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Melt organisation and strain partitioning in the lower crust

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Abstract

Partial melts can form as a result of crustal thickening due to orogenesis. Even small melt fractions weaken the crust, so that partially molten volumes should accumulate significant amounts of strain. However, relatively little is known of how strain partitions in partial melts, and how effective the melt expulsion processes from the partially molten crust are. Using examples from the Western Gneiss Region (WGR), Norway, we consider a case of co-existing migmatites and shear zones. Field, image analysis, and microanalytical methods allow (semi)quantification of melt volume, rock mineralogy and mineral chemistry, and microstructures. Integration of these analyses implies effective syn-melt strain partitioning and subsequent freezing of both the shear zone and migmatite texture. We propose a mechanism that allows i) syn-melt strain localisation at an outcrop scale through stress-driven melt organisation, resulting in significant relative competence differences in a partially molten rock volume; and ii) formation of fine-grained rocks at outcrop

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that is entirely or mostly syn-melt, without subsequent mylonitic shearing in the solid-state. Syn-melt shear zones that have not acted as effective melt transport channels and/or that have not accumulated post-melt deformation may be more common than conventionally assumed.

Keywords: Melt, Migmatite, Shear zone, Microstructures, Strain, Western Gneiss Region

1. Introduction

In-situ partial melting is known to cause potentially dramatic strength decreases in the crust, even for small melt volumes, constraining styles of orogenic deformation and exhumation (e.g. Beaumont et al., 2001; Rosenberg and Handy, 2005; Jamieson and Beaumont, 2013; Levine et al., 2013). Such partial melting adds to the already heterogeneous nature of most rocks (e.g. based on grain size, mineralogy, microstructure, etc.). Lithological heterogeneities are significant factors in controlling strain partitioning on all scales (Fossen and Cavalcante, 2017). However, relatively little is known about how strain partitions in partially molten rock volumes. For example, the co-existence of partially molten rock together with regions of high strain (i.e. shear zones) is a common feature of many orogenic belts (e.g. the Himalaya) but the mechanism(s) and relative timing(s) of their formation remain poorly understood.

There are known theoretical feedback relationships between melting, rheological weakening (depending on melt fraction and melt connectivity), shear zone nucleation, and melt transport so that syn-melt shear zones are expected to function as effective transport channels for crustal partial melts

19 (e.g. Brown and Solar, 1998; Rosenberg and Handy, 2005; Brown, 2007, and
20 references therein). Mid- and lower crustal partial melts are indeed seen
21 to infiltrate many shear zones and source many large intrusive bodies, with
22 some areas even showing possible direct evidence of melt removal (e.g. Brown,
23 1994; Johannes et al., 2003; Stuart et al., 2016).

24 Despite the basic relationships being known, the behaviour of partially
25 molten crust as observed at outcrop is not always easily explained by the
26 models and experiments (Lee et al., 2017; Rosenberg and Handy, 2005),
27 meaning that many aspects of how partially molten crust actually deforms
28 remain unknown. For example, it is unclear why very large volumes of melts
29 are seen to remain approximately in-situ within the crust in the form of
30 migmatites, despite their sometimes immediate proximity to one or several
31 shear zones (e.g. Labrousse et al., 2004). Conversely, it remains unclear in
32 many cases what caused the strain partitioning into a shear zone within the
33 partially molten volume in the first place. Non-expulsion of melts might be
34 explained by the shear zones forming post-crystallization (post-melt), but
35 this contradicts with the theoretical predictions of inevitable formation of
36 syn-melt shear zones (e.g. Holtzman et al., 2003; Walte et al., 2005). Another
37 option is that the shear zones formed syn-melt but did not act as effective
38 melt transport channels. If this is true, there may be significant implications
39 to how partially molten crust deforms at a large scale.

40 In order to begin investigating the possibility of non-expulsion of partial
41 melts through shear zones, we first need to demonstrate that such shear
42 zones may exist. In this paper, we address this using representative rock
43 samples from an extensively migmatized crust of the Western Gneiss Re-

44 gion (WGR), Norway. We use image analysis, optical microscope, Electron
45 Backscatter Diffraction (EBSD) and electron microprobe to i) quantify leu-
46 cosome fraction, ii) semi-quantify melt fraction, iii) quantify mineral geo-
47 chemistry, and iv) quantify crystallographic preferred orientation (CPO) in
48 representative rock samples from the WGR. We demonstrate that exten-
49 sive partial melting and syn-melt deformation of a geochemically relatively
50 homogeneous granitoid protolith resulted in strong strain partitioning into
51 a syn-melt fine-grained shear zone with no melt expulsion, and no or very
52 little post-crystallization plastic deformation.

53 **2. Geological setting**

54 The WGR is the deepest structural level of the Scandinavian Caledonides
55 (Figure 1a; Andreasson and Lagerblad, 1980). It is dominated by tonalites
56 of 1686 to 1650 Ma, subsequently intruded by granite, gabbro and diabase
57 from 1640 to 900 Ma (Tucker et al., 1990). At 950 Ma, the igneous basement
58 underwent granulite facies metamorphism at 900°C and 1 GPa associated
59 with extensive plutonism (Tucker et al., 1990; Krabbendam et al., 2000; Corfu
60 and Andersen, 2002).

61 *Insert Figure 1*

62 During the early Palaeozoic (480-430 Ma), the Caledonian Orogeny ini-
63 tiated, causing deformation and metamorphism of the Proterozoic basement
64 gneiss and oceanic allochthons to 725°C and 1.2 GPa (Hacker et al., 2010).
65 The final stage of the Caledonian Orogeny, the Scandian, resulted in the
66 closure of the Iapetus Ocean and emplacement of oceanic allochthons onto
67 Baltica between 430 and 410 Ma (Tucker et al., 2004; Hacker and Gans,

68 2005). Later collision of Baltica and Laurentia between 425 and 400 Ma
69 resulted in the westward subduction of the Proterozoic Baltican basement
70 and portions of the allochthon to ultrahigh-pressures (UHP) of 1.8-3.6 GPa
71 and temperatures of 600-800°C (Figure 2; Andersen et al., 1991; Schärer and
72 Labrousse, 2003; Tucker et al., 2004; Hacker and Gans, 2005; Kylander-Clark
73 et al., 2008; Hacker et al., 2010).

74 From 400 to 385 Ma the WGR was exhumed to shallow crustal levels
75 (Andersen, 1998; Terry et al., 2000; Tucker et al., 2004; Hacker, 2007; Walsh
76 et al., 2007; Hacker et al., 2010). During the exhumation event, an E-W hori-
77 zontal stretching was imprinted alongside in-situ partial melting of the gneiss
78 via post-UHP decompression-related retrograde amphibolite metamorphism
79 as the pressure decreased from 2.8 to 0.5 GPa at temperatures of 600 to
80 800°C (Figure 2; Krogh, 1980; Chauvet et al., 1992; Andersen, 1998; Straume
81 and Austrheim, 1999; Hacker et al., 2003; Labrousse et al., 2002; Schärer and
82 Labrousse, 2003; Labrousse et al., 2004; Walsh and Hacker, 2004; Root et al.,
83 2005; Engvik et al., 2007; Gordon et al., 2013, 2016). Evidence for (U)HP
84 metamorphism was almost completely overprinted during the open-system
85 partial melting event, as the UHP rocks were exhumed from 100 km depth
86 to 15-20 km (Schärer and Labrousse, 2003; Root et al., 2005). Exhumation
87 occurred from 394 to 389 Ma at a rate of 5 mm/year, followed by rapid cool-
88 ing to reach 300°C by 357 Ma (Schärer and Labrousse, 2003; Root et al.,
89 2005).

90 *Insert Figure 2 (1 column figure)*

91 This study focuses on the Nupen peninsula in the southwest of Gurskøy
92 (Figure 1b,c). The primary lithology is amphibolite-facies quartzofeldspathic

93 gneiss that has undergone partial melting. The gneisses show layers of
94 melanosomes and leucosomes that were stretched and sheared at a later stage,
95 indicating that the migmatization commenced early in the exhumation-related
96 deformation history (Labrousse et al., 2002). Gurskøy was exhumed via
97 thrusting and formed a NW-SE trending isoclinal fold verging southwest
98 on the peninsula (Figure 1; Labrousse et al., 2004). Geothermometry from
99 nearby Vanylven migmatized gneiss (approx. 5 km south of Nupen) indi-
100 cate an influx of H₂O-rich fluids, allowing decompression melting to begin at
101 600-650°C and not exceeding 800°C (Labrousse et al., 2002, 2004; Ganzhorn
102 et al., 2014).

103 Over a 1.2 km section (Figure 1c), there is a diverse deformation sequence
104 of migmatized gneiss, mylonitic shear zones, sillimanite bearing garnet-mica
105 schists, augen gneiss and boudinaged amphibolite dykes. The strongly de-
106 formed mylonitic shear zones extend from 5 to over 100 meters in width,
107 but deformation is also high in the migmatitic layers, indicated by S-C fab-
108 rics and isoclinal folding of leucosome, mesosome and melanosome showing
109 deformation is widespread over the peninsula.

110 Sixteen samples representing leucosome:melanosome ratio at the outcrop
111 were taken from the different lithologies. The leucosome abundance and the
112 degree of strain varies between different sample locations (Figure 1c).

113 **3. Melt characteristics**

114 In general, migmatitic melting falls into two melting reactions: granitic
115 wet melt and dehydration melt. Granitic wet melting occurs via the simple
116 reaction of quartz + plagioclase + K-feldspar + H₂O = melt, and begins in

117 migmatites with a granitic protolith at approximately 650°C (Wyllie, 1977).
118 Although the melt reaction is simple, the crystallisation products are the
119 same as the reactants (i.e. quartz, plagioclase and K-feldspar), making iden-
120 tification of melt phases difficult. However, some distinctive textures may be
121 present that are diagnostic of crystallisation from melt. These include cus-
122 pate or serrated grain boundaries with low dihedral angles (Harte et al., 1991;
123 Sawyer, 2001; Holness and Sawyer, 2008), pseudomorphs of melt along grain
124 boundaries and at triple/multiple junctions (Rosenberg and Riller, 2000; Hol-
125 ness and Sawyer, 2008), and grains with straight crystal faces that indicate
126 crystallisation from a melt (Vernon and Collins, 1988). Dehydration melt-
127 ing reactions are easier to identify because of the peritectic products and
128 melt textures. These consist of solid products of melting reactions form-
129 ing euhedral crystal faces against the melt (Sawyer, 1999, 2001), reactant
130 phases exhibiting rounded or corroded boundaries surrounded by melt films
131 (Mehnert et al., 1973; Büsch et al., 1974), small cusplate-shaped melt pools
132 similar to those formed in experimental studies (Harte et al., 1991; Rosenberg
133 and Riller, 2000; Holness and Sawyer, 2008), and also inter-growths between
134 quartz and solid products of melting (Waters, 2001; Barbey, 2007).

