cl-

754 enriched biotite commonly being formed by halogen-enriched aqueous fluid-rock interaction

755 in high-grade metamorphic rocks where fluid flux was limited (Markl & Bucher, 1998; Markl

756 & Piazolo, 1998; Aranovich & Safonov, 2018). In these case studies of metamorphic fluids, a

757 correlation between halogen-enriched fluids and halogen-enriched biotite reaction formation

758 could be made.

759 The discrimination of biotite formed at supra-solidus versus sub-solidus

760 conditions is not straightforward. Here, the discrimination diagram of Nachit *et al*. (1985) is

761 used, based on the inference of melt-present conditions (Fig. 10b). The diagram is potentially

762 useful to inform on the type of melt involved in the formation of glimmerite. The

763 peraluminosity of the melt is inferred from the aluminium content of the equilibrated

764 crystalised biotite (Nachit *et al*., 1985; Stussi & Cuney, 1996), and our data suggest that the

765 glimmerite biotite shares similarities with magmatic biotite formed from and partially

766 reequilibrated with peraluminous melt.

767 The highly dispersed and overall lower concentrations of Al, Ti, Fe and Mn of

768 biotite in the L-Qtz sample likely reflects variable inheritance from the precursor quartzite over

769 the externally-derived melt, compared to H-Qtz and M-Qtz/GL biotite that are dominated by

770 the chemistry of the melt (i.e., are melt buffered; Fig. 10). About half of the L-Qtz biotite

771 analyses trend towards the formation conditions observed for H-Qtz and M-Qtz/GL biotite,

772 showing an increase in the apparent temperature of formation and peraluminosity of

773 crystalising fluid, concordant with the igneous biotite data compiled by Stussi & Cuney (1996).

774 This wide range of L-Qtz biotite compositions likely records variable but low local melt

775 volumes in the quartzite domains and lower degrees of equilibration with the fluxing melt.

776   The trace element analyses of biotite in all samples shows lower chondrite-

777 normalised concentrations of REE, Zr and Hf, and higher chondrite-normalised concentrations

778 of Nb and Ta (Figs. 11 and 12). The latter likely resulted from the high partition coefficient

779 between biotite and granitic melt. In their experiments and analysis of natural examples,

780 Stepanov & Herman (2013) observed that biotite in partial melting environments preferentially

781 incorporates Nb over Ta, concluding that biotite plays an important role for the missing Nb-

782 paradox observed in the continental crust. Such behaviour of Nb incorporation over Ta in

783 biotite is observed in our data, with most of the L-Qtz biotite grains having lower Nb

784 concentrations and overall lower Nb/Ta ratios when compared to H-Qtz, M-Qtz and GL biotite

785 grains (Figs. 11 and 12).

786   Concentrations of Zr and Hf in biotite for all samples are similar, although lower

787 than biotite from the partially melted metapelite enclaves of El Hoyazo (Acosta-Vigil *et al*.,

788 2010), and lower by 2–3 orders of magnitude when compared to rhyodacite biotite (Nash &

789 Crecraft, 1985) (Fig. 11). The broad similarity in Zr and Hf concentrations between GDSZ and

790 El Hoyazo biotites can be explained by the presence of abundant zircon grains in equilibrium

791 with biotite, possibly allowing the zircon grains to buffer Zr and Hf in the melt (Acosta-Vigil

792 *et al*., 2010; Nash & Crecraft, 1985). Conversely, for rhyodacite biotite it is proposed that Zr

793 and Hf melt buffering did not occur during rhyodacite formation due to the relative timing of

794 zircon crystallisation.

795   The GDSZ biotites are characterised by low REE values, with chondrite-

796 normalised concentrations below unity and slightly higher LREE over HREE (Fig. 11a–c).

797 Such low REE concentrations are similar to the biotite REE chemistry of (i) partial melting

798 environments of the El Hoyazo enclave; (ii) migmatite leucosomes of the Peña Negra Complex

799 (Acosta-Vigil *et al*., 2010; Bea *et al*., 1994*b*); and (iii) closed system fractional crystallisation

800 from peraluminous melt of the Pedrobernardo pluton (Bea *et al*., 1994*a*) (Fig. 11d). The REE-

801   depleted patterns observed for El Hoyazo, Peña Negra and Pedrobernardo biotites (Table S4

802   and Fig. 11d) are consistent with the presence of accessory minerals with higher REE

803   mineral/melt partitioning ratios (i.e., monazite, zircon and apatite), or by REE-depleted melt

804   due to REE retention in residual accessory minerals in the palaeosome (e.g., Prent *et al.,* 2019;

805   Gonçalves *et al.,* 2021). Either of these mechanisms, or a combination of both, can be

806   considered viable explanations for the biotite REE compositions of H-Qtz, M-Qtz and GL

807   samples in our study, due to the presence of monazite and apatite in equilibrium with the

808   glimmerite mineral assemblage, or the externally-derived melt becoming depleted in REE

809   following crystallisation and fractionation of mineral phases rich in REE during prior melt

810   flow.

