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1   **Resubduction of lawsonite eclogite within a serpentinite-filled subduction channel**

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9

10   **Abstract**

11   Translating burial and exhumation histories from the petrological and geochronological

12   record of high-pressure assemblages in subduction channels is key to understanding subduction

13   channel processes. Convective return flow, either serpentinite or sediment hosted, has been suggested

14   as a potential mechanism to retrieve rocks from significant depths and exhume them. Numerical

15   modelling predicts that during convective flow, subducted material can be cycled within a

16   serpentinite-filled subduction channel. Geochronological and petrological evidence for such cycling

17   during subduction is preserved in lawsonite eclogite from serpentinite mélange in the Southern New

18   England Orogen, eastern Australia. Ar–Ar, Rb–Sr phengite and U–Pb titanite geochronology,

19   supported by phase equilibrium forward modelling and mineral zoning, suggest Cambro-Ordovician

20   eclogite underwent two stages of burial separated by a stage of partial exhumation. The initial

21   subduction of the eclogite at ca. 490 Ma formed porphyroblastic prograde-zoned garnet and lawsonite

22   at approximate $P–T$ conditions of at least 2.9 GPa and 600 °C. Partial exhumation to at least 2.0 GPa

23   and 500 °C is recorded by garnet dissolution. Reburial of the eclogite resulted in growth of new Mg-

24   rich garnet rims, growth of new prograde-zoned phengite and recrystallization of titanite at $P–T$

25   conditions of approximately 2.7 GPa and 590 °C. U–Pb titanite, and Ar–Ar and Rb–Sr phengite ages

26   constrain the timing of reburial to ca. 450 Ma. This was followed by a second exhumation event at

27   approximately 1.9 GPa and 520 °C. These conditions fall along a cold approximate geotherm of 230

28   °C/GPa. The inferred changes in pressure suggest the lawsonite eclogite underwent depth cycling

29   within the subduction channel. Geochronological data indicates that partial exhumation and reburial

30 occurred over ca. 50 M.y., providing some estimation on the timescales of material convective cycling

31 in the subduction channel.

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1. INTRODUCTION

37 Metamorphism along cool geothermal gradients in subduction zones produces eclogite-facies

38 mineral assemblages. These assemblages often preserve an abundance of prograde, peak and

39 retrograde petrological relationships. For example, elemental zoning in garnet and micas can be

40 preserved due to slow rates of diffusion, and temperature-sensitive minerals such as lawsonite can

41 also be preserved (Tsujimori et al., 2006; Tsujimori & Ernst, 2014). Given this potential to preserve

42 extensive petrological relationship histories, methods such as phase equilibrium modelling coupled

43 with new and emerging geochronological techniques open avenues to explore the pressure–

44 temperature–time (P–T–t) histories of these high-pressure low-temperature assemblages, and make

45 inferences about the behaviour of the subduction systems that host them.

46

47 The application of sophisticated petrological techniques to eclogite and blueschist assemblages

48 has highlighted the potential for complex P–T–t histories. This is particularly apparent in serpentinite

49 mélanges. Mixing, accumulation, long-lived residence and intrachannel circulation of high-pressure

50 lithologies have started to emerge as common phenomena within subduction channels (e.g. Krebs et

51 al., 2008; Lázaro et al., 2009). Numerical modelling has sought to explain the dynamics of material

52 within subduction channels, and has demonstrated that complex P–T–t histories are possible, arising

53 from cycling and corner flow within the channel (Gerya et al, 2002, Stöckhert & Gerya, 2005; Roda et

54 al., 2019). Examples of high-pressure rocks that have seen two or more cycles of burial have been

55 recognised through geochronological and mineralogical evidence, for example, dating zones within

56 metamorphic zircon (e.g. Liati et al., 2016). However, in lieu of such useful petrochronological

57 methods, recognising multiple burial cycles can prove difficult.

58

59   Lawsonite eclogite is globally rare, despite the range of bulk compositions that support the

60 formation of lawsonite-garnet-omphacite (Wei & Clarke, 2011). Therefore, their occurrence probably

61 reflects unusually good preservation of deeply buried, cold subducted material. Consequently,

62 lawsonite eclogites potentially preserve comprehensive records of the physical conditions of

63 subduction channels. This study presents detailed petrological results from eclogite in the Southern

64 New England Orogen (SNEO), that supports partial exhumation and reburial of a single block of

65 lawsonite eclogite during on-going subduction. Phase diagrams constrain the approximate pressure–

66 temperature (P–T) conditions of partial exhumation and reburial within the subduction channel, which

67 is supported by U–Pb titanite, and Rb–Sr and Ar–Ar phengite geochronological data. The cycle of

68 exhumation and reburial records an approximate total pressure and temperature fluctuation of at least

69 1.6 GPa and 190 °C and is constrained to have occurred over ca. 50 M.y.. If the pressure fluctuations

70 recorded by high-pressure mineral assemblages reflect oscillations in burial depths, such records

71 support the notion of convective flow within the subduction channel.

72

73   **2. GEOLOGICAL SETTING**

74

75   **2.1 The Southern New England Orogen**

76   The Southern New England Orogen (SNEO) in eastern Australia is the youngest and most

77 outboard orogen of the Tasmanides (Fig. 1a; Kemp et al., 2009; Glen, 2013; Phillips et al., 2015). The

78 Tasmanides are interpreted to have formed as a result of long-lived subduction on the East

79 Gondwanan margin during the Cambrian–Triassic, which resulted in a series of eastwards-younging

80 orogens. The oldest and most westerly of these is the Delamerian Orogen, which formed at ca. 515–

81 490 Ma on the margin of cratonic Australia, and was post-dated by the Lachlan-Thompson Orogen at

82 ca. 484–340 Ma (Fig. 1a). The SNEO formed between the upper Devonian-Triassic, and consists of a

83 volcanic arc, forearc basin and accretionary complex (Fig. 1b; Jenkins et al., 2002; Jessop et al.,

84 2019). The Devonian-Carboniferous volcanic arc outcrops, but is generally obscured under the

85 younger Permian-Triassic Sydney-Gunnedah basin (Jenkins et al., 2002). The Tamworth Belt contains

86 a forearc basin, consisting of felsic volcanic rocks and volcaniclastics (Roberts & James, 2010). The

87 Tablelands Complex consists of deformed sediments and oceanic basalt that accumulated within an

88 accretionary complex. These are separated by the serpentinite-bearing Peel Manning Fault System

89 (Fig. 1b). The SNEO also contains numerous felsic magmatic rocks of the Permian-Triassic New

90 England Batholith, which intrude the Tablelands Complex and the Early Permian basins (Leitch,

91 1975; Shaw and Flood, 1981; Phillips et al., 2011). These are interpreted to have formed as the orogen

92 evolved from an accretionary system to a back-arc position (Jenkins et al., 2002; Phillips et al., 2011).

93

94 **2.2 Tectonic framework of the high-pressure rocks and the Tasmanides**

95 There has been significant debate surrounding the evolution of the Tasmanides, including the

96 tectonic significance of the high-pressure rocks. Notably, all the geochronology reported from

97 serpentinite-hosted high-pressure exotic blocks in the SNEO is significantly older than the enclosing

98 geology (Fig. 1b; e.g. Fukui et al., 1995, Sano et al., 2004; Phillips et al., 2015; Manton et al., 2017;

99 Tamblyn et al., 2019a). The most coherent views about the evolution of the Tasmanides involve

100 eastward retreat of a subduction system that formed close to the cratonic margin during the mid-late

101 Cambrian (e.g. Collins, 2002; Phillips & Offler, 2011; Moresi et al., 2014; Phillips et al., 2015). The

102 general consensus is the high-pressure rocks found in the SNEO were formed in a subduction channel

103 on the cratonic margin of Australia during the Delamerian Orogen in the Cambrian–Ordovician.

104 Subsequent cooling and transport of the high-pressure rocks occurred during eastward-directed slab

105 rollback, until the high-pressure rocks were exhumed in their current location in the Permian-Triassic

106 SNEO (Phillips & Offler, 2011; Phillips et al., 2015; Tamblyn et al., 2019a). During this rollback, the

107 Lachlan/Thompson Orogens developed in a backarc setting on the upper plate (e.g. Collins, 2002;

108 Kemp et al., 2009; Moresi et al., 2014). Eastward younging arcs track the rollback of the subduction

109 zone, accompanied by voluminous I and S-type magmatism formed from arc and back-arc rifting

110 processes (Collins, 2002; Collins & Richards, 2008). This eastward migration is also revealed in ε_{Nd}

111 isotopes of magmatic rocks, which demonstrate an increasingly juvenile signature as the subduction

112 zone migrated away from the cratonic margin (Kemp et al., 2009).

113

2.3 High-pressure metamorphism and mélange formation in the SNEO

High-pressure metamorphic rocks hosted in serpentinite occur throughout the SNEO at Attunga, Gleneden, Glenrock, Pigna Barney and Port Macquarie (Fig. 1b; Phillips et al., 2015). The Port Macquarie mélange contains by far the largest lithological variety of high pressure-rocks, which are described in detail by Och et al. (2003; 2007) and Tamblyn et al. (2019a). The mélange consists of exotic high-pressure blocks encased in chlorite-actinolite schist, which is in turn encased within serpentinite (Fig. 1c). These exotic blocks consist of variably metamorphosed omphacitite, low-grade basaltic rocks, sedimentary rocks, marble, blueschist facies rocks, retrogressed eclogite and lawsonite-bearing eclogite. In-situ U–Pb geochronology on micro-zircon from the lawsonite eclogite returned a range of ages from ca. 560–440 Ma (Tamblyn et al., 2019a). The zircons occur as inclusions in garnet and also occur in the surrounding matrix. Lu–Hf geochronology on lawsonite porphyroblasts gives an age of 506 ± 15 Ma, and a Lu–Hf garnet-whole rock isochron gives an age of 489.7 ± 5 Ma. These ages are interpreted to date prograde metamorphism, as garnet cores are enriched in Lu compared to rims, biasing the age to early garnet growth. Pressure-temperature conditions obtained from phase equilibrium modelling are approximately 2.7 GPa and 590 °C for the lawsonite eclogite, and 2.0 GPa and 550 °C for a garnet-omphacite-bearing blueschist block within the same mélange. Lu–Hf, Sm–Nd, K–Ar and Ar–Ar ages from adjacent blueschist blocks and fuchsite in serpentinite give ages between ca. 472 and 420 Ma (Fukui et al., 1995; Och et al., 2010; Tamblyn et al., 2019a). This geochronology, along with the presence of low-grade sedimentary blocks, confirms the accumulation of rocks with different *P–T* histories in the mélange.

3. PETROGRAPHY

Mineralogy of the lawsonite eclogite block (sample RB11) has been described in detail by Tamblyn et al. (2019a) and Och et al. (2003; 2007). The eclogite contains porphyroblastic garnet, lawsonite and omphacite, in a well-foliated matrix dominated by phengite and containing minor titanite (Fig. 2). However, it is clear from optical petrography and elemental mapping that there are two distinct mineral assemblages recorded in the sample, summarized in Supplementary Table 1, which are referred to as stage 1 and stage 2. Garnet is euhedral and forms crystals up to 7 mm in

142 diameter, which have red-coloured cores and paler rims, referred to as garnet 1 and garnet 2

143 respectively (Fig. 2a-d). The inclusion relationships in garnet are summarised in Supplementary

144 Figure 1. Garnet 1 has inclusion-rich cores which contain lawsonite, titanite, omphacite, phengite,

145 glaucophane, chlorite, epidote, stilpnomelane, quartz and zircon (Supp. Fig. 1). Very rarely, quartz

146 inclusions in the outermost part of garnet 1 are surrounded by prominent radial fractures, suggesting

147 the possibility that they were once coesite (Fig. 2e). Garnet 2 forms as ~ 400 µm wide rims on garnet

148 1 that contain chlorite, glaucophane, titanite, omphacite, lawsonite, quartz and zircon (Fig. 2d), and in

149 some instances overgrow matrix phengite. Enclosed inclusions of chlorite, phengite, omphacite,

150 quartz, titanite and pyrite occur at the boundary between garnet 1 and 2 (Supp. Fig. 1). Sigmoidal

151 inclusion trails of titanite are also commonly truncated at the boundary between garnet 1 and garnet 2,

152 with the trails forming at a high angle to the matrix foliation (Supp. Fig. 1). Titanite included in garnet

153 2 is coarser-grained than in garnet 1. Lawsonite forms euhedral crystals up to 3 mm (Fig. 2b,f). It has

154 darker sections containing abundant fine-grained inclusions and clearer sections with coarser-grained

155 inclusions. The inclusions are predominately titanite, epidote, glaucophane and phengite. In places the

156 lawsonite grain boundaries are mantled by small garnet grains, which have been overgrown by a

157 second generation of lawsonite. In other places the clearer lawsonite occurs as an overgrowth on

158 darker lawsonite crystals (Fig. 2f). The darker, older lawsonite is referred to as lawsonite 1 and the

159 clearer overgrowths are referred to as lawsonite 2. The foliated matrix is part of the stage 2 mineral

160 assemblage, and is defined by coarse phengite (up to 1 mm but commonly ~ 300 µm) and pervasive

161 but modally minor titanite (< 100 µm; Fig. 2b,d). The latter is associated with fine grained allanite

162 and zircon (< 20 µm). Titanite and phengite are coarser in the matrix relative to the inclusions in

163 garnet cores, and in places titanite forms aggregates of grains that rim phengite grains. Omphacite

164 forms monomineralllic patches up to 10 mm which are partially replaced by glaucophane and chlorite,

165 as well as relic fractured porphyroblasts up to ~ 500 µm in size (Fig. 2a,b). Omphacite is partially

166 replaced by chlorite and glaucophane. Garnet is partially replaced by chlorite along its grain

167 boundaries or along fractures which cross-cut the garnet (Fig. 2c; Supp. Fig. 1). Where retrograde

168 chlorite occurs with lawsonite, lawsonite is euhedral and unretrogressed, suggesting it was stable

169 during the retrogression of the eclogite. In the matrix, glaucophane also forms foliated euhedral

170 crystals (Fig. 2c). These crystals, as well as coarse titanite crystals, are overgrown by stage 2

171 lawsonite rims. These petrological observations denote two distinct stages of peak metamorphism in

172 the eclogite, separated by a period of retrogression. A schematic representation of the petrologic

173 observations made from optical petrography, phase mapping and elemental mapping can be found in

174 Figure 3.

175

4. METHODS

177

4.1 Electron Probe Micro Analyses and mapping

179 Spot analyses on minerals were obtained using a Cameca SX-5 microprobe at Adelaide

180 Microscopy, using an accelerating voltage of 15 kV and a beam current of 20 nA. Element maps used

181 an accelerating voltage of 15 kV and a 200 nA beam current, Ca, Fe, Mn and Mg were mapped using

182 Wavelength Dispersive Spectrometers (WDS), whereas Al, Si, Ti, K and Na were mapped using

183 Energy Dispersive Spectrometry (EDS). Maps were colour scaled in ImageJ.

184

4.2 Phase equilibrium forward modelling

186 Full phase equilibrium modelling methods are in Supplementary File 1. The rock composition

187 was calculated for the lawsonite eclogite by combining modal proportions of each mineral with its

188 measured chemical composition (Supp. Table 2; 3). Thin sections were mapped with a Quanta600

189 scanning electron microscope (SEM) with mineral liberation analysis (MLA) software to calculate

190 modal proportions of minerals.

191

192 The phase equilibrium models were calculated using THERMOCALC (TC340i) using the

193 internally-consistent thermodynamic dataset 'ds5' (filename tc-ds55.txt; November 2003 updated

194 version of the Holland & Powell, 1998 data set) and activity–composition (a–x) models (Supp. Table

195 4; Holland & Powell, 1988; Holland & Powell, 2003; White et al., 2007; Green et al., 2007; Holland

196 & Powell, 2011; Diener et al., 2012) in the chemical system NCKFMASHO. A model for the local

197 rock composition was calculated with water in excess, and a model for the local rock composition was

198 calculated with a set water content, derived from the abundance of hydrous minerals (Supp. Table 2;

199 3).

200

4.3 U–Pb titanite geochronology

202 Titanite grains were analysed in-situ for U–Pb isotopes using an ASI m50 LA–ICP with an

203 attached 7700 MS at Adelaide Microscopy. Grains were ablated using a spot size of 51 μm, a

204 frequency of 5 Hz, and a surface energy density of 3.5 Jcm^{-2}, with an acquisition time of 80 s for each

205 analysis including 30 s of background measurement and 50 s of ablation. The primary reference

206 titanite MKED-1 was used to correct for elemental fractionation, mass bias and instrument drift over

207 the courses of analyses (Spandler et al., 2016), and a titanite grain from Mt Painter with an age of

208 442.6 ± 1.8 Ma (Elburg et al., 2003) was used as the secondary standard. Corrections were done using

209 the software Iolite (Paton et al., 2011) and age calculations were done using IsoplotR (Vermeesch,

210 2018). The secondary standard from Mt Painter returned a weighted mean $^{206}Pb/^{238}U$ age of 440.3±

211 2.5 Ma (MSWD = 1.5, n = 22) throughout the course of the analyses.

212

4.4 Ar–Ar phengite geochronology

214 Phengite $^{40}Ar/^{39}Ar$ isotope analyses utilized conventional furnace step-heating techniques at

215 the University of Melbourne, after the analytical procedures of Phillips et al. (2007). Phengite grains

216 were selected from mineral separates. Due to the possibility of contamination by excess argon in high-

217 pressure rocks (Kelley, 2002), a two-increment step-heating laser approach was employed on single

218 grains of phengite. The laser analyses were conducted at The University of Melbourne, employing the

219 analytical procedures described by Phillips and Harris (2008). This approach was used to screen for

220 any significant intrasample inconsistency in argon content that could be an artefact of (i) multiple

221 generations of phengite growth or (ii) varied amounts of isotopic resetting or excess argon

222 contamination within single grains. Using the single-grain laser approach, spurious isotopic results

223 from a single-grain analysis can be removed from the mean age calculations, using Isoplot software

224 (Ludwig, 2003).

225

4.5 Rb–Sr phengite geochronology

226

Phengite was analysed in-situ for Rb and Sr isotopes and major and trace elements using an

227

ESI 213 NWR (TwoVol2) laser ablation system with an attached Agilent 8800QQQ ICP-MS/MS in

228

the Microgeochemistry Laboratory at the University of Gothenburg, Sweden, following the methods

229

of Zack & Hogmalm (2016) and Hogmalm et al. (2017). Full methods are in Supplementary File 1.

230

Laser parameters were a fluence of ~ 6 Jcm^{-2} and a frequency of 10 Hz. Phengite was analysed with a

231

spot size of 40 μm, both perpendicular and parallel to cleavage. Titanite was also analysed to obtain

232

initial $^{87}Sr/^{86}Sr$ ratios, with a spot size of 50 μm. The primary standard for $^{87}Rb/^{86}Sr$ calibration was a

233

pressed nanopowder MicaMg ($^{87}Rb/^{86}Sr$ ratio of 154.6; Hogmalm et al., 2017). Other primary

234

standards for $^{87}Sr/^{86}Sr$ ratios were the synthetic glass NIST610 ($^{87}Sr/^{86}Sr$ ratio: 0.709699; Woodhead

235

& Hergt, 2001) and the natural basaltic glass BCR2G ($^{87}Sr/^{86}Sr$ ratio: 0.705003; Elburg et al., 2005).