135 At the outcrop scale it is often assumed that melt is present when there is
136 a high-temperature mineral assemblage alongside quartzofeldspathic lenses,
137 layers or patches (i.e. the leucosome; Mehnert, 1968). However, at the mi-
138 croscale the leucocratic material may show solid-state solution recrystallisation
139 and deformation microstructures as opposed to melt microstructures. Here
140 we use microstructures, image analysis and electron backscatter diffraction
141 (EBSD) to determine melt volume and properties of a section of migmatized

142 and sheared gneiss from the WGR where partial melting occurred during the
143 formation of the Scandinavian Caledonides (Hacker et al., 2010, and refer-
144 ences therein). Through analysis of textures in the migmatites and shear
145 zones we demonstrate that even in volumes with high melt percentage, de-
146 formation may partition away from the highly molten volumes into discrete
147 shear zones.

148 *3.1. Macroscale evidence for melt*

149 Nupen is comprised of migmatized gneiss, mylonitic shear zones, silliman-
150 ite bearing garnet-mica schists and augen gneiss. All lithologies have under-
151 gone varying degrees of partial melting. Figure 3 shows the characteristic
152 outcrop scale variation across the peninsula. The migmatites are typically
153 stromatic in structure with intense folding but have a varied leucosome con-
154 tent (Figure 3a,d). The next most common lithology is the mylonitic shear
155 zones, which are strongly lineated with segregation of felsic and mafic lay-
156 ers (Figure 3b,c). Foliation parallel zones of garnet schist are common in
157 the north and south (Figure 3e). The schist contains garnet porphyroblasts
158 (0.5 to 5 cm) with the foliation defined by bands of biotite and sillimanite,
159 hornblende, and felsics. A zone of augen gneiss is observed in the south of
160 the Nupen peninsula; the augen (3 to 10+ cm) are mostly K-feldspar and
161 occasionally plagioclase (Figure 3f).

162 *Insert Figure 3*

163 Figure 4 shows examples of the variation in the migmatized gneiss at
164 the outcrop scale. Stromatic migmatites are most common but the leuco-
165 somes can also appear patchy, indicating the migmatite texture is surreitic
166 or ophthalmic. Stromatic-type migmatites have the highest leucosome con-

167 tent (Figure 4a) and surreitic migmatites have the lowest (Figure 4c). The
168 variation in melt fractions and migmatite structures are expected to reflect
169 strain partitioning due to internal rheological variations.

170 *Insert Figure 4*

171 Macroscale leucosome content is determined via image analysis using Im-
172 ageJ of digitised outcrop photos (Schneider et al., 2012). Digitising pho-
173 tographs and correcting for shadows or cracks on the outcrop helps to con-
174 strain leucosome vs. restite proportions. Examples of the digitised outcrop
175 drawings are shown in figure 4. The leucosome content in the Nupen area
176 varies from 0 to 90%. However, this does not mean up to 90% of the crust
177 was melt, as the leucosome fraction does not necessarily equal the melt frac-
178 tion. The leucosome content gives the maximum melt proportion left in-situ.
179 It is necessary to use microstructures to distinguish if the leucosome formed
180 from melting or solid-state deformation and recrystallisation processes.

181 *3.2. Microstructural evidence for melt*

182 At the microscale we are able to qualitatively determine how much of the
183 leucosome is representative of melt and the type of migmatitic melting that
184 occurred. The typical melt microstructures observed in the migmatites are
185 peritectic melt products, euhedral crystal faces, cusped-shaped melt pools,
186 felsic compositional zoning (quartz and feldspar bands), low dihedral angles
187 (e.g. $<60^\circ$), disequilibrium grain boundaries, ‘string of beads’ texture, inter-
188 stitial quartz, and myrmekite lobes (Figure 5; Sawyer, 1999, 2001; Holness
189 et al., 2005, 2011). On their own, these microstructures would be weak in-
190 dicators of melting, but when found together significantly strengthen the
191 interpretation of melt (Vernon, 2011).

192 Fluid-present granitic eutectic melting starts to occur at lower tempera-
193 tures ($\sim 650^\circ\text{C}$; Wyllie, 1977) than biotite-dehydration melting ($760\text{-}800^\circ\text{C}$;
194 Spear et al., 1999). Decompression related partial melting in the WGR be-
195 gan at $600\text{-}650^\circ\text{C}$ and did not exceed 800°C (Labrousse et al., 2002, 2004;
196 Ganzhorn et al., 2014). Thin quartz films along grain boundaries between
197 plagioclase and K-feldspar and occasionally biotite and K-feldspar are ob-
198 served in the migmatized gneiss (Figure 5f). When the presence of quartz
199 films is combined with the lack of garnet, the peritectic product from biotite-
200 dehydration melting, and absence of chessboard extinction within quartz
201 grains, they together suggest melting of the migmatites occurred at the lower
202 end of the temperature range via granitic wet melting rather than biotite-
203 dehydration melting. Thus, the reaction systems are:



204



205 However the garnet schists contain both the reactants and products in
206 Equation 2 (Figure 5i). This suggests there were different protoliths of the
207 partially melted rocks, most likely a granitoid protolith for the migmatized
208 gneiss and a pelitic protolith for the garnet schists (Bucher and Grapes, 2011).
209 The abundance of biotite and presence of minor sillimanite suggests that
210 not all the biotite and sillimanite were consumed during biotite-dehydration
211 melting and/or the biotite and sillimanite are retrograde. Here we focus
212 on the fluid-present granitic eutectic melting of the migmatitic gneiss to
213 determine the melt fraction and styles from microstructures.

214 *Insert Figure 5*

215 If melting occurs under static conditions and absence of deformation in
216 an igneous setting, the grains crystallised from melt will reach a textural
217 equilibrium with a uniform grain size and grain boundary angles relative
218 to interfacial angles (e.g. 103 to 115° for quartz, plagioclase or K-feldspar;
219 Vernon, 1968). This does not apply to the migmatites from the WGR as par-
220 tial melting occurred under a deformation regime via decompression related
221 retrograde metamorphism. As a result we mainly observe irregular grain
222 boundaries and low dihedral angles (<60°) as the migmatite crystallised in
223 textural disequilibrium (Figure 5a).

224 Dihedral angles play a key role in melt connectivity: melt is likely to
225 be interconnected if the melt-solid dihedral angle is less than 60°, but is
226 likely to form isolated pockets when greater than 60° (Holness, 2006; Holness
227 et al., 2011). Thus, if there is a high dihedral angle the melt connectivity is
228 low and the strength of the partially melted rock would be greater than if
229 there is a low dihedral angle where the melt connectivity is high. Figure 5b
230 shows examples of the high and low dihedral angles observed in our samples.
231 Following the method of Holness et al. (2005) a sample of 50 dihedral angles
232 were measured for each thin section. Figure 6 shows results of solid-solid-melt
233 and solid-solid-solid dihedral angles, where the relationship between median
234 angle and standard deviation indicate melt-present or solid-state equilibrium.
235 We observe a median dihedral angle of 53° for all boundaries, 37° for solid-
236 solid-melt boundaries and 88° for solid-solid-solid boundaries (Figure 6).
237 Where melt is present it is observed to be connected and thus the strength
238 of the migmatite during melting is low.

239 *Insert Figure 6*

240 The ‘string of beads’ texture of quartz grains is also observed in some
241 WGR samples (Figure 5a). This texture usually forms during concentration
242 of melt from thick films on grain boundaries into ‘beads’ (Holness et al.,
243 2011). The texture suggests a slow crystallisation rate to allow nucleation
244 of individual grains from melt films (Holness et al., 2011). Interstitial melt
245 can appear similar to the ‘string of beads’ texture. This is more commonly
246 observed and may represent larger collections of crystallised melt that did
247 not separate into individual beads. Irregular grains of quartz with lobate
248 (Figure 5c) and cusped (Figure 5b,d,e) grain boundaries are also commonly
249 present. Figure 5d shows embayed biotite and plagioclase where quartz has
250 penetrated into the non-melt grains. Fingers of inferred former melt with
251 small dihedral angles are shown in Figure 5b and commonly align parallel to
252 foliation. The shear zone samples also show interstitial quartz infilling pore
253 space implying they have also melted (Figure 5e).

254 Lobes of myrmekite are very common in the migmatized gneisses and are
255 also present in small amounts in the shear zone samples; however, their
256 origin is controversial. It has been proposed that they originated either
257 from a quenched fluid-rich melt (Hibbard, 1979, 1987) or that they represent
258 solid-state reactions in the presence of hydrous fluid (e.g. Ashworth, 1972;
259 Phillips, 1974, 1980; Simpson and Wintsch, 1989; Vernon, 1991; Yuguchi and
260 Nishiyama, 2008; Vernon, 2011). Figure 5h shows a large K-feldspar grain
261 mantled by myrmekite lobes, which may be of melt origin as they co-exist
262 with other melt microstructures. Myrmekite is also observed in the shear
263 zone, figure 5g shows a coarse myrmekite band as well as finer myrmekite

264 mantling K-feldspar grains. Conversely if they are a solid-state texture, they
265 indicate deformation occurred post-melt in a fluid-rich system.

266 *Insert Figure 7*

267 *Insert Figure 8 (1 column figure)*

268 To understand the effects of melt on rheology, it is important to quan-
269 tify the melt fraction. Microscale melt determination is qualitative as we
270 use microstructures indicative of melt or the former presence of melt. With
271 the use of ImageJ we are able to isolate the melt and solid fractions of the
272 rock. Plane polarised (PPL) and cross-polarised (XPL) light photomicro-
273 graphs with and without gypsum plate are used to construct the melt-solid
274 interpretations. Figure 7 shows examples of the quantification of the migma-
275 tized gneiss via photomicrographs and corresponding interpretations of melt
276 (white) vs. solid (grey). Figure 8 shows the comparison between leucosome
277 volume at the macroscale and melt volume at the microscale. In the Nupen
278 area, the leucosome volume is more than twice the interpreted microstruc-
279 tural melt content, leading to a possible overestimation and an unreliable
280 method of calculating accurate melt volumes in the field. Nevertheless the
281 leucosome and the interpreted melt percentages are systematically higher in
282 the migmatite gneisses outside the shear zones. The calculated melt percent-
283 age assumes all melt was liquid at the same time, which is unlikely.

284 **4. Mineral assemblages and compositions**

285 In the WGR we see evidence for fine grained shear zones in a migmatized
286 gneiss, but what is the timing of melting with respect to the shear zones?
287 To investigate how the shear zones are related to the migmatites, we use mi-

288 croanalytical techniques to compare the migmatites and shear zone samples.
289 WGR12 is a migmatized gneiss with 18-20% melt in the microstructure and
290 WGR13 is a mylonitized gneiss with $\sim 1\%$ melt in the microstructure. Figure
291 9 shows backscattered electron maps of both samples where there is a grain
292 size variation of 0.2 to 5 mm in the migmatite and 0.05 to 0.3 mm in the
293 shear zone.