811   Another explanation for the REE-depleted signature of GDSZ biotites may be a

812   smaller equilibration volume and/or variable composition of the fluxing melt, as proposed by

813   Stuart *et al.* (2018*a*) for variable igneous-like mineral REE patterns measured in metasomatic

814   reactions of the Pembroke Granulite (lower crust of a magmatic arc). Similar to this example,

815   the presence of melt in the GDSZ is implied by the overall enrichment of incompatible elements

816   in biotite from sample L-Qtz to the group of samples H-Qtz, M-Qtz and GL (Fig. 12). The

817   abundance of incompatible elements follows an approximately linear trend with the apparent

818   increase in temperature informed by the increase in biotite Ti concentrations. This pattern can

819   be attributed to a more fractionated melt enriched in incompatible elements (Cerny, 1991;

820   O'Connor *et al.,* 1991; Bea *et al.,* 1994*a*), or an increasing melt equilibration volume from

821   sample L-Qtz to the most reacted samples, possibly also explaining the apparent temperature

822   difference between sample L-Qtz and sample H-Qtz.

823

5. 4. Estimates of reactive hydrous melt volume

825   In the absence of field evidence for upper crustal melt products that would have

826 formed as a consequence of melt flux (i.e., igneous intrusions), we use the modal percentage

827 of hydrous phases in the studied samples to estimate the reactive hydrous melt volume. We

828 consider that the hydrous phases present are a consequence of the time-integrated volume of

829 reactive melt flux (Daczko *et al*., 2016; Stuart *et al*., 2018*a*). Arrested melt flux may occur

830 upon draining of source magma or cooling and freezing of the system. Regardless, the cessation

831 of reactive melt flux formed the microstructures observed in the studied samples, and their

832 patterns suggest that melt migration occurred mainly via intergranular mechanisms and was

833 focused along structurally-controlled planes, such as the main rock foliation and mode 1

834 fractures potentially formed by melt pathway fluid overpressure (Sibson, 1988). In the

835 glimmerite domain, the extreme metasomatism implies higher reactive melt flux, where the

836 protolith is partially dissolved by the reactive melt, enhancing porosity and permeability (e.g.,

837 Daczko *et al*., 2016) and providing a positive feedback that allows a higher volume of melt

838 flow and consequent increase in the formation of hydrous minerals. In this section, the

839 minimum volume of reactive hydrous melt needed to form glimmerite is estimated for each of

840 the Gough Dam shear zone samples. The observed increase in modal biotite and muscovite of

841 the GDSZ can be related to a melt flow-derived hydrous metasomatism mechanism (following

842 the method of Stuart *et al*., 2018*a*). With this interpretation in mind, the calculation of a

843 minimum volume of reactive melt flux can be derived.

844   The minimum volume of reactive melt flux is calculated considering a

845 hypothetical cube of quartzite rock containing no hydrous phases and an H_2O content close to

846 0%. The proportions and average H_2O content of the hydrous phases formed during the melt-

847 rock hydrous metasomatism (i.e., biotite and muscovite in this study) are used to calculate the

848 final H_2O composition of our hypothetical quartzite cube. Calculated H_2O content for biotite

849 and muscovite from EMP analyses are 3.97 wt% (n = 55) and 4.44 wt% (n = 13), respectively

850   (Table 1 and Table S1). Totalling the proportion of biotite and muscovite in each sample gives

851   the final H_2O content of 0.3 wt% and 0.7 wt% for samples L-Qtz and H-Qtz, respectively

852   (Table 3).

853   The reactive melt flux is calculated by using the volume of melt necessary to

854   drive the increase in H_2O content. Four melts of variable H_2O content are considered (with 3,

855   6, 9 and 12 wt% H_2O), which represent a typical range of water content in granitic melts (Holtz

856   *et al*., 2001). Two simplifying theoretical assumptions are made: (1) the totality of the H_2O

857   present in the melt is partitioned into the new hydrous phases; and (2) melt flux is homogeneous

858   throughout the rock cube. In nature, this perfect behaviour of H_2O partitioning and

859   homogeneous flux would not necessarily be expected. The model is also restricted by

860   disregarding some anhydrous phases (plagioclase, sillimanite, K-feldspar) into the H_2O

861   partition calculation, leading to an underestimation of the amount of melt flux for sample H-

862   Qtz. Further support for an underestimation of melt flux is based on the fact that there is the

863   possibility of armoured and unreactive melt flux following formation of glimmerite layers,

864   analogous to hornblendite channels observed in arc environments (Daczko *et al*., 2016; Stuart

865   *et al*., 2018*a*) or dunite channels observed in mantle rocks (Kelemen *et al*., 1995). Calculated

866   minimum melt flux volumes range between 0.03 to 0.23 m^3 of melt per m^3 of rock (Table 3).

867   Values of melt flux volume vary according to the interplay of the initial H_2O present in the

868   fluxing melt and the proportion of reactive hydrous phases observed in the samples. Sample L-

869   Qtz shows the lowest values of melt flux (0.03–0.12 m^3), with H-Qtz approximately doubling

870   the L-Qtz melt flux values (0.06–0.23 m^3). Unfortunately, upper crustal rocks that may contain

871   the crystallised products of the unreacted flowing melt are not preserved in the GDSZ (Fig. 1).

872   Attempting a volume estimation of the igneous bodies formed from the channelised melt

873   flowing through the armoured shear zone is thus not possible, making the obtained values of

874   melt flux the absolute minimum necessary for the observed glimmerite hydration reactions.