236

Secondary standards included La Posta biotite grains, with Rb–Sr isochron age of 91.6 ± 1.2 Ma and

237

an initial $^{87}Sr/^{86}Sr$ ratio of 0.7049 ± 5 (Zack & Hogmalm, 2016), and MDC biotite grains with an

238

isochron age of 519.4 ± 6.5 Ma and an initial $^{87}Sr/^{86}Sr$ ratio of 0.72 ± 0.002 (unpublished data). Over

239

the course of the analyses, La Posta grains returned an isochron age of 90.5 ± 4.2 Ma, and MDC

240

grains returned an isochron age of 501.2 ± 7.2 Ma.

241

242

4.6 LA–ICP–QQQ mapping

243

Phengite was mapped in-situ for Rb and Sr using an ASI m50 LA–ICP–MS with attached

244

Agilent 8800 triple quadropole MS at Adelaide Microscopy, Australia. The reaction gas used was O_2,

245

with laser parameters of a fluence of 7 Jcm^{-2} and a frequency of 10 Hz. The primary standard used

246

was a pressed nanopowder MicaMg, and NIST610 was used as a secondary glass standard. The maps

247

were processed in Iolite (Paton et al., 2011). Results of the mapping are semi-quantitative, and are

248

presented in counts per second (CPS).

249

250

5. RESULTS

251

252

5.1 Electron Probe Micro Analyses and mapping

Mineral compositions and end member proportions are reported in Supplementary Table 2 and Tamblyn et al. (2019a), additional phengite, chlorite, omphacite and glaucophane compositional data is in Supplementary Table 5.

EPMA traverses and mapping of whole garnet grains (Fig. 4; 5; Supp. Fig. 2) show complex zoning in the major elements. In garnet 1, cores show a bell-shaped Mn enrichment, consistent with prograde zoning (Fig. 4a; 5a). In the outer part of garnet 1 and prominently in garnet 2 there is a series of thin concentric Mn-rich rings or oscillations (Fig. 5b-c). These narrow Mn-rich zones are often regular, forming hexagonal garnet crystal shapes (Fig. 5c), but can also be highly irregular, appearing to form on embayments in the garnet crystal (Fig. 5b). There is an overall average increase in Mn content from the outer part of garnet 1 to the garnet 2 rim (Fig. 4b). The grossular content of the entire garnet increases slightly from core to rim, from 0.23 to 0.3, and also shows slight enrichments in thin concentric rings in the outer garnet that are correlated with the patterns exhibited by Mn (Fig. 5). Garnet 1 has almandine-poor cores (0.45) increasing to higher almandine content in the outer garnet (0.65; Fig. 4a), with the exception of slight almandine-poor concentric rings which correlate with the Mn-rich rings. Garnet 2 is slightly depleted in almandine (Fig. 4b). Pyrope contents are low in garnet 1, ~0.03, with the innermost core showing a very slight depletion compared to the rest of the garnet (Fig. 5a). Garnet 2 shows distinct step increase in pyrope (0.06) that defines an outer rim ~ 200 μm in width (Fig. 4b).

High resolution EPMA maps (Fig. 5b-c) reveal that garnet 2 forms on a regular to irregular-shaped original surface of the garnet, as indicated by the white dashed line. The growth of garnet 2 is marked by a sharp increase in Mn, followed by several variations in Mn, and a final Mn-poor rim which has been partially replaced by chlorite. Notably, these Mn zoning patterns only correspond to slight changes in MnO (<0.5 wt%). Fe is slightly depleted in garnet 2. This outer-rim transition is also marked by a sharp increase in Mg. The oscillations in Mn commonly correlate with slight increases in Ca.

281

282   EPMA X-ray maps were also obtained from phengite grains in the matrix and in garnet (Fig.

283   6; Supp. Fig. 3). Matrix phengites display cores that are depleted in Fe, enriched in Al and slightly

284   enriched in Na, surrounded by a rim enriched in Si, with enriched Fe and slight enrichment in Mg

285   (Fig. 6). Phengite grains in garnet show phengite included in garnet 1 is enriched in Si, Mg and Fe,

286   and depleted in Al and Na (Fig. 6). Phengite included in garnet 2 is enriched in Al and Na, and

287   depleted in Si and Fe with a slight depletion in Mg, similar to cores of phengite grains in the matrix

288   (Fig. 6). Compositional zoning in the phengite is sharp and well defined. Overall, the phengite zoning

289   is defined by an increase in celadonite proportion from core (40%) to outer rim (45%), accompanied

290   by a decrease in X_{Mg} (0.71–0.63) and an increase in Si (3.43–3.56 p.f.u.; Fig. 7).

291

292   Omphacite occurs as inclusions in garnet 1 and garnet 2, and as coarser grains in the matrix,

293   which show subtle differences in composition (Fig. 7). Omphacite in the matrix and as inclusions in

294   garnet 1 is generally 0.37–0.44 proportion jadeite, 0.36–0.43 proportion diopside, 0.08–0.19

295   proportion hedenbergite with a minor acmite component. Omphacite as inclusions in garnet 2 is

296   similar in composition, with the exception of occasional diopsodic grains, which are 0.15–0.23

297   proportion jadeite, 0.47–0.50 proportion diopside, and 0.19–0.26 proportion hedenbergite. Coarse

298   grains in the matrix and as inclusions in garnet 2 are subtly zoned in X_{Fe}, from ~ 0.27 in the core to ~

299   0.23 in the rim (Supp. Fig. 4). Inclusions in garnet 1 have an average X_{Fe} of 0.28, and inclusions in

300   garnet 2 have an average X_{Fe} of 0.3.

301

302   Chlorite forms grains in the matrix that break down garnet and omphacite, along fractures

303   which cross-cut garnet, and as inclusions at the boundary between garnet 1 and garnet 2 (Supp. Fig.

304   1). Chlorite as inclusions between garnet 1 and garnet 2 has an average X_{Mg} of 0.46 (Supp. Table 5). It

305   shows elevated MnO contents compared to the adjacent garnet (0.54–0.73 wt%). Chlorite throughout

306   the matrix has an average X_{Mg} of 0.38. EPMA X-ray maps show that chlorite is often zoned or patchy

307   in composition (Supp. Fig. 5).

308

309   Retrograde amphibole is ferro-glaucophane in composition after the nomenclature of Leake et

310 al. (1997), and also shows sharply defined compositional zonation (Supp. Fig. 6). X_{Mg} increases from

311 core to rim, accompanied by a slight decrease in Al (1.72–1.39 p.f.u.). There is a slight enrichment of

312 Ca in the cores of the glaucophane (0.17 p.f.u.), however, the grains are dominantly sodic. Some

313 grains have thin outer rims with slight enrichment in MnO (0.14–0.27 wt%, Supp. Fig. 6).

314

315   Titanite differs slightly from ideal formula, with incorporation of Al_2O_3 (up to 2.27 wt%),

316 FeO (up to 0.66 wt%) and F (up to 0.6 wt%) into its crystal structure (Supp. Table 5). Al_2O_3, FeO and

317 F contents show positive correlations (Fig. 7).

318

5.2 Phase equilibrium forward modelling

320   The results of the phase equilibrium forward modelling and corresponding mineral modal and

321 compositional isopleths are shown in Figures 8, 9 and 10. Figure 8 shows the phase equilibrium

322 model for the total bulk rock composition after Tamblyn et al. (2019a). Figure 9 shows the phase

323 equilibrium model for the local rock composition with water in excess, and garnet 1 and lawsonite 1

324 removed. This was done to simulate the effective local composition for the stage 2 mineral

325 assemblage, as garnet 1 and lawsonite 1 were chemically isolated from the reactive rock composition

326 (e.g. Spear, 1988; Marmo et al., 2002; Evans, 2004; Gaides et al., 2008; Konrad-Schmolke et al.,

327 2008; Lanari & Engi, 2017). Figure 10 shows the phase equilibrium model with garnet 1 and

328 lawsonite 1 removed, but with a set water content, calculated from the modes of hydrous minerals

329 (H_2O = 4.65 wt%). This was done as the current water content in the rock is most likely a measure of

330 the minimum amount of water available in the local rock composition. Figures 11 and 12 can only be

331 used for the mineral relations during the secondary assemblage formation after garnet 1 and lawsonite

332 1 have been formed. It is likely that the local rock composition had a water content between that

333 which is measured in the rock currently, and water in excess. Therefore, the *P–T* points during

334 retrograde 1, prograde 2 and peak 2 have been interpreted from both of the models. A summary of the

335 phase equilibrium models used for each stage of the metamorphic evolution is in Supplementary

336 Table 6. The interpreted stages of the *P–T* path are indicated as grey arrows.

337

5.2.1 Stage 1

The prograde path and peak conditions for the formation of the first mineral assemblage are shown on Figure 8b, after Tamblyn et al. (2019a). Garnet 1 nucleated at the 'garnet cores in' line at ~ 0.8–1.1 GPa and approximately 350 °C, based on the core inclusion assemblage of phengite + omphacite + lawsonite + quartz + chlorite + epidote. The path is then interpreted to track up-pressure into the + glaucophane field to satisfy the inclusion relationship of glaucophane in lawsonite and in the core domains of garnet 1 porphyroblasts.

The peak conditions of stage 1 metamorphism are determined from inclusions in the outer part of garnet 1 crystals. We acknowledge that the record of peak 1 has been obliterated by subsequent mineral growth, and as such the interpreted $P–T$ conditions are tentative. We interpret that the lawsonite eclogite reached the phengite + garnet + lawsonite + omphacite + coesite field (based on radial fractures around quartz inclusions), at around 2.9 GPa and 600 °C. This field is bound up-pressure by the conversion of omphacite to jadeite, bound up-temperature by the destabilization of lawsonite, and is bound down-temperature and down-pressure by the loss of coesite or the conversion of coesite to quartz. No inclusions of jadeite were detected, supporting that clinopyroxene remained omphacitic during the metamorphic evolution. Lawsonite is interpreted to have been stable during stage 1 burial as lawsonite preserves ages synchronous with garnet growth (ca. 500 Ma; Tamblyn et al., 2019a), and is ubiquitously found as inclusions in garnet 1. Rare examples of radial fractures around quartz in the outer parts of garnet 1 (Fig. 2e) could indicate the former presence of coesite, further supporting these inferred $P–T$ conditions.

These interpreted peak conditions are supported by the modal proportion of garnet 1. The interface between garnet 1 and 2 is marked by a step increase in X_{Mg} and a prominent increase in X_{Mn} (Fig. 4; 6), together with truncation of core-hosted inclusion trails. The microstructural relationships and change in X_{Mn} and X_{Mg} between the core and the rim strongly imply there was a hiatus in garnet growth. The overall average MnO of the garnet rim is higher than the MnO content of the outer garnet

365 core (Fig. 4), suggesting that there was a period of garnet 1 breakdown prior to the formation of

366 garnet 2. Making the assumption that garnet growth effectively sequesters all available MnO in the

367 rock, particularly at high *P–T* conditions when chlorite is no longer stable, the higher average MnO

368 concentrations in garnet 2 compared to the outer part of garnet 1 suggests that it originally occupied a

369 larger modal proportion of the rock (Fig. 3a). The original size of the garnets during stage 1 can be

370 calculated using the amount of MnO stored in the MnO dissolution rings in garnet 2. If the outer part

371 of the garnet that formed during stage 1 is assumed to have had a flat MnO profile whose MnO is now

372 incorporated into garnet 2, stage 1 garnet at one point occupied ~ 1.05 times its current modal volume,

373 equivalent to approximately 25 mol% (1-oxide-normalised; Fig. 4a). This is a minimum estimate, as

374 the Mn-rich garnet 2 has been partially broken down and replaced by stage 2 Mn-rich chlorite. The

375 calculated modal proportion of stage 1 garnet porphyroblasts plots at *P–T* conditions of ~ 2.9 GPa and

376 600 °C (Fig. 8c), within lawsonite stability. The modal proportion of garnet at this stage of the rock's

377 evolution may be overestimated in the phase equilibrium model, by perhaps up to 5 vol % (e.g. Lanari

378 & Engi, 2017), as garnet growth fractionates major elements. This uncertainty in the garnet mode

379 estimate is presented on the phase equilibrium model (Fig. 8b). The calculated modal proportion of

380 garnet passes through the phengite + lawsonite + omphacite + coesite field. The *P–T* path linking the

381 interpreted prograde assemblages with the peak *P–T* point passes through talc-bearing *P–T* space,

382 however the calculated modal proportion of talc is low (< 1 vol%), and as such it is unlikely that talc

383 would have been preserved in the eclogite.

384

385   The retrograde conditions following peak 1 must be interpreted from the phase equilibrium

386 models calculated from the local bulk rock composition, where garnet 1 and lawsonite 1 cores are not

387 part of the chemical system. However, the *P–T* conditions are difficult to assess, as the water content

388 in the rock at this stage in the lawsonite eclogite's evolution is unknown. Figure 9 shows the phase

389 equilibrium model for the local rock composition with water set to excess (i.e. water saturation), and

390 Figure 10 shows the same composition but with a water content defined by the current mineralogy in

391 the rock. The amount of water currently in the rock is relevant to the assemblages formed during stage

392 2 retrogression, however its relation to the assemblages formed during stage 1 retrogression is

393 unknown. Water content affects the stable mineral assemblages at lower pressures and temperatures,

394 particularly the stability of garnet. The primary mineralogical evidence for stage 1 retrogression is the

395 addition of chlorite + glaucophane to the mineral assemblage (as found as inclusions in garnet and

396 lawsonite rims), and the consumption of garnet 1 rims. As such, the 'garnet-in' line is an important

397 indicator of stage 1 retrograde *P–T* conditions, as it represents the modal volume of garnet in the rock

398 before new stage 2 rims formed. In Figure 9, the garnet-in line occurs in the retrograde assemblage at

399 ~ 1.8–2.1 GPa and 490–500 °C. However, it is unknown if the mineral assemblage was water

400 saturated during retrogression. If the water content was below saturation, we would expect the garnet-

401 in line to move further down pressure and temperature. This can be seen in Figure 10. The difference

402 in position of the garnet-in line is due to water content limiting the formation of chlorite at the

403 expense of garnet. Essentially, this means the garnet-in line in Figure 9 represents the *maximum*

404 pressures and temperatures the rock could have reached during stage 1 retrogression. As water

405 saturation during retrogression seems unlikely, it appears the lawsonite eclogite may have been

406 exhumed to lower pressures and temperatures than can be constrained. In Figure 9, the amount of

407 water in the local bulk composition necessary to destabilize garnet is approximately 6 wt%, 1.35 wt%

408 more than is currently held in hydrous minerals in the rock.

409

5.3.2 Stage 2

411 Renewed burial leading to the growth of the new garnet rims (garnet 2) is interpreted to start at

412 the 'garnet rims in' line (Fig. 9). This is notably only an estimate of the maximum *P–T* conditions for

413 the start of the prograde path for stage 2, as the water content in the eclogite was unknown as this

414 time, and it may have been exhumed to shallower conditions within the subduction channel but failed

415 to record it. The early stages of the prograde path can be determined using compositions of the cores

416 of stage 2 minerals. Useful compositional parameters are x(g), z(g) and y(phe), where x(g) is the X_{Fe}

417 in garnet, z(g) is the grossular component of garnet, and y(phe) is the proportion of Al on the M2 site

418 in phengite (i.e. dioctahedral Al), a pressure-sensitive parameter. In the *P–T* model with water in

419 excess, the x(g) and z(g) parameters from the interior of garnet 2 plot at approximately 1.9 GPa and

420 500–510 °C within the prograde field. Notably, the x(phe) and y(phe) parameters from the cores of

421 phengite intersect at approximately 2.2 GPa and 540°C, possibly indicating the initiation of

422 recrystallisation of phengite at this *P–T* point. The x(o) parameter from cores of omphacite grains,

423 which reflects the X_{Fe}, also plots close to these conditions, however as the textural timing of

424 omphacite growth in this sample is not well known, it is not clear if its compositions reflect stage 1

425 burial or stage 2 burial. In the *P–T* model with set water content, these same compositional parameters

426 plot over a larger range of *P–T* conditions, however converge towards 1.95 GPa and 450 °C. The

427 x(phe) and y(phe) parameters from the cores of phengite intersect at the same point as Figure 9: 2.2

428 GPa and 540°C. As such, an alternative prograde 2 *P–T* path may have begun at lower pressures and

429 temperatures and tracked up towards ~ 2.2 GPa and 540 °C.

430

431   The peak assemblage during stage 2 is interpreted to have been phengite + garnet + omphacite +

432 lawsonite + quartz, which is modelled to occur at conditions of ~ 2.7 GPa and 590 °C. This field

433 exists identically on both Figure 9 and Figure 10, and is bound up-temperature by the loss of

434 lawsonite, down-temperature by the presence of talc and/or glaucophane, and up-pressure by the

435 conversion of quartz to coesite and/or omphacite to jadeite.

436

437   Stage 2 retrograde *P–T* conditions are constrained from Figure 10. It is characterised by the

438 growth of chlorite and retrograde-zoned glaucophane at the expense of garnet and omphacite. It also

439 includes a modal proportion increase in lawsonite (lawsonite 2), which overgrows garnet 2 (Fig. 2f).

440 The current modal proportions of all stage 2 minerals – minus the garnet 1 and lawsonite 1 cores

441 which are not included in the modelling – intersect in the phengite + garnet + lawsonite + omphacite

442 + quartz + chlorite + glaucophane + H_2O field (Fig. 10c). The modelled isopleths track parallel to

443 each other down temperature in the H_2O absent field, spanning between 1.8–2.1 GPa (Fig. 10c). Final

444 recorded retrogression is therefore interpreted to occur down to ~ 2.0 GPa and 520 °C.

445

446   **5.3 U–Pb titanite geochronology**

447   In-situ U–Pb isotopic analyses of titanite in the matrix and in the garnet 2 rims define a

448 common lead trend with a lower intercept of 455.1 ± 9.6 Ma (MSWD = 3.3, *n* = 163, Fig. 11a). The

449 upper intercept has a ^{206}Pb/^{207}Pb ratio of 0.8166 ± 0.0096. During data processing, 16 analyses were

450 discarded due to low or no signal, and a further 23 analyses were discarded due to contamination from

451 detectable or visible micro-zircons. Incidental ablation of these zircons was visible in the signals as

452 large Zr spikes. U–Pb geochronological data can be found in Supplementary Table 7.

453

454

5.4 Ar–Ar phengite geochronology

456 Step-heating results of two aliquots of phengite from the lawsonite eclogite taken from ~ 50

457 cm apart are shown in Figure 12. The mineral grains come from the rock matrix. Ar–Ar data can be

458 found in Supplementary Table 8. For aliquot 1, an age of 458.4 ± 3.7 Ma was obtained from 72.4% of

459 released ^{39}Ar. Aliquot 2 shows a rise in age with increasing temperature steps. An age of 462.6 ± 3.6

460 Ma was calculated from 61.9% of released ^{39}Ar.

461

5.5 Rb–Sr phengite geochronology

463 In-situ LA–ICP–MS Rb–Sr analyses of phengite cores and rims produce almost identical

464 isochrons (Fig. 13). Phengite rims have a wide range of ^{87}Rb/^{86}Sr ratios and produce an age of 448 ±

465 13 Ma (MSWD = 0.61, n = 18, Fig. 13a), in many cases with favourably high ^{87}Rb/^{86}Sr ratios (up to

466 2000). Phengite cores generally have lower ^{87}Rb/^{86}Sr ratios and produce an age of 437 ± 37 Ma

467 (MSWD = 0.84, n = 16, Fig. 13b). When combined, all phengite data give an isochron age of 449 ±

468 19 Ma (MSWD = 0.55, n = 34, Fig. 13c). Based on textural evidence that titanite occurs as inclusions

469 in phengite, as matrix grains and locally as grains rimming phengite (Fig. 11), it was assumed to have

470 grown synchronously with matrix phengite. Therefore, titanite was analysed to obtain an initial

471 ^{87}Sr/^{86}Sr ratio to anchor the isochron, producing an average ^{87}Sr/^{86}Sr ratio of 0.70986 (n = 6, Fig. 13d).