294 *Insert Figure 9*

295 If the shear zone has not undergone partial melting or was originally a
296 different layer of rock, we would expect to see different mineralogies due
297 to melt reactions that would have occurred in the migmatite. However, if
298 the protolith was fairly homogeneous and the mylonite-like fabric formed
299 via syn-melt shearing, the melt reactions will produce the same mineralogy
300 for the migmatized and mylonitized gneisses. In detail, there may be small
301 differences between the migmatite and shear zone as more melt loss is likely
302 to have occurred via transport through the shear zone than in the migmatized
303 wall rock as indicated in Figure 8.

304 *Insert Table 1*

305 The mineral compositions in the migmatites and shear zones are the same;
306 quartz, K-feldspar, plagioclase, biotite and minor hornblende. The modal
307 proportions are calculated via the number of points indexed from EBSD and
308 shown in Table 1. Quartz is the most abundant mineral in both samples
309 followed by K-feldspar. K-feldspar makes up more of WGR12 than WGR13
310 but this is due to large K-feldspar grains dominating the sample. As the
311 modal compositions are similar it suggests there has not been a significant
312 change in mineralogy due to melt reactions or melt loss between the two

313 samples. Both observed leucosome and interpreted melt volume are much
314 lower than in surrounding migmatites, mineralogical proportions suggest the
315 leucosome is underestimated greatly in the shear zone due to the small grain
316 size.

317 Spot analyses for plagioclase (plag), K-feldspar (kfs), biotite (bte) and
318 hornblende (hbl) were measured with the University of Leeds, JEOL JXA8230
319 electron microprobe operated at 15kV and 15nA with a defocused beam (Ta-
320 ble 1). The results show there are no significant variations in the major or
321 minor elements. Plagioclase and some K-feldspar represent the melt phase
322 and the element totals are similar between the migmatite and shear zone
323 samples. Both samples indicate the plagioclase is andesine in composition
324 (An_{32-48}), although some analyses indicate a more sodic composition, due
325 to the perthitic relationship between plagioclase and K-feldspar where albite
326 forms the perthite. Mineral compositions and abundances suggest a pro-
327 toolith of a granitic composition such as felsic gneiss (Bucher and Grapes,
328 2011). The similarities in element totals suggest the shear zone was involved
329 in melting and did not permit increased melt loss when compared to the
330 migmatites.

331 5. Crystallographic Preferred Orientations

332 If the shear zones have not undergone melting they should preserve a
333 record of pre-, syn- and post-melt deformation(s). In particular, the lack of
334 pervasive melting means the crystal microstructure has not been ‘reset’ in
335 non-melt zones and hence should be observed in crystallographic preferred
336 orientations (CPO). In contrast, if all of the gneiss is involved in melting

337 and the shear zone forms syn-melting, any pre-existing fabric will have been
338 ‘reset’ with crystallisation under the same stresses for the migmatite and
339 shear zone. Subsequent deformation may occur but the post-melt CPO from
340 the migmatite should be recognisable in the shear zone CPO which would
341 be strong to reflect the deformation. We compare the CPO for the adjacent
342 migmatite, WGR12, and shear zone, WGR13, to identify timing of deforma-
343 tion relative to melting.

344 *Insert Figure 10*

345 CPO results were measured via EBSD using a FEI Quanta 650 FEG-
346 ESEM with AZtec software and an Oxford/HKL Nordlys S EBSD system
347 at the University of Leeds (mapped areas are shown in Figure 9). Figure
348 10 shows the CPO for biotite and quartz in samples WGR12 and WGR13
349 (plagioclase and K-feldspar CPO is available in supporting information). The
350 biotite represents a solid phase whereas the quartz represents one of the melt
351 phases.

352 Figure 10a,b shows the biotite CPO for both samples, they are somewhat
353 similar both exhibiting strong c-axes maxima parallel (\parallel) to Z, although more
354 dispersed in WGR12 (e.g. Ji et al., 2015). Similarly, single girdles parallel
355 to basal plane for $\langle a \rangle$ and $\langle b \rangle$, again, less well-defined for WGR12
356 with distinct a-max \parallel X and b-max \parallel Y in WGR13. The presence of the girdles
357 and a/b-max suggest gliding on the $\langle 001 \rangle$ parallel to $\langle 100 \rangle$ or perhaps
358 $\langle 110 \rangle$, which is the average of $\langle 100 \rangle$ and $\langle 010 \rangle$ in WGR13. However,
359 the dispersion of a and b in a girdle parallel to XY for WGR12 plane could
360 indicate ‘floating’ in a melt. The dispersed c-max is compatible with this idea
361 as it indicates slight ‘floating’ induced undulation. As a result, the WGR12

362 biotite CPO is more likely the effect of a syn-melt fabric and controlled by
363 the shape preferred orientation of biotite aligned by flow of the melt instead
364 of a deformation induced CPO present in the shear zone.

365 The quartz CPO are generally weak for both the shear zone and espe-
366 cially migmatite samples (Figure 10c,d). If deformation to form the shear
367 zone occurred pre- or post-melt producing a mylonite, we would expect to
368 see a strong CPO in WGR13. However, the weak CPO is atypical for a my-
369 lonite (Toy et al., 2008; Barth et al., 2010) suggesting there was not much
370 deformation in solid-state (Figure 10d). Furthermore, whilst both samples
371 indicate $r||Y$ as a possible vorticity axis in quartz, the biotite CPO's imply
372 no such rotation. This suggests any deformation is preferentially irrotational
373 (i.e. 'pure shear') for biotite whereas it is rotational (i.e. 'simple shear') for
374 quartz. By comparison with VPSC models (Morales et al., 2014), the quartz
375 CPO correlates with amphibolite facies r -slip simple shear simulations ($\gamma =$
376 1.73). This implies both melt and crystallisation occurred within the amphi-
377 bolite facies (Chauvet et al., 1992; Andersen, 1998; Schärer and Labrousse,
378 2003; Labrousse et al., 2011). In summary, the similarities between the CPO
379 for the migmatite, WGR12, and the shear zone, WGR13, suggest the CPO
380 were reset during the melting event; the lack of strength in the shear zone
381 CPO and similarities with the migmatite CPO suggests deformation was
382 transient or even absent post-melting.

383 6. Discussion

384 Searle (2013) posed the question "do shear zones control the generation
385 and ascent of magmas (Brown and Solar, 1998; Brown, 2007) or do magmas

386 trigger nucleation of shear zones?”. The Nupen peninsula has fine grained
387 shear zones within migmatized gneiss. However, there is no evidence for
388 significant mineral chemistry variations or post-crystallisation strain in the
389 shear zone. We propose a hypothesis that grain size reduction is a result
390 of initial syn-melt shearing which results in the observed geometry of the
391 bodies (thin laminae/sheets of melt). The layering constrained grain growth
392 and increased heterogeneous nucleation rates to maintain a small grain size.

393 Low dihedral angles, interstitial quartz infilling pore space and myrmekite
394 are present in shear zone samples from Nupen (Figure 5e). These microstruc-
395 tures are not strong evidence of melting, but when taken together and con-
396 sidering the close proximity to the migmatites it is likely melt was present
397 in the shear zones. If the shear zones were active post-melting, the melt
398 microstructures are likely to have been erased by solid-state processes. Here
399 we do not see deformation of the melt microstructures in solid-state (Figure
400 5e), suggesting the shear zone was active pre- or syn-melt. As well as the
401 shear zone microstructures, the alignment of the foliation, shear zones and
402 melt ‘fingers’ in the migmatite is evidence for a single pattern of strain for
403 the formation of all three features, suggesting a syn-melt fabric.

404 The absence of significant variations in mineralogy or mineral chemistry
405 between the migmatite and shear zone suggest the protolith was the same
406 (Figure 1), most likely a felsic gneiss to allow for fluid-present granitic melting
407 (Equation 1). This suggests melt was pervasive through the migmatite and
408 shear zone. The shear zone may have been used as a pathway for melt loss
409 but the mineralogy suggests no evidence for melt loss or gain in the shear
410 zone. This suggests the shear zone was short-lived as the migmatite was not

411 ‘drained’ of melt relative to the shear zone.

412 The strong biotite CPO is expected for the shear zone sample with strong
413 c-axes maximum and normal to foliation in $\langle a \rangle$ and $\langle b \rangle$. The biotite
414 CPO for the migmatite is similar to the CPO from the shear zone but more
415 diffuse. The similarity between the CPO’s suggests deformation occurred
416 under the same stress field resulting in passive rotation of the biotite grains
417 sub-perpendicular to the maximum normal stress component. Although the
418 shear zone sample appears to have a mylonitic fabric, the quartz CPO is
419 not typical for a mylonite (Toy et al., 2008; Barth et al., 2010). We suggest
420 syn-melt deformation of the quartz with progressive crystallisation results in
421 a weak CPO regardless of deformation intensity. As there is little evidence
422 for post-melt deformation in the CPO, we suggest grain size reduction gives
423 the appearance of a mylonite that formed syn-melt.

424 *Insert Figure 11 (1 column figure)*

425 As a result of the microstructural and petrology data, we propose a model
426 for melt organisation, strain localisation and formation of a fine grain size
427 during partial melting. The model starts with a homogeneous solid gneiss
428 where strain is distributed evenly (Figure 11a). Melting is generally evenly
429 distributed throughout the leucosome; however, where melt connectivity is
430 slightly higher, the melt organises into layers whereas elsewhere it remains
431 as disorganised ‘pools’ (Figure 11b,c).

432 The organised melt system results in a viscosity reduction compared to
433 the disorganised melt system and in turn results in a higher strain for a lower
434 shear stress as shown by Rosenberg and Handy (2005). Increased shearing
435 of the melt thins the melt layers (Figure 11d). Strain localises into the melt

436 zones resulting in a stress-driven organisation of melt (Brown and Solar,
437 1998; Rosenberg and Handy, 2005; Vanderhaeghe, 2009). The melt drains
438 from the nearby migmatite into the shear zones and forms the initial stages
439 of a melt pathway and the mylonite-like fabric. We suggest the shear zones
440 at Nupen are short-lived and do not develop melt pathways to expel melt
441 from the system.

442 The shear zones formed thin melt bands due to higher strain accommo-
443 dation when the rocks were partially molten. This layering constrained grain
444 growth and increased heterogeneous nucleation rates in the fine grained rocks.
445 When the melt crystallised the grains do not have the space (or surface en-
446 ergy) to grow. This resulted in the small grain size in the organised melt
447 layers, than the disorganised melt layers giving the mylonitic appearance
448 seen at outcrop scale today (Figure 11e). This process induces a mylonitic
449 macroscale appearance of the shear zone without the solid-state deformation
450 fabric expected of a mylonite. The melt organisation process generates (or
451 preserves) grain size heterogeneity without significant solid-state deformation
452 in shear zones after solidification of the migmatite. This means a mechanism
453 for later strain localisation is preserved in a system which has not undergone
454 later deformation. The resulting heterogeneity allows formation of shear
455 zones in areas such as South Armorcian Shear Zone, Brittany (Brown and
456 Dallmeyer, 1996), Wet Mountains, Colorado (Levine et al., 2013), and the
457 Himalayas and Karakoram (Searle, 2013) where there is a longer history of
458 post solidification deformation.

