The microstructure of layered ultramafic cumulates: Case study of the Bear Creek intrusion, Trinity ophiolite, California, USA

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22   **Abstract**

23   In the Trinity ophiolite, California, USA, several mafic-ultramafic plutons intruded a

24   peridotitic host 435 to 405 m.y. ago in a tectonic setting interpreted as an arc-related

25   spreading centre. One of these intrusions, in the Bear Creek area, exposes basal ultramafic

26   cumulates with igneous layering comprising an alternation of uncommonly thin (down to a

27   few mm) layers of dunite, peridotite and pyroxenite that might be specific to this tectonic

28   setting. These layers offer an excellent opportunity to characterise the microstructure of

29   uncommon cumulates from the lower crust using EBSD (Electron Backscatter Diffraction).

30   This "high-frequency" layering rests on underlying lherzolites and grades upward to more

31   massive pyroxenites (i.e. clinopyroxenites with minor olivine-rich layers). Our field

32   observations and data from the Bear Creek cumulates together with the preservation of

33   magmatic features suggest the environment was tectonically stable after the emplacement of

34   the cumulates. A detailed microstructural investigation of all minerals from the Bear Creek

35   cumulates allows us to decipher their magmatic and plastic deformation history. In a

36   structural reference frame defined by the compositional layering and the elongation direction

37   of the surrounding host peridotites, olivine in the cumulates presents a [010]-fibre fabric and

38   rarely a [001](010) fabric. Clinopyroxene shows a concentration of [010] axes normal to the

39   layering plane with [100] and [001] defining girdles. Orthopyroxene mostly has a fabric with

40   [100] and/or [010] subnormal to the layering plane and [001] scattered along a girdle in the

41   plane of layering. All minerals show a strong fabric. We interpret the formation of the

42   developed planar microstructures as a result of magmatic processes, with high contribution of

43   crystal settling. To a lesser extent, compaction could have been operating and may be linked

44   to the rare evidence of plastic deformation. Clusters of axes within the girdles of olivine and

45   pyroxene CPOs preferentially appear close to the direction of elongation of the surrounding

46   peridotites (i.e. N115°). EBSD analysis of the shape-preferred orientation of Bear Creek's

minerals revealed a preferential alignment of the olivine and cpx long axis with the N°115 direction. This magmatic lineation and the preferred direction in the CPOs girdles are both consistent with the stretching lineation acquired during solid-state deformation by the mantle peridotite of the Trinity ophiolite. We suggest that a weak magma flux early on and/or an ongoing but limited regional stress could be responsible for these clusters. Although a direct coupling between asthenospheric flow and magmatic flow cannot be invoked in this context of melt intrusion in the lithosphere, this result highlights that the stress field applied on the mantle could have been still active and similar during the formation of Bear Creek intrusion. Our new field and microstructural data, together with previously presented petrological data, fit a scenario for the evolution of the Trinity ophiolite in which a mantle segment was intruded by a single large batch of primitive boninitic-andesitic melt. Our results emphasise the importance of considering the initial magmatic microstructures and the original shape anisotropy when investigating later deformation in ultramafic rocks.

1. Introduction

In the past decades, pyroxenites have been the subject of numerous studies as their geochemical composition and microstructure provide important clues to petrological processes related to the development of heterogeneities in the Earth's upper mantle (e.g. Bodinier and Godard, 2003; O'Reilly and Griffin, 2013). Mineral fabrics in upper-mantle peridotites and pyroxenites are commonly interpreted as clues to solid-state ductile deformation that occurred by mechanisms such as dislocation and diffusion creep (e.g. Avé Lallemant and Carter, 1970; Frets et al., 2012; Nicolas and Christensen, 1987). When formed by dislocation creep, mineral fabrics will reflect the dominant slip system (or a combination of multiple slip systems each having a specific activity) that was active during the deformation episode (e.g. Ismaïl and Mainprice, 1998; Karato, 2008). The dependence of the activation of a slip system on parameters such as temperature, pressure, chemical composition or the presence of water in the mineral structure, which has been demonstrated by experimental studies, can be used to better characterise the deformation history of a rock and better constrain regional and global geodynamic models.

In contrast to pyroxenite layering in the mantle, the microstructures of pyroxene-rich intervals in lower-crustal mafic cumulates have received less attention from petrologists (e.g. Boorman et al., 2004; Feinberg et al., 2006; Higgins, 1998, 2015; Williams et al., 2006). More specifically, the crystallographic preferred orientation (CPO) of pyroxenes in crustal pyroxenites has not yet been investigated using electron backscatter diffraction (EBSD) methods (e.g. Prior et al., 1999; Wright et al., 2011). Pyroxenitic cumulates are mostly present in the lower-crustal section of ophiolites and in layered magmatic intrusions (e.g.

90   Benn and Allard, 1989; Cawthorn and Walraven, 1998; Ceuleneer and Le Sueur, 2008;

91   Clénet et al., 2010; Ying et al., 2013).

92   In magmatic intrusions, the rocks can record both magmatic and deformational processes that

93   generate preferred grain orientation. The difficulty when studying this kind of lithologies is to

94   differentiate between the two processes, which could be overprinted by one another. Previous

95   studies reported that the microstructure of cumulates may result from a sequence of processes

96   including crystal nucleation, crystal settling and the subsequent compaction of the mush

97   (Cheadle and Gee, 2017; McBirney and Nicolas, 1997; Nicolas, 1992). Fabric generated

98   during magma flow aligns grains in a specific direction, thus conferring a planar-linear fabric

99   on the rock, which then shows both a shape- and a crystallographic preferred orientation

100   (CPO) (e.g. Boudreau and McBirney, 1997; Charlier et al., 2015; Namur et al., 2015 and

101   references therein; Vukmanovic et al., 2018; Williams et al., 2006). In this case, olivine

102   crystals are generally elongated on [001] and show a correspondence between (010) and

103   macroscopic markers like layering and igneous foliation (e.g. Benn and Allard, 1989;

104   Boudier, 1991). A similar behaviour has been reported for the crystallographic axis of

105   clinopyroxenes (cpx) with their [001] axes clustering along the magmatic lineation, which is

106   itself contained in the magmatic foliation coinciding with the (010) plane (e.g. Benn and

107   Allard, 1989). During the subsequent magmatic history of the sample, the grains' SPO can be

108   modified by late overgrowth that could change the primary fabric (e.g. Brothers, 1964;

109   Cheadle and Gee, 2017). With decreasing proportions of melt, the deformation regime

110   approaches the solid state and possible grain boundary sliding assisted by diffusion-

111   precipitation processes at low stress will take place (e.g. Paterson, 2001). Finally, when the

112   porosity becomes low enough, compaction occurs and tends to form a planar fabric; at this

113   point the dominant deformation process becomes dislocation creep and leads to the formation

114   of a CPO driven by plastic deformation .

115 Depending on the relative contribution of the processes involved, the respective strengths of

116 the planar and linear component of the magmatic fabrics may vary. Consequently, the

117 symmetry of the resulting fabric will range from an orthorhombic symmetry to a fibre

118 symmetry (e.g. Benn and Allard, 1989; Biedermann et al., 2016; Feinberg et al., 2006). The

119 symmetry of a magmatic fabric can thus be used as an indicator of the magmatic-flow

120 geometry and can be related to magma-chamber processes (e.g. Higgins, 2015).

121 Unfortunately, in natural samples, it is common that the initial magmatic fabric is overprinted

122 by a later event of solid-state flow. This overprint results in a new fabric that carries different

123 information from the original magmatic fabric (e.g. Satsukawa et al., 2013).

124

125 In the Klamath Mountains, California (USA), the Trinity ophiolite crops out as a large

126 peridotite body intruded by a dozen ultramafic-mafic plutons of sizes from a few km to 10

127 km, exposed at different structural levels from the lower crust to the middle crust) Boudier et

128 al., 1989; Cannat and Lécuyer, 1991; Ceuleneer and Le Sueur, 2008). Among them, the Bear

129 Creek pluton exposes cumulates that crystallised in the vicinity of the mantle/crust boundary

130 (Ceuleneer and Le Sueur, 2008). As the Trinity ophiolite did not experience extensive

131 regional metamorphism after its formation, it provides a unique opportunity to study the

132 macro- and microstructure of pyroxene-rich cumulates and to observe magmatic and

133 deformational processes occurring in magma chambers in a supra-subduction zone. In this

134 work, field observations and EBSD data are used to characterise the microstructure and CPO

135 of cumulate minerals from the Bear Creek pluton and its host peridotite. This constitutes the

136 first EBSD study of crustal pyroxenites from a magmatic intrusion. These results are then

137 used to discuss the interplay of magmatic and deformational processes in the development of

138 mineral fabric in this context.

139

2. Geological background

The Trinity ophiolite is the largest (3,000 km^2) and best-preserved exposure of mantle mafic-ultramafic rocks cropping out in California. It is included in the Paleozoic Terranes accreted to the North American Cordillera during early Devonian times (e.g. Lindsley-Griffin, 1977). Although its precise tectonic setting and emplacement history are poorly constrained, its association with arc-derived volcano-sedimentary formations together with geochemical and geochronological data suggest that it is a dismembered fragment of a former arc-related ocean basin that may have opened between Rodinia and Laurentia; the Trinity ophiolite itself could represent the youngest remnant of this oceanic domain (Brouxel et al., 1988; Coleman, 1986; Mankinen et al., 2002; Wallin et al., 1988; Wallin and Metcalf, 1998).

The Trinity ophiolite consists of an incomplete ophiolite sequence comprising mantle peridotites pockmarked by different plutons, which range from mafic-ultramafic to gabbroic and granodioritic in composition and several kilometres in size (Fig. 1A, Cannat and Lécuyer, 1991; Ceuleneer and Le Sueur, 2008). The mantle peridotites have a porphyroclastic texture that is typical of so-called mantle tectonites and that was developed under high temperature (close to the solidus) and low deviatoric stress, which have been attributed to asthenospheric flow (Le Sueur et al., 1984). An image of the flow geometry at the time it stopped was obtained by mapping the foliation and lineation imprinted by plastic deformation. The global survey of Le Sueur et al. (1984) revealed a rather organised pattern characterised by a sub-vertical foliation bearing a sub-horizontal stretching lineation, both trending NW-SE, i.e. oriented parallel to the inferred azimuth of the paleo-spreading axis. Local deviations from this regional pattern are frequent; some of them can be attributed to the development of late mylonitic shear zones.

165

166   The Trinity peridotites are mostly of lherzolitic composition and resulted from a moderate

167   degree of decompression melting which started in the spinel lherzolite facies and ended in the

168   plagioclase lherzolite facies (Quick, 1981). Patches of plagioclase lherzolite are quite

169   represent about 10% of the surface of the Trinity peridotite (Quick, 1981). Local

170   concentrations of plagioclase (+/- cpx and spinel) attest to melt migration and segregation

171   contemporaneous with the end of plastic deformation (Ceuleneer and Le Sueur, 2008).

172

173   Abundant channels and dykes of pyroxenite (clinopyroxenite and cpx-rich websterites)

174   crosscut the Trinity peridotites. The pyroxenites crystallised from a boninitic-andesitic melt

175   (Mg / Si-rich and Ti-poor) derived from another mantle source that is not the Trinity

176   lherzolite as presently exposed (Ceuleneer and Le Sueur, 2008). These boninitic-andesitic

177   melts were in disequilibrium with the plagioclase lherzolite at low pressure as attested by the

178   melt/rock reaction features exposed. Harzburgites and dunites crop out at the margins of the

179   pyroxenite porous flow channels and dykes, to distances that can reach several meters

180   (Dygert et al., 2016; Kelemen et al., 1992). Reactive dunite with minor harzburgite also

181   occurs as bodies a few tens of meters thick and hundreds of meters in lateral extent. They

182   have been called "tabular dunites" by Quick (1982) in reference to their shape and their

183   subhorizontal orientation. The large compositional diversity of the Trinity peridotite at

184   various scales can be ascribed to the cryptic effects of the interaction between the moderately

185   depleted mantle lherzolite and the ultra-depleted melt that crystallised as pyroxenite dykes

186   (Ceuleneer and Le Sueur, 2008). The overall tilt of the Trinity massif, at the regional scale,

187   was found to be relatively moderate according to the orientation of the contacts between the

188   different units making up the ophiolite and to the sub-vertical dip of the diabase dykes at

189   upper-crustal levels (Le Sueur et al., 1984).

190

191 Among the plutons intrusive into the Trinity peridotite, granites and granodiorites were

192 emplaced during Mesozoic times and are not related to the accretion history of the ophiolite.

193 The mafic and ultramafic plutons are much older (405 to 435 Ma) (Wallin et al., 1995;

194 Wallin and Metcalf, 1998) and their composition shows clear affinities with that of the

195 pyroxenite dykes intruding into and reacting with the peridotite, which most likely belong to

196 the same magmatic episode (Ceuleneer and Le Sueur, 2008). They are differentiated, with a

197 lithological succession consistent with the classical ophiolite igneous stratigraphy i.e., from

198 the base to the top, primitive layered cumulates, more evolved cumulates devoid of

199 developed layering and diabase dykes interpreted as the root of a sheeted dyke complex that

200 is not preserved due to erosion. Contrasting with the classical ophiolitic model, these crustal

201 units present a limited lateral extent (a few km at most) and were emplaced at various depths

202 within the peridotite (Cannat and Lécuyer, 1991; Wallin and Metcalf, 1998).

203

204 The Bear Creek area exposes the most complete and continuous section through one of these

205 mafic and ultramafic plutons (Ceuleneer and Le Sueur, 2008). Its lower contact with the

206 mantle-derived rocks (i.e. of lherzolitic composition) is affected by a late fault where the

207 Trinity River flows (Fig. 1B). The lateral contact with the host rock is underlined by a screen

208 of pegmatite made up essentially of large crystals of cpx (up to a few decimetres in length)

209 partly replaced by magmatic hornblende and lower-temperature alteration minerals. In the

210 pluton, the basal cumulate section is ~500 m thick and comprises an alternation of cm-thick

211 layers of quite variable composition (dunite, wehrlite, pyroxenite, lherzolite…). Plagioclase

212 is typically absent from this horizon; the modal abundance of olivine gradually decreases

213 from bottom to top and it is virtually absent from the uppermost part of the cumulate pile.

214 Plagioclase appears above this ultramafic sequence, in the upper half of the cumulate section,

215   in a diffuse way at first (plagioclase pyroxenites), becoming increasingly abundant toward the

216   top of the section. Its crystallisation always coincides with that of hornblende. The top of the

217   cumulate sequence is characterised by abundant magmatic breccias crosscut by diabase

218   dykes, and is not shown in Fig. 1B (Le Sueur et al., 1984).

219

220   Ceuleneer and Le Sueur (2008) attributed the aforementioned crystallisation sequence and the

221   bottom-to-top evolution in modal and mineral compositions to the intrusion of a single batch

222   of boninitic-andesitic melt (high MgO, silica and water content, low Al_2O_3 and low TiO_2)

223   into the Trinity peridotites in a supra-subduction setting. Furthermore, the occurrence of

224   diabase dykes at the top of the cumulate sequence were interpreted as representing the roots

225   of a sheeted dyke complex and suggests that the magma injection occurred at relatively low

226   pressure (<< 1 GPa) in a spreading setting (Le Sueur et al., 1984).