472 The rims are interpreted to form during peak metamorphism and thus were used in the final age

473 calculations. When the phengite rims are anchored with the titanite analyses, the isochron age

474 produced is 450 ± 11 Ma (MSWD = 1.09, n = 26, Fig. 13e). Rb–Sr geochronological data can be

475 found in Supplementary Table 9.

476

Phengite was mapped for ^{85}Rb and ^{87}Sr (Fig. 14). The qualitative results reveal the phengites are zoned, showing enrichments in Sr in the cores of the grains with slight Rb depletion, and Rb enrichments in the rims which correlates to comparatively low levels of Sr. The boundary between these cores and rims is sharp, and in larger grains is aligned along cleavage (Fig. 14). This zoning correlates to zoning seen in the major elements; Sr enriched cores correspond to low Si/high Al cores, and Rb enriched rims correspond to high Si/low Al rims. High Sr grains are titanite. The outline of phengite grains is shown in white.

6. DISCUSSION

6.1 Petrological evidence for reburial of the lawsonite eclogite

Petrographical observations from the block of lawsonite eclogite in the Port Macquarie serpentinite mélange denote at least two interpretable burial events, summarized in Supplementary Table 1 and schematically outlined in Figure 3.

Evidence for stage 1 prograde (or burial 1) is preserved in the cores of garnet porphyroblasts, which are prograde-zoned, showing bell-shaped Mn profiles. Lawsonite porphyroblasts grew in equilibrium with garnet, as evidenced by their ca. 500 Ma age, synchronous with garnet growth (Tamblyn et al., 2019a). Stage 1 burial is also recorded by phengite inclusions in the outer part of garnet 1, which are high in Si and low in Al, consistent with crystallization at high pressure. It is possible that rare coesite was formed and captured in the outer parts of garnet 1, interpreted from the presence of radial fractures around quartz.

The petrologic evidence for a second burial event after partial exhumation is recorded as inclusions within garnet rims, lawsonite rims, and the compositions of matrix minerals. The step increase in X_{Mg} and a Mn oscillation marks the boundary of new garnet 2 growth, and the general

504 increased Mn throughout the new garnet rim suggests that it was formed from the breakdown of

505 previously larger garnet porphyroblasts. The presence of chlorite on the boundary between garnet 1

506 and garnet 2 is key evidence that lower temperatures and pressures were reached before the growth of

507 garnet 2. Omphacite grains in the matrix are prograde zoned (Supp. Fig. 4), and possibly grew from

508 the breakdown of glaucophane during the second burial event, however, may also be relic from stage

509 1. Phengite and titanite (which make up the strongly foliated rock matrix) underwent recrystallization

510 during prograde 2 deformation (e.g. Spencer et al., 2013; de Meyer, 2014; Itaya & Tsujimori, 2015).

511 This interpretation is supported by petrological observation that the pervasive foliation in the matrix is

512 at an oblique angle to the foliation preserved as inclusion trails in garnet 1 (Fig. 3c). The prograde

513 zoning in the phengite is consistent with recrystallization with increasing pressure and temperature,

514 however the sharp boundary between zones could either be interpreted as i) progressive prograde

515 zoning during total recrystallization of the phengite grains or ii) partial recrystallization of the

516 phengite rims at higher pressure, preserving an older low-pressure core. Phengite with low Si/high Al,

517 identical to the cores of phengite in the matrix, is preserved in the garnet 2 rims (Fig. 6), providing

518 further evidence that garnet 2 rims grew at lower P–T conditions than the outer parts of garnet 1.

519 Evidence for stage 2 retrograde includes partial replacement of garnet by chlorite, and partial

520 replacement of omphacite by glaucophane and chlorite (Fig. 3d). Retrograde zoning in glaucophane

521 (Supp. Fig. 6) supports that it grew during decreasing pressure and temperature. New lawsonite

522 growth on existing crystals captured coarse titanite, garnet and glaucophane, suggesting that lawsonite

523 increased its modal proportion during retrograde 2 (Fig. 3d).

524

525 **6.2 The origin of oscillatory garnet zoning and the possible role of fluids**

526

527 Garnets from the lawsonite eclogite show oscillatory zoning in Ca, Mn, Fe and Mg (Fig. 5;

528 Supp. Fig. 2), which is subtle in the outer parts of stage 1 garnet but marked throughout the stage 2

529 garnet rims. Narrow annuli enriched in Mn and Ca are depleted in Fe and Mg, and vice versa. The

530 oscillations are sharp; they occasionally form epitaxial hexagonal crystal shapes on pre-existing

531 garnet (Fig. 5c), or form on irregular-shaped embayed surfaces of pre-existing garnet (Fig. 5b). These

532 features suggest that garnet was resorbed and regrown several times during the metamorphic history

533 of the lawsonite eclogite (aside from the major garnet resorption event between burial 1 and burial 2).

534

535   Garnet resorption and regrowth have been suggested to occur via changes in pressure and

536 temperature which influence garnet stability (e.g. García-Casco et al., 2002; Kohn, 2004). This

537 mechanism has been interpreted as the cause of oscillatory garnet zoning from high-pressure rocks in

538 subduction systems, where seismic instability and/or serpentinite hosted channel circulation cause

539 cyclic $P–T$ changes (García-Casco et al., 2002; Li et al., 2016, Viete et al., 2018). However, garnet

540 resorption has also been suggested to be a result of external fluid influx, which dissolves and

541 reprecipitates garnet in equilibrium with the fluid (Angiboust et al., 2011; Hyppolito et al, 2018).

542 Additionally, the influx of fluids may enrich or deplete cations incorporated into the garnet during its

543 growth (e.g. Martin et al., 2011; Angiboust et al., 2014; Park et al., 2017). Such fluid influxes may not

544 be easy to trace, however the lack of oscillatory zoning in the major elements of other minerals (e.g.

545 phengite, omphacite) suggests that fluid influxes did not occur on a major scale, or at least were only

546 recorded in garnet. As there is no petrologic evidence supporting either case, the major element

547 oscillatory zoning in garnet is interpreted to have formed by subtle fluctuations in $P–T$ and/or influxes

548 of fluid.

549

550 **6.3 *P–T* conditions of reburial**

551   The record of the $P–T$ conditions during the first burial event is contained as inclusions in

552 lawsonite and garnet, while the record of the second burial event is found in garnet and lawsonite rims

553 and the matrix mineralogy. The $P–T$ path proposed is summarised in Figure 15, the interpreted

554 conditions are summarised in Supplementary Table 6. The proposed $P–T$ path is tentative, due to the

555 difficulty in reconstructing the mineralogical evolution of the sample.

556

557   The $P–T$ path during stage 1 burial is recorded from 0.8–1.1 GPa at approximately 350 °C,

558 and then tracks up to peak conditions of at least 2.9 GPa and 600 °C based on lawsonite stability,

559 inclusion assemblages in lawsonite and garnet and the estimated modal proportion of garnet (Fig. 8).

560 The potential former presence of coesite, now recorded by radial fractures around quartz inclusions in

561 the outer parts of garnet 1, supports these peak conditions. High Si/low Al phengite inclusions in the

562 outer parts of garnet 1 also support that high-pressures were reached during this stage (Fig. 6).

563 Subsequent retrogression to maximum conditions of ~ 2.0 GPa and 500 °C is recorded by the

564 consumption of garnet, and growth of chlorite and glaucophane (Fig. 9; 10). It is likely the lawsonite

565 eclogite reached lower pressures and temperatures than this during retrogression, however this part of

566 its evolution is difficult to reconstruct.

567

568 Stage 2 reburial began at maximum conditions of ~ 2.0 GPa and 500 °C. Growth of new

569 garnet rims with elevated Mn and X_{Mg} captured the stage 1 chlorite and glaucophane. Stage 2 peak

570 conditions of 2.7 GPa and 590 °C were reached, based on the preserved peak mineral assemblage.

571 Retrogression to approximately 1.9 GPa and 520 °C is based on the current modal mineralogy of the

572 rock (Fig. 10c), and is supported by the formation of retrograde-zoned glaucophane and the partial

573 break down of garnet to form chlorite.

574

575 If these tentative estimates are taken at face value, the earliest stages of the prograde

576 evolution of the rock occurred on a geothermal gradient of approximately 370 °C/GPa, while the peak

577 and retrograde stages of both stage 1 and 2 occurred on a cooler geothermal gradient of approximately

578 210–265 °C/GPa. These gradients are in line with a subduction channel which is cooling over time,

579 potentially in response to the refrigeration effects of on-going subduction (Gerya, 2002; Agard et al.,

580 2018). If this is correct, it suggests the eclogite records the transition from subduction initiation (e.g.

581 Tamblyn et al., 2019a) to subduction maturity. These inferred geothermal estimates are comparable to

582 the geothermal gradient experienced by garnet blueschist from the same mélange that hosts the

583 lawsonite eclogite at Port Macquarie (~ 275 °C/GPa; Tamblyn et al., 2019a). The complete $P–T$ path

584 suggested in Figure 15 is two hair-pin loops, which are constrained by the presence of lawsonite

585 throughout the recorded history. Such $P–T$ evolutions have been previously suggested for high-

586 pressure rocks in oceanic subduction systems (e.g. Gerya et al, 2002; Krebs et al., 2011; Pourteau et

587 al., 2019).

588

6.4 Timing of reburial

Tamblyn et al. (2019a) obtained Lu–Hf ages of 489.7 ± 5 Ma for garnet–whole rock and 505 ± 15 Ma for lawsonite–whole rock from the eclogite. The cores of stage 1 garnet are highly enriched in Lu compared to the stage 2 garnet, indicating the garnet Lu–Hf age is essentially the age of stage 1 garnet growth. The lawsonite Lu–Hf age is less precise but within uncertainty of the garnet age, suggesting that stage 1 garnet and stage 1 lawsonite grew within a single isotopic reservoir at ca. 500–490 Ma. The lawsonite grains contain patches that are enriched in Lu (Tamblyn et al., 2019a), and in places are overgrown by volumetrically minor second generation lawsonite (Fig. 2f). The Lu-rich domains in stage 1 lawsonite formed prior to garnet nucleation. The *P–T* path during the stage 1 burial tracks up to tentative peak conditions of at least 2.9 GPa and 600 °C (section 6.3). This occurred at ca. 505–490 Ma as dated by garnet and lawsonite Lu–Hf (Tamblyn et al., 2019a; Fig. 15).

Matrix phengite that encloses garnet, lawsonite and omphacite porphyroblasts shows an increasing Si and celadonite content from core to rim (Fig. 6; Carswell et al., 2000). This compositional component is consistent with increasing pressure, which in the absence of information about changes in non-lithostatic stress, equates to increasing burial depth. The plateau ages obtained from Ar–Ar geochronology from phengite from the rock matrix give ages of ca. 458 and ca. 462 Ma. The suggested closure temperature of argon diffusion in phengite is variable (e.g. Warren et al., 2011; Fornash et al., 2016), and is dependent on grain size and fluid availability (e.g. Glodny et al., 2002). Numerical diffusion modelling suggests that phengite grown at blueschist to eclogite facies (up to 550 °C) can retain argon from its prograde history (Warren et al., 2011). This has been supported by step-heating experiments which suggest phengite can be highly retentive of argon (Forster and Lister, 2014). These findings are in line with samples from high-pressure terranes where Ar–Ar ages are similar to peak metamorphic ages obtained from zircon or garnet geochronology (Fornash et al., 2016), or can be texturally related to mineral growth during sequential phases of deformation (Putlitz et al., 2005).

616   In-situ LA–ICP–MS Rb–Sr dating of mica is still in its infancy as a geochronological

617   technique. However, it has the potential to rapidly constrain ages of metamorphism and deformation

618   in metamorphic rocks. Rb–Sr analyses of phengite rims, coupled with an initial $^{87}Sr/^{86}Sr$ value from

619   texturally intergrown titanite, give an isochron age of ca. 450 Ma. The rims of phengite were selected

620   for the final age calculation, as they are interpreted to grow at peak conditions of stage 2.

621

622   Rb and Sr maps also support that the geochronologic results are growth ages (Fig. 14); the

623   zoning is sharply defined, suggesting minimal post-growth diffusion within the grains (e.g. Konrad-

624   Schmolke et al., 2011). Rb–Sr maps show Sr enrichments in the low-Si cores of phengite, and Rb

625   enrichments in the high-Si rims, a pattern that has been noted in high-pressure rocks (Di Vincenzo et

626   al., 2006). The incorporation of Sr into phengite may therefore be related to the breakdown of another

627   Sr-bearing mineral in the system at lower pressures, or, a preference for Sr in low-Si less-phengitic

628   white mica, which declines with increasing phengitic component up-pressure. The Rb–Sr age of the

629   phengite rims is within uncertainty of the age obtained from phengite Ar–Ar analyses and titanite U–

630   Pb age of ca. 455 Ma (below). This interpretation is in line with similar high-pressure terranes, where

631   Rb–Sr dates record crystallization of phengite on the prograde-peak path during metamorphism (e.g.

632   Hetzel & Romer, 2000; Glodny et al., 2002; Bosse et al., 2005; Di Vincenzo et al., 2006). In some of

633   these examples, Rb–Sr phengite growth ages are synchronous with Ar–Ar phengite ages

634   (Anczkiewicz et al., 2000; Hetzel & Romer, 2000; Glodny et al., 2002; Rodríguez et al., 2003). Rb–Sr

635   has also been shown to date continuous recrystallization of white mica during retrograde deformation

636   in polymetamorphic terranes, at peak temperatures of ~ 500–550 °C (Bröcker et al., 2013),

637   demonstrating the ability for the Rb–Sr white mica ages to record recrystallization events.

638

639   The U–Pb data from titanite in the matrix and as inclusions in the stage 2 garnet gives an age

640   of ca. 455 Ma. Although there is some dispersion in the data, it is evident the matrix titanites are

641   significantly younger than the ca. 490 Ma age for stage 1 lawsonite and garnet. Phase equilibrium

642   modelling suggests the rock experienced *P–T* conditions below the closure temperature of titanite

643   (~600–800 °C; Hartnady et al., 2019). Additionally, there is no evidence of a rutile-bearing

644 assemblage in the lawsonite eclogite. Therefore, it is unlikely the titanite formed from breakdown of

645 rutile on the retrograde path. Consequently, we interpret the titanite U–Pb data to reflect the timing of

646 titanite recrystallisation.

647

648 Titanite occurs throughout the matrix and forms around the edges of prograde-zoned phengite

649 (Fig. 11b-d). It occurs as fine-grained inclusion trails in garnet cores and lawsonite, and as coarse-

650 grained inclusions in stage 2 garnet and lawsonite rims (Fig. 11b-c). These textural relationships

651 indicate that titanite was stable during the entire metamorphic evolution of the eclogite, first growing

652 during the stage 1 burial event, and recrystallizing during the prograde evolution of the stage 2 burial

653 event to form coarser-grained titanite. This recrystallization was possibly tracked by increasing F,

654 Al_2O_3 and FeO content in the titanite (Fig. 7).

655

656 The inclusions in titanite are limited to fine-grained apatite, phengite and zircon (< 10 µm).

657 Zircon is either euhedral and 5–10 µm in size, or occurs as fine-grained clusters or web-like

658 intergrowths which are 2 µm or less (Fig. 11b,c). The 5–10 µm sized apatite, phengite and zircon

659 inclusions are interpreted to predate the crystallization of titanite, and were avoided during analyses.

660 However, the extremely fine-grained web-like micro-zircon was difficult to avoid, and was possibly

661 incorporated into analyses. This is suggested by Zr concentrations in the titanite analyses of up to ~

662 18800 ppm (Supp. Table 7). Texturally, these extremely fine-grained zircon dustings only occur in

663 sections of titanite grains, and are absent in phengite, garnet, lawsonite and omphacite. This suggests

664 they are probably exsolution of Zr from the titanite, rather than inclusions (Fig. 11b,c). Titanite can

665 substitute Zr^{4+} for Ti^{4+} in its crystal lattice, as evidenced by development of the Zr-in-titanite

666 thermobarometer (Hayden et al., 2008). However, in quartz-absent rocks, titanite can take in

667 significant amounts of ZrO_2 (up to 15 wt%; Della Ventura et al., 1999; Seifert & Kramer, 2003;

668 Seifert; 2005; Vuorinen & Hålenius, 2005; Liferovich & Mitchell, 2005; Chakhmouradian, 2004).

669 While these studies are from igneous rocks, they demonstrate the ability for significant Zr to

670 substitute into titanite.

671

672   The phase equilibrium modelling of the stage 2 effective bulk composition suggests the

673   prograde-peak path taken by the lawsonite eclogite during reburial was quartz-free until absolute peak

674   conditions were obtained (590 °C, 2.7 GPa). Zircon exsolution from titanite has not been directly

675   reported before, but in lieu of other explanations, we suggest these fine-grained clusters of zircon

676   within titanite represent high-Zr titanite which formed in a quartz absent matrix assemblage, which

677   then exsolved zircon after crystallization of the parent mineral. As such, while we tried to avoid areas

678   of obvious weblike zircon in titanite, we suggest that if small amounts of fine-grained zircon

679   exsolution were incorporated into the titanite U–Pb analyses, we are dating U–Pb isotopic ratios

680   related to titanite recrystallization at ca. 455 Ma.

681

682   The above interpretations suggest recrystallization of phengite and titanite at ca. 450 Ma dates

683   the timing of burial 2, which reached peak P–T conditions of 2.7 GPa and 590 °C. The two burial

684   events were separated by a period of exhumation to maximum conditions of approximately 2.0 GPa

685   and 500 °C, however the lawsonite eclogite may have been exhumed further up the subduction

686   channel. The two stages of metamorphism and burial in the lawsonite eclogite occurred over a period

687   of at least ca. 50 M.y. (Fig. 15). The timing of final exhumation after burial 2 is unknown, but had

688   occurred by the Permian, based on serpentinite detritus in nearby early Permian basins in the SNEO

689   (Aitchison et al., 1994).

690

6.5 Examples of other burial cycles

692   The suggestion that the block of lawsonite eclogite at Port Macquarie underwent two stages

693   of burial, each marked by garnet growth and separated by a stage of garnet breakdown, is similar to

694   P–T histories proposed from other studies on subducted rocks in oceanic settings, and is supported by

695   results of numerical modelling (e.g. Gerya et al., 2002). Despite the difficulties in recognizing and

696   constraining multiple cycles of burial and exhumation within single high-pressure samples, workers

697   have had success using various geochronological and mineralogical tools. Dating zones within

698   metamorphic zircon, combined with REE chemistry and inclusion assemblages, have allowed two or

699   more subduction cycles to be identified in Rhodope, Greece (Liati et al., 2016) and the Italian Alps

700 (Rubatto et al., 2011), however the latter was within a continental subduction system. Three burial

701 cycles in the Sambagawa belt in Japan have been identified from mineral assemblages and mineral

702 chemistry of garnet and amphibole in eclogite by Kabir & Takasu (2010). Two burial cycles are

703 recorded in eclogite from the Akeyazi terrane in Tianshan, China, interpreted from petrological and

704 mineral zoning evidence, with $P–T$ paths constrained from psuedosections (Li et al., 2016). García-

705 Casco et al. (2002) modelled oscillatory zoning in garnet and amphibole chemistry from serpentinite

706 mélange in Cuba, concluding that they formed in near-equilibrium and record fluctuations in $P–T$

707 conditions during subduction. Blanco-Quintero et al. (2011) also reported reburial during interpreted

708 large-scale convective flow from this mélange, interpreted from garnet zoning and $P–T$ pseudosection

709 modelling. Geochronological constraints on the development of the mineral assemblages in the Port

710 Macquarie eclogite point to a long history of high-pressure metamorphism, with a significant age

711 difference between the formation of stage 1 and stage 2 assemblages (Fig. 15). Based on time scales

712 of material movement in subduction systems from numerical modelling (Gerya et al., 2002), the

713 geochronological data from the eclogite supports the likelihood that multiple burial and partial

714 exhumation paths may have occurred.