5. 5. Model for glimmerite schist belt formation and its geotectonic significance

876

877   Our proposed model for glimmerite schist belt formation is summarised in Fig.

878 15. It evolves from an initial stage of melt influx advancing upward within the GDSZ along

879 pre-existing anisotropies within the original rock packages (Fig. 15a), to a final stage of high-

880 strain zone metasomatism resulting in glimmerite formation as a consequence of hydrous

881 peraluminous melt interaction with adjacent anhydrous rocks (Fig. 15b, c). During the initial

882 stage, channelised melt flow within the high-strain zone causes strength contrasts between rock

883 packages that drive the formation of orthogonal extensional fractures, further increasing the

884 surface area for melt-rock interaction and promoting the formation of glimmerite (Fig. 15c).

885 The growth of extensional fractures towards the centre of the high-strain zone allows for the

886 creation of additional interconnected melt pathways, enhancing the glimmerite formation

887 process in the high-strain zone and increasing the size of granite lenses and sheets towards the

888 centre of the shear zone (Fig. 15b).

889   The interpreted evolution of the GDSZ is distinct to the seminal examples of

890 glimmerite formation driven by metasomatism of mantle rocks (Becker *et al*., 1999; Grégoire

891 *et al*., 2002; Rajesh *et al*., 2004). The "classic" glimmerite rocks generally show a spatial

892 relationship in outcrop between mantle rocks and either a K-rich metasome with or without a

893 third component involving felsic igneous intrusions (Fyfe & McBirney, 1975; Wyllie &

894 Sekine, 1982; Rakotondrazafy *et al*., 1997; Fuertes-Fuente *et al*., 2000). The K-rich mantle

895 metasomes are characterised by an assemblage composed usually of biotite + amphibole ±

896 pyroxene (Menzies & Murthy, 1980; Smart *et al*., 2019). These characteristics of "classic"

897 glimmerites are not observed in sample GL. However, similarities between our proposed mode

898 of glimmerite formation and the classic examples include melt-rock interaction involving a

899 felsic melt, and modification of a protolithic rock (felsic versus ultramafic) by melt-mediated

900 coupled dissolution-precipitation, forming abundant biotite (Figs. 4–7). Genetically and

901 mineralogically, sample GL shows a higher degree of similarity to the formation of "back

902 reactive melanosome" in anatectic migmatites described by Kriegsman (2001). In that model,

903 locally-derived in-situ melt is argued to interact with the mesosome. This diverges from our

904 proposed mode of protolith (quartzite) modification and glimmerite formation, where we

905 invoke an externally-sourced melt to drive dissolution of protolith minerals and precipitation

906 of glimmerite minerals, mainly quartz and biotite, respectively.

907 We now put our model of melt migration and deformation within GDSZ into

908 the context of crustal evolution. A key feature of the Earth's crust is its layered character mainly

909 based on chemical composition, with the upper crust having a higher silicic bulk composition

910 and concentration of granitic material compared with the more ferromagnesian bulk

911 composition of the lower crust (Vigneresse, 1995; Brown & Rushmer, 2006). Deeply seated

912 high-strain zones in the lower to middle crust are considered part of the crustal system for melt

913 migration (Brown & Solar, 1998; Daczko *et al*., 2016). Zones of dominantly flattening strain

914 function as major pathways for the ascent of melt originating from anatectic processes in the

915 lower to middle crust and are partially responsible for batch mass transfer of melt to the upper

916 crust (Hutton, 1988; Brown & Solar, 1999; Sawyer *et al*., 2011; Daczko *et al*., 2016; Etheridge

917 *et al*., 2020), as exemplified by the proposed transfer of melt throughout high-strain zones in

918 our model of glimmerite schist belt formation (Fig. 15).

919 In the dehydrated middle to lower crust, the influx of locally- or externally-

920 derived fluids forms pronounced local chemical gradients between fluid and rock, which when

921 paired with high differential stresses, enhances metamorphic reactions (Putnis & Austrheim,

922 2010; Wheeler, 2014; Stuart *et al*., 2016, 2017). If the fluid has high enough water fugacity,

923 exothermic hydration (or retrograde) metamorphic reactions of anhydrous minerals will occur.

924 The intracontinental Alice Springs Orogeny, involving the GDSZ in particular and possibly

925 other equivalent shear zones in central Australia, are good examples of melt flow pathways in

926 high-strain zones that lead to melt-rock interaction and hydration of protolith granulites. This

927 process is proposed as the source of peraluminous melt that initially fluxed through pre-existing

928 rock anisotropies leading to the formation of glimmerite schist belts (Fig. 15).

929 The physical presence of melt and metamorphic reactions that produce

930 rheologically weak minerals in high-strain zones (i.e., reaction softening) favour a decrease in

931 the strength of the affected rocks. This increases the accommodation of strain transmitted from

932 the surrounding and more competent host rocks, and facilitates strain localisation to enhance

933 the displacement across the high-strain zone (Hollister & Crawford, 1986; Brown & Solar,

934 1998; Molnar & Dayem, 2010; Cunningham, 2013; Piazolo et al., 2020). Reaction softening

935 such as that presented in this study, characterised by the formation of layers with high

936 proportions of phyllosilicates within a competent anhydrous granulite terrane (Figs. 2 and 15),

937 can be considered a textbook example of decreased rock strength, enabling deformation

938 accommodation and consequent strain localisation within structures such as the GDSZ.