227

228   **3. Field observations**

229

230   In the Bear Creek plutonic section, the basal level (~500 m thick) is made up of serpentinised

231   dunite layers alternating with pyroxene-rich and serpentine-free layers varying in

232   composition between websterites, wehrlites and lherzolites. The thickness of these layers

233   ranges from several hundred micrometres (i.e. similar to the grain size) to several

234   centimetres. We will refer to this compositional alternation between dunite and pyroxene-rich

235   layers as "finely layered cumulates" (Figs. 2, 3A, 3B and Ceuleneer and Le Sueur, 2008).

236   The compositional layering is marked and is the unique macroscopic structural feature visible

237   in the field. Its strike ranges from N115° to N85° with an associated dip of 30° to 40°S. The

238   lateral extent of the layers exceeds several tens of metres and oblique structures made of

239   clinopyroxenitic intrusions, generally with a coarser-grained texture, were observed.

240   Peridotitic enclaves and pyroxene megacrysts (e.g. sample T3BC3A, Ceuleneer and Le

241   Sueur, 2008) also occur throughout this cumulate lower section, locally interrupting the

242   compositional layering (i.e. at the hand specimen scale and thin section scale respectively;

243   Supplementary Material B).

244   Towards higher stratigraphic levels, the compositional layering becomes increasingly

245   difficult to observe due to the progressive disappearance of olivine-rich layers (Figs. 1B and

246   3E). This results in seemingly more massive layers (up to ~250 m thick) that display a

247   variable but always high cpx / orthopyroxene (opx) ratio and that we will refer to as ¨massive

248   cumulates¨ (Fig. 3C and D). The strike of the layering, where still measurable, remains in the

249   same range as that of the finely layered cumulates but its dip is more variable, ranging from

250   60 to 20°S.

251

252   The country rock in which the cumulates are embedded is of lherzolitic composition (Figs.

253   2A, 3E and 3F). The lherzolite is highly serpentinised and contains relatively fresher

254   subparallel 1-5 cm thick pyroxenite veins highlighting a foliation which is also visible thanks

255   to pyroxene grains (Fig. 3E). The foliation of the lherzolites ranges between N115º and N45°

256   30-40°S. In contrast to the cumulates, a mineral elongation was identified in the lherzolite

257   and remains sub-horizontal. Figure 2A shows the contact between the lherzolite and the

258   finely layered cumulates which is unfortunately always hidden by loose blocks. A clear

259   contact between the finely layered cumulates and the lherzolite foliation was only observed

260   on a loose block, and was discordant. It is noteworthy that we did not observe any plagioclase

261   within the Bear Creek finely layered cumulates or lherzolite.

262

263   In some locations along the section that we sampled, clinopyroxenite sills and dykes locally

264   crosscut both the lherzolite foliation and the compositional layering of the cumulates (Fig. 3A

265 and 3E). They appear to be dendritic extensions of meter-scale pockets of clinopyroxenitic

266 intrusions that crop out locally in the lherzolite. The clinopyroxenitic dikes and sills can

267 exhibit (1) a pegmatitic texture with idiomorphic, centimetric grains or (2) a finer-grained

268 texture comparable to that of the massive cumulates. The two textures can coexist within a

269 single clinopyroxenite intrusion with the pegmatitic texture making up the core of the dyke

270 and the finer-grained texture appearing at the rim.

271

272

273 **4. Sample collection and analytical methods**

274

275 **4.1. Sample collection and preparation**

276

277 Our sampling of the Bear Creek intrusion and its host rock covers a range of lithologies

278 including the host mantle section (i.e. lherzolite) and the finely layered and massive

279 cumulates that crop out in the basal section of the intrusion. At Bear Creek, sampling was

280 done along a section roughly perpendicular to the compositional layering of the cumulates

281 (Fig. 1B and Tab. 1). Two lherzolites, ten finely layered pyroxene-rich cumulates, two

282 massive pyroxenites and two clinopyroxenitic sills were selected for microstructural

283 characterisation by EBSD. The samples are described in detail in Table 1 and Supplementary

284 Material A, including GPS coordinates and microstructural data.

285

286 For samples in which no lineation could be observed in the field nor during sample

287 preparation, thin sections were systematically cut in the plane normal to the foliation and

288 containing the N115° direction. This direction corresponds to the azimuth of the regional

289 stretching elongation in the mantle peridotites and to the strike of the intrusive diabase dykes,

290 as inferred from the structural studies of Le Sueur et al. (1984) and Boudier et al. (1989).

291 This direction was eventually used to relate the microstructure of the cumulate with that of

292 the neighbouring peridotites. A suite of samples (i.e. 01T12A, 01T13A, 01T13B, 01T13C,

293 03T46B and 02T117A) collected in a previous field expedition was cut perpendicular to the

294 compositional layering but was not oriented relative to the Trinity structural framework. For

295 all samples, mineral fabrics were interpreted after checking the coincidence of the azimuth of

296 the crystallographic-axis directions of at least two phases; petrofabrics were not interpreted

297 where this criterion of internal consistency could not be verified.

298

299 **4.2. Analytical methods**

300

301 During sample preparation, a colloidal silica-water solution (mixed in a 70:30 proportion)

302 was used for 5 min following mechanical polishing, in order to reduce surface damage. Thin

303 sections were then coated with a ~ 6-nm-thick carbon layer to prevent charging during

304 analysis. Crystallographic orientations of minerals were acquired using an HKL NordlysNano

305 EBSD detector attached to a Zeiss EVO MA15 scanning electron microscope in the

306 Geochemical Analysis Unit (GAU) at GEMOC/CCFS, Macquarie University, Australia. For

307 all analyses, a 15-μm step size and a 20-kV acceleration voltage were used; working distance

308 ranged between 15 and 20 mm. Automatic indexing was performed using the AZtec software

309 (Oxford Instruments). The HKL software was used to perform a standard noise reduction and

310 to extrapolate missing data with at least 8, 7, 6 and then 5 identical neighbours. Resulting

311 EBSD data were then processed in MATLAB using the MTex toolbox (Bachmann et al.,

312 2010b, 2011). A value of 10° as the cut-off angle for the determination of the grains was

313 used. Further details on the data processing protocol in MTex can be found in Henry et al.

314 (2017) and Henry (2018).

315

To avoid oversampling of large grains, pole figures were plotted with one point per grain on a

lower-hemisphere, equal-area stereographic projection. For strongly serpentinised samples (3

samples), pole figures were produced using available measurements, which proved to yield

results comparable to those of a standard one-point-per-grain model (Achenbach et al., 2011)

(see Supplementary material A for the number of grains and the type of representation).

When twinning in cpx was detected during data processing, pole figures of cpx were drawn

using all the available measurements.

Orientation density functions (ODF) have been calculated with a 10° Gaussian half-width for

samples with > 100 grains/individual measurements. The strength of the crystallographic

preferred orientation was estimated using both the J- and M-indices (Bunge, 1982; Skemer et

al., 2005) and the appropriate ODF (i.e. calculated after all the measurements if twinning or

serpentinisation occurred otherwise using one point per grain). In addition, the misorientation

of each pixel was calculated relative to the mean orientation of its corresponding grain, and

then combined as maps of misorientation relative to mean grain orientation. Any

misorientation value greater than 10° away from the mean orientation was equated to 10° to

allow for a clearer observation. Orientation tensors were also calculated for each

crystallographic axis of each phase of interest using the method of Bachmann et al. (2010a).

For each orientation tensor, eigenvalues for [100], [010] and [001] of olivine, cpx and opx

were extracted and used to quantitatively describe the fabric symmetry using the BA- and

LS-indices (Mainprice et al., 2015; Ulrich and Mainprice, 2005). In order to quantify a

possible shape-preferred orientation (SPO) of the grains that may not have been obvious in

thin section, we used EBSD data to quantify their shape. To do so, we fitted an ellipse on

every grain and extracted the angle between the long axis of that ellipse and the X direction

340   (i.e. N115 direction in the layering plane). Results are then plotted as a multiple rose diagram

341   with each data point being weighted by the surface area of the grain.

342

343   **5. Petrography and microstructures of the Bear Creek intrusion and its country rocks**

344

345   **5.1. Lherzolitic country rock**

346

347   The lherzolites are highly serpentinised, mostly preserving opx porphyroclasts (~ 2 mm),

348   which are surrounded by fine-grained (~50 µm) olivine and minor cpx (Fig. 3F, samples

349   T4BC2 and T4BC3). Undulose extinction, sub-grain boundaries and cpx exsolution along

350   cleavages are observed in the large grains. Olivine is coarse-grained (~3 mm) away from

351   opx-rich aggregates, and free from observable deformation markers. No obvious shape-

352   preferred orientation of the grains is visible in thin section.

353

354   **5.2. Finely layered cumulates**

355

356   In some of our samples (e.g. T3BC1A, Fig. 3B, 4A and Table 1), olivine, opx and cpx are

357   relatively homogeneously distributed. However, in other samples high-frequency alternation

358   between olivine-rich and pyroxene-rich millimetre-thick layers is visible (Fig. 4B and 4C).

359   The transition between layers is sharp and no gradual change in the mode of olivine is seen.

360   Mineral modes can vary strongly within each compositional layer, within the sample and

361   from one sample to the others (Fig. 4 and Ceuleneer and Le Sueur, 2008). Grain-size

362   distribution is rather homogeneous with a mean of ~ 400 µm for all phases. Grain boundaries

363   between phases are curved/lobate and 120° contacts are common. Grain boundaries can be

364   either straight or curvilinear between grains of same phase and 120° triple junctions are

365   common. We did not preferentially observe either convex or concave grain boundaries of a

366   specific phase at contacts between two phases. Some olivine and cpx grains can be elongated

367   on the layering plane but no clear shape-preferred orientation is seen for any phase. Only rare

368   deformation markers can be observed in the finely layered cumulates (i.e. straight sub-grain

369   boundaries in olivine, and to a lesser extent in cpx and opx; Fig. 5A). Cpx often presents

370   twins with continuous to stepped twin-boundaries crosscutting the whole grain in most cases.

371   At the thin section scale, macrocrysts of opx (~ 1 cm) can bend the layer contacts and can

372   cross over multiple layers (e.g. T3BC3A, Supplementary Material B). Amphibole appears as

373   small interstitial grains (~150 μm), often associated with cpx. An amphibole-rich vein

374   crosscutting the cumulates was observed in sample 01T13A (Fig. 4C).

375

376   **5.3. Massive cumulates**

377

378   In contrast to the finely layered cumulates, this lithology is dominated by coarse-grained

379   minerals (~ 500 μm to 1 mm) made up essentially of cpx alternating with rare olivine with

380   minor opx (i.e. compositional layering visible in the field, Fig. 3E). Cpx grain boundaries are

381   mostly lobate, but can also be locally straight (i.e. Fig 5B). In the finer-grained fraction (~ 50

382   μm grain size, <5% of the sample surface area), olivine and opx assemblages appear with

383   mainly lobate/curved grain boundaries. Triple junctions are visible between most cpx grains

384   and indicate a local textural equilibrium. Internal deformation in the form of subgrains and

385   undulose extinction is observed in olivine, and sometimes in cpx grains. EBSD-derived maps

386   reveal that concentric misorientation is present in some cpx grains (Fig. 5B). As in the finely

387   layered cumulates, cpx macrocrysts (~ 1 cm) locally occur and twinned cpx are also observed

388   (sample T4BC4A, Supplementary Material B).

389

5.4. Shape-preferred orientation of the Bear Creek cumulates

The shape-preferred orientation (SPO) of the grains in the Bear Creek cumulates was investigated by EBSD mapping. In the finely layered cumulates, our results show a clear elongation of the cpx and olivine grains in the layering plane (Fig. 6A). Opx present an SPO with a preferred direction of elongation normal to the layering plane.

In the massive cumulates, the quantification of cpx SPO shows a range of orientations from sub-parallel to olivine and cpx in the finely layered cumulate (sub-parallel to N115 direction) to a maximum at a small angle to the lineation (Fig. 6B).

5.5. Clinopyroxenite intrusions

Clinopyroxenite intrusions are composed dominantly of sub-idiomorphic to idiomorphic coarse-grained cpx (ranging in size from ~5mm to ~2 cm, Fig. 7) with interstitial assemblages of olivine, cpx, opx and hornblende (1-5 mm). Grain boundaries appear linear and 120° triple junctions are common among cpx grains. Rare macrocrysts of euhedral to subhedral cpx locally exhibit twinning with straight to stepped twin boundaries (Fig. 7B, 7C and Ceuleneer and Le Sueur, 2008). In contrast, olivine and opx are xenomorphic, with mostly lobate or curvilinear grain boundaries; triple junctions and deformation markers are absent. Locally, these intrusions grade into finer-grained layers (e.g. sample 02T117-A) where their texture becomes similar to that of the massive cumulates (Figs. 7A and 8A).

6. EBSD results

6.1. Lherzolitic country rock

415

416   The mineral fabrics of olivine, cpx and opx from the two analysed lherzolites are presented in

417   Fig. 9. The olivine fabric shows a strong concentration of [010] axes perpendicular to the

418   foliation plane, while either [100] or [001] axes are sub-parallel with the lineation direction

419   depending on the sample. The cpx fabric has [001] axes parallel to the lineation and [010]

420   axes perpendicular to the foliation plane. Opx exhibits a cluster of [001] axes around the

421   lineation direction and a weak point maximum for [100] and [010] axes, both fitting the

422   foliation pole (Fig. 9). Amphibole, when present, has a fabric mimicking that of cpx

423   (Supplementary Material D). The fabric symmetry for olivine and opx in the lherzolite plots

424   within the orthorhombic domain while cpx is spread between the orthorhombic and planar

425   domains (Fig. 10).