459 7. Conclusions

460 In this study we have interpreted the mechanisms for strain partitioning
461 at Nupen in the WGR, Norway. We observe a feedback process where reor-
462 ganisation of the melt leads to strain localisation and grain size reduction,
463 which in turn results in increased strain partitioning.

464 Whereas mylonite-like fine grained rocks are normally interpreted to have
465 formed due to shearing in the solid-state, here we observe a mylonite-like
466 rock that probably formed while partially molten and, therefore, lack the
467 deformed microstructure of a mylonite. This situation implies that while
468 partially molten volumes are weaker than completely solid rock, internal het-
469 erogeneities may result in significant differences in relative, effective rheology
470 and therefore strain partitioning. An important implication is syn-melt shear
471 zones do not necessarily lead to melt expulsion, here we see no evidence of
472 increased melt loss or melt accumulation as the migmatite transitions into
473 the shear zones.

474 Fine grained shear zones are not necessarily post-melt or retrograde my-
475 lonites. The once-molten rocks do not necessarily accumulate further strain
476 once crystallised. However, the organisation of partial melt forming syn-
477 melt shear zones could be the origin of the mechanical heterogeneity needed
478 to allow later strain localisation. If the shear zones continued to deform
479 post-melting in the solid-state we would expect to see strain localise into the
480 shear zones with evidence for solid-state deformation and the loss of melt
481 microstructures. This process would produce the mylonitic fabrics observed
482 in shear zones.

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492 **References**

- 493 Andersen, T. B., 1998. Extensional tectonics in the Caledonides of southern Norway, an overview. *Tectono-*
494 *physics* 285 (3-4), 333–351.
- 495 Andersen, T. B., Jamtveit, B., Dewey, J. F., Swensson, E., 1991. Subduction and exhumation of continental
496 crust: major mechanisms during continent-continent collision and orogenic extensional collapse, a model
497 based on the south Norwegian Caledonides. *Terra Nova* 3 (3), 303–310.
- 498 Andreasson, P. G., Lagerblad, B., 1980. Occurrence and significance of inverted metamorphic gradients
499 in the western Scandinavian Caledonides. *Journal of the Geological Society* 137 (3), 219–230.
- 500 Ashworth, J. R., 1972. Myrmekites of exsolution and replacement origins. *Geological Magazine* 109 (1),
501 45–62.
- 502 Ashworth, J. R. (Ed.), 1985. *Migmatites*. Blackie, Glasgow.
- 503 Auzanneau, E., Vielzeuf, D., Schmidt, M. W., 2006. Experimental evidence of decompression melting
504 during exhumation of subducted continental crust. *Contributions to Mineralogy and Petrology* 152 (2),
505 125–148.
- 506 Barbey, P., 2007. Diffusion-controlled biotite breakdown reaction textures at the solid/ liquid transition
507 in the continental crust. *Contributions to Mineralogy and Petrology* 154 (6), 707–716.

- 508 Barth, N. C., Hacker, B. R., Seward, G. G. E., Walsh, E. O., Young, D., Johnston, S., 2010. Strain within
509 the ultrahigh-pressure Western Gneiss region of Norway recorded by quartz CPOs. Geological Society,
510 London, Special Publications 335, 663–685.
- 511 Beaumont, C., Jamieson, R. A., Nguyen, M. H., Lee, B., 2001. Himalayan tectonics explained by extrusion
512 of a low-viscosity crustal channel coupled to focused surface denudation. *Nature* 414 (December), 738–
513 742.
- 514 Brown, M., 1994. The generation, segregation, ascent and emplacement of granite magma: the migmatite-
515 to-crustally-derived granite connection in thickened orogens. *Earth Science Reviews* 36 (1-2), 83–130.
- 516 Brown, M., 2007. Old Crustal melting and melt extraction, ascent and emplacement in orogens: mecha-
517 nisms and consequences. *Journal of the Geological Society* 164 (4), 709–730.
- 518 Brown, M., Dallmeyer, R. D., may 1996. Rapid Variscan exhumation and the role of magma in core
519 complex formation: southern Brittany metamorphic belt, France. *Journal of Metamorphic Geology*
520 14 (3), 361–379.
- 521 Brown, M., Solar, G. S., 1998. Granite ascent and emplacement during contractional deformation in
522 convergent orogens. *Journal of Structural Geology* 20 (9-10), 1365–1393.
- 523 Bucher, K., Grapes, R., 2011. Metamorphic Rocks. In: *Petrogenesis of Metamorphic Rocks*. Springer-
524 Verlag, Berlin, Ch. 2, pp. 21–56.
- 525 Büsch, W., Schneider, G., Mehnert, K. R., 1974. Initial melting at grain boundaries. Part II: melting
526 in rocks of granodioritic, quartzdioritic and tonalitic composition. *Neues Jahrbuch für Mineralogie,*
527 *Monatshefte* 8, 345–370.
- 528 Carswell, D. A., Brueckner, H. K., Cuthbert, S. J., Mehta, K., Brien, P. J. O., 2003. The timing of
529 stabilisation and the exhumation rate for ultra-high pressure rocks in the Western Gneiss Region of
530 Norway. *Journal of Metamorphic Geology* 21 (6), 601–612.
- 531 Chauvet, A., Kienast, J. R., Pinardon, J. L., Brunel, M., 1992. Petrological Constraints and Pt Path of
532 Devonian Collapse Tectonics within the Scandian Mountain Belt (Western Gneiss Region, Norway).
533 *Journal of the Geological Society* 149, 383–400.
- 534 Corfu, F., Andersen, T. B., 2002. U-Pb ages of the Dalsfjord complex, SW Norway, and their bearing on
535 the correlation of allochthonous crystalline segments of the Scandinavian Caledonides. *International*
536 *Journal of Earth Sciences* 91 (6), 955–963.
- 537 Engvik, A. K., Andersen, T. B., Wachmann, M., 2007. Inhomogeneous deformation in deeply buried
538 continental crust, an example from the eclogite-facies province of the Western Gneiss Region, Norway.
539 *Norsk Geologisk Tidsskrift* 87 (4), 373–389.

- 540 Fossen, H., Cavalcante, G. C. G., 2017. Shear zones A review. *Earth-Science Reviews* 171 (May), 434–455.
- 541 Ganzhorn, A. C., Labrousse, L., Prouteau, G., Leroy, C., Vrijmoed, J. C., Andersen, T. B., Arbaret, L.,
542 aug 2014. Structural, petrological and chemical analysis of syn-kinematic migmatites: insights from
543 the Western Gneiss Region, Norway. *Journal of Metamorphic Geology* 32 (6), 647–673.
- 544 Gordon, S. M., Whitney, D. L., Teyssier, C., Fossen, H., 2013. U-Pb dates and trace-element geochemistry
545 of zircon from migmatite, Western Gneiss Region, Norway: Significance for history of partial melting
546 in continental subduction. *Lithos* 170-171, 35–53.
- 547 Gordon, S. M., Whitney, D. L., Teyssier, C., Fossen, H., Kylander-Clark, A. R. C., 2016. Geochronology
548 and geochemistry of zircon from the northern Western Gneiss Region: Insights into the Caledonian
549 tectonic history of western Norway. *Lithos* 246-247, 134–148.
- 550 Hacker, B. R., 2007. Ascent of the ultrahigh-pressure Western Gneiss Region , Norway. *Geological Society
551 of America, Special Papers* 419, 171–184.
- 552 Hacker, B. R., Andersen, T. B., Johnston, S., Kylander-Clark, A. R. C., Peterman, E. M., Walsh, E. O.,
553 Young, D., 2010. High-temperature deformation during continental-margin subduction & exhumation:
554 The ultrahigh-pressure Western Gneiss Region of Norway. *Tectonophysics* 480 (1-4), 149–171.
- 555 Hacker, B. R., Andersen, T. B., Root, D. B., Mehl, L., Mattinson, J. M., Wooden, J. L., 2003. Exhuma-
556 tion of high-pressure rocks beneath the Solund Basin, Western Gneiss Region of Norway. *Journal of
557 Metamorphic Geology* 21 (6), 613–629.
- 558 Hacker, B. R., Gans, P. B., 2005. Continental collisions and the creation of ultrahigh-pressure terranes:
559 Petrology and thermochronology of nappes in the central Scandinavian Caledonides. *Bulletin of the
560 Geological Society of America* 117 (1-2), 117–134.
- 561 Harte, B., Pattison, D. R. M., Linklater, C. M., 1991. Field relations and petrography of partially melted
562 pelitic and semi-pelitic rocks. In: *Equilibrium and Kinetics in Contact Metamorphism: the Ballachulish
563 Igneous Complex and its Aureole*. Springer, Heidelberg, pp. 181–210.
- 564 Hibbard, M. J., 1979. Myrmekite as a marker between preaqueous and postaqueous phase saturation in
565 granitic systems. *Geological Society of America Bulletin* 90 (11), 1047–1062.
- 566 Hibbard, M. J., 1987. Deformation of incompletely crystallized magma systems: granitic gneisses and
567 their tectonic implications. *Journal of Geology* 95 (4), 543–561.
- 568 Holness, M. B., 2006. Melt-solid dihedral angles of common minerals in natural rocks. *Journal of Petrology*
569 47 (4), 791–800.

- 570 Holness, M. B., Cesare, B., Sawyer, E. W., 2011. Melted rocks under the microscope: Microstructures and
571 their interpretation. *Elements* 7 (4), 247–252.
- 572 Holness, M. B., Cheadle, M. J., McKenzie, D. P., 2005. On the use of changes in dihedral angle to decode
573 late-stage textural evolution in cumulates. *Journal of Petrology* 46 (8), 1565–1583.
- 574 Holness, M. B., Sawyer, E. W., 2008. On the pseudomorphing of melt-filled pores during the crystallization
575 of migmatites. *Journal of Petrology* 49 (7), 1343–1363.
- 576 Holtzman, B. K., Groebner, N. J., Zimmerman, M. E., Ginsberg, S. B., Kohlstedt, D. L., 2003. Stress-
577 driven melt segregation in partially molten rocks. *Geochemistry, Geophysics, Geosystems* 4 (5), 1–26.
- 578 Jamieson, R. A., Beaumont, C., 2013. On the origin of orogens. *Bulletin of the Geological Society of*
579 *America* 125 (11), 1671–1702.
- 580 Ji, S., Shao, T., Michibayashi, K., Oya, S., Satsukawa, T., Wang, Q., Zhao, W., Salisbury, M. H., 2015.
581 Magnitude and symmetry of seismic anisotropy in mica-and amphibole-bearing metamorphic rocks and
582 implications for tectonic interpretation of seismic data from the southeast Tibetan Plateau. *Journal of*
583 *Geophysical Research: Solid Earth* 160 (9), 6404–6430.
- 584 Johannes, W., Ehlers, C., Kriegsman, L. M., Mengel, K., jun 2003. The link between migmatites and
585 S-type granites in the Turku area, southern Finland. *Lithos* 68 (3-4), 69–90.
- 586 Kildal, E. S., 1970. Geologisk kart over Noreg Måløy: Trondheim. Norges Geologiske Undersøking scale
587 1:25.
- 588 Krabbendam, M., Wain, A., Andersen, T. B., 2000. Pre-Caledonian granulite and gabbro enclaves in
589 the Western Gneiss Region, Norway: indications of incomplete transition at high pressure. *Geological*
590 *Magazine* 137 (3), 235–255.
- 591 Krogh, E. J., 1980. Geochemistry and petrology of glaucophane-bearing eclogites and associated rocks
592 from Sunnfjord, Western Norway. *Lithos* 13 (4), 355–380.
- 593 Kylander-Clark, A. R. C., Hacker, B. R., Mattinson, J. M., 2008. Slow exhumation of UHP terranes:
594 Titanite and rutile ages of the Western Gneiss Region, Norway. *Earth and Planetary Science Letters*
595 272 (3-4), 531–540.
- 596 Labrousse, L., Jolivet, L., Agard, P., Hebert, R., Andersen, T. B., aug 2002. Crustal-scale boudinage and
597 migmatization of gneiss during their exhumation in the UHP Province of Western Norway. *Terra Nova*
598 14 (4), 263–270.