939 However, we do not observe classic solid-state deformation microstructures in any of our

940 samples, suggesting that either (1) melt was involved in any/all reactivation events, forming

941 microstructures indicative of the former presence of melt each time; or (2) differential stresses

942 at the continental interior were never high enough to reactivate reaction softened high-strain

943 zones.

944 Weakening of deforming rocks is proposed by various authors as an important

945 mechanism to enhance orogenesis, especially true for far-field contractional intracontinental

946 orogenies (e.g., Alice Springs and Altai orogenies; Hand & Sandiford, 1999; Raimondo et al.,

947 2014; Cunningham, 2005; Silva et al., 2018; Piazolo et al., 2020), where differential stresses

948 are comparatively lower compared to those observed in areas proximal to collisional plate

949 boundaries (Coblentz et al., 1998; Heidbach et al., 2010). Moreover, weakened rocks are

950 described by various authors to be significant in the multiple reactivation events documented

951 for the ASO (Raimondo *et al*., 2014; Silva *et al*., 2018; Piazolo *et al*., 2020), because they

952 facilitate the accommodation of lithospheric horizontal stresses during increased early

953 Palaeozoic dextral rotation of the northern Australian block (Roberts & Houseman, 2001; Silva

954 *et al*., 2018). The formation and presence of meter-scale glimmerites, such as we document

955 here in the GDSZ and characterise hydrated high-strain zones elsewhere in the SMC (Fig. 1),

956 indicates that this specific rock type, characterised by its intrinsic low competence and

957 enhanced by the physical presence of melt, was an ideal candidate to intensify intracontinental

958 orogenesis in central Australia.

959

6. CONCLUSIONS

961   Glimmerite schists in the study area of the Gough Dam shear zone (GDSZ)

962 formed during the Palaeozoic Alice Springs Orogeny (ASO) by the migration of a hydrous

963 peraluminous melt and its interaction with granulite protoliths of quartzo-feldspathic

964 composition. In addition to melt migration through high-strain rocks, fractures parallel and

965 perpendicular to the shear zone foliation created melt pathways that enhanced melt infiltration

966 into the protolith. High-temperature melt-rock interaction at conditions close to the melt

967 liquidus promoted dissolution of the protolith and precipitation of abundant biotite. Localised

968 temperature-dependent quartz grain growth recovery is observed in highly organised modified

969 quartzite mylonite, whereas a random quartz grain orientation is observed in glimmerite. The

970 mode of hydrous minerals in the various rock types is inferred to relate to variable time-

971 integrated volumes of reactive melt flux. Zones of higher time-integrated melt flux through the

972 shear zone show increased equilibration and homogenisation of the neoformed hydrous

973 minerals. For example, a calculated melt flux between 0.06–0.23 m^3 per m^3 of rock allowed an

974 increase in hydrous minerals within the host rock of 8–17 vol%, concomitantly increasing the

concentration of trace elements (Ti, Nb, Zr, Hf, Sc, Ba and Li) and REE (mostly Eu). Inherited

monazite with ages aligned with the *c*. 1735–1690 Ma Strangways Event is variably reset to

early-ASO ages (*c*. 450 Ma) by coupled dissolution-precipitation processes, indicating that the

mobility of Pb was low and leading to high dispersion in the age population. The formation of

glimmerite schist in anhydrous terranes may be an important mechanism for stress localisation

and by consequence strain localisation, facilitating orogenesis in intracontinental regions as we

propose for the ASO.

SUPPLEMENTARY MATERIAL

Supplementary materials are available at Gondwana Research online.

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1500

FIGURE CAPTIONS

Fig. 1. Generalised geological map of the southeastern Arunta Region, emphasising the distribution of regional schistose high strain zones (yellow structures numbered on map) that cross-cut the dominantly granulite and amphibolite facies rocks comprising this basement terrane. Also represented are the regional anastomosing high-strain zones and thrust faults. Simplified structural cross-section A–A' (modified after Raimondo *et al*., 2014) shows the principal crustal discontinuities and high-strain zones in the Arunta Region associated with the Alice Springs Orogeny (Collins & Teyssier, 1989; Ballèvre *et al*., 2000; Maidment, 2005; Raimondo *et al*., 2011; Scrimgeour, 2013). Regional high strain zones: #1 Gough Dam shear zone (GDSZ); #2 West Bore shear zone; #3 Wallaby Knob shear zone; #4 Yambah shear zone; #5 Southern Cross shear zone; #6 Harry Creek shear zone (HCSZ); #7 Erontonga/Two Mile Bore shear zone; #8 Illogwa shear zone; #9 Delny shear zone; #10 Yalbadjandi shear zone. AB, Amadeus Basin; GB, Georgina Basin; HRG, Harts Range Group; SMC, Strangways Metamorphic Complex; RAR, Reynolds-Anmatjira Ranges.