426

427   **6.2. Finely layered cumulates**

428

429   The finely layered cumulates were investigated using nine samples; results are shown in

430   figure 11 (see also Figs. 4, 5A and Table 1). The grain sizes calculated using EBSD data do

431   not show any variation across this part of the section (Supplementary Material C). The

432   olivine aspect ratio is always slightly higher than that of the pyroxenes, for a smaller grain

433   size. The olivine fabric is dominated by a strong concentration of [010] axes perpendicular to

434   the layering plane. Olivine [100] and [001] axes are distributed as girdles in the layering

435   plane (Fig. 10). This observation is consistent with the BA-indices (Fig. 10), suggesting

436   [010]-axial symmetries for the olivine fabrics. Other samples exhibit clustering of the [001]

437   axes (e.g. samples T4BC7B, 01T12A) or the [100] axes (e.g. samples T3BC3A and 01T13A,

438   Fig. 11) around the N115° direction. Cpx displays a clustering of [010] axes, and to a lesser

439   extent of [100], normal to the layering plane. The cpx [001] axes mostly form girdles in the

440 layering plane, as indicated by their low LS-indices (Fig. 10, Sup. Material A), but may also

441 define a sub-maximum within the girdle, close to the N115° direction. Cpx fabrics are

442 dominated by a planar symmetry, as again indicated by their low LS-indices (Fig. 10). The

443 opx fabric usually has a weak concentration of both [100] and [010] axes next to the layering

444 pole, while the [001] axis forms a girdle with a sub-maximum aligned in the N115° direction

445 (Fig. 11). Orientation maps reveal that two evenly-spread populations of opx may coexist and

446 are responsible for the [100] and [010] clusters near the pole of the layering plane

447 (Supplementary Material B). Opx appears to have higher BA-indices than LS-indices,

448 suggesting a greater contribution of [010] and [001] of opx to the planar structure (Figs. 10

449 and 11). When amphibole is present, its fabric is similar to that of cpx (Fig. 10). As described

450 earlier, a thin amphibole vein crosscutting the magmatic layering is observed in one sample

451 (01T13A, Fig. 4C). The fabric of the amphibole making up the vein is similar to that of the

452 amphibole in the finely layered cumulates that appears as interstitial grains between olivine,

453 opx and cpx grains. Twinning of the cpx in the finely layered cumulates is characterised by

454 EBSD and represents a rotation of about 180° around [001].

455

456 **6.3. Massive cumulates**

457

458 Two samples (T3BC4A, T4BC4A) were used to investigate the fabric of massive cumulates

459 (Fig. 12). In both samples, the cpx [010] axes form a cluster normal to the layering plane

460 while [001] axes are scattered and form a girdle in the layering plane (Fig. 12). In sample

461 T3BC4A, olivine yields a fabric with the [001] axes clustering weakly along the N115°

462 direction and the [010] axis normal to the layering. The number of olivine grains in sample

463 T4BC4A does not allow for a clear interpretation of the fabric. In terms of fabric symmetry,

464 cpx in our two samples has a low LS-index, suggesting a strong planar component for the

465   fabric, while the only occurrence of olivine plots in the orthorhombic domain (Fig. 10). In the

466   massive cumulates, the occurrence of opx and amphibole is so low that no reliable

467   observation of the mineral fabric can be made.

468

469   **6.4. Clinopyroxenitic intrusions**

470

471   The pegmatitic microstructure in the clinopyroxenitic intrusions was investigated using

472   sample 03T16A (Figs. 7B and 7C, Supplementary mat. E). The determination of fabrics is

473   hampered by the lack of structural framework inherent to this lithology; internal deformation

474   is lacking, and the large grain-size limits the number that can be analysed. In sample

475   03T16A, cpx and amphibole are the only phases with sufficient grains to calculate an ODF.

476   The cpx [001] axes exhibit point maxima for sample 03T16A (Supplementary mat. E). In

477   sample 03T16A, amphibole displays a fabric similar to that for cpx.

478

479   The microstructure of the finer-grained section of the clinopyroxenite intrusion was studied

480   in sample 02T117-A, which represents the contact between a clinopyroxenitic intrusion and

481   the finely layered cumulates (Fig. 8A, B). In the clinopyroxenite part, only cpx was observed

482   while cpx, opx and olivine are all visible in the finely layered cumulates. The cpx in the

483   clinopyroxenite intrusion shows [010] normal to the intrusion plane with [001] axes forming

484   a girdle on that plane. The fabrics and textures observed in the finely layered cumulate are

485   comparable to that previously described. Similarly, the amphibole fabric matches that of cpx

486   (Supplementary mat. E).

487

488

489   **7. Discussion**

490   Field observations, textural and microstructural data are integrated from the bottom to the top

491   of the Bear Creek intrusion and its associated mantle domain in order to unravel the solid

492   state and magmatic processes that took place in the Bear Creek. Implications of the

493   deformation episodes and the related conditions are then considered in the context of the

494   evolution of the Trinity ophiolite.

495

496   **7.1. Conditions of deformation in the host Trinity mantle**

497

498   In this section, we first discuss the EBSD data in the light of the field and microstructural

499   observations and then use them to estimate the conditions under which plastic deformation

500   occurred prior to the intrusion.

501

502   Lherzolitic cpx displays a SL-type fabric ([010] and [001] axes of clustering along the

503   foliation pole and the lineation direction respectively), suggesting deformation at high

504   temperature, and a combination of pure and simple shear (Bascou et al., 2002; Helmstaedt et

505   al., 1972). Opx displays a ABC fabric (i.e. [001] axis along the lineation direction and weak

506   cluster of [100] and [010] normal to the layering, Jung et al., 2010) while olivine shows an A-

507   type fabric (T4BC2A), consistent with deformation of the lherzolite by dislocation creep at

508   high temperature and low stress under "asthenospheric" conditions in the upper mantle (Jung

509   and Karato, 2001; Michibayashi and Oohara, 2013). Sample T4BC3A (Figs. 3E and 9) was

510   collected less than 2 m below the finely layered cumulates of the Bear Creek intrusion in the

511   cross section and is the closest lherzolite sample to the contact between the intrusion and the

512   host rock. The olivine fabric can be interpreted as a [001](010) (B-type, Jung and Karato,

513   2001). We suggest that the possible change in the olivine fabrics between A- and B-type

514   could be related to ongoing plastic deformation conditioned by infiltration of melt from the

21

515   intrusion, which could trigger a fabric switch either by the integration of structural water *via*

516   diffusion and /or the presence of interstitial melt (Jung and Karato, 2001; Kohlstedt and

517   Holtzman, 2009; Qi et al., 2018).

518

519   **7.2 Magmatic and plastic processes recorded in the Bear Creek intrusion**

520

521   Ceuleneer and Le Sueur (2008), based on the igneous stratigraphy, proposed that the Bear

522   Creek intrusion evolved in a closed system following the injection of a single batch of

523   magma in the shallow lithosphere. Studying the microstructure of cumulates could thus

524   potentially provide valuable information on the dynamics of crystallisation and direction of

525   transport in this specific magmatic environment. Below, we decipher the progression of the

526   deformation in the section.

527

528   **7.2.1. Constraints from the finely layered cumulates**

529

530   At the base of the cumulate section, our field observations revealed that the finely layered

531   cumulates display a well-defined planar structure corresponding to the compositional

532   layering (Fig. 2). On the layering plane, no lineation or foliation (i.e. crystal shape-preferred

533   orientation) were visible at the outcrop or hand specimen scale. In the finely layered cumulate

534   microstructure, we observed the occurrence of growth twins in cpx. The quantification of

535   mineral SPOs shows that the long axie of the larger population of grains of olivine and cpx

536   are subparallel to N°115, while opx grains are preferentially aligned at about 90 degrees to

537   cpx and olivine (i.e. subparallel to the layering pole, Fig. 6A). The orthogonal shape-

538   preferred orientation of opx relative to the orientation of cpx and olivine (note the slight

539   elongation of opx in Fig. 4B), and very scarce internal deformation (Fig. 5A) are not

540   consistent with extensive plastic deformation. All the above suggest that the macro- and

541   microstructures of the finely layered cumulates were not overprinted by extensive plastic

542   deformation and probably preserve magmatic features. Regarding the mineral fabrics, olivine,

543   opx and cpx CPOs have [001] axes forming girdles on the compositional layering. This has

544   been observed in a magmatic context (e.g. Holness et al., 2017) and is consistent with the

545   preservation of magmatic features in the finely layered cumulates that was hinted at by field

546   and microstructural observations.

547

548   Although no macroscopic lineation was detected, opx, cpx [001] and ol [001] or [100]

549   crystallographic axes define sub-maxima aligned along the regional N°115 orientation, which

550   is also the stretching lineation in the mantle peridotites (Fig 11). Le Sueur et al. (1984) and

551   Boudier et al. (1989) interpreted this lineation in the peridotite as reflecting asthenospheric

552   flow channelised parallel to the azimuth of the former spreading centre. The olivine and

553   pyroxene clusters can have two origins. (1) Formed through mechanical alignment of crystals

554   by rotation around the vorticity axis of minerals during magmatic flow, when the proportion

555   of melt is still high or (2) formed by limited plastic deformation at low melt volume probably

556   driven by the regional stress field in the case of the Bear Creek intrusion (Boudier et al.,

557   1989; Le Sueur et al., 1984). The limited evidence of plastic deformation reported here, such

558   as the few subgrains and undulose extinction in olivine, cpx and opx (Fig. 5A) could be

559   linked to plastic deformation under low stress (i.e. regional stress) or alternatively could be

560   the result of the compaction process.

561   We suggest that the texture and CPO observed in the finely layered cumulates may have been

562   formed by the following processes. (1) Rock-forming minerals crystallising with a shape

563   anisotropy would be deposited and a preferred orientation was induced by a moderate

564   magmatic flow; (2) compaction of the cumulates leading to limited plastic deformation and to

565   the accentuation of the planar feature together with a reduction of the linear feature (Figs. 2A,

566   4, 6 and 11). Alternatively, an additional and limited episode of deformation could have

567   occurred and could have been responsible for the weak clusters of [001] in pyroxenes and

568   olivine contained in the girdles.

569

570   If an initial shape anisotropy was present during crystal settling, our results suggest that cpx

571   crystals are likely to settle mainly on the (010) plane, and to a minor extent on the (100)

572   plane. In EBSD-derived orientation maps, we observe two populations of opx SPO. The first

573   population displays a long axis contained in the compositional plane and has the [010] axis

574   parallel to the layering plane pole. The second population has a long axis perpendicular to the

575   layering plane and has the [001] axis parallel to the layering pole (Supplementary Material

576   B). Based on these observations, we suggest that (010) is more likely to be the settling plane

577   for opx than (100) as (010) and the crystal long axie are contained within the compositional-

578   layering plane. In Bear Creek, olivine crystals tended to preferentially settle on the (010)

579   plane. For all three minerals, the [001] axis commonly is within the layering plane, consistent

580   with the fact that this axis coincides with the direction of crystal elongation.

581

582   These results are consistent with previously-reported fabrics in layered intrusions from the

583   crustal section of the Oman ophiolite. (Benn and Allard, 1989) showed that the (010) planes

584   of olivine and cpx coincide with the compositional layering. Similar observations were made

585   in recent studies of layered intrusions from different geological settings (e.g. Holness et al.,

586   2012; Vukmanovic et al., 2018). Our observations are also consistent with recent modelling

587   and the study of adcumulates from a mafic-ultramafic intrusion in the Tarim Craton, which

588   showed that compaction can lead to weak B-type fabrics and formation of subgrains in

589   olivine (Yao et al., 2019).

590

7.2.2. Constraints from the massive cumulates

592

In a previous study of the Bear Creek intrusion, Ceuleneer and Le Sueur (2008) used the

continuous gradational lithological change and the continuous and correlated variations in

major elements (i.e. Mg#) to interpret the finely layered cumulates and the massive cumulates

as the crystallisation product of a single crystallisation sequence. The massive cumulates are

essentially mono- to bi-mineralic in the olivine-poor level, and exhibit a coarser grain size

and straighter grain boundaries than the finely layered cumulates (Figs. 3D and 5B). Our

results show that the fabrics observed in the massive cumulates with development of girdles

on pyroxene [001] axes are comparable to those of their finely layered counterparts (Fig. 12).

The fabric of olivine in the massive cumulates plots into the orthorhombic symmetry (Fig.

10) and contrasts with that of cpx. Based on the change in fabric symmetry and the likely

preservation of magmatic microstructures in the cpx-rich layers, we suggest that the olivine-

rich layers may have preferentially recorded and accommodated any deformation subsequent

to the magmatic stage (i.e. compaction or solid state deformation). The quantification of cpx

SPO shows a range of orientations from sub-parallel to olivine and cpx in the finely layered

cumulates (parallel to the lineation) to a maximum at small angle to the lineation (Fig. 6).

This fan of orientations could be due to the nearly monomineralic composition of the layer

triggering further crystal growth.

The lack of deformation markers observed during the field investigation, the relatively scarce

internal deformation of large cpx crystals (Fig. 5B), and the CPO of cpx showing a girdle on

[001] with weak clusters (Fig. 12) mimic what was observed in the finely layered cumulates

(Fig. 5A and 11). Considering the above and the concentric subgrain boundaries observed in

the cpx in the massive cumulates, our data suggest that the microstructure of the massive

615   cumulates is mostly magmatic. A similar combination of processes was at play in the finely

616   layered cumulates. This includes a weak magmatic flow and a limited episode of plastic

617   deformation (probably due to compaction or to regional stress). In this case, the coarser cpx

618   grain size and straighter grain boundaries in the massive pyroxenites can be interpreted as a

619   combined consequence of the rock composition and the crystallisation processes. The

620   crystallisation of a single liquidus phase would favour growth of early grains (rather than

621   high nucleation rate), possibly by grain-boundary migration at high temperature, as suggested

622   by the complex grain-boundary shapes of some grains in the massive pyroxenites (e.g. Fig.

623   5B). Conversely, in the finely layered cumulates, the homogeneous distribution of the phases

624   will result in a smaller grain-size and can satisfactorily account for the curvilinear grain

625   boundaries at the contact between different phases (Figs. 3 and 5).

626

627   **7.2.3. Constraints from the clinopyroxenitic intrusions**

628

629   Clinopyroxenite intrusions cross-cut the layered pyroxenitic cumulates (Fig. 3A and 3C), and

630   have a pegmatitic texture (i.e. grain size from 1 mm to multiple centimetres). The grain size

631   of the clinopyroxenitic intrusions can decrease towards the rims of the dykes and may

632   approach the grain size of the host cumulate layers. This can make such intrusions difficult to

633   distinguish from the massive pyroxenites. The presence of numerous cpx twins (Fig. 7B and

634   7C) and the low internal deformation suggest a purely magmatic origin for their

635   microstructure. However, and contrary to the finely layered cumulates and massive

636   cumulates, the relatively small size (~ 5-10 cm) of the clinopyroxenitic intrusions does not

637   allow compaction and crystal settling to operate. The textures and fabrics observed in the

638   sample of a clinopyroxenitic intrusion that preserves the contact with finely layered

639   cumulates (02T117-1; Fig. 8A and 8B) are indistinguishable from those observed in the

640   massive pyroxenites (Fig. 5B and 12). In the finer-grained part of the clinopyroxenitic

641   intrusions (i.e. part shown in Fig 8), epitaxial crystallisation of new cpx grains on the cpx

642   already present in the finely layered cumulates or massive pyroxenites can explain the

643   development of the observed fabric. In the pegmatitic parts of the clinopyroxenitic intrusions

644   (sample 03T16B, Fig. 7B, 7C and Supplementary Material E), the orthorhombic symmetry of

645   the cpx fabric could be interpreted as a clue to a magmatic flow geometry within the dyke,

646   which would not be purely linear. However, the presence of twined cpx leads to an over-

647   sampling of the large grains during data collection, thus creating an impression of

648   orthorhombic symmetry while the actual symmetry could be linear (which would be expected

649   within a dyke). Although further investigation is required (in particular, analysis of their

650   respective chemical compositions), it cannot be excluded that some intrusions and massive

651   pyroxenites could be cogenetic, especially in sections where no olivine-rich levels are visible.