- 599 Labrousse, L., Jolivet, L., Andersen, T. B., Agard, P., Maluski, H., 2004. Pressure-temperature-time
600 deformation history of the exhumation of ultra-high pressure rocks in the Western Gneiss Region,
601 Norway. *Geological Society of America, Special Papers* 380, 155–183.
- 602 Labrousse, L., Prouteau, G., Ganzhorn, A. C., 2011. Continental exhumation triggered by partial melting
603 at ultrahigh pressure. *Geology* 39 (12), 1171–1174.
- 604 Lee, A. L., Walker, A. M., Lloyd, G. E., Torvela, T., 2017. Modeling the impact of melt on seismic
605 properties during mountain building. *Geochemistry, Geophysics, Geosystems* 18 (3), 1090–1110.
- 606 Levine, J. S. F., Mosher, S., Siddoway, C. S., 2013. Relationship between syndeformational partial melting
607 and crustal-scale magmatism and tectonism across the Wet Mountains, central Colorado. *Lithosphere*
608 5 (5), 456–476.
- 609 Lutro, O., Robinson, P., Solli, A., 1997. Proterozoic geology and Scandian overprinting in the Western
610 Gneiss Region. *Norges Geologiske Undersøkelse* 97, 86.
- 611 Lutro, O., Tveten, E., 1998. Geologisk kart over Noreg Årdal: Trondheim. *Norges Geologiske Undersøking*
612 scale 1:25.
- 613 Mehnert, K. R., 1968. *Migmatites and the origin of granitic rocks*. Elsevier, Amsterdam.
- 614 Mehnert, K. R., Büsch, W., Schneider, G., 1973. Initial melting at grain boundaries of quartz and feldspar
615 in gneisses and granulites. *Neues Jahrbuch für Mineralogie, Monatshefte* 4, 165–183.
- 616 Morales, L. F., Lloyd, G. E., Mainprice, D., 2014. Fabric transitions in quartz via viscoplastic self-
617 consistent modeling part I: Axial compression and simple shear under constant strain. *Tectonophysics*
618 636, 52–69.
- 619 Phillips, E. R., 1974. Myrmekite - one hundred years later. *Lithos* 7 (3), 181–194.
- 620 Phillips, E. R., 1980. On polygenetic myrmekite. *Geological Magazine* 117 (1), 29–36.
- 621 Root, D. B., Hacker, B. R., Gans, P. B., Ducea, M. N., Eide, E. A., Mosenfelder, J. L., 2005. Discrete
622 ultrahigh-pressure domains in the Western Gneiss Region, Norway: Implications for formation and
623 exhumation. *Journal of Metamorphic Geology* 23 (1), 45–61.
- 624 Rosenberg, C. L., Handy, M. R., 2005. Experimental deformation of partially melted granite revisited:
625 implications for the continental crust. *Journal of Metamorphic Geology* 23 (1), 19–28.
- 626 Rosenberg, C. L., Riller, U., 2000. Partial-melt topology in statically and dynamically recrystallized
627 granite. *Geology* 28 (1), 7–10.

- 628 Sawyer, E. W., 1999. Criteria for the Recognition of Partial Melting. *Physics and Chemistry of the Earth*,
629 Part A: Solid Earth and Geodesy 24 (3), 269–279.
- 630 Sawyer, E. W., may 2001. Melt segregation in the continental crust: distribution and movement of melt
631 in anatectic rocks. *Journal of Metamorphic Geology* 19 (3), 291–309.
- 632 Schärer, U., Labrousse, L., 2003. Dating the exhumation of UHP rocks and associated crustal melting in
633 the Norwegian Caledonides. *Contributions to Mineralogy and Petrology* 144 (6), 758–770.
- 634 Schneider, C. A., Rasband, W. S., Eliceiri, K. W., 2012. NIH Image to ImageJ: 25 years of image analysis.
635 *Nature Methods* 9 (7), 671–675.
- 636 Searle, M., 2013. Crustal melting, ductile flow, and deformation in mountain belts: Cause and effect
637 relationships. *Lithosphere* 5 (6), 547–554.
- 638 Simpson, C., Wintsch, R. P., 1989. Evidence for deformation-induced K-feldspar replacement by
639 myrmekite. *Journal of Metamorphic Geology* 7 (2), 261–275.
- 640 Spear, F. S., Kohn, M. J., Cheney, J. T., 1999. P -T paths from anatectic pelites. *Contributions to*
641 *Mineralogy and Petrology* 134, 17–32.
- 642 Straume, Å. K., Austrheim, H., 1999. Importance of fracturing during retro-metamorphism of eclogites.
643 *Journal of Metamorphic Geology* 17 (6), 637–652.
- 644 Stuart, C. A., Piazzolo, S., Daczko, N. R., 2016. Mass transfer in the lower crust: Evidence for incip-
645 ient melt-assisted flow along grain boundaries in the deep arc granulites of Fiordland, New Zealand.
646 *Geochemistry, Geophysics, Geosystems* 17, 3733–3753.
- 647 Terry, M. P., Robinson, P., Hamilton, A., Jercinovic, M. J., 2000. Monazite geochronology of UHP and HP
648 metamorphism, deformation, and exhumation, Nordøyane, Western Gneiss Region, Norway. *American*
649 *Mineralogist* 85 (11-12), 1651–1664.
- 650 Toy, V. G., Prior, D. J., Norris, R. J., 2008. Quartz fabrics in the Alpine Fault mylonites: Influence of pre-
651 existing preferred orientations on fabric development during progressive uplift. *Journal of Structural*
652 *Geology* 30 (5), 602–621.
- 653 Tucker, R. D., Krogh, T. E., Råheim, A., 1990. Proterozoic evolution and age-province boundaries in the
654 central part of the Western Gneiss Region, Norway: Results of U-Pb dating of accessory minerals from
655 Trondheimsfjord to Geiranger. *Mid-Proterozoic Laurentia-Baltica* 38, 149–173.
- 656 Tucker, R. D., Robinson, P., Solli, A., Gee, D. G., Thorsnes, T., Krogh, T. E., Nordgulen, Ø., Bickford,
657 M. E., 2004. Thrusting and extension in the Scandian Hinterland, Norway: New U-Pb ages and
658 tectonostratigraphic evidence. *American Journal of Science* 304 (6), 477–532.

- 659 Tveten, E., Lutro, O., Thorsnes, T., 1998. Geologisk kart over Noreg Ulsteinvik: Trondheim. Norges
660 Geologiske Undersøkelse scale 1:25.
- 661 Vanderhaeghe, O., nov 2009. Migmatites, granites and orogeny: Flow modes of partially-molten rocks and
662 magmas associated with melt/solid segregation in orogenic belts. *Tectonophysics* 477 (3-4), 119–134.
- 663 Vernon, R. H., 1968. Microstructures of high-grade metamorphic rocks at Broken Hill, Australia. *Journal*
664 *of Petrology* 9 (1), 1–22.
- 665 Vernon, R. H., 1991. Questions about myrmekite in deformed rocks. *Journal of Structural Geology* 13 (9),
666 979–985.
- 667 Vernon, R. H., 2011. Microstructures of melt-bearing regional metamorphic rocks. *GSA Memoirs* 207 (01),
668 1–11.
- 669 Vernon, R. H., Collins, W. J., 1988. Igneous microstructures in migmatites. *Geology* 16 (January 2009),
670 1126–1129.
- 671 Walsh, E. O., Hacker, B. R., 2004. The fate of subducted continental margins: Two-stage exhumation
672 of the high-pressure to ultrahigh-pressure Western Gneiss Region, Norway. *Journal of Metamorphic*
673 *Geology* 22 (7), 671–687.
- 674 Walsh, E. O., Hacker, B. R., Gans, P. B., Grove, M., Gehrels, G., 2007. Protolith ages and exhumation
675 histories of (ultra)high-pressure rocks across the Western Gneiss Region, Norway. *Bulletin of the*
676 *Geological Society of America* 119 (3-4), 289–301.
- 677 Walte, N. P., Bons, P. D., Passchier, C. W., 2005. Deformation of melt-bearing systems - Insight from in
678 situ grain-scale analogue experiments. *Journal of Structural Geology* 27 (9), 1666–1679.
- 679 Waters, D. J., 2001. The significance of prograde and retrograde quartz-bearing intergrowth microstructures
680 in partially melted granulite-facies rocks. *Lithos* 56 (1), 97–110.
- 681 Wyllie, P. J., 1977. Crustal anatexis: An experimental review. *Tectonophysics* 43 (1-2), 41–71.
- 682 Yuguchi, T., Nishiyama, T., 2008. The mechanism of myrmekite formation deduced from steady-diffusion
683 modeling based on petrography: Case study of the Okueyama granitic body, Kyushu, Japan. *Lithos*
684 106 (3-4), 237–260.

Table 1: Microprobe data of plagioclase (plag), K-feldspar (kfs), biotite (bte) and hornblende (hbl) compositions for migmatite sample WGR12 and shear zone sample WGR13^a.