715

716   **6.6 Subduction channel dynamics**

717

718   A corner or return flow model of material within the subduction channel (e.g. Shreve & Cloos,

719 1986, Gerya et al., 2002; Agard et al., 2009; Roda et al., 2019), is suggested for repeated burial and

720 exhumation of the lawsonite eclogite at Port Macquarie. Numerical modelling predicts that during

721 oceanic subduction, hydration of the mantle wedge creates a buoyant low viscosity serpentinite-filled

722 subduction channel which can pluck high-pressure rocks from the subducting slab and circulate them

723 within the channel (e.g. Gerya et al., 2002). In these models, the high-pressure fragments can become

724 deeply subducted, but can also experience partial exhumation and reburial, resulting in $P–T$ loops

725 (Gerya et al., 2002). Rocks with different $P–T–t$ histories can also accumulate within the channel and

726 be exhumed together to form mélanges that contain a variety of subducted products (e.g. Krebs et al.,

727 2008). At Port Macquarie, this behaviour is confirmed by the presence of garnet blueschist, which is

728 ca. 470 Ma in age, and has a different *P–T* path to the eclogite, yet is found as a tectonic block less

729 than 10 meters away from the lawsonite eclogite (Och et al., 2003; Tamblyn et al., 2019a).

730

731   The geochronological data presented above also supports the notion of long-lived residence of

732 high-pressure materials within oceanic subduction channels. The minimum duration spent in the

733 subduction channel by the lawsonite eclogite at Port Macquarie is ca. 50 M.y. (Fig. 15). Analogous

734 durations of high-pressure metamorphism and accumulation have been reported from oceanic

735 subduction systems. Exotic high-pressure rocks hosted in serpentinite, which have been interpreted to

736 have been buried and exhumed by convective flow processes, span an age range of ca. 55 M.y. in

737 Cuba (Lázaro et al., 2009), and ca. 40 M.y. in the Dominican Republic (Krebs et al., 2008). High-

738 pressure rocks exhumed in Turkey with different *P–T–t* histories span an age range of ca. 20 M.y.

739 (Pourteau et al., 2019), supported by in-situ Ar–Ar data (Fornash et al., 2016). In the Mariana

740 subdution system, the residence time for blueschist within the subduction channel is ca. 48 M.y.

741 (Tamblyn et al., 2019b).

742

743   With the caveat that the *P–T* conditions suggested from the phase equilibrium modelling of

744 the eclogite at Port Macquarie are taken as approximations, a *P–T–t* path can be constructed for the

745 lawsonite eclogite, and the rate of exhumation to reburial estimated. If the maximum possible *P–T*

746 conditions of stage 1 exhumation are taken as 2.0 GPa and 500 °C, the total pressure and temperature

747 change from peak stage 1 to peak stage 2 equates to ~ 1.6 GPa and 190 °C. If an average lithostatic

748 pressure of 3.2 km/kbar with no non-lithostatic loads is assumed, this pressure change is equivalent to

749 approximately 50 km of vertical depth change. Depending on subduction angle, the total travel

750 distance of the lawsonite eclogite within the channel may have been more than 100 km during the

751 looping *P–T* history.

752

753

754   **7. CONCLUSIONS**

755

756   Lawsonite eclogite in serpentinite mélange at Port Macquarie in the Southern New England

757 Orogen records P–T conditions of ~ 2.9 GPa and 600 °C, attained at ca. 500–490 Ma. The mineral

758 assemblages that define these P–T conditions were overprinted by a second high-pressure

759 assemblage that formed at ca. 450 Ma. Compositional zoning in phengite, and growth of a second

760 generation of garnet, indicate the overprinting mineral assemblage formed during up-pressure

761 metamorphism to ~ 2.7 GPa and 590 °C. Between these two stages of burial, partial exhumation

762 occurred to at least 2.0 GPa, but probably to lower P–T conditions. The P–T evolution defines hair-

763 pin looping P–T paths that developed along a similar geothermal gradient, suggesting the lawsonite

764 eclogite was cycled within a serpentinite-filled subduction channel during return flow of subducted

765 material. This cycling occurred over a minimum timeframe of ca 50 M.y., implying long-lived

766 residence of the eclogite within the subduction system. A number of studies are beginning to

767 discover that subducted material may undergo complex P–T–t evolutions that record multiple

768 episodes of burial and exhumation during prolonged residence within subduction systems. These

769 cyclic P–T evolutions support predictions from numerical models which suggest material can follow

770 convective flow paths within evolving serpentinite filled subduction channels.

771

782

783   *References*

784

785 Agard, P., Plunder, A., Angiboust, S., Bonnet, G., & Ruh, J. (2018). The subduction plate interface:

786   Rock record and mechanical coupling (from long to short time scales). *Lithos*.

787 Agard, P., Yamato, P., Jolivet, L., & Burov, E. (2009). Exhumation of oceanic blueschists and

788   eclogites in subduction zones: timing and mechanisms. *Earth-Science Reviews, 92*(1), 53-79.

789 Aitchison, J. C., Blake, M. C., Flood, P. G., & Jayko, A. S. (1994). Paleozoic ophiolitic assemblages

790   within the southern New England orogen of eastern Australia: Implications for growth of the

791   Gondwana margin. *Tectonics, 13*(5), 1135-1149. doi: 10.1029/93TC03550

792 Anczkiewicz, R., Burg, J., Villa, I. M., & Meier, M. (2000). Late Cretaceous blueschist

793   metamorphism in the Indus suture zone, Shangla region, Pakistan Himalaya. *Tectonophysics,*

794   *324*(1-2), 111-134.

795 Anczkiewicz, R., Platt, J. P., Thirlwall, M. F., & Wakabayashi, J. (2004). Franciscan subduction off to

796   a slow start: evidence from high-precision Lu–Hf garnet ages on high grade-blocks. *Earth*

797   *and Planetary Science Letters, 225*(1), 147-161.

798 Anczkiewicz, R., Szczepański, J., Mazur, S., Storey, C., Crowley, Q., Villa, I. M., Thirlwall, M. F., &

799   Jeffries, T. E. (2007). Lu–Hf geochronology and trace element distribution in garnet:

800   implications for uplift and exhumation of ultra-high pressure granulites in the Sudetes, SW

801   Poland. *Lithos, 95*(3-4), 363-380.

802 Angiboust, S., Agard, P., Raimbourg, H., Yamato, P., & Huet, B. (2011). Subduction interface

803   processes recorded by eclogite-facies shear zones (Monviso, W. Alps). *Lithos, 127*(1-2), 222-

804   238.

805 Angiboust, S., Pettke, T., De Hoog, J. C., Caron, B., & Oncken, O. (2014). Channelized fluid flow

806   and eclogite-facies metasomatism along the subduction shear zone. *Journal of Petrology,*

807   *55*(5), 883-916.

808 Baitsch-Ghirardello, B., Gerya, T. V., & Burg, J.-P. (2014). Geodynamic regimes of intra-oceanic

809   subduction: Implications for arc extension vs. shortening processes. *Gondwana Research,*

810   *25*(2), 546-560.

811   Blanco-Quintero, I. F., García-Casco, A., & Gerya, T. V. (2011). Tectonic blocks in serpentinite

812   mélange (eastern Cuba) reveal large-scale convective flow of the subduction channel.

813   *Geology, 39*(1), 79-82.

814   Bosse, V., Féraud, G., Ballevre, M., Peucat, J.-J., & Corsini, M. (2005). Rb–Sr and 40Ar/39Ar ages in

815   blueschists from the Ile de Groix (Armorican Massif, France): implications for closure

816   mechanisms in isotopic systems. *Chemical Geology, 220*(1-2), 21-45.

817   Bröcker, M., Baldwin, S., & Arkudas, R. (2013). The geological significance of 40Ar/39Ar and Rb–

818   Sr white mica ages from S yros and S ifnos, G reece: a record of continuous (re)

819   crystallization during exhumation? *Journal of metamorphic Geology, 31*(6), 629-646.

820   Carswell, D., Wilson, R., & Zhai, M.-g. (2000). Metamorphic evolution, mineral chemistry and

821   thermobarometry of schists and orthogneisses hosting ultra-high pressure eclogites in the

822   Dabieshan of central China. *Lithos, 52*(1-4), 121-155.

823   Castelli, D., & Rubatto, D. (2002). Stability of Al-and F-rich titanite in metacarbonate: petrologic and

824   isotopic constraints from a polymetamorphic eclogitic marble of the internal Sesia Zone

825   (Western Alps). *Contributions to Mineralogy and Petrology, 142*(6), 627-639.

826   Chakhmouradian, A. R. (2004). Crystal chemistry and paragenesis of compositionally unique (Al-,

827   Fe-, Nb-, and Zr-rich) titanite from Afrikanda, Russia. *American Mineralogist, 89*(11-12),

828   1752-1762.

829   Cherniak, D. (2006). Zr diffusion in titanite. *Contributions to Mineralogy and Petrology, 152*(5), 639-

830   647.

831   Collins, W., & Richards, S. (2008). Geodynamic significance of S-type granites in circum-Pacific

832   orogens. *Geology, 36*(7), 559-562.

833   Collins, W. J. (2002). Nature of extensional accretionary orogens. *Tectonics, 21*(4).

834   de Meyer, C. M., Baumgartner, L. P., Beard, B. L., & Johnson, C. M. (2014). Rb–Sr ages from

835   phengite inclusions in garnets from high pressure rocks of the Swiss Western Alps. *Earth and*

836   *Planetary Science Letters, 395*, 205-216.

837   Della Ventura, G., Bellatreccia, F., & Williams, C. (1999). Zr-and LREE-rich titanite from Tre Croci,

838   Vico volcanic complex (Latium, Italy). *Mineralogical Magazine, 63*(1), 123-130.

839 Di Vincenzo, G., Tonarini, S., Lombardo, B., Castelli, D., & Ottolini, L. (2006). Comparison of
840   40Ar–39Ar and Rb–Sr data on phengites from the UHP Brossasco–Isasca Unit (Dora Maira
841   Massif, Italy): implications for dating white mica. *Journal of Petrology*, *47*(7), 1439-1465.

842 Diener, J., & Powell, R. (2012). Revised activity–composition models for clinopyroxene and
843   amphibole. *Journal of metamorphic Geology*, *30*(2), 131-142.

844 Dodson, M. H. (1973). Closure temperature in cooling geochronological and petrological systems.
845   *Contributions to Mineralogy and Petrology*, *40*(3), 259-274.

846 Elburg, M., Bons, P., Foden, J., & Brugger, J. (2003). A newly defined Late Ordovician magmatic-
847   thermal event in the Mt Painter Province, northern Flinders Ranges, South Australia.
848   *Australian Journal of Earth Sciences*, *50*(4), 611-631.

849 Elburg, M., Vroon, P., van der Wagt, B., & Tchalikian, A. (2005). Sr and Pb isotopic composition of
850   five USGS glasses (BHVO-2G, BIR-1G, BCR-2G, TB-1G, NKT-1G). *Chemical Geology*,
851   *223*(4), 196-207.

852 Evans, T. (2004). A method for calculating effective bulk composition modification due to crystal
853   fractionation in garnet- bearing schist: implications for isopleth thermobarometry. *Journal of*
854   *metamorphic Geology*, *22*(6), 547-557.

855 Fornash, K. F., Cosca, M. A., & Whitney, D. L. (2016). Tracking the timing of subduction and
856   exhumation using 40 Ar/39 Ar phengite ages in blueschist-and eclogite-facies rocks
857   (Sivrihisar, Turkey). *Contributions to Mineralogy and Petrology*, *171*(7), 67.

858 Forster, M. A., & Lister, G. S. (2014). 40Ar/39Ar geochronology and the diffusion of 39Ar in
859   phengite–muscovite intergrowths during step-heating experiments in vacuo. *Geological*
860   *Society, London, Special Publications*, *378*(1), 117-135.

861 Fukui, S., Watanabe, T., Itaya, T., & Leitch, E. C. (1995). Middle Ordovician high PT metamorphic
862   rocks in eastern Australia: Evidence from K- Ar ages. *Tectonics*, *14*(4), 1014-1020.

863 Gaidies, F., De Capitani, C., & Abart, R. (2008). THERIA_G: a software program to numerically
864   model prograde garnet growth. *Contributions to Mineralogy and Petrology*, *155*(5), 657-671.

865 García- Casco, A., Torres- Roldán, R., Millán, G., Monié, P., & Schneider, J. (2002). Oscillatory

866   zoning in eclogitic garnet and amphibole, Northern Serpentinite Melange, Cuba: a record of

867   tectonic instability during subduction? *Journal of metamorphic Geology, 20*(6), 581-598.

868 Gerya, T. V., Stöckhert, B., & Perchuk, A. L. (2002). Exhumation of high- pressure metamorphic

869   rocks in a subduction channel: A numerical simulation. *Tectonics, 21*(6).

870 Glen, R. (2013). Refining accretionary orogen models for the Tasmanides of eastern Australia.

871   *Australian Journal of Earth Sciences, 60*(3), 315-370.

872 Glodny, J., Bingen, B., Austrheim, H., Molina, J. F., & Rusin, A. (2002). Precise eclogitization ages

873   deduced from Rb/Sr mineral systematics: the Maksyutov complex, Southern Urals, Russia.

874   *Geochimica et Cosmochimica Acta, 66*(7), 1221-1235.

875 Green, E., Holland, T., & Powell, R. (2007). An order-disorder model for omphacitic pyroxenes in the

876   system jadeite-diopsidehedenbergite-acmite, with applications to eclogitic rocks. *American

877   Mineralogist, 92*(7), 1181-1189.

878 Hartnady, M. I., Kirkland, C. L., Clark, C., Spaggiari, C. V., Smithies, R. H., Evans, N. J., &

879   McDonald, B. J. (2019). Titanite dates crystallization: Slow Pb diffusion during super-

880   solidus re- equilibration. *Journal of metamorphic Geology, 37*(6), 823-838.

881 Hayden, L. A., Watson, E. B., & Wark, D. A. (2008). A thermobarometer for sphene (titanite).

882   *Contributions to Mineralogy and Petrology, 155*(4), 529-540.

883 Hetzel, R., & Romer, R. L. (2000). A moderate exhumation rate for the high- pressure Maksyutov

884   Complex, southern Urals, Russia. *Geological Journal, 35*(3- 4), 327-344.

885 Hogmalm, K. J., Zack, T., Karlsson, A. K.-O., Sjöqvist, A. S., & Garbe-Schönberg, D. (2017). In situ

886   Rb–Sr and K–Ca dating by LA-ICP-MS/MS: an evaluation of N 2 O and SF 6 as reaction

887   gases. *Journal of Analytical Atomic Spectrometry, 32*(2), 305-313.

888 Holland, T., & Powell, R. (1998). An internally consistent thermodynamic data set for phases of

889   petrological interest. *Journal of metamorphic Geology, 16*(3), 309-343.

890 Holland, T., & Powell, R. (2003). Activity–composition relations for phases in petrological

891   calculations: an asymmetric multicomponent formulation. *Contributions to Mineralogy and

892   Petrology, 145*(4), 492-501.

893 Holland, T., & Powell, R. (2011). An improved and extended internally consistent thermodynamic

894 dataset for phases of petrological interest, involving a new equation of state for solids.

895 *Journal of metamorphic Geology*, *29*(3), 333-383.

896 Itaya, T., & Tsujimori, T. (2015). White mica K–Ar geochronology of Sanbagawa eclogites from

897 Southwest Japan: implications for deformation-controlled K–Ar closure temperature.

898 *International Geology Review*, *57*(5-8), 1014-1022.

899 Jenkins, R., Landenberger, B., & Collins, W. (2002). Late Palaeozoic retreating and advancing

900 subduction boundary in the New England fold belt, New South Wales. *Australian Journal of*

901 *Earth Sciences*, *49*(3), 467-489.

902 Jessop, K., Daczko, N., & Piazolo, S. (2019). Tectonic cycles of the New England Orogen, eastern

903 Australia: A review. *Australian Journal of Earth Sciences*, *66*(4), 459-496.

904 Kabir, M., & Takasu, A. (2010). Evidence for multiple burial–partial exhumation cycles from the

905 Onodani eclogites in the Sambagawa metamorphic belt, central Shikoku, Japan. *Journal of*

906 *metamorphic Geology*, *28*(8), 873-893.

907 Kelley, S. (2002). Excess argon in K–Ar and Ar–Ar geochronology. *Chemical Geology*, *188*(1), 1-22.

908 Kemp, A., Hawkesworth, C., Collins, W., Gray, C., & Blevin, P. (2009). Isotopic evidence for rapid

909 continental growth in an extensional accretionary orogen: The Tasmanides, eastern Australia.

910 *Earth and Planetary Science Letters*, *284*(3), 455-466.

911 Kohn, M. J. (2004). Oscillatory- and sector- zoned garnets record cyclic (?) rapid thrusting in central

912 Nepal. *Geochemistry, Geophysics, Geosystems*, *5*(12).

913 Konrad-Schmolke, M., O'Brien, P. J., de Capitani, C., & Carswell, D. A. (2008). Garnet growth at

914 high-and ultra-high pressure conditions and the effect of element fractionation on mineral

915 modes and composition. *Lithos*, *103*(3-4), 309-332.

916 Krebs, M., Maresch, W., Schertl, H.-P., Münker, C., Baumann, A., Draper, G., Idleman, B., & Trapp,

917 E. (2008). The dynamics of intra-oceanic subduction zones: a direct comparison between

918 fossil petrological evidence (Rio San Juan Complex, Dominican Republic) and numerical

919 simulation. *Lithos, 103*(1), 106-137.

920 Krebs, M., Schertl, H.-P., Maresch, W., & Draper, G. (2011). Mass flow in serpentinite-hosted
921   subduction channels: P–T–t path patterns of metamorphic blocks in the Rio San Juan mélange
922   (Dominican Republic). *Journal of Asian Earth Sciences*, *42*(4), 569-595.

923 Lanari, P., & Engi, M. (2017). Local bulk composition effects on metamorphic mineral assemblages.
924   *Reviews in mineralogy and geochemistry*, *83*(1), 55-102.

925 Lázaro, C., García- Casco, A., Rojas Agramonte, Y., Kröner, A., Neubauer, F., & Iturralde- Vinent,
926   M. (2009). Fifty- five- million- year history of oceanic subduction and exhumation at the
927   northern edge of the Caribbean plate (Sierra del Convento mélange, Cuba). *Journal of*
928   *metamorphic Geology*, *27*(1), 19-40.

929 Leake, B. E., Woolley, A. R., Arps, C. E., Birch, W. D., Gilbert, M. C., Grice, J. D., Hawthorne, F.
930   C., Kato, A., Kisch, H. J., & Krivovichev, V. G. (1997). Report. Nomenclature of
931   amphiboles: report of the subcommittee on amphiboles of the international mineralogical
932   association commission on new minerals and mineral names. *Mineralogical Magazine*, *61*(2),
933   295-321.

934 Leitch, E. (1975). Plate tectonic interpretation of the Paleozoic history of the New England Fold Belt.
935   *Geological Society of America Bulletin*, *86*(1), 141-144.