652

653

654   **7.2.4. Amphibole in the Bear Creek**

655

656   Small amounts of igneous amphiboles (brown hornblende and pargasite) occur in every

657   lithology from the Bear Creek area (Tab. 1). Amphibole is usually found as interstitial grains

658   (Fig. 4), is commonly associated with cpx and can also appear as veins discordant to the

659   layering (01T13A, Fig. 4C). In our samples, the amphibole CPO systematically mimics that

660   of the cpx (i.e. [001](010)]). This fabric contrasts with the previously-reported dislocation-

661   creep-induced fabrics in amphibole (e.g. Ko and Jung, 2015). The observation that an

662   amphibole-rich vein cross-cutting the pre-existing layering (Fig. 4C) shows a CPO similar to

663   that of disseminated amphiboles within the sample (Fig. 11 and Supplementary Material E) is

664   an indication of a possible melt/rock or fluid/rock-interaction origin for the amphiboles in

665 Trinity. The textural relationship between cpx and amphibole, together with the mimicking of

666 the crystallographic preferred orientations of cpx by the amphibole, also point to a possible

667 epitaxial growth of the amphibole as already observed in other localities (e.g. Henry et al.,

668 2017; Kaczmarek et al., 2016; Moghadam et al., 2020).

669

670 **7.3. A scenario for the emplacement of the Bear Creek intrusion**

671

672 The magmatic processes and the plastic deformation recorded in the Bear Creek intrusion

673 share similarities with other intrusions emplaced in extensional settings (e.g. Holness et al.,

674 2012; Holness et al., 2017; Vukmanovic et al., 2018 and reference therein). However, the

675 Bear Creek intrusion presents some features that could be typical for this type of

676 environment, and lead us to draw a scenario for its emplacement into the mantle domain

677 exposed in the ophiolites.

678

679 (1) An episode of solid-state flow occurred in ambient Trinity mantle wall-rock, mostly

680 under high-temperature and low-stress conditions in the dislocation creep regime. This

681 episode probably occurred as the "asthenospheric" mantle rose in an oceanic spreading

682 environment and was channelised at shallow depth along the spreading axis (e.g.

683 Boudier et al., 1989; Dygert et al., 2016; Le Sueur et al., 1984). Pressure probably was

684 below 1 GPa (Quick, 1981) and the temperature above 1000°C so that an A-type olivine

685 fabric could form as observed in the lherzolite.

686 (2) In the Bear Creek area, intrusion of a boninitic magma at shallow depth (<1 GPa) led to

687 the crystallisation of opx, cpx and olivine, producing finely layered cumulates, then with

688 further differentiation, massive pyroxenites dominated by cpx (Ceuleneer and Le Sueur,

689 2008). During that accumulation of ultramafic minerals, a dominantly planar mineral

690   fabric developed in the cumulates due to the combination of a crystal settling with weak

691   magmatic flux. With decreasing melt proportions, compaction probably operated; it

692   could have accentuated the planar component of the lithologies and formed the

693   subgrains observed in the minerals of the Bear Creek samples. In the fabrics, we

694   observe the presence of sub-maxima of preferred orientation for cpx, opx or olivine

695   crystallographic axes within girdles coinciding with the compositional layering (Fig. 9).

696   This may provide constraints on the geometry of the magmatic flow in a magma-

697   chamber setting, and/or it may reflect the persistence of the regional stress field. Our

698   observations suggest that the magmatic processes recorded are homogeneous over the

699   entire section. Later intrusions occurred, crystallising into coarse-grained to pegmatitic

700   clinopyroxenite-rich veins and dykes crosscutting the weakly-consolidated cumulate

701   layers. Finally, the crystallisation of amphibole was followed by epitaxial growth on

702   cpx.

703

704   **8. Conclusions**

705

706   New field and EBSD data have allowed us to characterise the deformation mechanisms and

707   the formation/deformation history of the layered pyroxenitic cumulates, and their

708   relationships with the host mantle peridotites (lherzolite) in the Trinity ophiolite. This

709   detailed microstructural study of the Bear creek intrusion suggests a scenario for the

710   formation of layered intrusions that first includes crystal settling, then magma flow and later

711   compaction. CPO, SPO and internal deformation of minerals suggest that crystal settling is

712   probably the major process creating the anisotropy of the cumulates, rather than magma flow,

713   crystal growth or compaction. Our data show that plastic deformation is present but not

714   widely recorded in any minerals, even at the base of the section. We interpret this as

715   indicating a limited contribution of compaction or alternatively for an ongoing regional stress

716   to the microstructure. From a purely microstructural aspect, this study emphasises the

717   importance of crystal-shape control during the development of microstructures in cumulate

718   rocks.

719

720   In the Trinity complex, magmatic processes gave rise to strong crystallographic preferred

721   orientations for olivine, cpx and opx, similar to those classically attributed to plastic flow in

722   the dislocation creep regime. Since these are likely to constrain the activation of favourably

723   oriented slip systems during subsequent deformation, the reliability of their use as proxies for

724   the conditions of deformation has to be regarded with caution.

725   Our results, along with those of previous structural and geochemical work, are consistent

726   with (1) the mantle section of Trinity ophiolite representing a fossil marginal slow-spreading

727   centre obducted onto the North American plate and (2) Bear Creek and similar intrusions

728   corresponding to the late- to post-tectonic emplacement and differentiation of boninitic

729   magmas in a subduction-related environment.

730   The microstructural data that we have presented may also have important implications for our

731   understanding of the development of the fabrics in pyroxene-rich lithologies, which could be

732   pre-conditioned by the initial magmatic fabric. The strong fabrics in the lithologies in the

733   Bear Creek cumulates would also have important implications in terms of seismic anisotropy

734   as they differ significantly from those of more classic mantle pyroxenites.

735

References

Achenbach, K.L., Cheadle, M.J., Faul, U., Kelemen, P., Swapp, S., 2011. Lattice-preferred orientation and microstructure of peridotites from ODP Hole 1274A (15°39'N), Mid-Atlantic Ridge: Testing models of mantle upwelling and tectonic exhumation. Earth and Planetary Science Letters 301, 199-212.

Avé Lallemant, H.G., Carter, N.L., 1970. Syntectonic Recrystallization of Olivine and Modes of Flow in the Upper Mantle. Geological Society of America Bulletin 81, 2203-2220.

Bachmann, F., Hielscher, R., Jupp, P.E., Pantleon, W., Schaeben, H., Wegert, E., 2010a. Inferential statistics of electron backscatter diffraction data from within individual crystalline grains. Journal of Applied Crystallography 43, 1338-1355.

Bachmann, F., Hielscher, R., Schaeben, H., 2010b. Texture Analysis with MTEX – Free and Open Source Software Toolbox. Solid State Phenomena 160, 63-68.

Bachmann, F., Hielscher, R., Schaeben, H., 2011. Grain detection from 2d and 3d EBSD data—Specification of the MTEX algorithm. Ultramicroscopy 111, 1720-1733.

Bascou, J., Tommasi, A., Mainprice, D., 2002. Plastic deformation and development of clinopyroxene lattice preferred orientations in eclogites. Journal of Structural Geology 24, 1357-1368.

Benn, K., Allard, B., 1989. Preferred Mineral Orientations Related to Magmatic Flow in Ophiolite Layered Gabbros. Journal of Petrology 30, 925-946.

Biedermann, A.R., Heidelbach, F., Jackson, M., Bilardello, D., McEnroe, S.A., 2016. Magnetic fabrics in the Bjerkreim Sokndal Layered Intrusion, Rogaland, southern Norway: Mineral sources and geological significance. Tectonophysics 688, 101-118.

Bodinier, J.-L., Godard, M., 2003. Orogenic, ophiolitic, and abyssal peridotites. Treatise on geochemistry 2, 568.

Boorman, S., Boudreau, A., Kruger, F.J., 2004. The Lower Zone–Critical Zone Transition of the Bushveld Complex: a Quantitative Textural Study. Journal of Petrology 45, 1209-1235.

Boudier, F., 1991. Olivine xenocrysts in picritic magmas. Contrib Mineral Petrol 109, 114-123.

Boudier, F., Le Sueur, E., Nicolas, A., 1989. Structure of an atypical ophiolite: The Trinity complex, eastern Klamath Mountains, California. Geological Society of America Bulletin 101, 820-833.

Boudreau, A.E., McBirney, A.R., 1997. The Skaergaard Layered Series. Part III. Non-dynamic Layering. Journal of Petrology 38, 1003-1020.

Brothers, R., 1964. Petrofabric analyses of Rhum and Skaergaard layered rocks. Journal of Petrology 5, 255-274.

Brouxel, M., Lapierre, H., Michard, A., Albaréde, F., 1988. Geochemical study of an early Paleozoic island-arc-back-arc basin system. Part 2: Eastern Klamath, early to middle Paleozoic island-arc volcanic rocks (northern California). Geological Society of America Bulletin 100, 1120-1130.

Bunge, H.J., 1982. Texture analysis in materials science: mathematical methods. Elsevier.

Cannat, M., Lécuyer, C., 1991. Ephemeral magma chambers in the Trinity peridotite, northern California. Tectonophysics 186, 313-328.

Cawthorn, R.G., Walraven, F., 1998. Emplacement and Crystallization Time for the Bushveld Complex. Journal of Petrology 39, 1669-1687.

797   Ceuleneer, G., Le Sueur, E., 2008. The Trinity ophiolite (California): the strange association of
798   fertile mantle peridotite with ultra-depleted crustal cumulates. Bulletin de la Societe
799   Geologique de France 179, 503-518.
800   Charlier, B., Namur, O., Latypov, R., Tegner, C., 2015. Layered intrusions, 1 ed.
801   Cheadle, M.J., Gee, J.S., 2017. Quantitative Textural Insights into the Formation of Gabbro in
802   Mafic Intrusions. Elements 13, 409-414.
803   Clénet, H., Ceuleneer, G., Pinet, P., Abily, B., Daydou, Y., Harris, E., Amri, I., Dantas, C., 2010.
804   Thick sections of layered ultramafic cumulates in the Oman ophiolite revealed by an
805   airborne hyperspectral survey: Petrogenesis and relationship to mantle diapirism. Lithos
806   114, 265-281.
807   Coleman, R.G., 1986. Ophiolites and accretion of the North American Cordillera. Bulletin de
808   la Societe Geologique de France II, 961-968.
809   Dygert, N., Liang, Y., Kelemen, P.B., 2016. Formation of Plagioclase Lherzolite and Associated
810   Dunite–Harzburgite–Lherzolite Sequences by Multiple Episodes of Melt Percolation and
811   Melt–Rock Reaction: an Example from the Trinity Ophiolite, California, USA. Journal of
812   Petrology 57, 815-838.
813   Feinberg, J.M., Wenk, H.-R., Scott, G.R., Renne, P.R., 2006. Preferred orientation and
814   anisotropy of seismic and magnetic properties in gabbronorites from the Bushveld layered
815   intrusion. Tectonophysics 420, 345-356.
816   Frets, E., Tommasi, A., Garrido, C.J., Padrón-Navarta, J.A., Amri, I., Targuisti, K., 2012.
817   Deformation processes and rheology of pyroxenites under lithospheric mantle conditions.
818   Journal of Structural Geology 39, 138-157.
819   Helmstaedt, H., Anderson, O.L., Gavasci, A.T., 1972. Petrofabric studies of eclogite, spinel-
820   Websterite, and spinel-lherzolite Xenoliths from kimberlite-bearing breccia pipes in
821   southeastern Utah and northeastern Arizona. Journal of Geophysical Research 77, 4350-
822   4365.
823   Henry, H., 2018. Mantle Pyroxenites: Deformation and Seismic Properties, Earth and
824   Planetary Department. Macquarie University and Université Toulouse III.
825   Henry, H., Tilhac, R., Griffin, W.L., O'Reilly, S.Y., Satsukawa, T., Kaczmarek, M.-A., Grégoire,
826   M., Ceuleneer, G., 2017. Deformation of mantle pyroxenites provides clues to geodynamic
827   processes in subduction zones: Case study of the Cabo Ortegal Complex, Spain. Earth and
828   Planetary Science Letters 472, 174-185.
829   Higgins, M.D., 1998. Origin of Anorthosite by Textural Coarsening: Quantitative
830   Measurements of a Natural Sequence of Textural Development. Journal of Petrology 39,
831   1307-1323.
832   Higgins, M.D., 2015. Quantitative Textural Analysis of Rocks in Layered Mafic Intrusions, in:
833   Charlier, B., Namur, O., Latypov, R., Tegner, C. (Eds.), Layered Intrusions. Springer
834   Netherlands, Dordrecht, pp. 153-181.
835   Holness, M.B., Sides, R., Prior, D.J., Cheadle, M.J., Upton, B.G., 2012. The peridotite plugs of
836   Rum: Crystal settling and fabric development in magma conduits. Lithos 134, 23-40.
837   Holness, M.B., Vukmanovic, Z., Mariani, E., 2017. Assessing the role of compaction in the
838   formation of adcumulates: a microstructural perspective. Journal of Petrology 58, 643-673.
839   Ismaïl, W.B., Mainprice, D., 1998. An olivine fabric database: an overview of upper mantle
840   fabrics and seismic anisotropy. Tectonophysics 296, 145-157.
841   Jung, H., Karato, S.I., 2001. Water-Induced Fabric Transitions in Olivine. Science 293, 1460-
842   1463.

843   Jung, H., Park, M., Jung, S., Lee, J., 2010. Lattice preferred orientation, water content, and
844   seismic anisotropy of orthopyroxene. J. Earth Sci. 21, 555-568.

845   Kaczmarek, M.A., Bodinier, J.L., Bosch, D., Tommasi, A., Dautria, J.M., Kechid, S.A., 2016.
846   Metasomatized Mantle Xenoliths as a Record of the Lithospheric Mantle Evolution of the
847   Northern Edge of the Ahaggar Swell, In Teria (Algeria). Journal of Petrology 57, 345-382.

848   Karato, S.I., 2008. Deformation of Earth Materials. Cambridge.

849   Kelemen, P.B., Dick, H.J.B., Quick, J.E., 1992. Formation of harzburgite by pervasive
850   melt/rock reaction in the upper mantle. Nature 358, 635-641.

851   Ko, B., Jung, H., 2015. Crystal preferred orientation of an amphibole experimentally
852   deformed by simple shear. Nature Communication 6.

853   Kohlstedt, D.L., Holtzman, B.K., 2009. Shearing Melt Out of the Earth: An Experimentalist's
854   Perspective on the Influence of Deformation on Melt Extraction. Annual Review of Earth and
855   Planetary Sciences 37, 561-593.

856   Le Sueur, E., Boudier, F., Cannat, M., Ceuleneer, G., Nicolas, A., 1984. The Trinity mafic-
857   ultramafic complex: first results of the structural study of an untypical ophiolite. Ofioliti 9,
858   487-498.

859   Lindsley-Griffin, N., 1977. The Trinity ophiolite, Klamath Mountains, California. Oregon
860   Department of Geology and Mineral Industries Bulletin 95, 107-120.

861   Mainprice, D., Bachmann, F., Hielscher, R., Schaeben, H., 2015. Descriptive tools for the
862   analysis of texture projects with large datasets usingMTEX: strength, symmetry and
863   components. Geological Society, London, Special Publications 409, 251-271.