Mineral	WGR12: Migmatite					WGR13: Shear zone				
	Qtz	Kfs	Plag	Bte	Hbl	Qtz	Kfs	Plag	Bte	Hbl
wt%	35%	33%	16%	14%	2%	46%	28%	17%	8%	1%
<i>N</i>		16	19	6	11		13	17	11	3
SiO ₂	No analyses	64.242	61.001	36.310	39.275	No analyses	64.647	60.828	35.806	38.879
TiO ₂		0.009	0.005	3.037	1.075		0.018	0.008	3.066	1.076
Al ₂ O ₃		18.835	24.165	17.704	11.609		18.656	24.539	18.552	11.565
Cr ₂ O ₃		0.006	0.005	0.025	0.015		0.007	0.009	0.031	0.016
FeO		0.049	0.114	19.243	21.173		0.034	0.104	19.096	20.949
MnO		0.003	0.008	0.506	0.827		0.008	0.007	0.297	0.834
MgO		0.005	0.007	8.619	7.105		0.004	0.006	8.993	7.012
CaO		0.051	5.987	0.018	11.484		0.058	6.177	0.033	11.452
BaO		0.699	0.015	0.090	0.008		0.671	0.024	0.144	0.000
Na ₂ O		0.973	8.310	0.075	1.455		1.014	8.074	0.052	1.460
K ₂ O		15.136	0.338	9.721	1.971		15.112	0.245	9.725	1.976
Total		100.009	99.955	95.348	95.997		100.228	100.021	95.794	95.218
No. of O		8	8	22	24		8	8	22	24
Si	2.974	2.740	5.539	6.640	2.984	2.729	5.432	6.630		
Ti	0.000	0.000	0.348	0.137	0.001	0.000	0.350	0.138		
Al	1.028	1.279	3.183	2.313	1.015	1.298	3.317	2.324		
Cr	0.000	0.000	0.003	0.002	0.000	0.000	0.004	0.002		
Fe	0.002	0.004	2.455	2.993	0.001	0.004	2.422	2.987		
Mn	0.000	0.000	0.065	0.118	0.000	0.000	0.038	0.120		
Mg	0.000	0.000	1.960	1.791	0.000	0.000	2.034	1.783		
Ca	0.002	0.224	0.002	1.619	0.002	0.231	0.004	1.628		
Ba	0.014	0.000	0.006	0.001	0.014	0.000	0.010	0.000		
Na	0.087	0.724	0.022	0.477	0.091	0.702	0.015	0.483		
K	0.894	0.019	1.892	0.425	0.890	0.014	1.882	0.430		
Total	5.002	4.992	15.476	16.516	4.998	4.980	15.507	16.525		

^aModal proportions for each mineral, including quartz (qtz) are shown by weight % (wt%). *N* is number of spot analyses per mineral, analyses were reject where totals are outside 99.0-101.0% for feldspars and below 95% for biotite and hornblende.

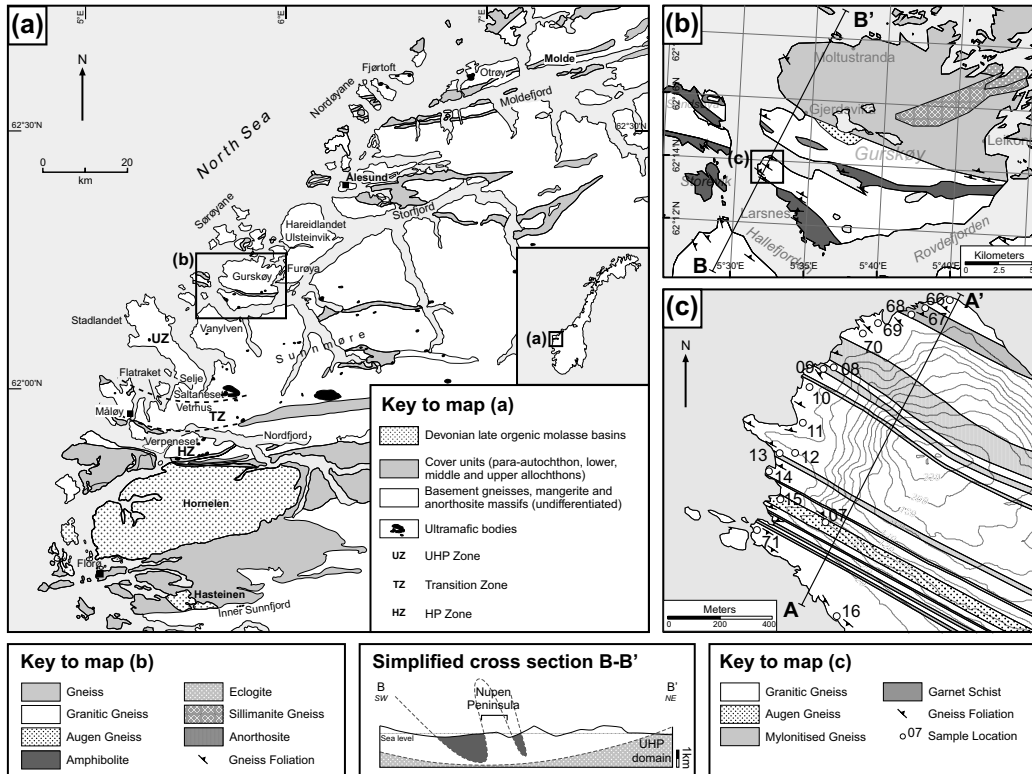


Figure 1: Geological map of (a) Western Gneiss Region with inset detail maps of (b) Gurskøy with section line B-B' and (c) Nupen Peninsula with sample locations and section line A-A' shown in Section 3.2 (Geological maps modified from Kildal, 1970; Lutro et al., 1997; Lutro and Tveten, 1998; Tveten et al., 1998; Carswell et al., 2003; Root et al., 2005).

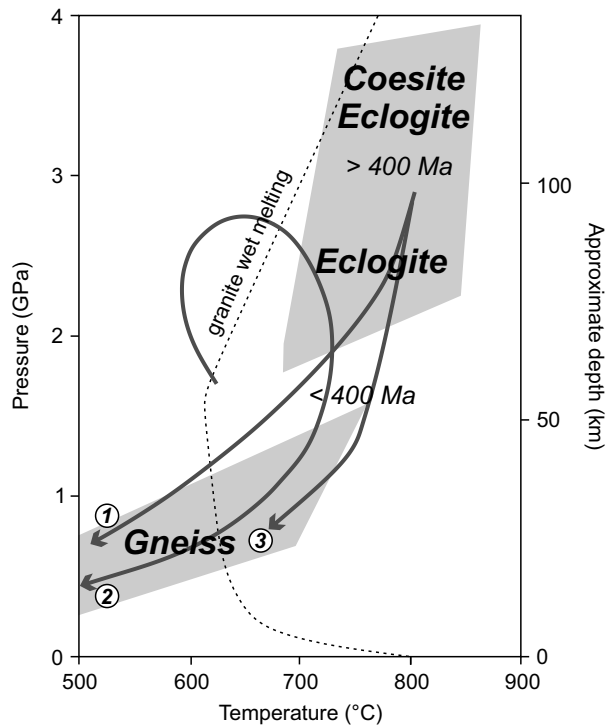


Figure 2: P-T diagram showing the relationship of WGR metamorphic conditions recorded by eclogite and gneiss in the Nordfjord-Stadlandet and Sørøyane UHP domains and surrounding HP domains. P-T paths from 1) Gordon et al. (2016), 2) Labrousse et al. (2002) and 3) Gordon et al. (2013), granite solidus from Auzanneau et al. (2006). Timing of peak UHP and HP metamorphism from Hacker et al. (2010). Note that although both eclogite and gneiss likely experienced UHP conditions, the gneiss equilibrated at much lower PT conditions during decompression.

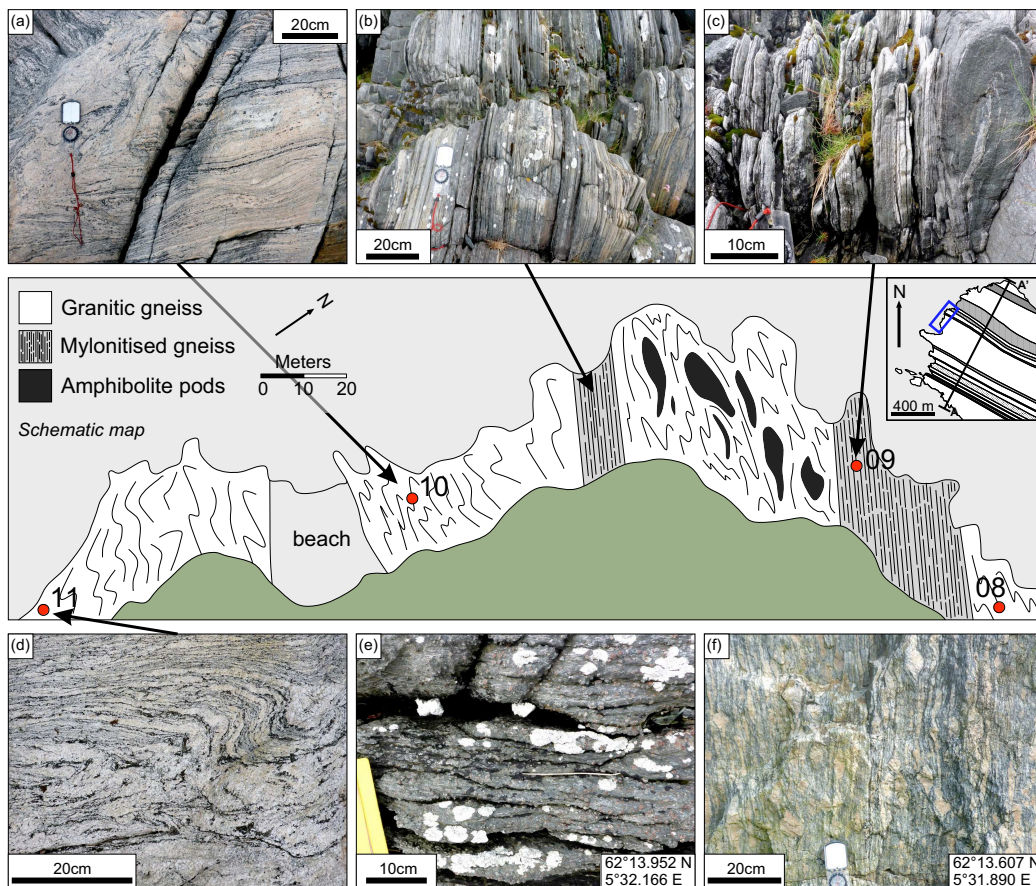


Figure 3: Detailed geological field map showing variation of the gneiss at Nupen, foliation pattern and amphibolite pod location is representative. Photos (a-f) are characteristic outcrop photos of the (a,d) migmatized gneiss and (b,c) mylonitized gneiss and not shown on detailed field map (e) garnet schist and (f) augen gneiss.