936 Li, J.-L., Klemd, R., Gao, J., & John, T. (2016). Poly-cyclic metamorphic evolution of eclogite:
937   evidence for multistage burial–exhumation cycling in a subduction channel. *Journal of*
938   *Petrology*, *57*(1), 119-146.

939 Li, P., Rosenbaum, G., Yang, J. H., & Hoy, D. (2015). Australian- derived detrital zircons in the
940   Permian- Triassic Gympie terrane (eastern Australia): Evidence for an autochthonous origin.
941   *Tectonics*, *34*(5), 858-874.

942 Liati, A., Theye, T., Fanning, C. M., Gebauer, D., & Rayner, N. (2016). Multiple subduction cycles in
943   the Alpine orogeny, as recorded in single zircon crystals (Rhodope zone, Greece). *Gondwana*
944   *Research*, *29*(1), 199-207.

945 Liferovich, R. P., & Mitchell, R. H. (2005). Composition and paragenesis of Na-, Nb-and Zr-bearing
946   titanite from Khibina, Russia, and crystal-structure data for synthetic analogues. *The*
947   *Canadian Mineralogist*, *43*(2), 795-812.

948 Manton, R. J., Buckman, S., Nutman, A. P., Bennett, V. C., & Belousova, E. A. (2017). U- Pb- Hf-
949   REE- Ti zircon and REE garnet geochemistry of the Cambrian Attunga eclogite, New
950   England Orogen, Australia: Implications for continental growth along eastern Gondwana.
951   *Tectonics, 36*(8), 1580-1613.

952 Marmo, B., Clarke, G., & Powell, R. (2002). Fractionation of bulk rock composition due to
953   porphyroblast growth: effects on eclogite facies mineral equilibria, Pam Peninsula, New
954   Caledonia. *Journal of metamorphic Geology, 20*(1), 151-165.

955 Martin, L. A., Rubatto, D., Brovarone, A. V., & Hermann, J. (2011). Late Eocene lawsonite-eclogite
956   facies metasomatism of a granulite sliver associated to ophiolites in Alpine Corsica. *Lithos,
957   125*(1-2), 620-640.

958 Moresi, L., Betts, P., Miller, M., & Cayley, R. (2014). Dynamics of continental accretion. *Nature,
959   508*(7495), 245-248.

960 Och, D., Leitch, E., Caprarelli, G., & Watanabe, T. (2003). Blueschist and eclogite in tectonic
961   melange, port macquarie, new south wales, australia.

962 Och, D., Zwingmann, H., Philips, G., & Leitch, E. (2010). K-Ar dating of serpentinisation using
963   fuchsite mica: the Rocky Beach metamorphic melange, PORT MACQUARIE.

964 Och, D. J. (2007). *Eclogite, serpentinite, mélange and mafic intrusive rocks: manifestation of long-
965   lived Palaeozoic convergent margin activity, Port Macquarie, eastern Australia*.

966 Park, C., Song, Y., Kang, I.-M., Shim, J., Chung, D., & Park, C.-S. (2017). Metasomatic changes
967   during periodic fluid flux recorded in grandite garnet from the Weondong W-skarn deposit,
968   South Korea. *Chemical Geology, 451*, 135-153.

969 Paton, C., Hellstrom, J., Paul, B., Woodhead, J., & Hergt, J. (2011). Iolite: Freeware for the
970   visualisation and processing of mass spectrometric data. *Journal of Analytical Atomic
971   Spectrometry, 26*(12), 2508-2518.

972 Phillips, D., & Harris, J. (2008). Provenance studies from 40 Ar/39 Ar dating of mineral inclusions in
973   diamonds: Methodological tests on the Orapa kimberlite, Botswana. *Earth and Planetary
974   Science Letters, 274*(1), 169-178.

975 Phillips, G., Landenberger, B., & Belousova, E. (2011). Building the New England Batholith, eastern

976 Australia—Linking granite petrogenesis with geodynamic setting using Hf isotopes in zircon.

977 *Lithos*, *122*(1-2), 1-12.

978 Phillips, G., & Offler, R. (2011). Contrasting modes of eclogite and blueschist exhumation in a

979 retreating subduction system: The Tasmanides, Australia. *Gondwana Research*, *19*(3), 800-

980 811.

981 Phillips, G., Offler, R., Rubatto, D., & Phillips, D. (2015). High- pressure metamorphism in the

982 southern New England Orogen: Implications for long- lived accretionary orogenesis in

983 eastern Australia. *Tectonics*, *34*(9), 1979-2010.

984 Phillips, G., Wilson, C. J., Phillips, D., & Szczepanski, S. K. (2007). Thermochronological

985 (40Ar/39Ar) evidence of Early Palaeozoic basin inversion within the southern Prince Charles

986 Mountains, East Antarctica: implications for East Gondwana. *Journal of the Geological*

987 *Society*, *164*(4), 771-784.

988 Pourteau, A., Scherer, E. E., Schorn, S., Bast, R., Schmidt, A., & Ebert, L. (2019). Thermal evolution

989 of an ancient subduction interface revealed by Lu–Hf garnet geochronology, Halilbağı

990 Complex (Anatolia). *Geoscience Frontiers*, *10*(1), 127-148.

991 Powell, R., & Holland, T. (1988). An internally consistent dataset with uncertainties and correlations:

992 3. Applications to geobarometry, worked examples and a computer program. *Journal of*

993 *metamorphic Geology*, *6*(2), 173-204.

994 Putlitz, B., Cosca, M., & Schumacher, J. (2005). Prograde mica 40Ar/39Ar growth ages recorded in

995 high pressure rocks (Syros, Cyclades, Greece). *Chemical Geology*, *214*(1-2), 79-98.

996 Roberts, J., & James, L. (2010). Stratigraphic relationships of Carboniferous volcanogenic

997 successions in the Clifton–Carroll block and Werrie syncline, northern Tamworth Belt,

998 southern New England Orogen. *Australian Journal of Earth Sciences*, *57*(2), 193-205.

999 Roda, M., Zucali, M., Regorda, A., & Iole Spalla, M. (2019). Formation and evolution of a

1000 subduction-related mélange: The example of the Rocca Canavese Thrust Sheets (Western

1001 Alps). *Geological Society of America Bulletin*.

1002 Rodrıguez, J., Cosca, M., Ibarguchi, J. G., & Dallmeyer, R. (2003). Strain partitioning and

1003 preservation of 40Ar/39Ar ages during Variscan exhumation of a subducted crust (Malpica–

1004 Tui complex, NW Spain). *Lithos*, *70*(3-4), 111-139.

1005 Rubatto, D., Regis, D., Hermann, J., Boston, K., Engi, M., Beltrando, M., & McAlpine, S. R. (2011).

1006 Yo-yo subduction recorded by accessory minerals in the Italian Western Alps. *Nature*

1007 *Geoscience*, *4*(5), 338.

1008 Sano, S., Offler, R., Hyodo, H., & Watanabe, T. (2004). Geochemistry and chronology of tectonic

1009 blocks in serpentinite mélange of the southern New England Fold Belt, NSW, Australia.

1010 *Gondwana Research*, *7*(3), 817-831.

1011 Schumacher, R., Rötzler, K., & Maresch, W. (1999). Subtle oscillatory zoning in garnet from regional

1012 metamorphic phyllites and mica schists, western Erzgebirge, Germany. *The Canadian*

1013 *Mineralogist*, *37*(2), 381-403.

1014 Seifert, W. (2005). REE-, Zr-, and Th-rich titanite and associated accessory minerals from a kersantite

1015 in the Frankenwald, Germany. *Mineralogy and Petrology*, *84*(3-4), 129-146.

1016 Seifert, W., & Kramer, W. (2003). Accessory titanite: an important carrier of zirconium in

1017 lamprophyres. *Lithos*, *71*(1), 81-98.

1018 Shaw, S., & Flood, R. (1981). The New England Batholith, eastern Australia: geochemical variations

1019 in time and space. *Journal of Geophysical Research: Solid Earth*, *86*(B11), 10530-10544.

1020 Shreve, R. L., & Cloos, M. (1986). Dynamics of sediment subduction, melange formation, and prism

1021 accretion. *Journal of Geophysical Research: Solid Earth*, *91*(B10), 10229-10245.

1022 Spandler, C., Hammerli, J., Sha, P., Hilbert-Wolf, H., Hu, Y., Roberts, E., & Schmitz, M. (2016).

1023 MKED1: a new titanite standard for in situ analysis of Sm–Nd isotopes and U–Pb

1024 geochronology. *Chemical Geology*, *425*, 110-126.

1025 Spear, F. S. (1988). Metamorphic fractional crystallization and internal metasomatism by diffusional

1026 homogenization of zoned garnets. *Contributions to Mineralogy and Petrology*, *99*(4), 507-

1027 517.

1028   Spencer, K., Hacker, B., Kylander-Clark, A., Andersen, T., Cottle, J., Stearns, M., Poletti, J., &

1029   Seward, G. (2013). Campaign-style titanite U–Pb dating by laser-ablation ICP: Implications

1030   for crustal flow, phase transformations and titanite closure. *Chemical Geology*, *341*, 84-101.

1031   Stöckhert, B., & Gerya, T. V. (2005). Pre- collisional high pressure metamorphism and nappe

1032   tectonics at active continental margins: A numerical simulation. *Terra Nova*, *17*(2), 102-110.

1033   Tamblyn, R., Hand, M., Kelsey, D., Anczkiewicz, R., & Och, D. (2019). Subduction and

1034   accumulation of lawsonite eclogite and garnet blueschist in eastern Australia. *Journal of*

1035   *metamorphic Geology*.

1036   Tamblyn, R., Zack, T., Schmitt, A., Hand, M., Kelsey, D., Morrissey, L., Pabst, S., & Savov, I.

1037   (2019). Blueschist from the Mariana forearc records long-lived residence of material in the

1038   subduction channel. *Earth and Planetary Science Letters*, *519*, 171-181.

1039   Tsujimori, T., & Ernst, W. (2014). Lawsonite blueschists and lawsonite eclogites as proxies for

1040   palaeo- subduction zone processes: a review. *Journal of metamorphic Geology*, *32*(5), 437-

1041   454.

1042   Tsujimori, T., Sisson, V. B., Liou, J. G., Harlow, G. E., & Sorensen, S. S. (2006). Very-low-

1043   temperature record of the subduction process: A review of worldwide lawsonite eclogites.

1044   *Lithos*, *92*(3), 609-624.

1045   Vermeesch, P. (2018). IsoplotR: a free and open toolbox for geochronology. *Geoscience Frontiers*,

1046   *9*(5), 1479-1493.

1047   Viete, D. R., Hacker, B. R., Allen, M. B., Seward, G. G., Tobin, M. J., Kelley, C. S., Cinque, G., &

1048   Duckworth, A. R. (2018). Metamorphic records of multiple seismic cycles during subduction.

1049   *Science advances*, *4*(3), eaaq0234.

1050   Vuorinen, J. H., & Hålenius, U. (2005). Nb-, Zr-and LREE-rich titanite from the Alnö alkaline

1051   complex: crystal chemistry and its importance as a petrogenetic indicator. *Lithos*, *83*(1-2),

1052   128-142.

1053   Warren, C., Sherlock, S., & Kelley, S. (2011). Interpreting high-pressure phengite 40Ar/39Ar

1054   laserprobe ages: an example from Saih Hatat, NE Oman. *Contributions to Mineralogy and*

1055   *Petrology*, *161*(6), 991-1009.

Wei, C., & Clarke, G. (2011). Calculated phase equilibria for MORB compositions: a reappraisal of

the metamorphic evolution of lawsonite eclogite. *Journal of metamorphic Geology, 29*(9),

939-952.

White, R., Powell, R., & Holland, T. (2007). Progress relating to calculation of partial melting

equilibria for metapelites. *Journal of metamorphic Geology, 25*(5), 511-527.

Woodhead, J. D., & Hergt, J. M. (2001). Strontium, neodymium and lead isotope analyses of NIST

glass certified reference materials: SRM 610, 612, 614. *Geostandards Newsletter, 25*(2- 3),

261-266.

Zack, T., & Hogmalm, K. J. (2016). Laser ablation Rb/Sr dating by online chemical separation of Rb

and Sr in an oxygen-filled reaction cell. *Chemical Geology, 437*, 120-133.

Declarations

Conflicts of interest/competing interests

The authors declare no conflicts of interest or competing interests

Availability of data and material

All data presented in this manuscript can be found in the supplementary materials

Code availability

Not applicable

1084

1085

1086 Figure 1: a) Simplified map of eastern Australia (east of the Tasman line) showing the main eastward-

1087 younging orogens. The SNEO is indicated in the black box. b) Geological map of the SNEO, showing

1088 other high-pressure rock localities and indicating the location of Port Macquarie in the black box.

1089 PMFS: Peel Manning Fault System. c) Geological map of Rocky Beach at Port Macquarie, showing

1090 the high-pressure mélange encased in serpentinite, modified from Och et al. (2003).

1091

1092 Figure 2: Mineral textural relationships in the Port Macquarie eclogite. a) Eclogite in outcrop,

1093 showing coarse-grained garnet, phengite and lawsonite with patches of omphacite, which is partially

1094 retrogressed by glaucophane. b) Photomicrograph showing the porphyroblastic nature of garnet,

1095 lawsonite and omphacite, in a finer-grained phengite and titanite matrix. c) Stage 2 retrograde features

1096 of the eclogite, showing chlorite replacement of garnet and glaucophane replacement of omphacite. d)

1097 Garnet 2 enclosing a pre-existing garnet 1 porphyroblast. Boundary between the stage 1 garnet and

1098 stage 2 garnet is shown by a white dashed line. The boundary between the two garnet domains

1099 contains entrapped chlorite and omphacite inclusions.. e) Radial fractures surrounding a quartz

1100 inclusion in the outermost part garnet 1. f) Large lawsonite porphyroblast, with an inclusion-rich

1101 euhedral crystal (lawsonite 1) which is overgrown by a rim (lawsonite 2). The new lawsonite growth

1102 contains inclusions of garnet 2 that developed on the margin of the older lawsonite, as well as stage 2

1103 glaucophane and titanite. Abbreviations: O: Omphacite; Law: Lawsonite; Phe: Phengite; Ttn: Titanite;

1104 G: Garnet; Chl: Chlorite; Gl: Glaucophane; Stlp: Stilpnomelane; Q: Quartz.

1105

1106 Figure 3: Schematic interpretation of the petrologic evolution of the lawsonite eclogite. a) Prograde–

1107 peak stage 1 burial: garnet cores nucleate and as garnet 1 grows it entraps prograde mineralogy and

1108 inclusion trails. Lawsonite porphyroblasts grow and entrap prograde minerals. b) Retrograde 1: garnet

1109 is partially consumed by chlorite, and glaucophane forms at the expense of omphacite. c) Prograde–

1110 peak 2: garnet forms new rims (garnet 2). In rare instances stage 2 garnet overgrows stage 1

1111 lawsonite. Phengite recrystallises in the matrix with coarse-grained titanite during deformation. d)
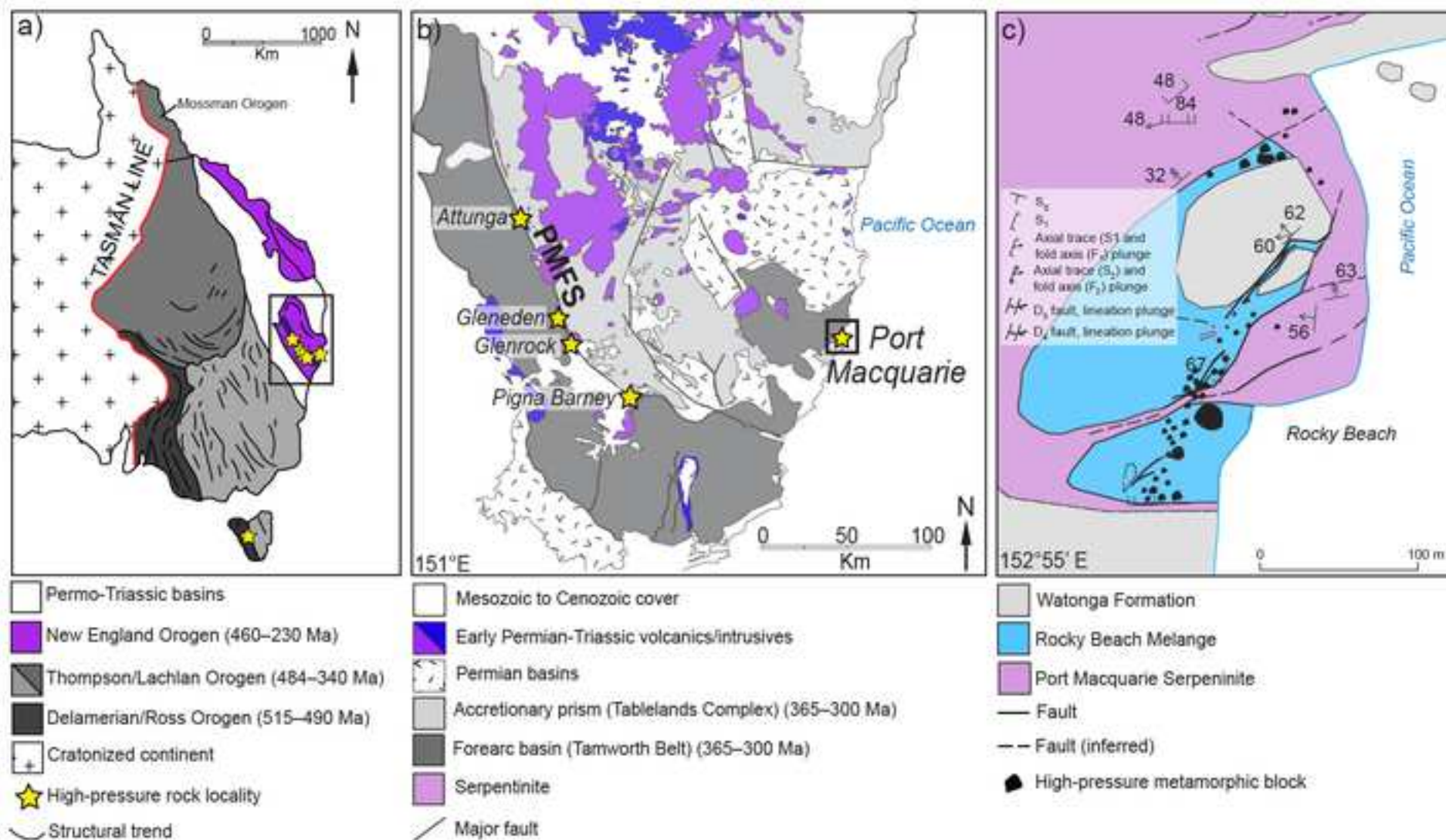
1112 Stage 2 retrograde: garnet is partially replaced by chlorite and omphacite is partially replaced by

1113 glaucophane. There is minor growth of lawsonite.