864   Mankinen, E.A., Lindsley-Griffin, N., Griffin, J.R., 2002. Concordant paleolatitudes for
865   Neoproterozoic ophiolitic rocks of the Trinity Complex, Klamath Mountains, California.
866   Journal of Geophysical Research: Solid Earth 107, EPM 11-11-EPM 11-18.

867   McBirney, A.R., Nicolas, A., 1997. The Skaergaard Layered Series. Part II. Magmatic flow and
868   Dynamic Layering. Journal of Petrology 38, 569-580.

869   Michibayashi, K., Oohara, T., 2013. Olivine fabric evolution in a hydrated ductile shear zone
870   at the Moho Transition Zone, Oman Ophiolite. Earth and Planetary Science Letters 377-378,
871   299-310.

872   Moghadam, H.S., Li, Q., Griffin, W., Stern, R., Ishizuka, O., Henry, H., Lucci, F., O'Reilly, S.,
873   Ghorbani, G., 2020. Repeated magmatic buildup and deep "hot zones" in continental
874   evolution: The Cadomian crust of Iran. Earth and Planetary Science Letters 531, 115989.

875   Namur, O., Abily, B., Boudreau, A.E., Blanchette, F., Bush, J.W.M., Ceuleneer, G., Charlier, B.,
876   Donaldson, C.H., Duchesne, J.-C., Higgins, M.D., Morata, D., Nielsen, T.F.D., O'Driscoll, B.,
877   Pang, K.N., Peacock, T., Spandler, C.J., Toramaru, A., Veksler, I.V., 2015. Igneous Layering in
878   Basaltic Magma Chambers, in: Charlier, B., Namur, O., Latypov, R., Tegner, C. (Eds.), Layered
879   Intrusions. Springer Netherlands, Dordrecht, pp. 75-152.

880   Nicolas, A., 1992. Kinematics in magmatic rocks with special reference to gabbros. Journal of
881   Petrology 33, 891-915.

882   Nicolas, A., Christensen, N.I., 1987. Formation of Anisotropy in Upper Mantle Peridotites - A
883   Review. Composition, structure and dynamics of the lithosphere-asthenosphere system,
884   111-123.

885   O'Reilly, S.Y., Griffin, W.L., 2013. Moho vs crust–mantle boundary: Evolution of an idea.
886   Tectonophysics 609, 535-546.

887   Paterson, M.S., 2001. A granular flow theory for the deformation of partially molten rock.
888   Tectonophysics 335, 51-61.

889   Prior, D.J., Boyle, A.P., Brenker, F., Cheadle, M.C., Day, A., Lopez, G., Peruzzi, L., Potts, G.,
890   Reddy, S., Spiess, R., Timms, N.E., Trimby, P., Wheeler, J., Zetterstrom, L., 1999. The
891   application of electron backscatter diffraction and orientation contrast imaging in the SEM
892   to textural problems in rocks. American Mineralogist 84, 1741-1759.
893   Qi, C., Hansen, L.N., Wallis, D., Holtzman, B.K., Kohlstedt, D.L., 2018. Crystallographic
894   Preferred Orientation of Olivine in Sheared Partially Molten Rocks: The Source of the "a-c
895   Switch". Geochemistry, Geophysics, Geosystems.
896   Quick, J.E., 1981. Petrology and petrogenesis of the Trinity peridotite, An upper mantle
897   diapir in the eastern Klamath Mountains, northern California. Journal of Geophysical
898   Research: Solid Earth 86, 11837-11863.
899   Quick, J.E., 1982. The origin and significance of large, tabular dunite bodies in the Trinity
900   peridotite, northern California. Contrib Mineral Petrol 78, 413-422.
901   Satsukawa, T., Ildefonse, B., Mainprice, D., Morales, L.F.G., Michibayashi, K., Barou, F., 2013.
902   A database of plagioclase crystal preferred orientations (CPO) and microstructures –
903   implications for CPO origin, strength, symmetry and seismic anisotropy in gabbroic rocks.
904   Solid Earth 4, 511-542.
905   Skemer, P., Katayama, I., Jiang, Z., Karato, S.-i., 2005. The misorientation index:
906   Development of a new method for calculating the strength of lattice-preferred orientation.
907   Tectonophysics 411, 157-167.
908   Ulrich, S., Mainprice, D., 2005. Does cation ordering in omphacite influence development of
909   lattice-preferred orientation? Journal of Structural Geology 27, 419-431.
910   Vukmanovic, Z., Holness, M., Monks, K., Andersen, J.Ø., 2018. The Skaergaard trough
911   layering: sedimentation in a convecting magma chamber. Contrib Mineral Petrol 173, 43.
912   Wallin, E.T., Coleman, D.S., Lindsley-Griffin, N., Potter, A.W., 1995. Silurian plutonism in the
913   Trinity terrane (Neoproterozoic and Ordovician), Klamath Mountains, California, United
914   States. Tectonics 14, 1007-1013.
915   Wallin, E.T., Mattinson, J.M., Potter, A.W., 1988. Early Paleozoic magmatic events in the
916   eastern Klamath Mountains, northern California. Geology 16, 144-148.
917   Wallin, E.T., Metcalf, Rodney V., 1998. Supra-Subduction Zone Ophiolite Formed in an
918   Extensional Forearc: Trinity Terrane, Klamath Mountains, California. The Journal of Geology
919   106, 591-608.
920   Williams, E., Boudreau, A.E., Boorman, S., Kruger, F.J., 2006. Textures of orthopyroxenites
921   from the Burgersfort bulge of the eastern Bushveld Complex, Republic of South Africa.
922   Contrib Mineral Petrol 151, 480-492.
923   Wright, S.I., Nowell, M.M., Field, D.P., 2011. A Review of Strain Analysis Using Electron
924   Backscatter Diffraction. Microscopy and Microanalysis 17, 316-329.
925   Yao, Z., Qin, K., Wang, Q., Xue, S., 2019. Weak B-type olivine fabric induced by fast
926   compaction of crystal mush in a crustal magma reservoir. Journal of Geophysical Research:
927   Solid Earth 0.
928   Ying, J., Zhang, H., Tang, Y., Su, B., Zhou, X., 2013. Diverse crustal components in pyroxenite
929   xenoliths from Junan, Sulu orogenic belt: Implications for lithospheric modification invoked
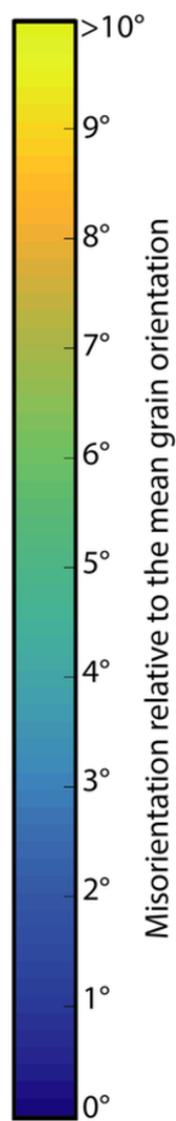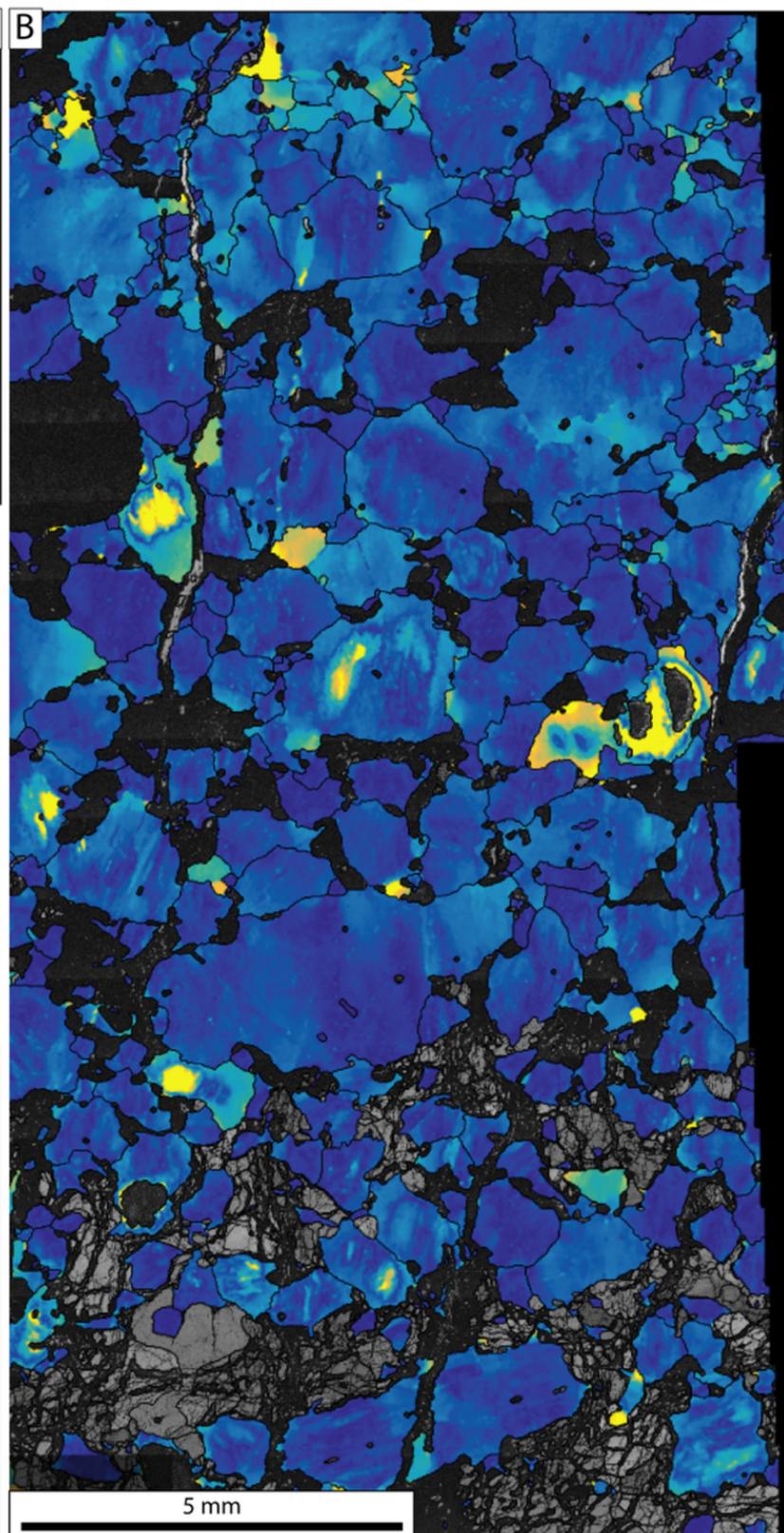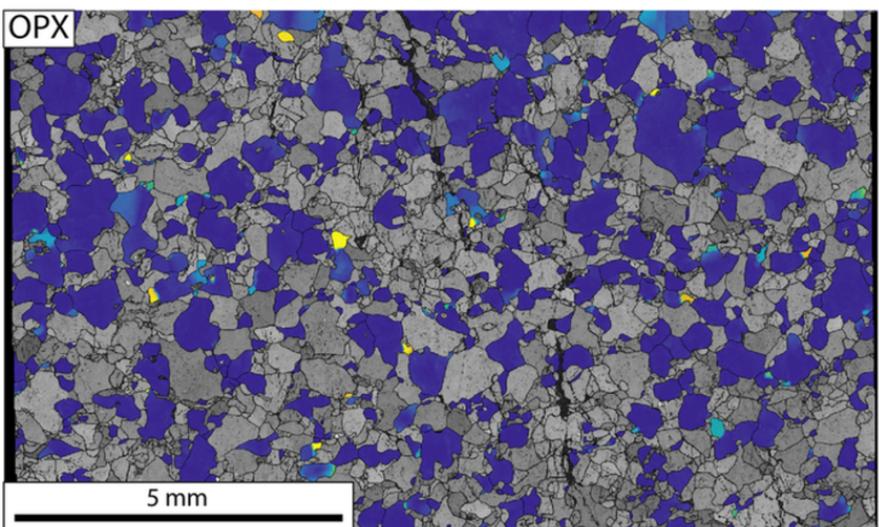930   by continental subduction. Chemical Geology 356, 181-192.
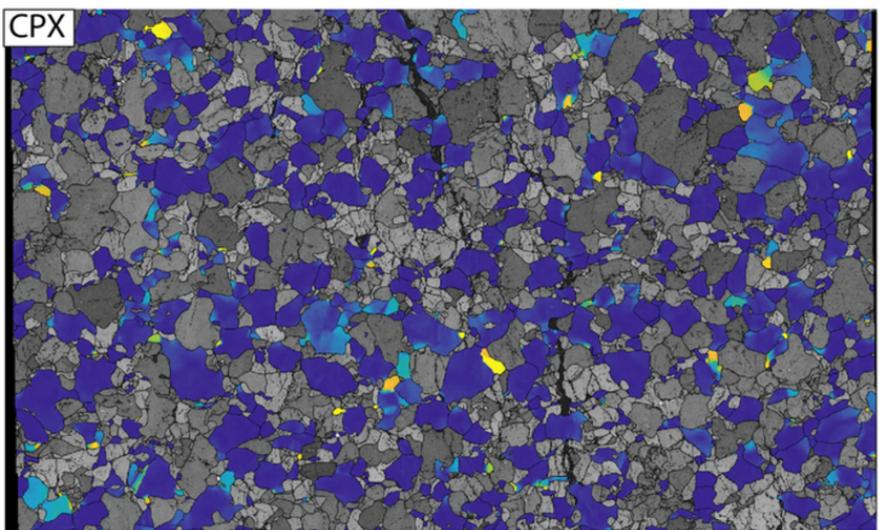931