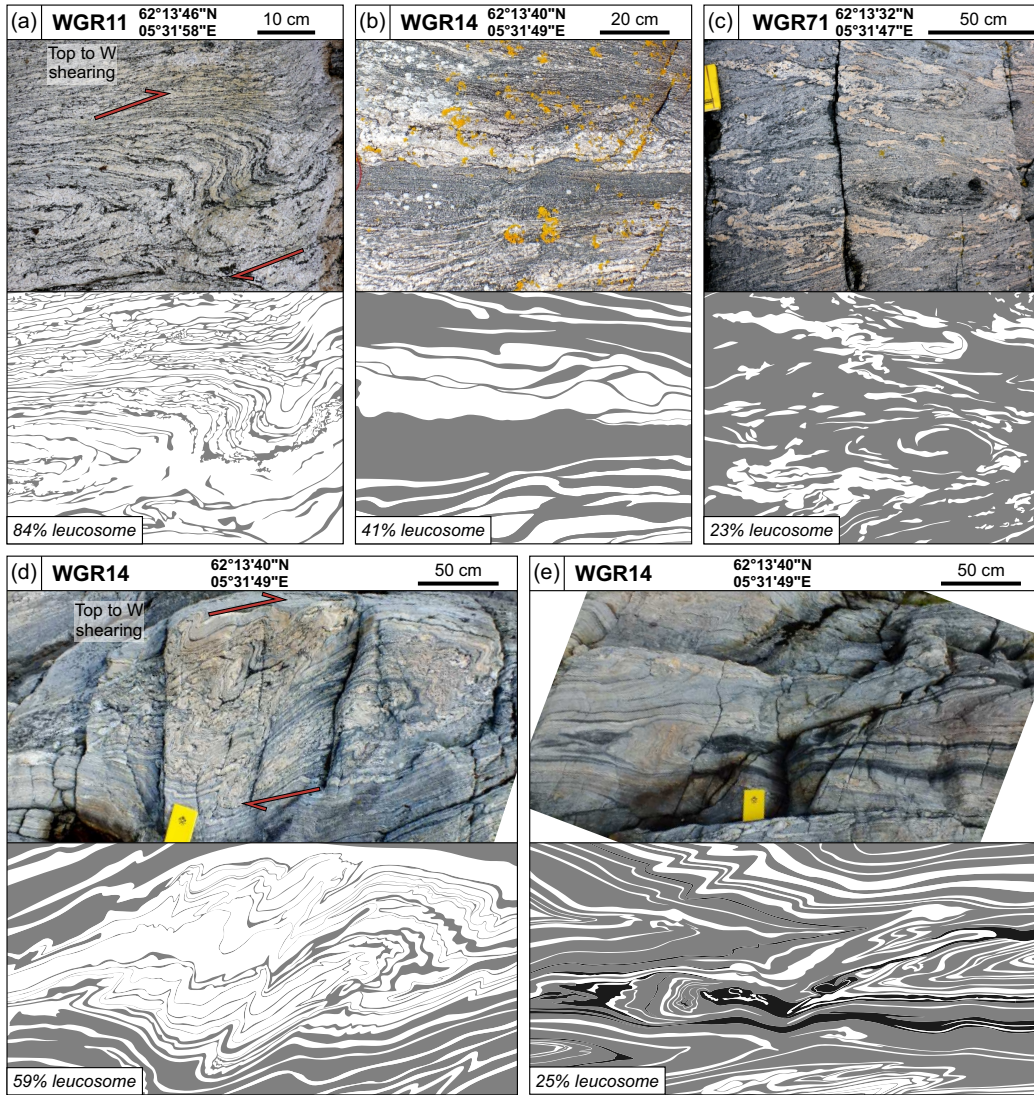


Figure 4: Field photographs with schematic drawings to emphasise the leucosome and melanosome segregation at outcrop scale. Examples shown from the WGR; (a) Leucosome-rich outcrop with top to the west shearing; (b) Ophthalmic layers in leucosome; (c) Surreitic/dilational migmatite; (d) Folded stromatic migmatite with top to the west shearing; (e) Folded stromatic migmatite with leucosome, mesosome and melanosome. Migmatite nomenclature after Ashworth (1985).

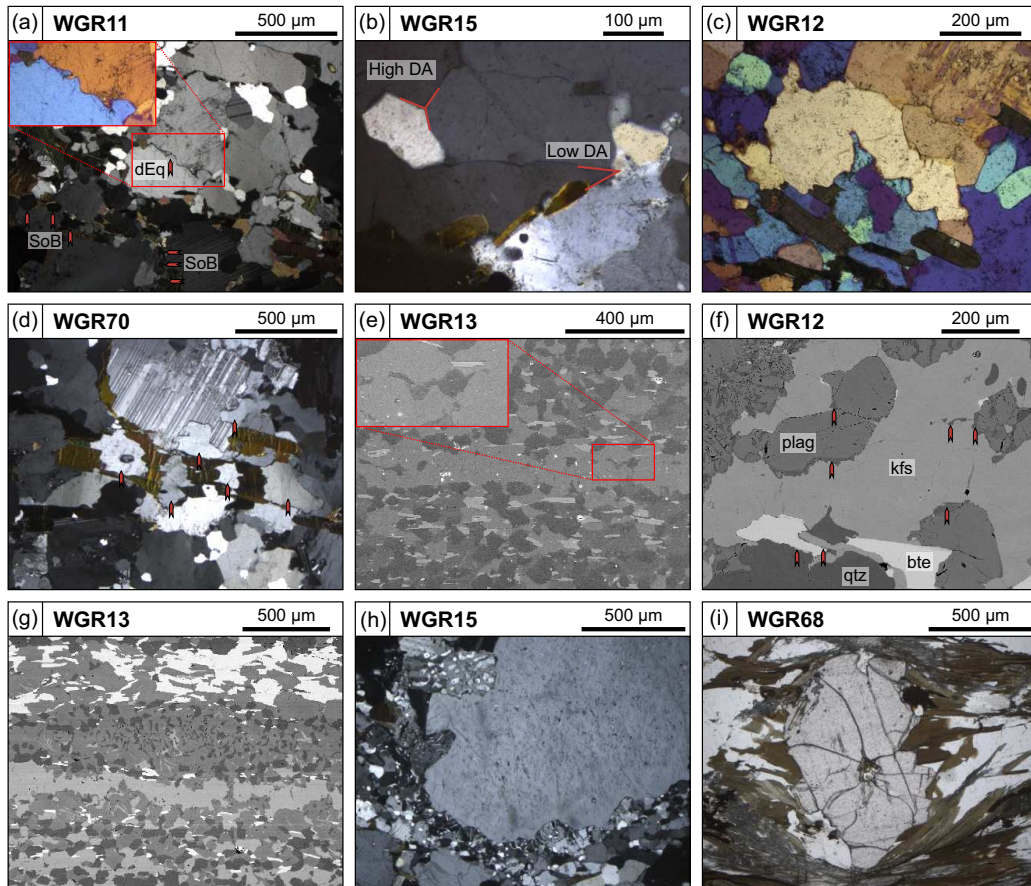


Figure 5: Melt microstructures in migmatitic gneiss from photomicrographs in plain-polarised light (i), cross-polarised light (a, b, d, h), cross-polarised light and gypsum plate (a-inset, c) and SEM-BSE image (e, f, g). (a) Migmatite with plagioclase and quartz grain boundaries in disequilibrium (dEq), string of beads texture (SoB) in quartz along plagioclase grain boundaries indicated by arrows. (b) Augen gneiss with high and low dihedral angles (DA) between quartz grains. (c-d) Migmatite with interstitial quartz melt infilling pore space. (e) Shear zone sample with interstitial quartz melt infilling pore space. (f) Migmatite with quartz (qtz) films along grain boundaries of plagioclase (plag), K-feldspar (kfs) and biotite (bte). (g) Shear zone sample with myrmekite and disequilibrium grain boundaries. (h) Augen gneiss with quartz-plagioclase myrmekite surrounding large K-feldspar grain. (i) Garnet schist with peritectic garnet surrounded by sillimanite, biotite and quartz.

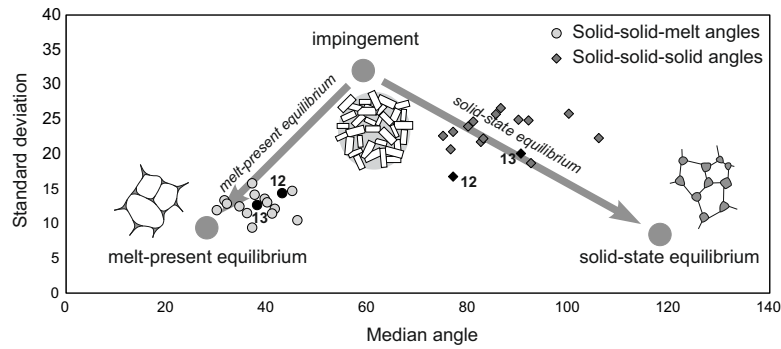


Figure 6: Dihedral angles from the WGR samples for solid-solid-melt boundaries (○) and solid-solid-solid boundaries (◇). Results for samples WGR12 (migmatite) and WGR13 (shear zone) are identified and shown with filled in black symbols. Also shown is a schematic diagram showing end-member solid-solid-melt dihedral angle populations (Adapted from Holness et al., 2005).

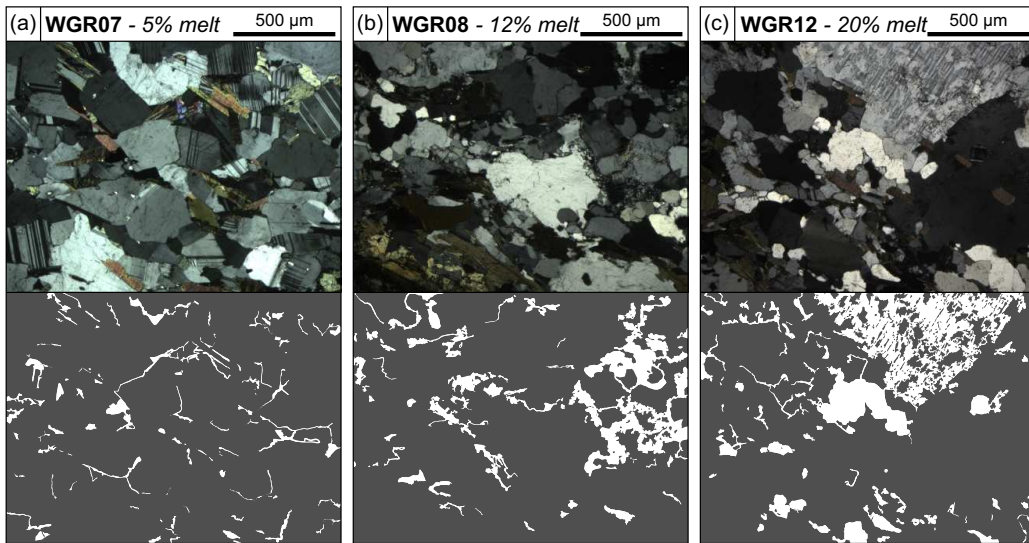


Figure 7: Top: Photomicrographs of melt on grain boundaries in SIP migmatites in plane polarised light. Bottom: Melt (white) vs. solid (grey) image analysis interpretations of each photomicrograph.