Fig. 2. Photographs of outcrops and collected samples representative of strained rocks in the Gough Dam shear zone, from (a) outer area of the shear zone; (b, c) adjacent to the central area of the shear zone; and (d–i) central area of the shear zone. (d) Outcrop looking west of meter-scale glimmerite schist with a well-developed foliation (S1, white line) that contains layers and lenses of felsic rock types (i.e., granitic gneiss and quartzite mylonite). Elongated quartzite mylonite (along S1) and granitic gneiss bodies are surrounded by glimmerite schist. Field sense of shear is north-up reverse (along a steeply-plunging L1); (e) Granitic gneiss elongated along S1 surrounded by glimmerite schist and granite lenses. Granite lenses in glimmerite schist are elongate parallel to S1 foliation. Glimmerite embayments (white arrows) are observed in the granitic gneiss body; (f) Magnification of granitic gneiss lenses 'invaded' by glimmerite along fractures at a high angle to the foliation; (g) Quartzite mylonite in glimmerite schist unit surrounded and cross-cut at a high angle to the foliation by glimmerite seams. Some K-feldspar grains are present in the glimmerite schist; (h) Magnification of the sampled quartzite mylonite

1529 hosting sample GD1609A (L-Qtz) within the glimmerite schist unit surrounded by glimmerite

1530 seams. Less glimmerite 'invaded' along high angle fractures to the foliation is observed; (i)

1531 Quartzite mylonite samples GD1609B–C (H-Qtz, M-Qtz and GL samples, respectively).

1532 Glimmerite seams warp around and 'invade' the quartzite mylonite, involving most of the H-

1533 Qtz sample.

1534

1535 **Fig. 3.** Back-scattered electron (BSE) images of thin sections from (a) to (d) granite mylonite

1536 outcrop in Fig 2a; (e) to (h) glimmerite schist outcrop in Fig 2b; and (i) to (l) glimmerite schist

1537 outcrop in Fig 2c, containing the following microstructures highlighted by coloured triangles:

1538 yellow – interstitial grains and/or finger-like protrusions; blue – grains terminating with

1539 apparent low dihedral angles; purple – myrmekitic intergrowth of quartz and plagioclase; green

1540 – triple junction of K-feldspar grains presenting straight grain boundaries and ~120° dihedral

1541 angles. Mineral abbreviations follow Whitney & Evans (2010).

1542

1543 **Fig. 4.** Petrography and X-ray elemental mapping of quartzite samples L-Qtz, H-Qtz and M-

1544 Qtz/GL. (a–c) Crossed-polarised light thin section photomicrographs in top image and thin

1545 section μ-XRF maps in bottom image. Magenta triangles highlight high grain mobility features

1546 in quartz grains; blue triangles highlight bands rich in sillimanite that are inferred to be former

1547 melt migration pathways and include pockets inferred to have pseudomorphed melt; (d) Plane-

1548 polarised light photomicrograph of sillimanite-rich bands in sample H-Qtz (blue triangles); (e)

1549 Synchrotron X-ray elemental concentration maps of Sr and Cu for area shown in image (d)

1550 highlighting the emplacement of a cryptic Si-Al alteration product (red arrows) relative to

1551 biotite (green arrows), plagioclase (white arrows) and quartz grains (light blue arrows). Mineral

1552 abbreviations follow Whitney & Evans (2010).

1553

1554 **Fig. 5.** BSE and cathodoluminescence images of inferred melt microstructures present in (a–c)

1555 H-Qtz and (d–j) M-Qtz/GL samples. (a) Inferred melt pocket displayed in Fig. 4b, d and e, with

1556 peraluminous granite mineral assemblage. Biotite has interlocking boundaries with

1557 surrounding quartz and finger-like protrusions (yellow arrows); (b, c) Biotite, muscovite and

1558 sillimanite microstructures; (b) Biotite and muscovite grains presenting elongated interstitial

1559 grains/finger-like protrusions (yellow arrows) and low apparent dihedral angles (blue arrows)

1560 between other grains; (c) Biotite located in sample H-Qtz showing interlocking boundary with

1561 quartz (yellow arrow); (d–f) Biotite, muscovite and sillimanite microstructures; (d) Biotite

1562 grain presenting elongated interstitial grains/finger-like protrusions (outside field of view of

thin section shown in Fig. 4c; yellow arrows); (e) Sillimanite grains in quartzite showing perpendicular fractures filled with a cryptic alteration product of Si-Al composition (labelled Alt.); (f) Biotite and quartz grains wrapping around a large, deformed muscovite grain (outside field of view of thin section shown in Fig. 4c); (g–j) Cathodoluminescence images for selected regions in sample M-Qtz, showing quartz and biotite grain contacts. Dashed lines highlight darker luminescent regions mostly at quartz grain boundaries when in contact with biotite grains. Yellow arrows highlight narrow dark bands inferred to be healed fractures within or at quartz grain boundaries. Mineral abbreviations follow Whitney & Evans (2010).