A

Finely layered cumulates

Sample T1BC1A

Lherzolite

Pyroxenite veins

B

Finely layered cumulates

Serpentinized olivine-rich levels

A Finely layered cumulates

Clinopyroxenitic intrusion

Finely layered cumulates

20 cm

B Finely layered cumulates

CPX

OPX

Ol

L

N115°

5 mm

C

Olivine-rich horizon

Massive pyroxenite

Massive cumulates

D

CPX

L

N115°

5 mm

E

Lherzolite

Pyroxenite

10 cm

Lherzolites

F

Ol

OPX

AMPH

CPX

F

L

5 mm

A

B

C

5 mm

5 mm

5 mm

Olivine

CPX

OPX

Subgrain

A

B

5 mm

5 mm

Misorientation relative to the mean grain orientation

>10°

9°

8°

7°

6°

5°

4°

3°

2°

1°

0°

Azimuth of olivine, cpx and opx grains' long axis in the Bear creek cumulates
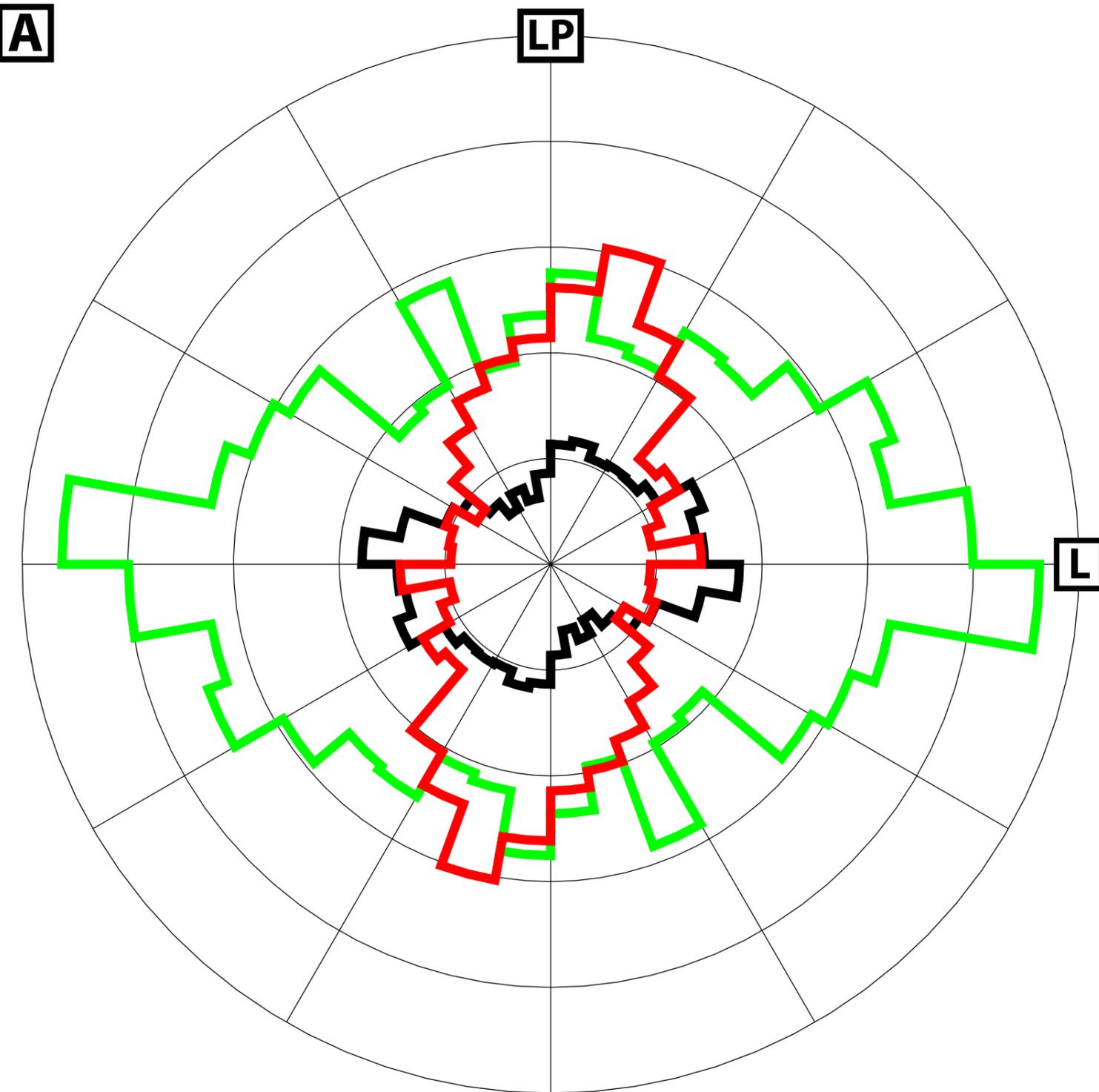
A — Finely layered cumulates

B — Massive cumulates

Olivine
Cpx
Opx

A — Coarse-grained clinopyroxenite intrusion; Massive pyroxenite

B — CPX; OPX; Ol; 5 mm

C — 10 mm; Twin boundary

M- vs BA-index for Olivine in Bear Creek (A)

M- vs LS-index for Cpx in Bear Creek (B)

BA- vs LS-index for Opx in Bear Creek (C)