1114

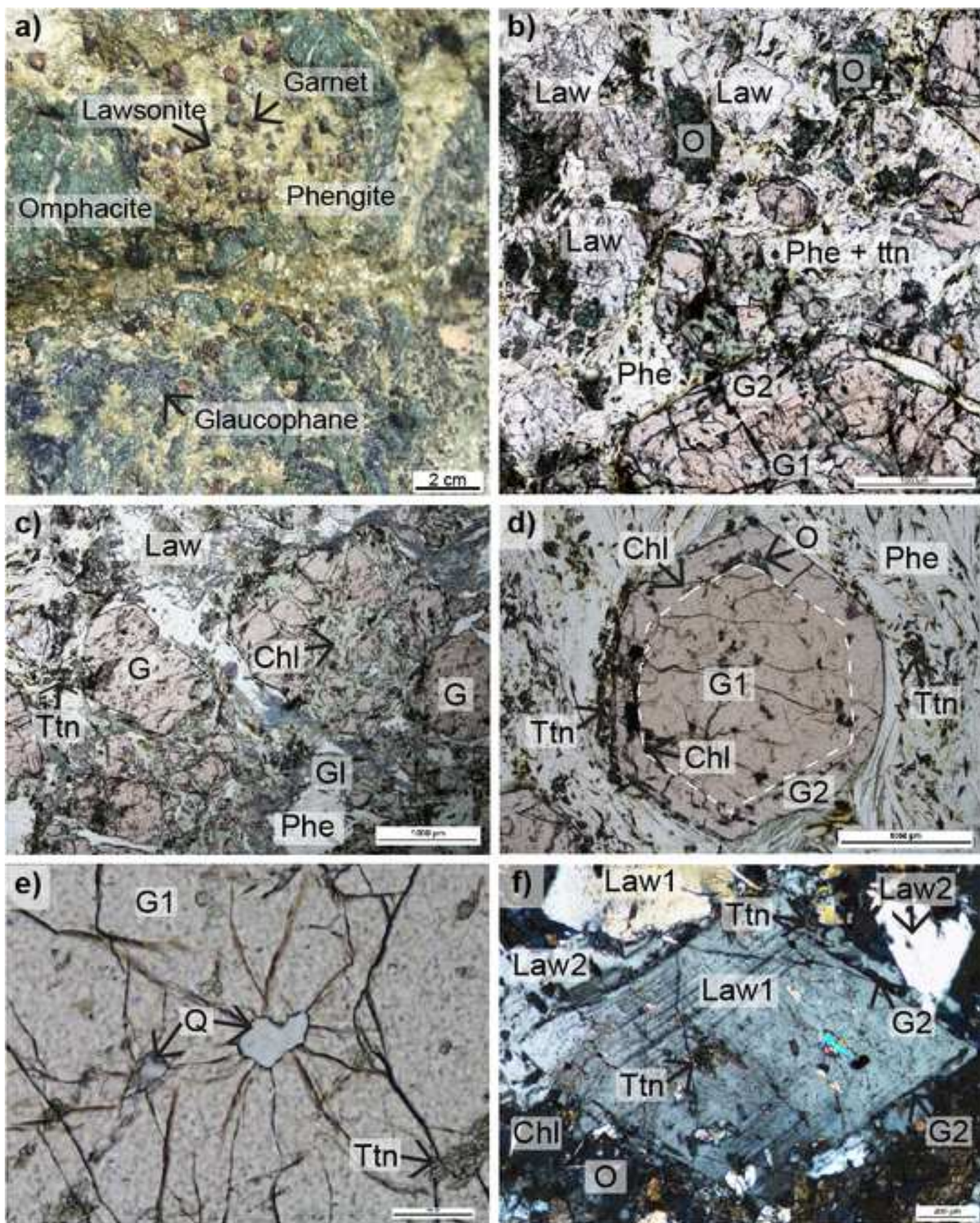1115 Figure 4: Representative EPMA traverse of garnet, showing the proportions of end-members and X_{Fe}.

1116 a) Traverse across entire garnet porphyroblast, showing a prograde bell-shaped zoning profile. b)

1117 Traverse across stage 1 garnet to stage 2 garnet, showing a step decrease in X_{Fe} and increase in

1118 spessartine content.

1119

1120 Figure 5: EPMA element maps of garnet, the white dashed line shows the interpreted boundary

1121 between garnet 1 and garnet 2. a) Garnet porphyroblasts with prograde zoning in Mn, along with

1122 several oscillatory rings in the stage 2 rim. Zoning in Fe and Mg show garnet 2 as a marked outer ring

1123 with elevated X_{Mg}. Subset maps are marked in grey boxes. b) Compositional oscillations in the garnet

1124 2 which define smaller hexagonal garnet crystals that nucleated on the older garnet 1 rim. c) Complex

1125 Mn zoning patterns and a marked X_{Mg} increase in the outer garnet rim. Embayments and subsequent

1126 overgrowths are highlighted in the Mn oscillations, with correlate with Ca oscillations. Late chlorite

1127 replaces garnet. The boundary between garnet 1 and garnet 2 contains Mn-rich chlorite inclusions.

1128

1129 Figure 6: EPMA maps across transition from garnet 1 to garnet 2 and into matrix phengite.

1130 White/black dashed line indicates the boundaries of garnet 1 and garnet 2. Solid white/black line

1131 indicates the boundaries of phengite included in garnet.

1132

1133 Figure 7: Mineral compositions of omphacite, phengite, chlorite and titanite measured by EPMA. a)

1134 Omphacite analyses from inclusions in garnet and omphacite porphyroblasts in the matrix. b)

1135 Phengite analyses showing increasing Si, Fe and Mg component from core to rim, and decreasing Na

1136 content. c) Chlorite analyses from stage 1 chlorite, from the boundary between garnet 1 and garnet 2,

1137 and stage 2 chlorite, from the matrix. d) Titanite analyses showing elevated Al_2O_3, FeO and F
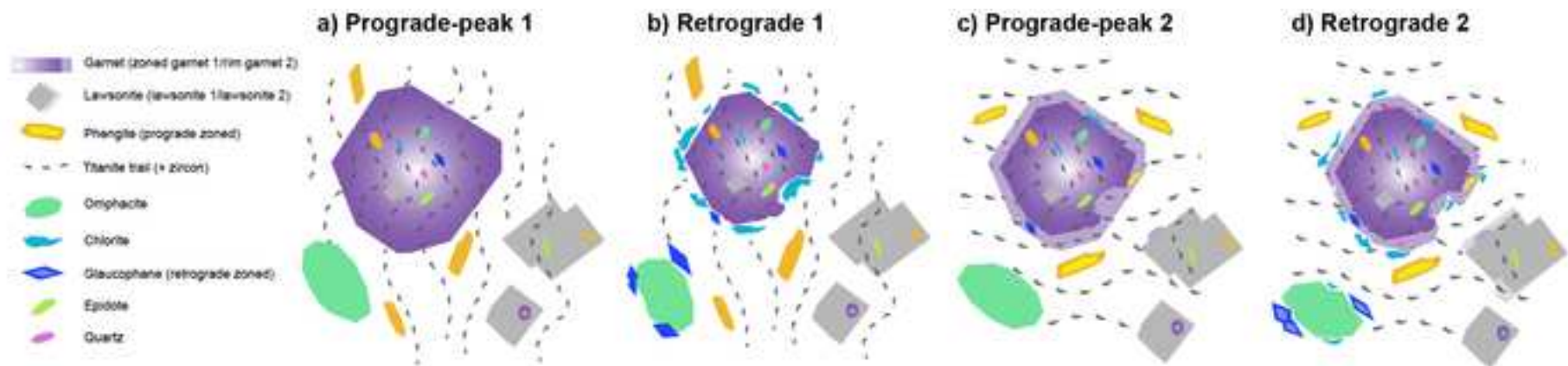
1138 contents with a positive correlation, suggesting increasing pressure during crystallization (e.g. Castelli

1139 & Rubatto, 2002).

1140

1141 Figure 8: a) Phase equilibrium model calculated with THERMOCALC, modified from Tamblyn et al.

1142 (2019a), calculated with garnet 1 and lawsonite 2 included in the bulk rock chemistry. Dashed thin

1143 grey lines indicate the position of solvi. Colour changes indicate changes in variance across fields. b)

1144 Model (a) showing the prograde assemblage and estimated garnet mode at stage 1 peak. Thick grey

1145 arrow indicates the interpreted *P–T* path of stage 1 prograde-peak. Abbreviations: Act: Actinolite, Bi:

1146 Biotite, Chl: Chlorite, Coe: Coesite, Ep: Epidote, G: Garnet, Gl: Glaucophane, Hb: Hornblende, Jd:

1147 Jadeite, Law: Lawsonite, O: Omphacite, Pa: Paragonite, Phe: Phengite, Pl: Plagioclase, Q: Quartz, Ta:

1148 Talc.

1149

1150 Figure 9: a) Phase equilibrium model calculated with THERMOCALC for the local rock composition

1151 with garnet 1 and lawsonite 1 removed, and water in excess. Dashed thin grey lines indicate the

1152 position of solvi. Colour changes indicate changes in variance across fields. Note the scale change

1153 from Figure 8. b) Model (a) showing the interpreted peak conditions of stage 1 peak from Figure 8b,

1154 and the interpreted stage 1 retrograde path in the thick grey arrow. Note these retrograde *P–T*

1155 conditions are considered the maximum possible pressures and temperatures obtained during

1156 retrogression. c) Model (a) showing the interpreted stage 2 prograde and peak assemblages, and

1157 mineral compositional isopleths. Thick grey arrow indicates the interpreted *P–T* path during stage 2

1158 prograde to peak. Abbreviations are listed in Figure 8.

1159

1160 Figure 10: a) Phase equilibrium model calculated with THERMOCALC for the local rock

1161 composition with garnet 1 and lawsonite 1 removed, and set water content. Dashed thin grey lines

1162 indicate the position of solvi. Colour changes indicate changes in variance across fields. Note the

1163 scale change from Figure 8. b) Model (a) showing a possible alternative path for stage 2 prograde,

1164 based on compositional isopleths from garnet 1, omphacite and phengite cores. Peak conditions
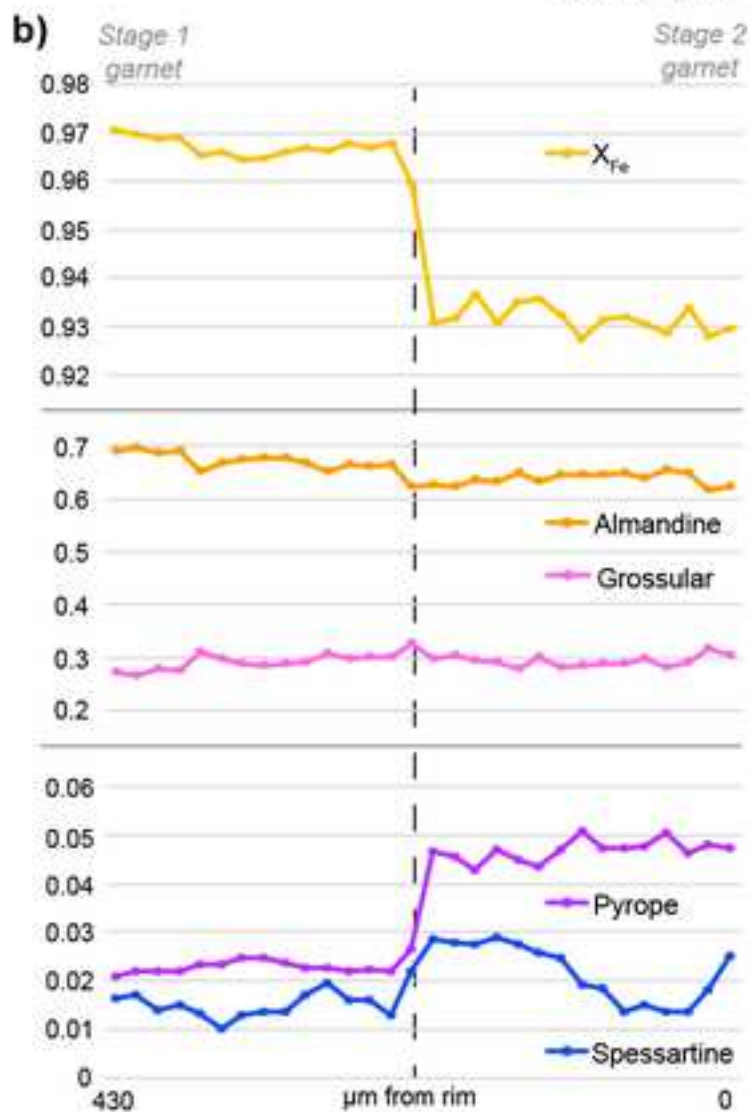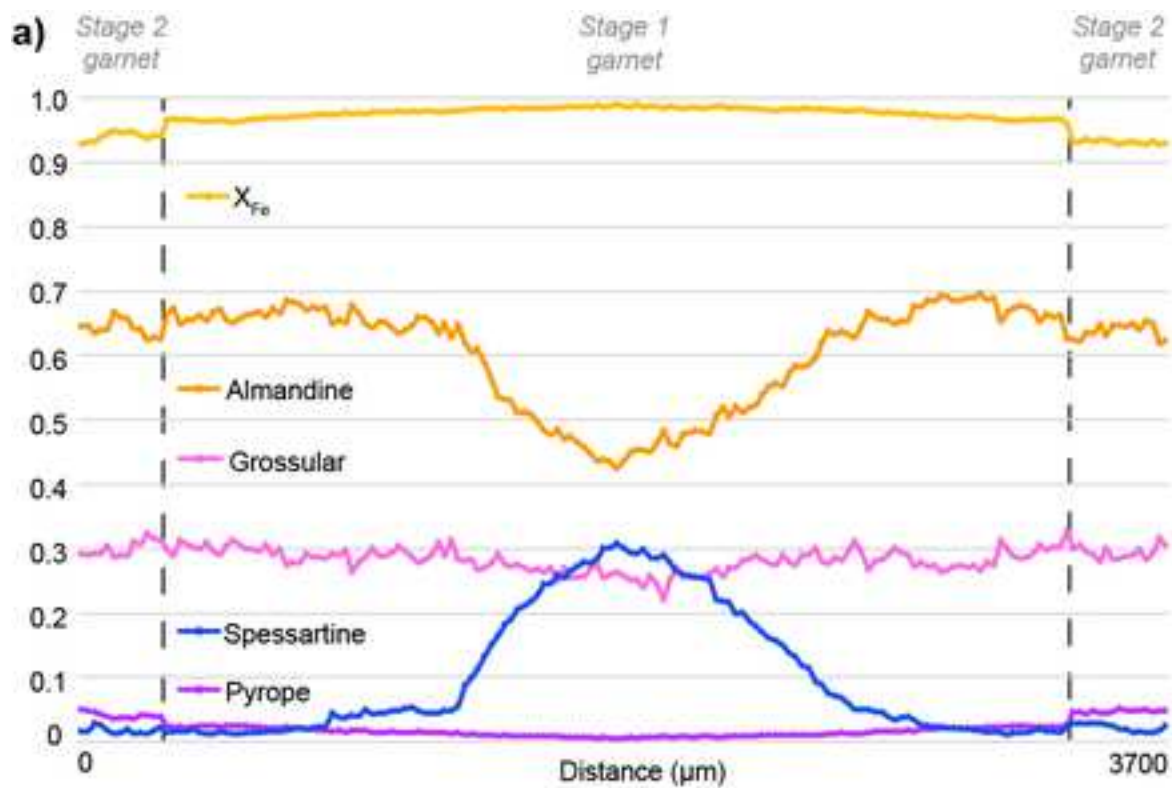
1165 remain the same. c) Model (a) showing the final stage 2 retrograde conditions reached, based on

1166 modal proportion isopleths. Abbreviations are listed in Figure 8.

1167

1168 Figure 11: Results of LA–ICP–MS U–Pb titanite geochronology and representative BSE images of

1169 analysed titanite grains. a) Terra-Wasserburg plot . b) Coarse-grained titanite in the rim of a stage 2

1170 lawsonite crystal (outlined in white dotted line) displaying a well-formed mesh-like intergrowth of

1171 exsolved zircon. c) Coarse titanite trails outlined in purple in the rims of stage 2 garnet, which is

1172 distinguished from the stage 1 garnet core by the black dotted line. d) Titanite grains rim a prograde

1173 zoned phengite, which has darker cores and brighter rims.

1174

1175 Figure 12: Step-heating Ar–Ar results from two phengite grains in the lawsonite eclogite.

1176

1177 Figure 13: In-situ LA–ICP–MS Rb–Sr geochronological results. a) Isochron constructed from

1178 analyses of phengite rims. b) Isochron constructed from analyses of phengite cores. c) Isochron using

1179 both phengite cores and rims. The isochrons produce similar ages within the uncertainties. d) Isotopic

1180 analyses of texturally intergrown titanite, which records the initial $^{87}Sr/^{86}Sr$ isotopic ratio. e) Isochron

1181 constructed from phengite rims and anchored with titanite analyses.

1182

1183 Figure 14: LA–ICP–MS Rb and Sr maps of phengite from the lawsonite eclogite. Maps are semi-

1184 quantitative. Phengites show Sr enrichments of all isotopes in their cores, and sometimes a slight Rb

1185 enrichment in their rims. These enrichments follow along cleavage planes of the phengite. The

1186 boundaries between these rims and cores are extremely sharp and well-defined. The Sr-rich mineral is

1187 titanite. The outline of phengite grains is shown in white. Rb-rich material without obvious grain

1188 boundaries traced is fine-grained phengite.

1189

1190 Figure 15: Summary of the *P–T–t* path of the lawsonite eclogite. Lu–Hf geochronology from

1191 lawsonite and garnet dates an earlier event during burial 1. Ar–Ar and Rb–Sr data from phengite, and
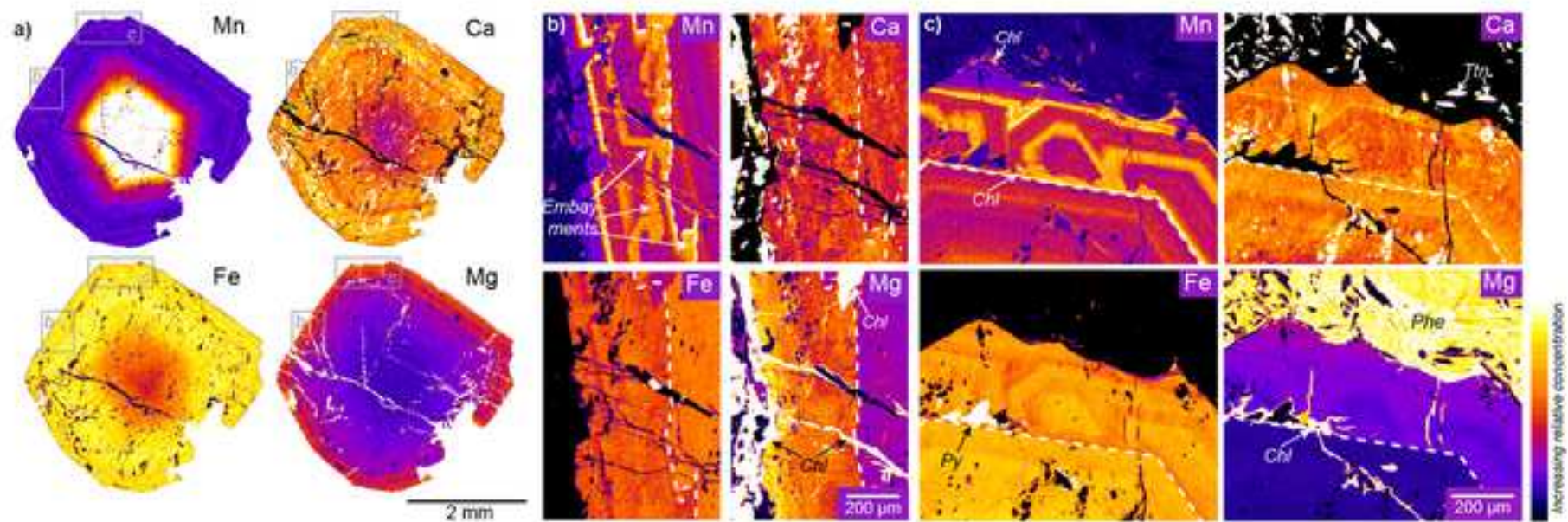
U–Pb data from titanite, date a younger recrystallization event associated with reburial of the

lawsonite eclogite within the subduction channel. This results in two hairpin P–T loops. While the

exact timing of exhumation and P–T conditions reached between these two burial events is unknown,

it must have occurred between approximately 485–465 Ma.