Fig. 6. Euler angles and inverse pole figure (IPF) maps describing the interconnectivity of neoformed quartz grains and quartzite mylonite dissolution in M-Qtz/GL samples. (a) Crossed-polarised light thin section photomicrograph of sample M-Qtz. Blue triangles indicate sillimanite-rich bands and red triangles indicate irregular quartz grain shapes; (b) Euler map of quartz grains for the area shown in panel (a), shown in more detail in (c) and (d) where similarly oriented intragranular quartz grains are identified in respective groups adjacent to the sillimanite-rich domain; (e) Quartz IPF map for M-Qtz and GL contact zone (see Fig. 4c for location), with colour coding relative to the Z-axis and other phases shown as band contrast. Euler angle rotation is coloured as ψ - red, θ - green and φ - blue.

Fig. 7. Description of quartz grains preferential orientation in H-Qtz, M-Qtz and GL samples; inverse pole figure (IPF) with colour coding relative to the Z-axis for (a) M-Qtz and GL samples, and (d) H-Qtz sample. Regions displayed in each sample represent individual quartz orientation calculation subsets; (b, c) and (e, f) EBSD-derived quartz crystallographic orientation pole figures and neighbouring quartz pair misorientation angle distributions (blue bars) for the respective samples and displayed regions. Note that scale bars for pole figure contouring vary between pole figures.

Fig. 8. Graph depicting the mean area weighted quartz grain size for the different regions shown in Fig. 7.

Fig. 9. Harker diagrams for whole rock major oxides (Al_2O_3, TiO_2, FeO, MgO and K_2O vs SiO_2) for selected rock types present in the Strangways Metamorphic Complex, sample GL (this study) and the average biotite composition for peraluminous granites (Neiva *et al.*, 2002).

1596 See Table S2 for all data references. Standard deviation for average biotite composition

1597 displayed only if larger than the symbol.

1598

Fig. 10. Graphical representation of biotite geochemical evolution between all samples from

1600 main study area. (a) Ti atoms per formula unit (a.p.f.u.) vs X_{Mg} diagram for biotite from all

1601 samples; (b) Al vs Mg a.p.f.u. classification diagram for biotite in igneous rocks (after Nachit

1602 *et al*., 1985 and Stussi & Cuney, 1996); (c) Ternary diagram TiO_2 x 10-(FeO + Mn)-MgO for

1603 discrimination between primary magmatic, reequilibrated and neoformed biotites (Nachit *et*

1604 *al*., 2005); (d) *P-T* diagram for Gough Dam shear zone samples, using Ti in biotite for

1605 temperature (Henry *et al*., 2005) and Si in phengite for barometry (Massonne & Schreyer,

1606 1987).

1607

Fig. 11. Chondrite-normalised trace elements and REE patterns for biotite in (a–c) Gough Dam

1609 shear zone samples and (d) literature compilation (migmatite leucosome, Bea *et al*., 1994b;

1610 pegmatite, Hulsbosch *et al*., 2014; mica schist, Laul & Lepel, 1987; peraluminous granite, Bea

1611 *et al*., 1994a; metapelite enclave, Acosta-Vigil *et al*., 2010; experimental Nb-Ta, Stepanov &

1612 Hermann, 2013; rhyodacite, Nash & Crecraft, 1985). Selected rocks are detailed in Table S4.

1613 Colours for samples L-Qtz, H-Qtz and M-Qtz/GL correspond to those used for symbols in other

1614 diagrams.

1615

Fig. 12. Bivariate plots for Ti vs Ba, Ti vs Sc, Ti vs Nb/Ta and Ti vs Li from Gough Dam

1617 shear zone samples and additional data from El Hoyazo enclaves (Table S4; Acosta-Vigil *et*

1618 *al*., 2010). Black error bars represent the measured uncertainty (2σ) of an individual analysis;

1619 red error bars represent the standard deviation (2σ) of multiple analyses.

1620

Fig. 13. Textural relationships for monazite grains in (a–b) sample GL with higher biotite

1622 modal composition, and (c–d) sample M-Qtz with lower biotite modal composition.

1623

Fig. 14. Monazite morphology, texture and U-Pb geochronology from sample M-Qtz/GL. (a)

1625 Morphology and textures of representative monazite grains from sample M-Qtz in back-

1626 scattered electron (BSE) imaging. $^{206}Pb/^{238}U$ ages with associated errors (2 SE) are displayed,

1627 and full analytical results are given in Table S5. White arrows indicate inward-penetrating

1628 boundary interfaces between dark and light BSE response and yellow arrows indicate thorite

1629 inclusions; (b) Tera-Wasserburg concordia plot showing U-Pb geochronological data from

1630 samples M-Qtz and GL and kernel density estimation for data with < 2% discordance. Linear

1631 regression age estimates are calculated using only sample M-Qtz analyses.

1632

1633 **Fig. 15.** Schematic block models showing the metasomatic (*s.l.*) evolution of the Gough Dam

1634 shear zone and formation of meter-scale glimmerite zones via melt-rock interaction. (a) Shear

1635 zone melt inflow at the inception stage (*c*. 450 Ma): south-directed thrust shear of granulite

1636 terranes in the Arunta Region. Melt sourced from lower crustal levels infiltrates and flows

1637 upward along lithological boundaries and pre-existing foliations. Melt flow intensity is higher

1638 towards the centre of the shear zone; (b) Shear zone metasomatism (*s.l.*) at the final stage (*c*.