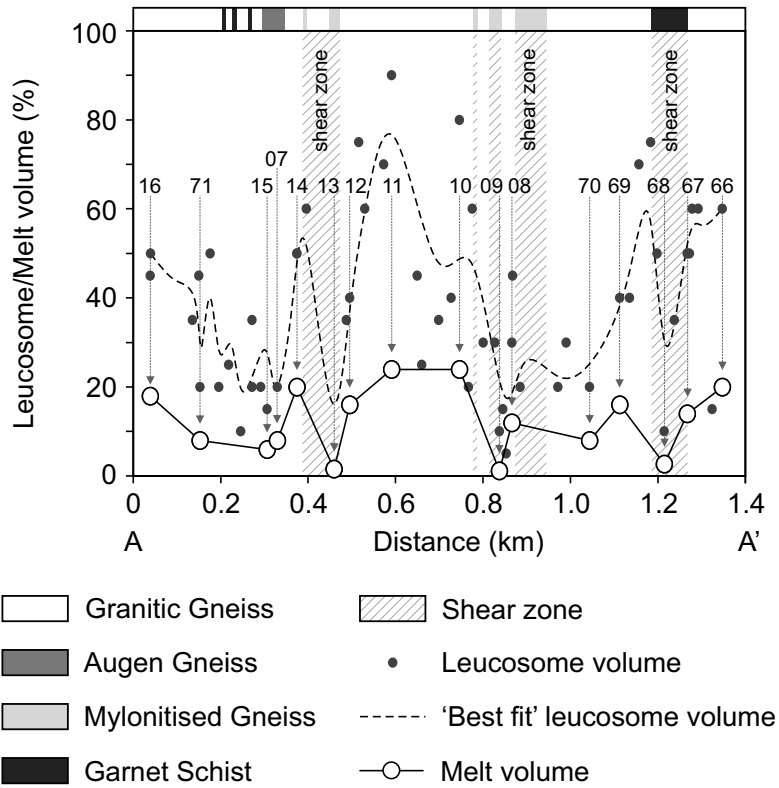


Figure 8: Leucosome and melt volume over the Nupen Peninsula (Figure 1 Map C, section A-A'). Macroscale leucosome volume calculated in the field and via image analysis at the outcrop scale and microscale melt volume calculated from image analysis of thin sections and SEM images.

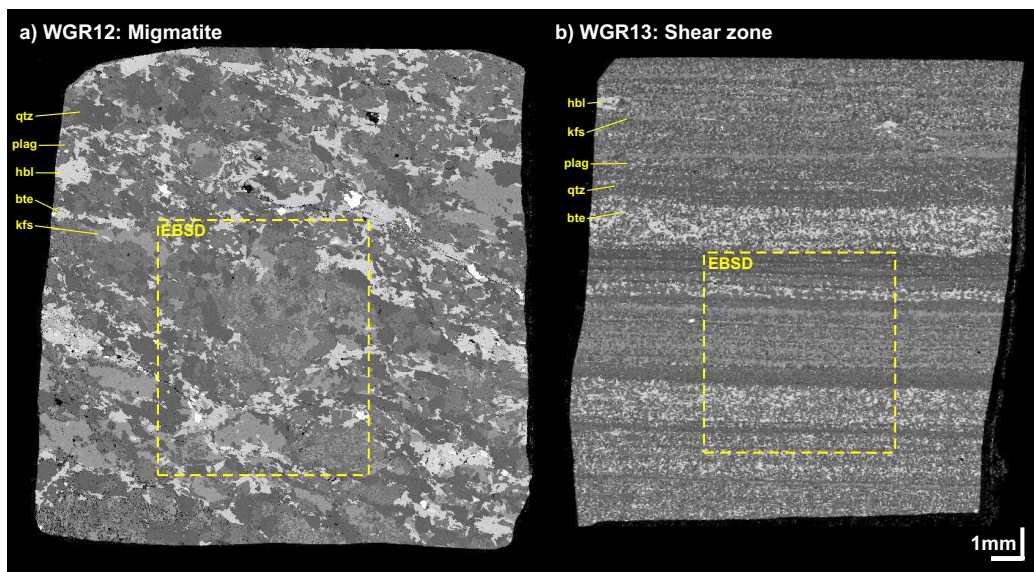
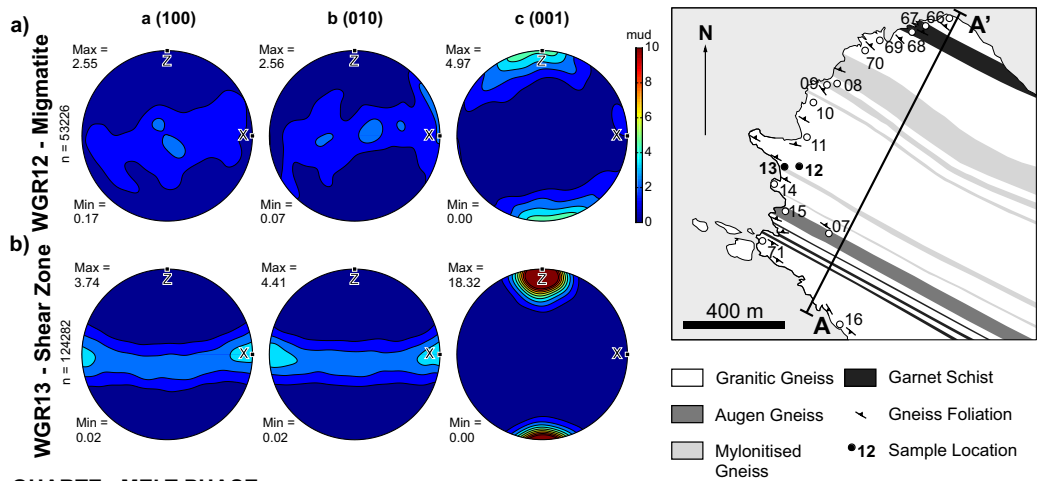


Figure 9: Backscattered electron maps of (a) migmatite sample WGR12 and (b) shear zone sample WGR13. Maps are at the same scale with 1 mm bar shown for scale. Dashed yellow boxes indicate area of EBSD analysis.

BIOTITE - SOLID PHASE



QUARTZ - MELT PHASE

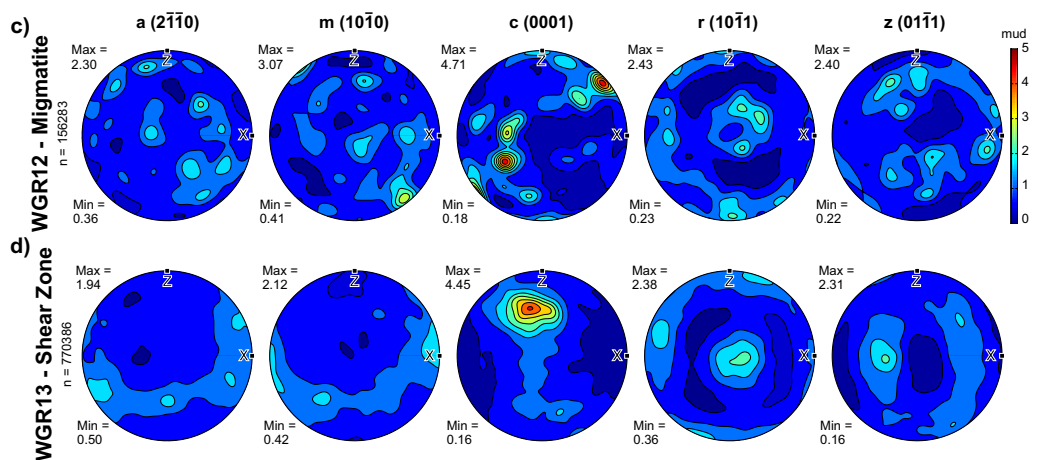


Figure 10: EBSD-derived CPO of biotite (a, b) and quartz (c, d) in kinematic coordinate system for migmatized gneiss sample WGR12 (a, c) and mylonitized gneiss sample WGR13 (b, d). All stereographic projections are lower hemisphere and CPO are contoured in terms of multiples of the uniform distribution (mud). Inset map shows locations of adjacent mylonitized and migmatized gneiss samples.

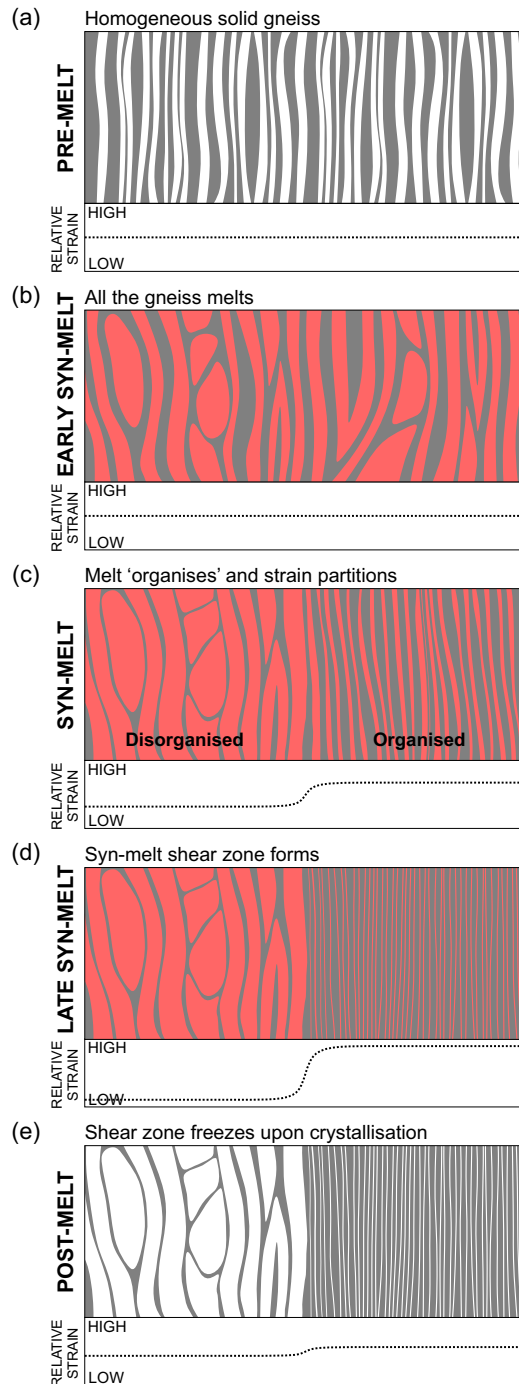


Figure 11: Schematic diagrams to show how strain localisation can vary during syn-melt shearing of a migmatized gneiss. Under each stage of the models are relative strain profiles indicating areas of strain localisation. (a) Homogeneous solid gneiss where strain is distributed evenly; (b) melting is pervasive throughout the leucosome; (c) melt organises into layers whereas elsewhere it remains as disorganised 'pools'; (d) runaway organisation effect produces syn-melt shear zones; (e) structure freezes upon crystallisation giving a mylonite-like outcrop style in the field.