Figure 1

Click here to access/download;Figure;Fig 1.jpg ±

a)

1000 N
Km

Mossman Orogen

TASMAN LINE

PMFS

Permo-Triassic basins

New England Orogen (460–230 Ma)

Thompson/Lachlan Orogen (484–340 Ma)

Delamerian/Ross Orogen (515–490 Ma)

Cratonized continent

High-pressure rock locality

Structural trend

b)

Attunga

PMFS

Gleneden

Glenrock

Pigna Barney

Pacific Ocean

Port Macquarie

151°E

0   50   100   N
Km

Mesozoic to Cenozoic cover

Early Permian-Triassic volcanics/intrusives

Permian basins

Accretionary prism (Tablelands Complex) (365–300 Ma)

Forearc basin (Tamworth Belt) (365–300 Ma)

Serpentinite

Major fault

c)

48
84
48

32

62
60
63
56
67

S2
S3
Axial trace (S1 and fold axis (F3) plunge
Axial trace (S2) and fold axis (F2) plunge
D3 fault, lineation plunge
D4 fault, lineation plunge

Pacific Ocean

Rocky Beach

152°55' E

0   100 m

Watonga Formation

Rocky Beach Melange

Port Macquarie Serpeninite

Fault

Fault (inferred)

High-pressure metamorphic block

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

Figure 3

a) Prograde-peak 1   b) Retrograde 1   c) Prograde-peak 2   d) Retrograde 2

Garnet (zoned garnet 1/rim garnet 2)
Lawsonite (lawsonite 1/lawsonite 2)
Phengite (prograde zoned)
Titanite trail (+ zircon)
Omphacite
Chlorite
Glaucophane (retrograde zoned)
Epidote
Quartz

Figure 4

Click here to access/download;Figure;Fig 4.jpg

Figure 5

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

Figure 6

Figure 7

Click here to access/download;Figure;Fig 7.jpg ⬇

Figure 7

Figure 8   Click here to access/download;Figure;Fig 8.jpg ⬇

Figure 8

Figure 9

Figure 10   Click here to access/download;Figure;Fig 10.jpg

Figure 11

Click here to access/download;Figure;Fig 11.jpg

Figure 12

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Figure 13

Click here to access/download;Figure;Fig 13.jpg ⬇

Figure 13

Figure 14

Figure 15

Click here to access/download;Figure;Fig 15.jpg ⬇

Figure 15

a) Band contrast
b) Phase map
c) Mn
d) Mg
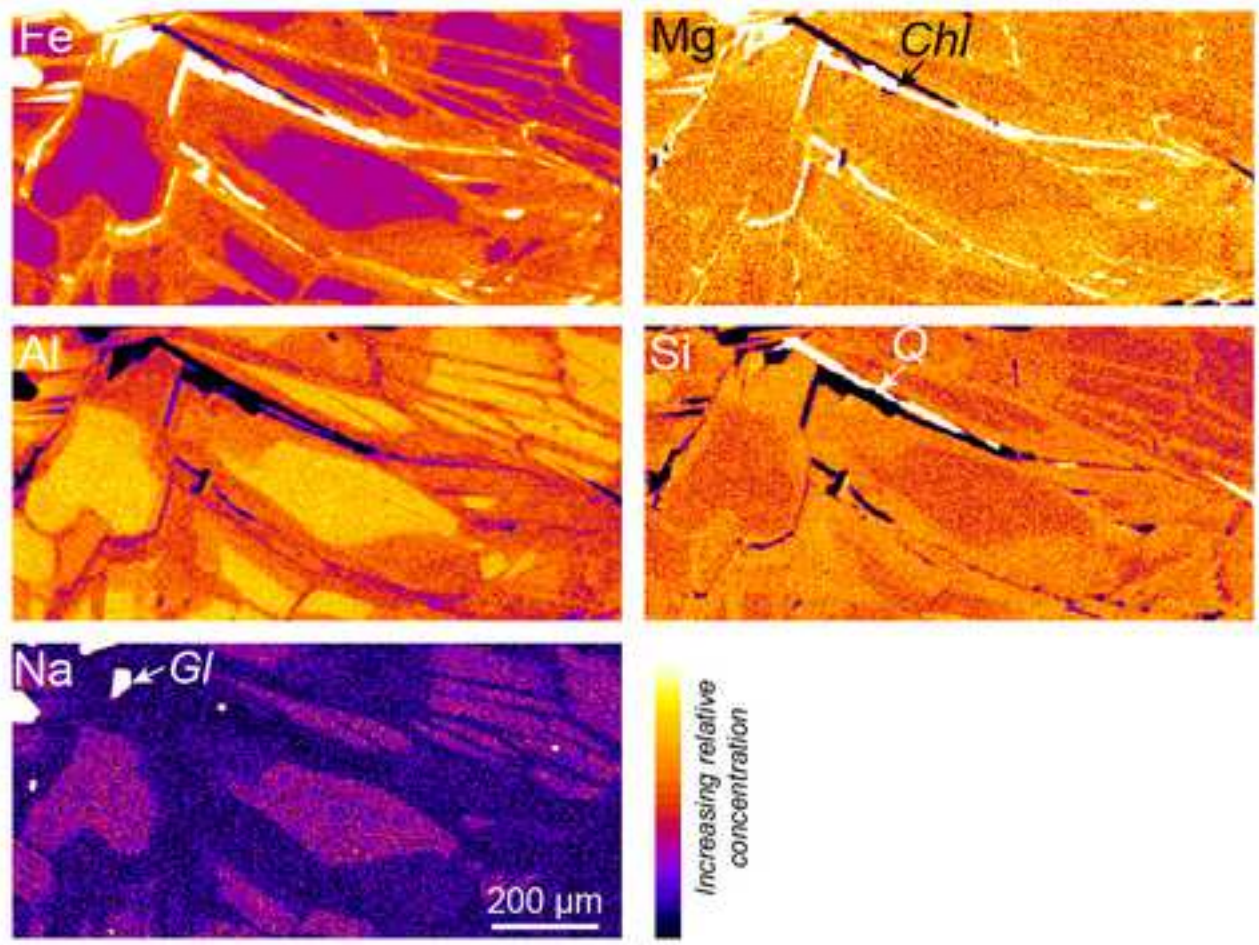e) Interpreted inclusion assemblages
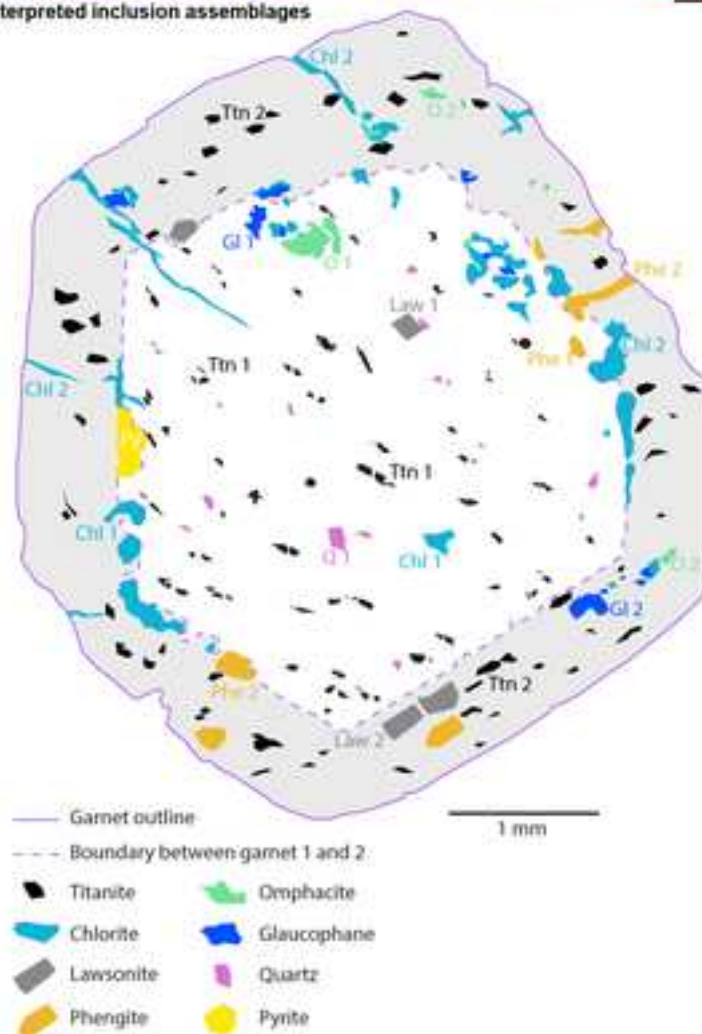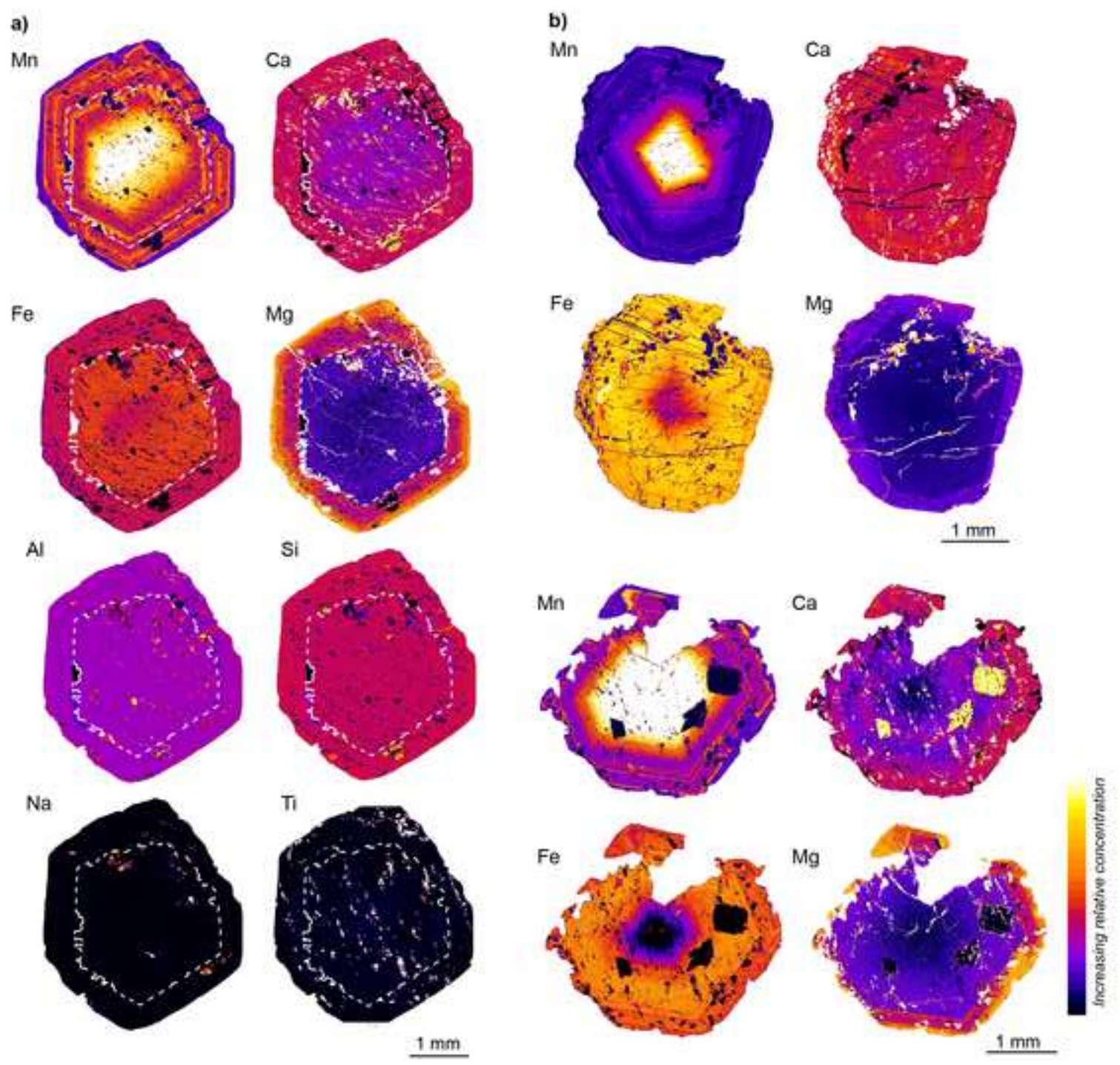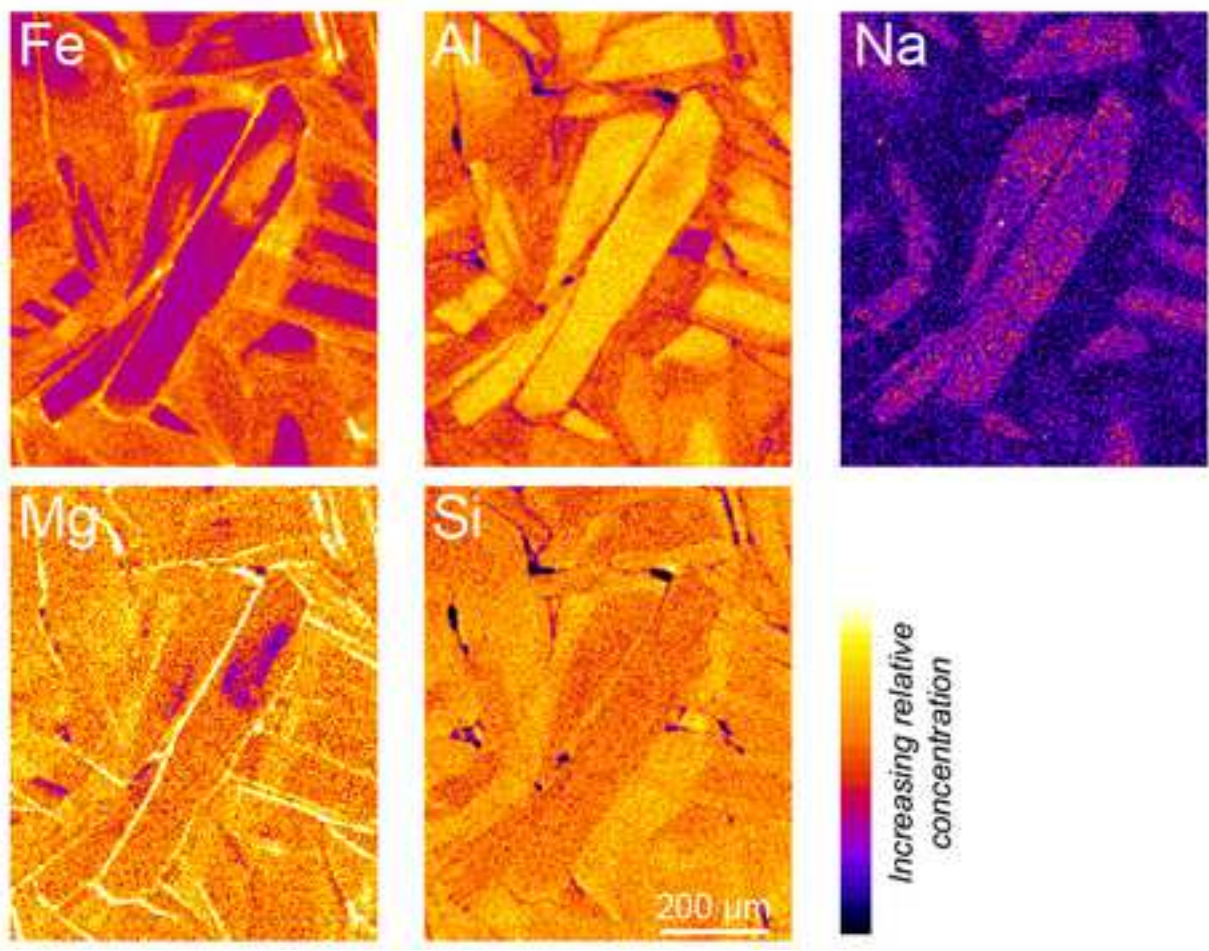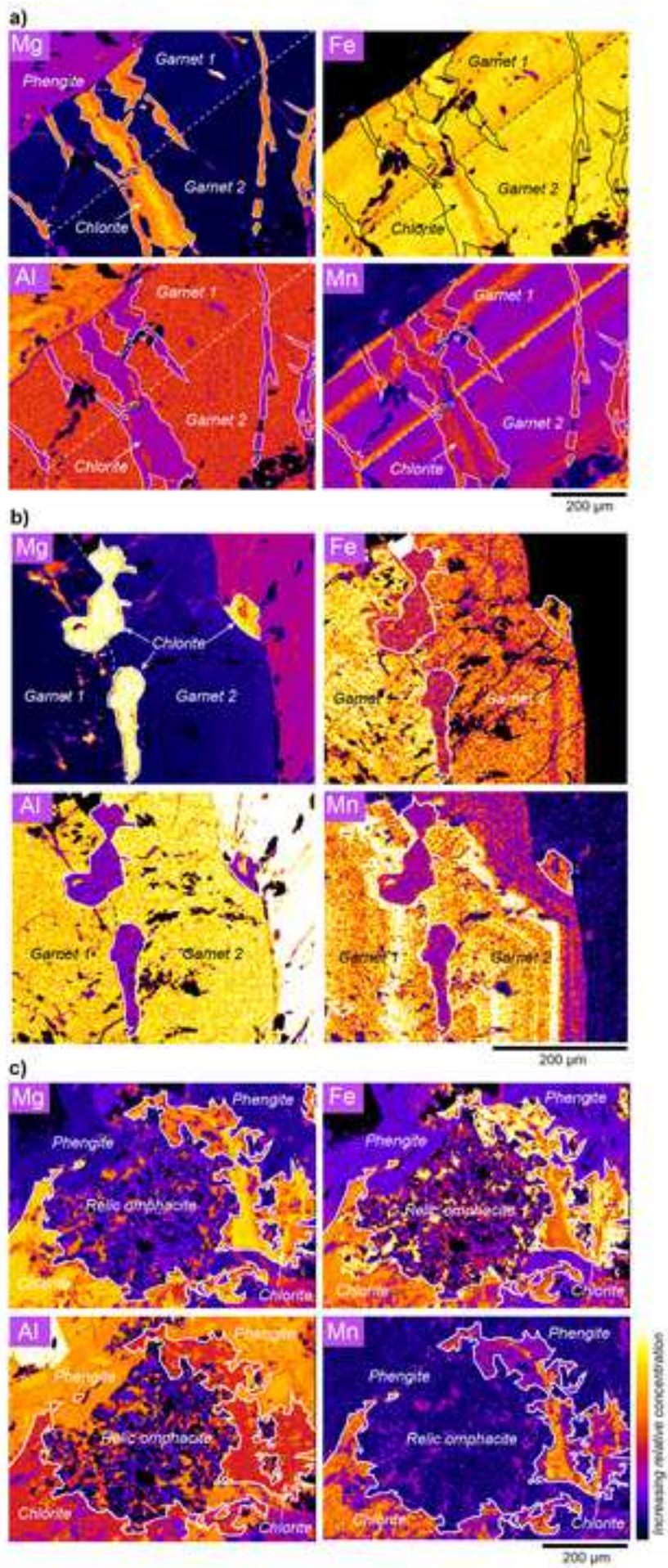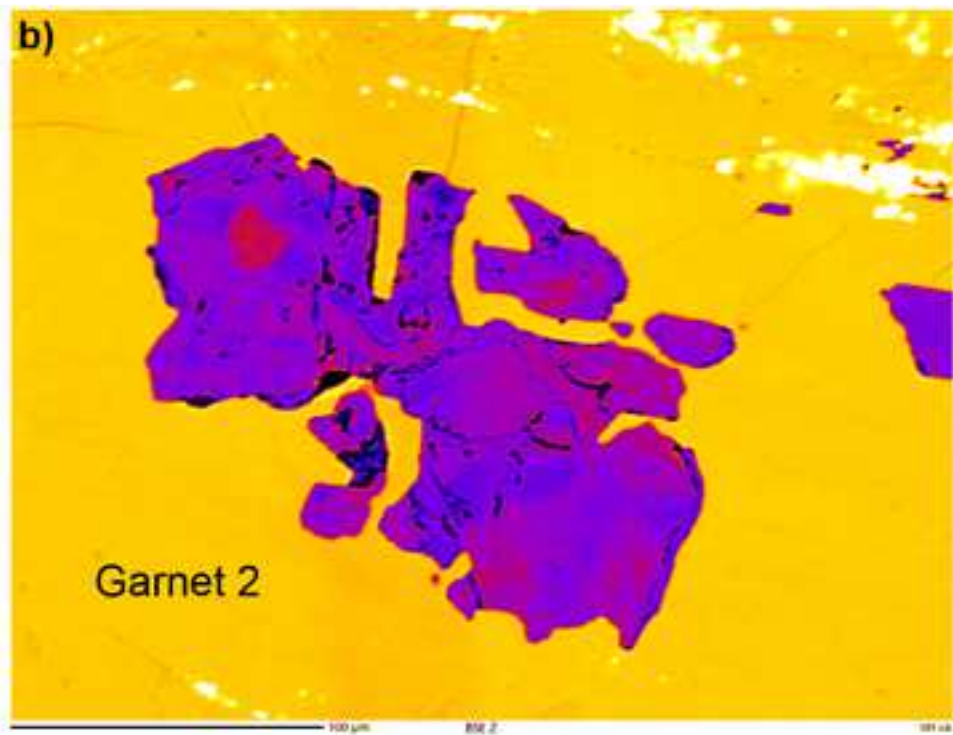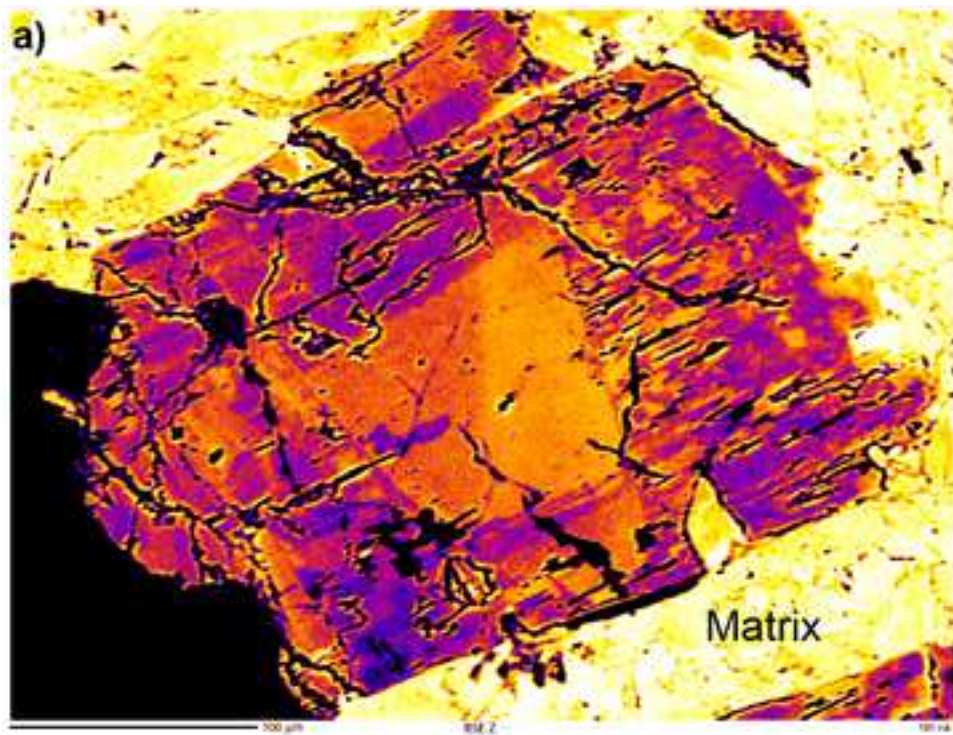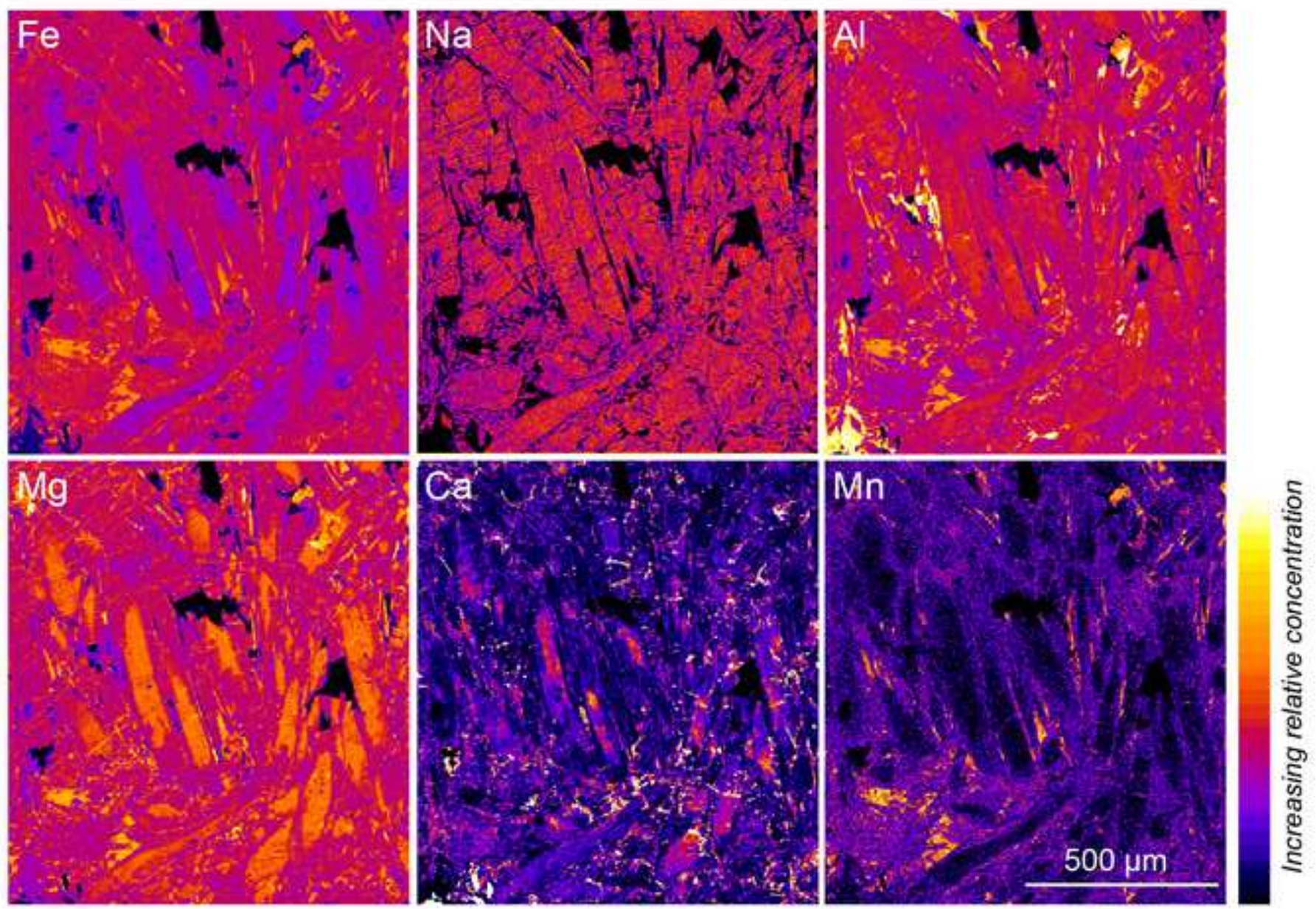
Supplementary File 1 and supplementary figure captions:

Phase mapping methods

To identify inclusions assemblages in garnet, phase maps were constructed in the Tescan Integrated Mineral Analyser (TIMA) software using data collected on a Tescan Mira3 TIMA with four PulsTor SDD X-ray detectors at the John de Laeter Center, Curtin University. Analysis was conducted in dot-mapping mode with 1 µm BSE pixel spacing. EDS analyses of 1000 counts were acquired at a regular spacing of 5 µm, and additionally at every change in BSE contrast. TIMA software was used to compare BSE intensity and EDS peaks to a custom mineral reference library and automatically classify individual mineral grains. Analytical conditions were an accelerating voltage of 25 kV, beam intensity of 19, probe current of 5.48 nA, spot size of 80 nm, and a working distance of 15 mm.

EBSD map data were collected using a Tescan Mira3 VP-FESEM with Oxford Instruments X-Max 150 SDD X-ray detector, Symmetry EBSD detector and Aztec Synergy software at the John de Laeter Center, Curtin University. Analytical conditions were an accelerating voltage of 20 kV, beam intensity of 18, and working distance of 25 mm. Stage tilt was 70° and step size was 1.22 µm. The EBSD camera was set to binning mode "Speed 1", camera exposure time was 10 ms, and the camera gain was 2. Auto background correction was used with a Hough Resolution of 40, band detection mode was "Centers", number of band detected was 11, and indexing mode was "Optimized - TDK". Band contrast and EDS maps were exported from Aztec and EDS maps were colour scaled in ImageJ.

Extended phase equilibrium modelling methods

The phase equilibria model was calculated using THERMOCALC (TC340i, Powell and Holland, 1988; Holland and Powell, 2011) using the internally-consistent thermodynamic dataset 'ds5' (filename tc-ds55.txt; November 2003 updated version of the Holland & Powell, 1998 data set) and activity–composition (a–x) models (Supp. Table 2; Holland and Powell, 1988; Holland and

Powell, 2003; White et al., 2007; Green et al., 2007; Holland and Powell, 2011; Diener et al., 2012) in the chemical system NCKFMASHO. The latest thermodynamic dataset ds62 (Green et al., 2016) was not used, to allow direct comparison with the phase equilibria models of Tamblyn et al. (2019a). The Mn-rich interiors of prograde-zoned garnet were not included in the calculated local rock composition for the stage 2 mineral assemblage, so Mn was not considered in the modelled system. Ti was also not considered, due to limitations in modelling the stability of titanite at high pressure. Titanite has been reported to be the stable Ti-bearing phase rather than rutile in systems due to the low activity of CO_2, or minor amounts of Al and Fe^{3+} in its crystal lattice (Enami et al., 1993; Castelli & Rubatto; 2002: Brovarone et al., 2011). As we did not investigate rutile/titanite stability as a function of X_{CO2}, and the current a–x model does not allow Al or Fe^{3+} substitution in titanite, we did not include it in the calculation of bulk compositions or the phase equilibria models. Stilpnomelane was also not predicted in the model as currently there is no a–x model available for it.

Water content was handled in two ways. A model for the local rock composition was calculated with water in excess, and a model for the local rock composition was calculated with a set water content, derived from the abundance of hydrous minerals (Table 2; Supp. Table 1). This was done to approximate the two 'end-members' of hydration the eclogite may have experienced, with the water content calculated from the abundance of hydrous minerals representing the minimum amount of water available in the rock system. We acknowledge this assumes essentially closed-system behaviour, i.e. does not model water flux into the rock from an external source. As we have no indication that this process occurred, we did not model such behaviour.

Oxidation state (Fe_2O_3) for the sample was constrained from the measured microprobe chemical analyses from each mineral used to calculate the metamorphic local rock composition, by using mineral stoichiometry in the calculation of cations from raw weight % oxides (Table 2; Supp. Table 1; Droop, 1987).

Extended Rb–Sr LA-QQQ-MS methods

Phengite was analysed in-situ for Rb and Sr isotopes and major and trace elements using an ESI 213 NWR (TwoVol2) laser ablation system with an attached Agilent 8800QQQ ICP-MS/MS in the Microgeochemistry Laboratory at the University of Gothenburg, Sweden, following the methods of Zack & Hogmalm (2016) and Hogmalm et al. (2017). Laser parameters were a fluence of $\sim$ 6 Jcm^{-2} and a frequency of 10 Hz. Phengite was analysed with a spot size of 40 μm, both perpendicular and parallel to cleavage. Titanite was also analysed to obtain initial $^{87}Sr/^{86}Sr$ ratios, with a spot size of 50 μm. The ICP-MS/MS allows separating $^{87}Sr+$ from $^{87}Rb+$ where Sr+ reacts with a reaction gas (here N_2O) to form SrO+ while Rb+ does not react. Complete separation is achieved by first allowing only one specific M/Z through the first quadrupole (Q1) into the reaction cell and then collect the reacted ion through the second quadrupole (Q2). In detail, to measure ^{86}Sr, Q1 is set to M/Z 86 and Q2 to M/Z 102, while for ^{87}Sr, Q1 is 87 and Q2 is 103. Since not all Sr reacts with N_2O (only about 80%; Hogmalm et al., 2017), ^{87}Rb cannot be analysed interference-free. Instead, ^{85}Rb can be used as a proxy for ^{87}Rb, as $^{87}Rb/^{85}Rb$ can be assumed to be constant on Earth within the precisions required for this approach. As Rb+ does not react with N2O, Q1 is set to M/Z 85 and Q2 is set to M/Z 85 as well. A range of other isotopes were collected during each analysis together with ^{85}Rb, ^{86}Sr and ^{87}Sr in order to detect and exclude contamination by other minerals, such as: ^{23}Na, ^{24}Mg, ^{27}Al, ^{28}Si, ^{39}K, ^{43}Ca, ^{48}Ti, ^{55}Mn, ^{56}Fe, ^{89}Y, ^{133}Cs, ^{137}Ba, and ^{140}Ce. Laser parameters chosen were a fluence of $\sim$ 6 Jcm^{-2} and a frequency of 10 Hz.

Supplementary Figure captions:

Supplementary Figure 1: Phase map and corresponding images showing an example of inclusion assemblages from garnet porphyroblast in Figure 2d. White dashed line indicates the boundary between garnet 1 and garnet 2. a) Band contrast image, showing that there are no cracks or fractures leading to the minerals included on the boundary of garnet 1 and garnet 2. b) Phase map, showing chlorite 1 trapped at the boundary of garnet 1 and garnet 2, and chlorite 2 cross-cutting the garnet porphyroblast. Phengite, omphacite, titanite, quartz and pyrite are also trapped at the boundary of garnet 1 and garnet 2. c) Mn EDS map of the garnet porphyroblast, showing the boundary between

garnet 1 and garnet 2 but also deflections in the Mn zoning around inclusions, indicating they were trapped during garnet growth. d) Mg EDS map, confirming the high-Mg chlorite grains, and also showing the boundary between garnet 1 and garnet 2. e) Schematic diagram of the inclusion assemblages in garnet, traced from phase maps and EPMA compositional maps (Supp. Fig. 1; 2). Chlorite 1 is from both stage 1 prograde and stage 1 retrograde. See figure 2 for abbreviations.

Supplementary Figure 2: Additional EPMA element maps of garnets from lawsonite eclogite. a) Maps of a different garnet that was used to trace the inclusion assemblages in Figure 4 of the main text. b) Maps of two garnets showing prograde zoning in stage 1 garnet. Stage 2 garnet is evident as rims with elevated X_{Mg}. Multiple embayments can be seen in Mn-rich oscillatory zoning.

Supplementary Figure 3: Additional EPMA maps of phengite from the lawsonite eclogite, showing prograde zoning, defined by increasing Si, Fe and Mg.

Supplementary Figure 4: BSE images of omphacite, showing zoning related to Fe content. a) Coarse omphacite crystal in the matrix. b) Omphacite crystals captured in garnet 2, showing similar zoning to omphacite in the matrix.

Supplementary Figure 5: EPMA maps of chlorite in textural positions throughout the lawsonite eclogite, showing that chlorite composition appears related to the minerals it is forming in association with. Continuous white/black line indicates the boundary of a chlorite crystals. Dotted white/black line indicates the boundary between garnet 1 and garnet 2. a) Zoned chlorite, forming in stage 2 retrograde fractures which cross-cut both garnet 1 and garnet 2. b) Chlorite 1, trapped as inclusions between garnet 1 and garnet 2, and chlorite 2, which is consuming garnet 2. The two different stages of chlorite have different compositions. c) Chlorite 2 forming from the breakdown of omphacite in the matrix. The chlorite zoning is patchy, showing strong differences in composition throughout the maps.

Supplementary Figure 6: EPMA maps of texturally late glaucophane from the lawsonite eclogite, showing retrograde zoning that developed during exhumation following stage 2 burial.

References:

Brovarone, A. V., Groppo, C., Hetényi, G., Compagnoni, R., & Malavieille, J. (2011). Coexistence of lawsonite‐ bearing eclogite and blueschist: phase equilibria modelling of Alpine Corsica metabasalts and petrological evolution of subducting slabs. *Journal of metamorphic Geology*, *29*(5), 583-600.

Diener, J., & Powell, R. (2012). Revised activity–composition models for clinopyroxene and amphibole. *Journal of metamorphic Geology*, *30*(2), 131-142.

Droop, G. (1987). A general equation for estimating Fe3+ concentrations in ferromagnesian silicates and oxides from microprobe analyses, using stoichiometric criteria. *Mineralogical Magazine*, *51*(361), 431-435.

Enami, M., Suzuki, K., Liou, J., & Bird, D. K. (1993). Al-Fe3+ and F-OH substitutions in titanite and constraints on their PT dependence. *European Journal of Mineralogy*, *5*(2), 219-231.

Green, E., Holland, T., & Powell, R. (2007). An order-disorder model for omphacitic pyroxenes in the system jadeite-diopsidehedenbergite-acmite, with applications to eclogitic rocks. *American Mineralogist*, *92*(7), 1181-1189.

Green, E., White, R., Diener, J., Powell, R., Holland, T., & Palin, R. (2016). Activity–composition relations for the calculation of partial melting equilibria in metabasic rocks. *Journal of metamorphic Geology*, *34*(9), 845-869.

Hogmalm, K. J., Zack, T., Karlsson, A. K.-O., Sjöqvist, A. S., & Garbe-Schönberg, D. (2017). In situ Rb–Sr and K–Ca dating by LA-ICP-MS/MS: an evaluation of N 2 O and SF 6 as reaction gases. *Journal of Analytical Atomic Spectrometry*, *32*(2), 305-313.

Holland, T., & Powell, R. (1998). An internally consistent thermodynamic data set for phases of petrological interest. *Journal of metamorphic Geology, 16*(3), 309-343.

Holland, T., & Powell, R. (2003). Activity–composition relations for phases in petrological calculations: an asymmetric multicomponent formulation. *Contributions to Mineralogy and Petrology, 145*(4), 492-501.

Holland, T., & Powell, R. (2011). An improved and extended internally consistent thermodynamic dataset for phases of petrological interest, involving a new equation of state for solids. *Journal of metamorphic Geology, 29*(3), 333-383.

Zack, T., & Hogmalm, K. J. (2016). Laser ablation Rb/Sr dating by online chemical separation of Rb and Sr in an oxygen-filled reaction cell. *Chemical Geology, 437*, 120-133.