1639 370 Ma): shear zone glimmerite formation with increasing biotite content towards the centre

1640 of the deformation zone. Granitic lenses close to the shear zone boundaries increase in size and

1641 morph into sheets towards the centre of the shear zone; (c) Inset of main glimmerite domain in

1642 the Gough Dam shear zone, summarising the melt-rock interaction structures from this study.

1643 Orthogonal fragmentation of quartzite mylonite lenses due to rheological contrast enhances

1644 melt flow and subsequent melt-rock interaction to form glimmerite. HRG, Harts Range Group;

1645 SMC, Strangways Metamorphic Complex.

1646

1647 **Table 1:** Electron microprobe data of key minerals from the reacted quartzites and

1648 glimmerite seams in the Gough Dam shear zone.

1649

1650 **Table 2:** Representative LA-ICP-MS trace element and REE compositional data for biotite

1651 grains from the Gough Dam shear zone, and comparison to published data from multiple

1652 locations and formation settings exogenous to the study area.

1653

1654 **Table 3:** Calculation of the melt volume forming hydrous minerals in a high-strain zone for

1655 samples L-Qtz and H-Qtz.

1656

1657

1658 **SUPPLEMENTARY DATA**

1659

1660 **Fig. S1.** Back-scattered electron (BSE) image showing the spatial distribution of monazite ages

1661 in sample M-Qtz/GL.

1662

1663   **Table S1:** Electron microprobe data of additional minerals from reacted quartzites and
1664 glimmerite seams in the Gough Dam shear zone.

1665

1666   **Table S2:** Whole rock major oxide concentrations for sample GL in comparison to average
1667 peraluminous granite from the literature and major rock types of the Strangways Metamorphic
1668 Complex.

1669

1670   **Table S3:** LA-ICP-MS biotite trace element data.

1671

1672   **Table S4:** Description of biotite samples from the literature for comparison with those sampled
1673 from the Gough Dam shear zone.

1674   **Table S5:** LA-ICP-MS monazite U-Pb data.

Figure 1

Legend:
- **Schistose high strain zones** (yellow)
- **Strangways Metamorphic Complex (SMC)** (blue)
- **Amphibolite facies rocks** (orange)
- **Granulite facies rocks** (red/blue)
- **Anastomosing shear zones** (black line)
- **Proterozoic–Palaeozoic intracontinental basins** (dotted)
- **Archean–Proterozoic terranes** (grey)
- **Arunta Region (inset)** (red)

Figure 2

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Figure 2

Figure 4   Click here to access/download;Figure;Figure 4.tif ⬇

Figure 4

Figure 6 Click here to access/download;Figure;Figure 6.tif

(a) M-Qtz sample (b) 500 μm (c) (d) Groups of similarly oriented grains (=<10°) — Group 1 — Group 2 — Group 3

Grain boundary (>10°) — Dauphiné twin (60°) — Quartz dissolution predominantly replaced by biotite

(e) IPF-Z [1010] [0001] [0110] 1 mm IPF map of M-Qtz/GL boundary

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

Figure 7

Figure 8

Area weighted mean grain size (mm²)

Region 2 (M-Qtz) and Region 1 (H-Qtz)

Region 3 (M-Qtz) and Region 2 (H-Qtz)

Sample GL

■ M-Qtz quartz grains
▲ H-Qtz quartz grains
⬢ GL quartz grains

Figure 9

Figure 10

Figure 11   Click here to access/download;Figure;Figure 11.pdf ⬇

Figure 12

Figure 12

Figure 13

Click here to access/download;Figure;Figure 13.tif ⬇

Figure 14

Click here to access/download;Figure;Figure 14.tif

(a) Monazites in sample M-Qtz

(b)

age = 482.08 ± 2.45 | 4.80 | 16.67 Ma (n=36)
$(^{207}Pb/^{206}Pb)_o$ = 0.11881 ± 0.00030 | 0.00058 | 0.00202
MSWD = 11, p(χ^2) = 0

Figure 15

(a)

N →

SMC

HRG

Granulite   Granulite   Granulite   Granulite

Quartzite
mylonite

Not to scale

c. 450Ma

**Peraluminous melt flow
intensity**

(b)

SMC

HRG

Granulite   Granulite

Remnant of quartzite
mylonite

Not to scale

**Glimmerite
schist**

**Granitic
lenses and
sheets**

Glimmerite

c. 370 Ma

Bt content

(c)

Neoformed
quartz grains

Granitic lenses

K-feldspar grains

Biotite
selvedge

Original quartzite
mylonite boundary

Remnant quartzite
mylonite

Glimmerite

Table 1

Table 1: Electron microprobe data of key minerals from the reacted quartzites and glimmerite seams in the Gough Dam shear zone.