Legend:
- Finely-Layered cumulates (green)
- Massive Pyroxenites (red)
- Host Lherzolite (yellow)
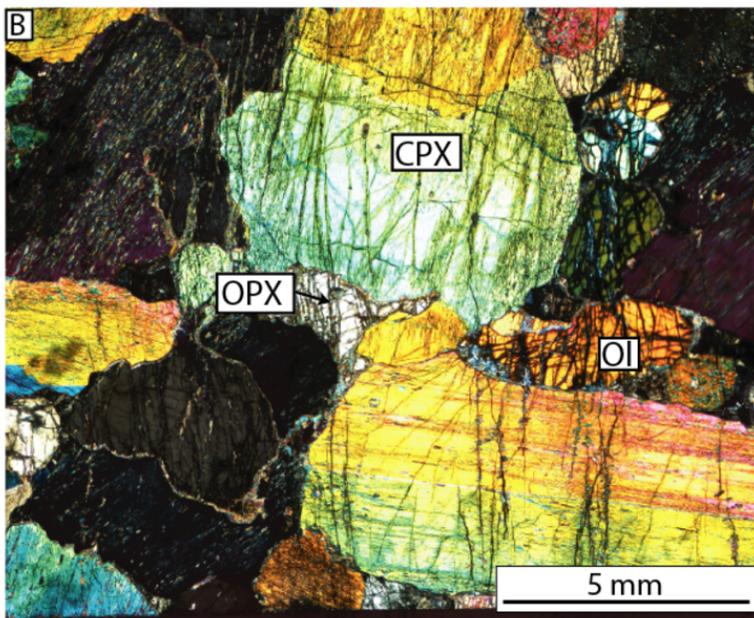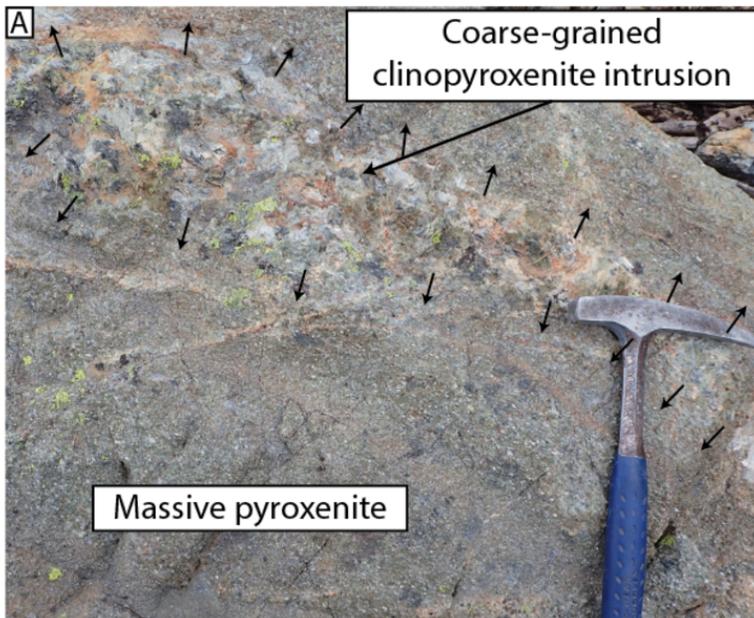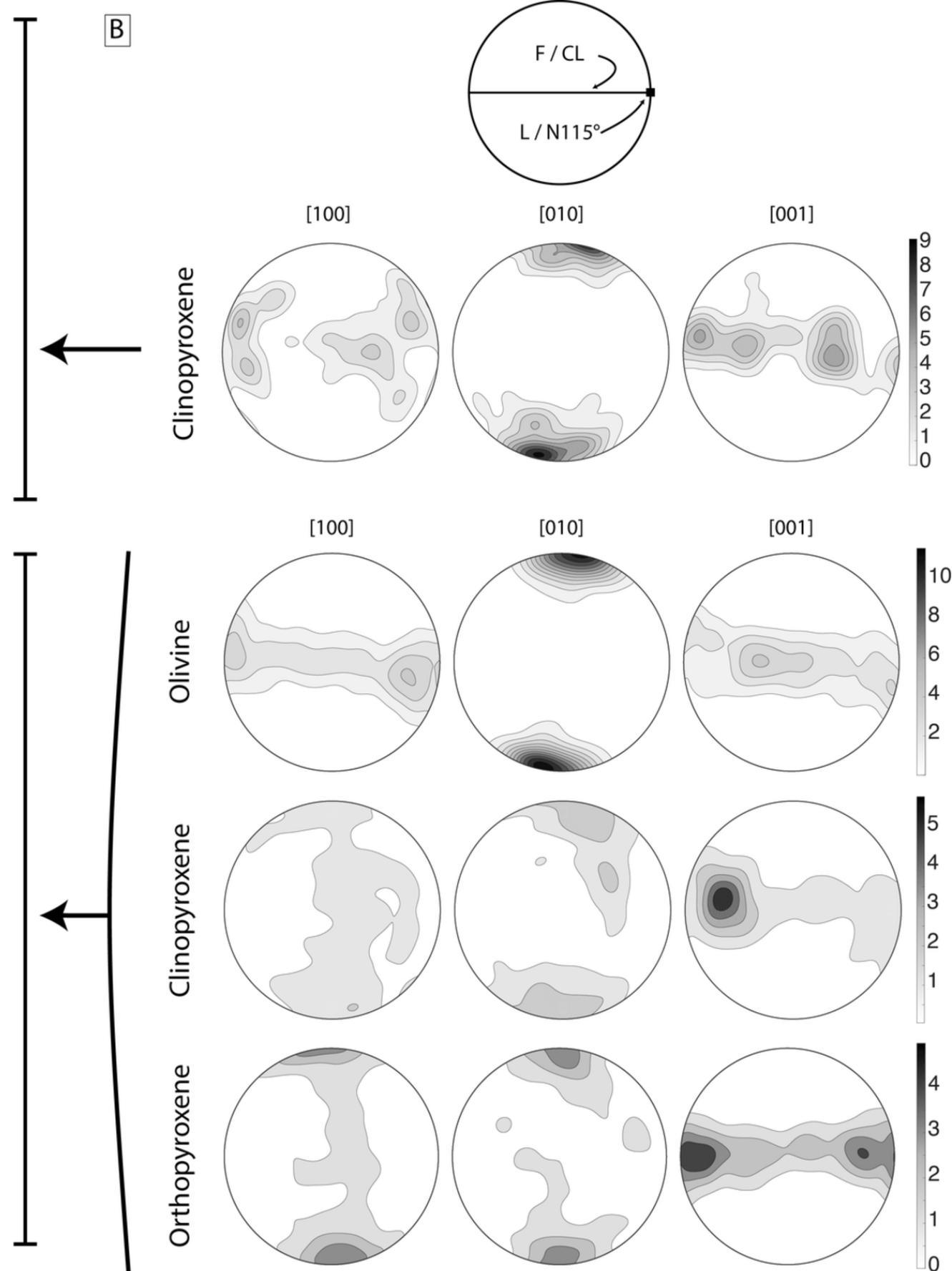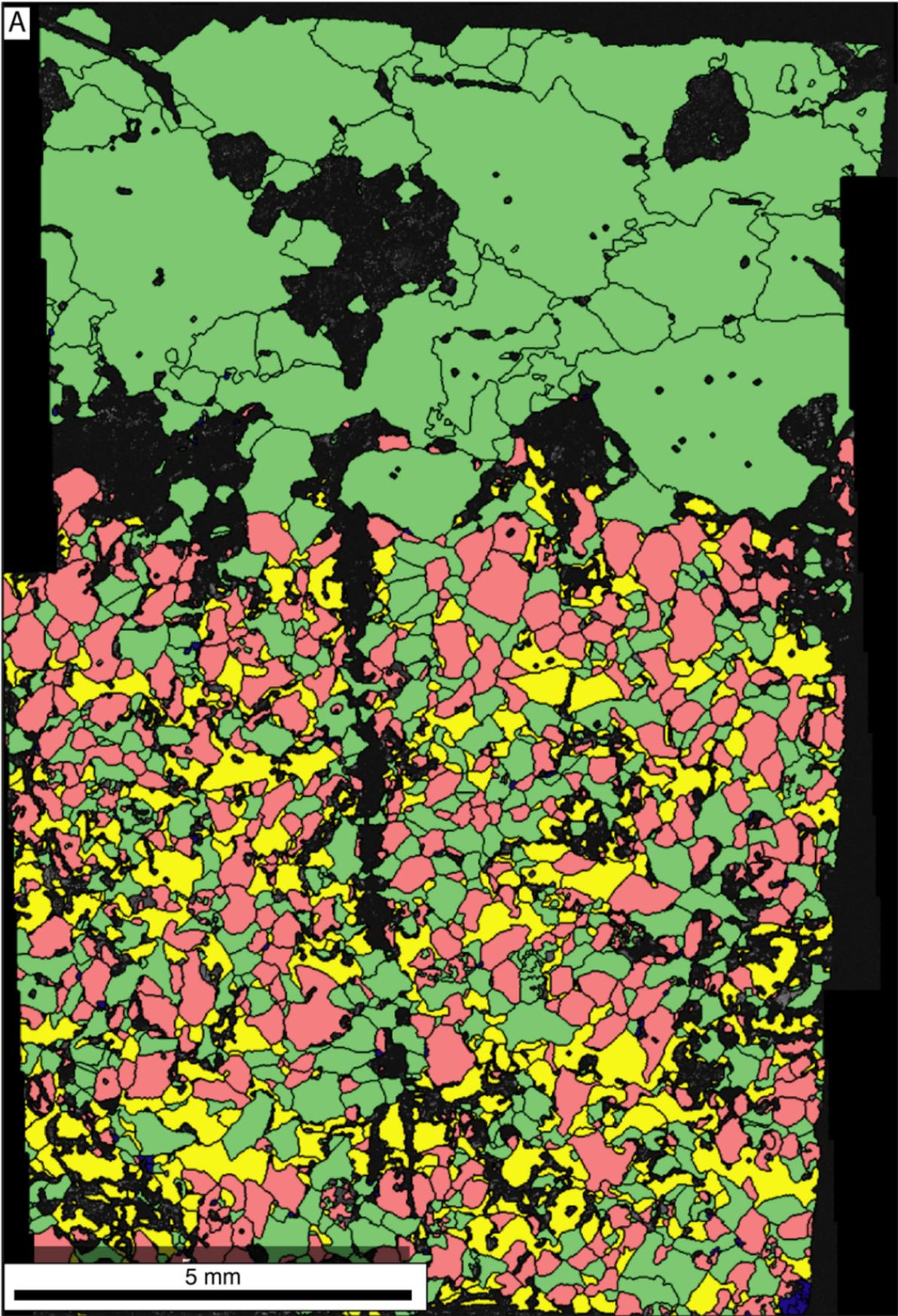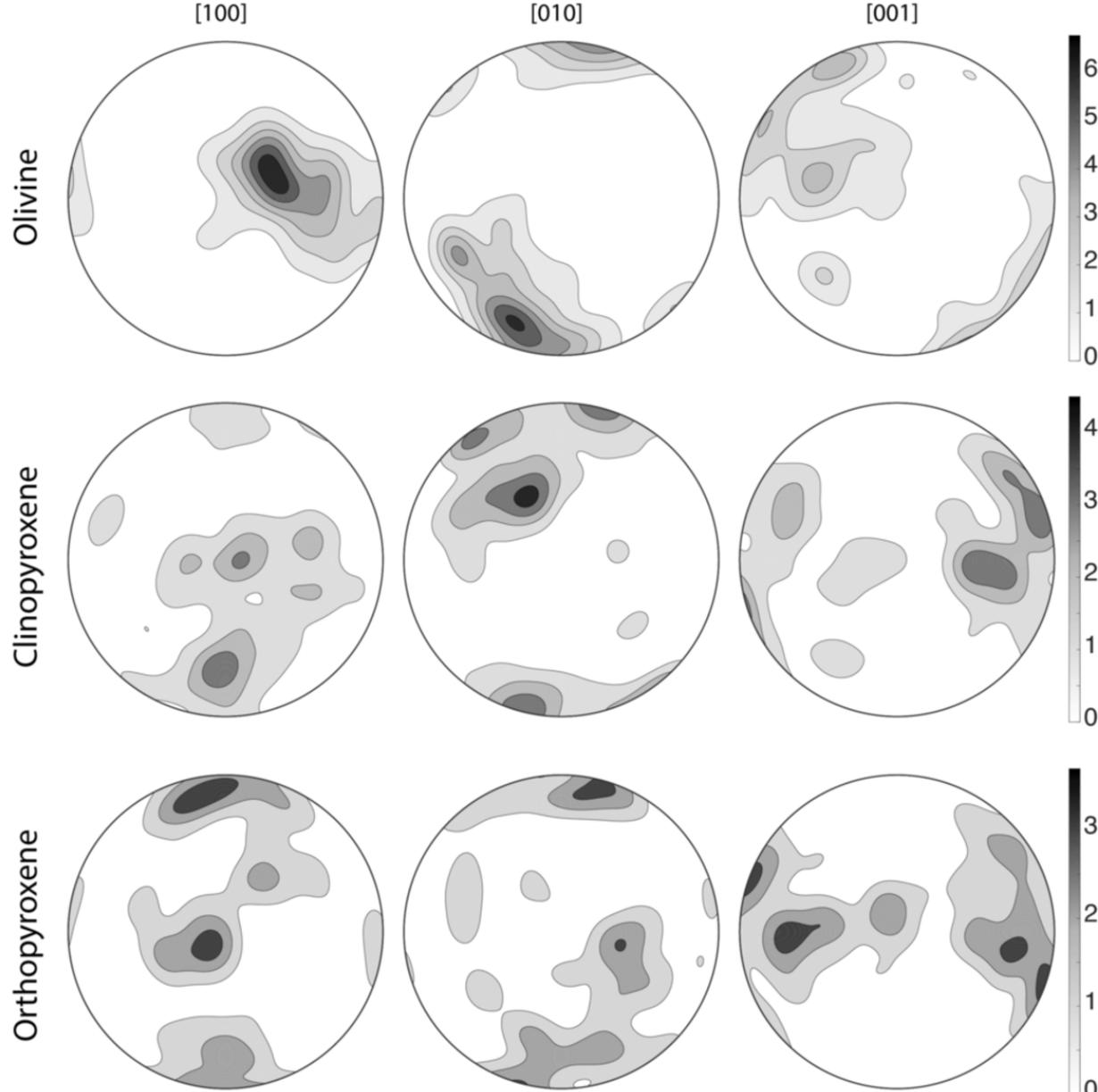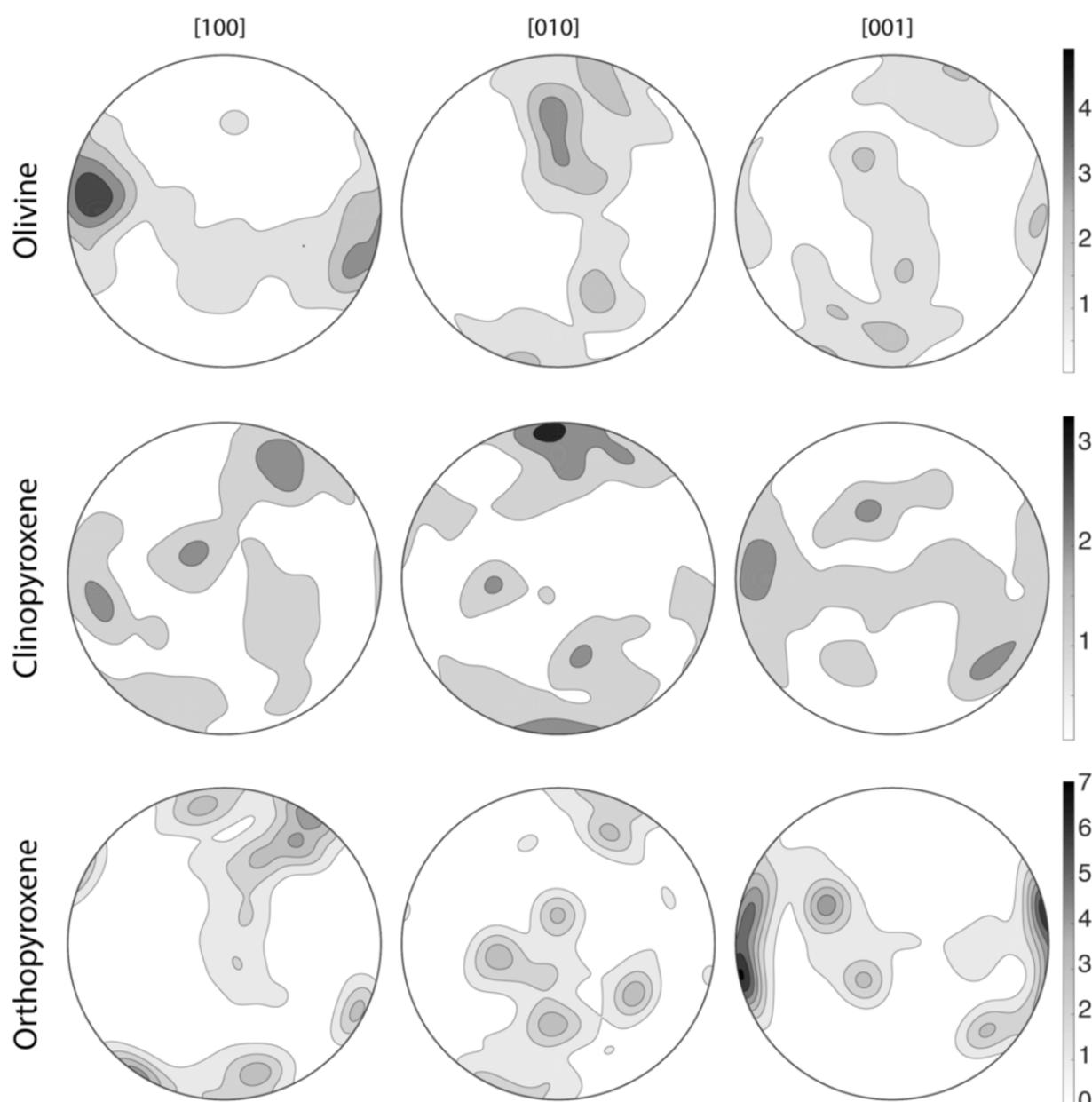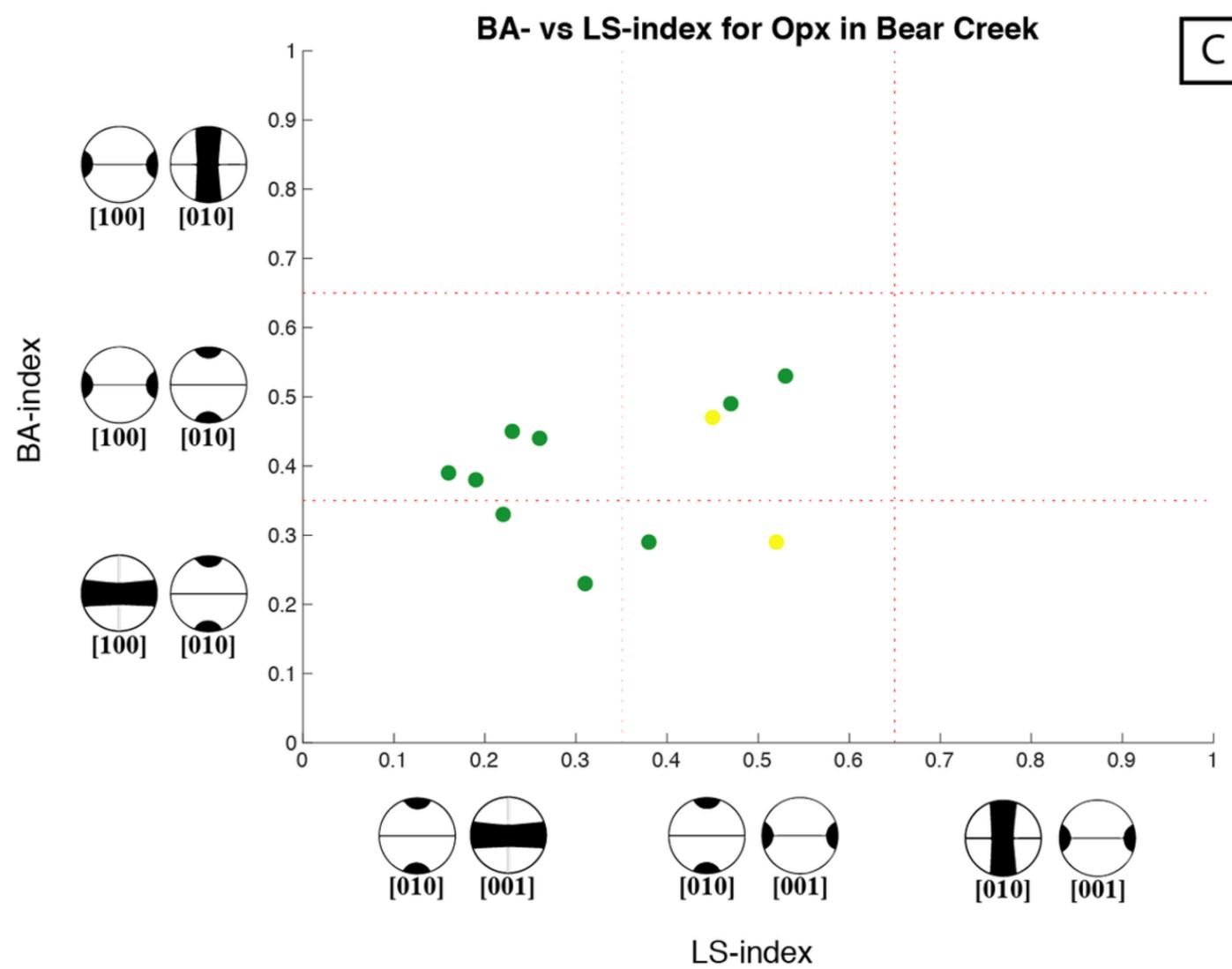