Sample	L-Qtz		H-Qtz			M-Qtz			GL		
Mineral	*Bt*	*Mag*	*Bt*	*Ms*	*Pl*	*Bt*	*Ms*	*Sil*	*Bt*	*Ms*	*Sil*
SiO_2 wt%	36.74	–	35.70	44.49	62.89	35.89	44.55	36.03	35.64	44.56	36.06
TiO_2	0.83	0.03	2.35	0.86	–	1.86	0.60	–	1.97	0.60	–
Al2O3	17.49	0.43	18.23	32.82	23.00	18.28	33.32	63.23	18.36	33.07	63.41
V_2O_3	–	0.06	0.05	–	–	0.06	0.03	–	0.04	0.03	–
Cr_2O_3	–	–	–	–	–	–	–	–	0.04	–	–
FeO_{total}	13.92	92.19	16.22	3.24	0.05	15.70	3.36	0.84	15.20	3.21	0.83
NiO	–	–	–	–	–	–	–	–	0.06	–	–
MnO	0.05	0.07	0.07	–	–	0.06	–	–	0.05	–	–
MgO	15.32	0.06	12.15	0.97	–	12.53	0.96	–	12.66	0.90	–
CaO	0.03	–	–	–	3.67	–	0.02	–	–	–	–
Na_2O	0.16	–	0.19	0.51	9.36	0.26	0.63	–	0.23	0.56	–
K_2O	9.72	–	9.86	10.53	0.19	9.82	10.00	0.01	9.79	10.63	–
P_2O_5	–	–	–	–	0.07	0.03	–	–	–	–	–
SO_3	–	–	–	–	–	–	–	–	–	–	–
Cl	0.26	–	0.18	–	0.02	0.18	0.06	–	0.19	0.02	–
F	0.92	–	0.67	0.06	–	0.66	0.12	–	0.71	0.08	–
O	-0.45	0.04	-0.32	-0.03	-0.02	-0.32	-0.07	0.00	-0.34	-0.04	-0.01
H_2O	3.99	0.00	3.96	4.45	0.00	3.97	4.45	0.00	3.98	4.45	0.00
Total	*98.99*	*92.89*	*99.30*	*97.89*	*99.25*	*99.00*	*98.04*	*100.11*	*98.57*	*98.07*	*100.30*

Mineral abbreviations follow Whitney & Evans (2010). (–): Below detection limit. The complete dataset is available in Supplementary Material, Table S1.

Table 2: Representative LA-ICP-MS trace element and REE compositional data for biotite grains from the Gough Dam shear zone, and comparison to published data from multiple locations and formation settings exogenous to the study area.

| Sample | Gough Dam shear zone | | | Compilation of published biotite compositions | | | | | | |
	L-Qtz (GD1609A)	H-Qtz (GD1609B)	M-Qtz & GL (GD1609C)	Two-mica granite [1]	Diatexite migmatite [2]	Metapelite xenolith [3]	Pegmatite [4]	Mica schist [5]	Rhyolite/ Rhyodacite [6]	Experiment [7]
Zr	0.21	0.35	0.28	- -	- -	1.41	- -	- -	40.05	- -
Hf	0.66	0.96	0.66	- -	- -	2.52	- -	- -	27.77	- -
Nb	85.13	221.75	188.88	- -	197.08	341.67	- -	- -	283.33	208.33
Ta	134.71	216.18	183.09	- -	462.50	323.53	- -	- -	119.85	1397.06
La	–	0.016	0.009	0.717	0.169	0.253	7.890	107.173	342.194	- -
Ce	0.013	0.002	0.002	0.277	0.196	0.016	5.726	81.240	274.062	- -
Pr	0.018	0.008	–	0.970	0.108	0.216	5.065	- -	- -	- -
Nd	0.014	0.009	0.009	0.700	0.088	0.131	4.136	52.516	133.479	- -
Sm	–	–	0.107	0.405	0.068	0.068	8.851	24.662	53.243	- -
Eu	0.357	0.702	0.444	0.533	0.355	2.664	0.710	11.368	7.105	- -
Gd	–	–	0.068	0.452	0.050	0.653	13.920	16.332	- -	- -
Dy	–	–	–	0.122	0.033	0.081	19.553	10.163	17.195	- -
Er	0.018	–	–	0.125	0.044	0.063	11.000	- -	- -	- -
Yb	–	–	–	0.124	0.050	0.124	10.932	6.832	15.466	- -
Lu	0.043	0.067	0.031	- -	0.081	- -	9.756	69.106	15.041	- -

REE and trace elements values were normalised using McDonough & Sun (1995) chondritic values.

(1) Bea *et al*., 1994a; (2) Bea *et al*., 1994b; (3) Acosta-Vigil *et al*., 2010; (4) Hulsbosch *et al*., 2014; (5) Laul & Lepel, 1987; (6) Nash & Crecraft, 1985; (7) Stepanov & Herman, 2013. The complete dataset is available in Supplementary Material, Table S3. (–): Below detection limit; (- -): Not available.

Table 3

Table 3: Calculation of the melt volume forming hydrous minerals in a high-strain zone f

Sample	Hydrous minerals wt%	
	L-Qtz	*H-Qtz*
Biotite (3.97 H_2O wt%)	8.0	15.0
Muscovite (4.44 wt%)	0.0	2.0
Total sample H_2O (wt%)	*0.32*	*0.68*

Melt H_2O (wt%)	**Volume of melt per m^3 rock**	
3.0	0.11	0.23
6.0	0.05	0.11
9.0	0.04	0.08
12.0	0.03	0.06

for samples L-Qtz and H-Qtz.

Figure S1

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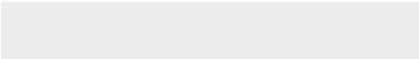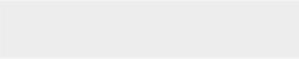
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