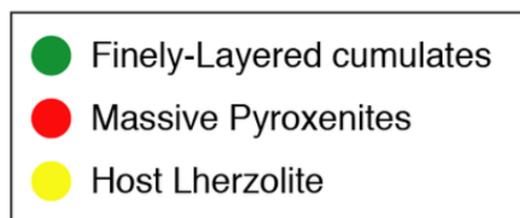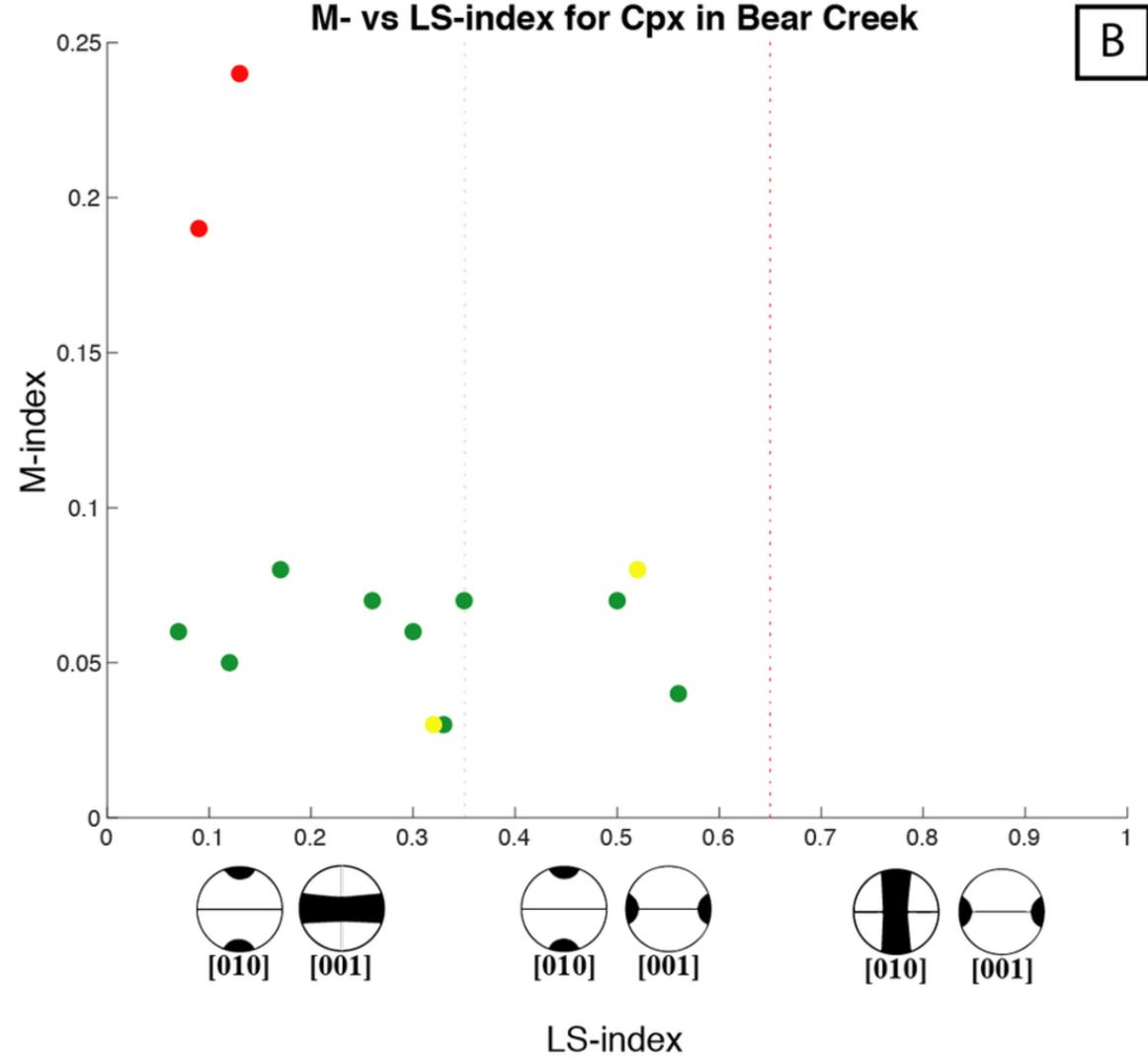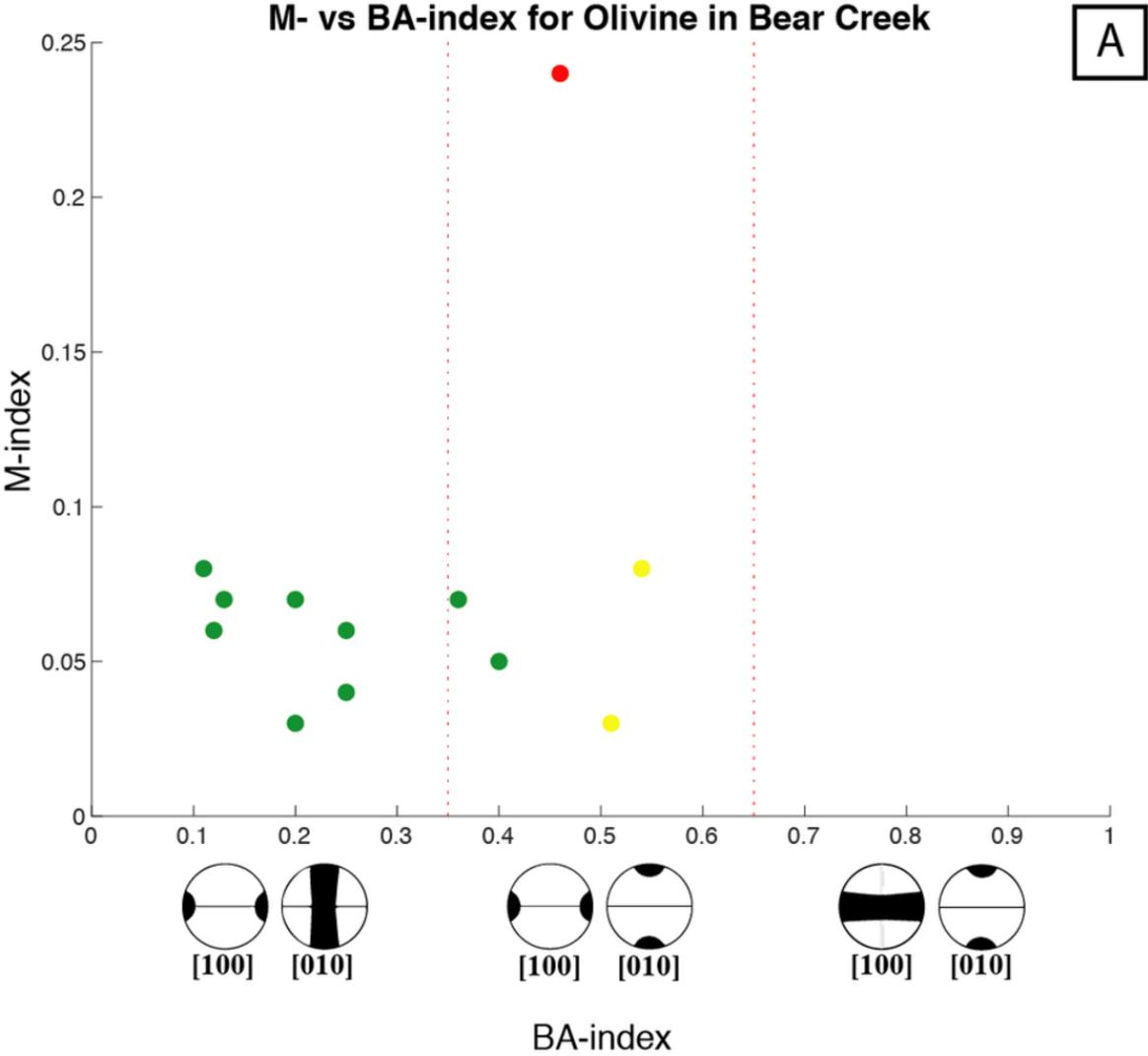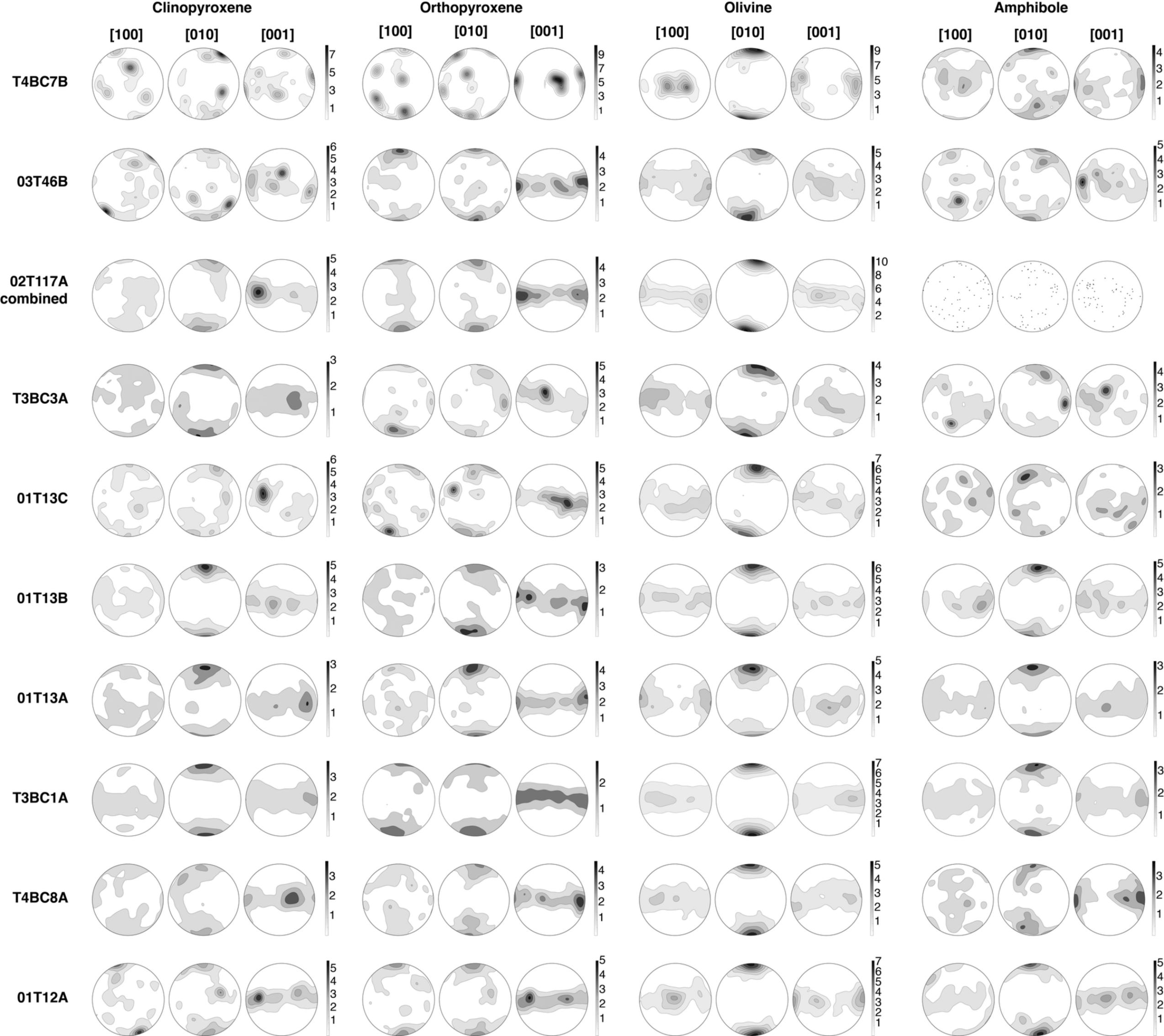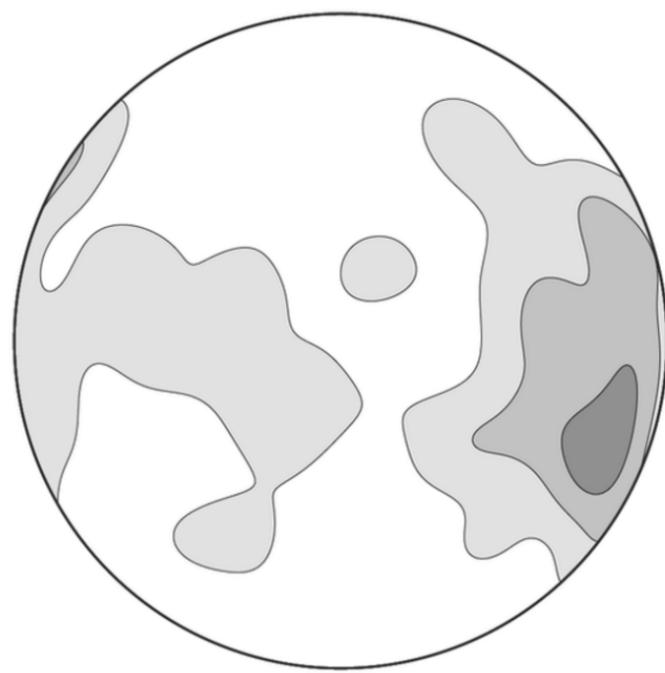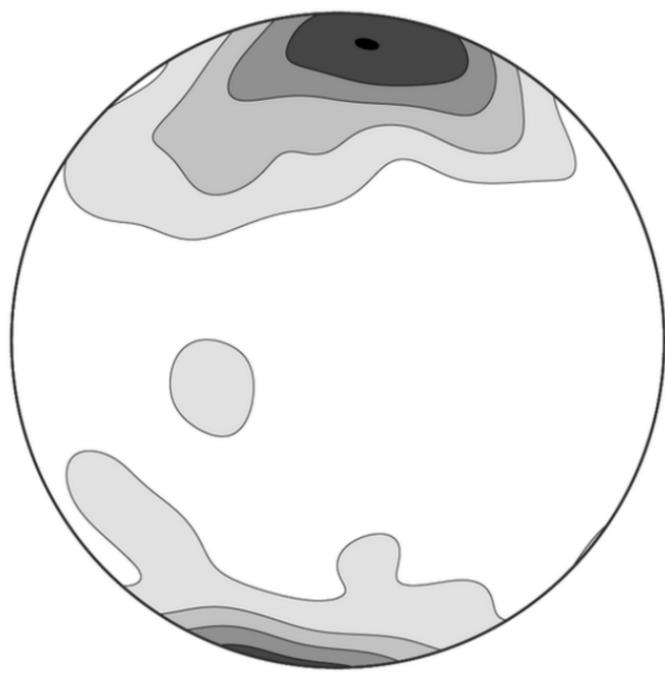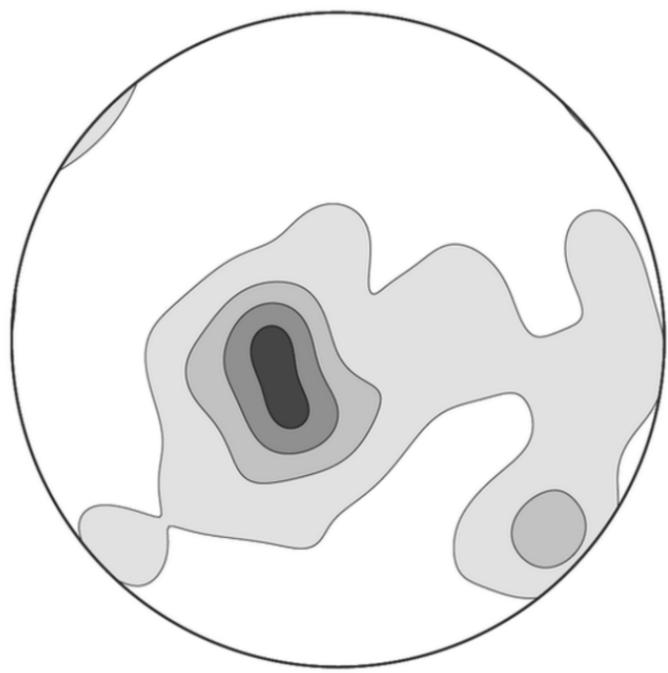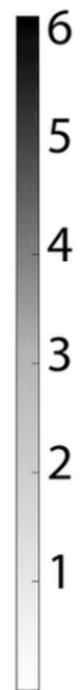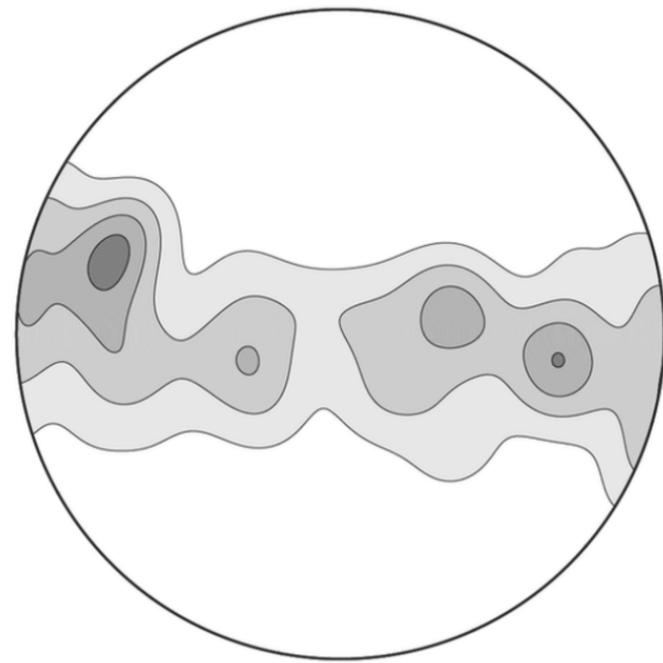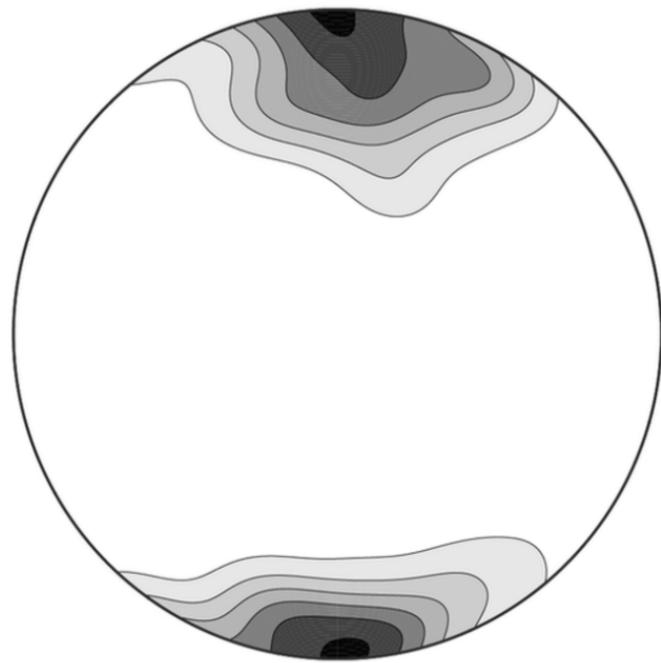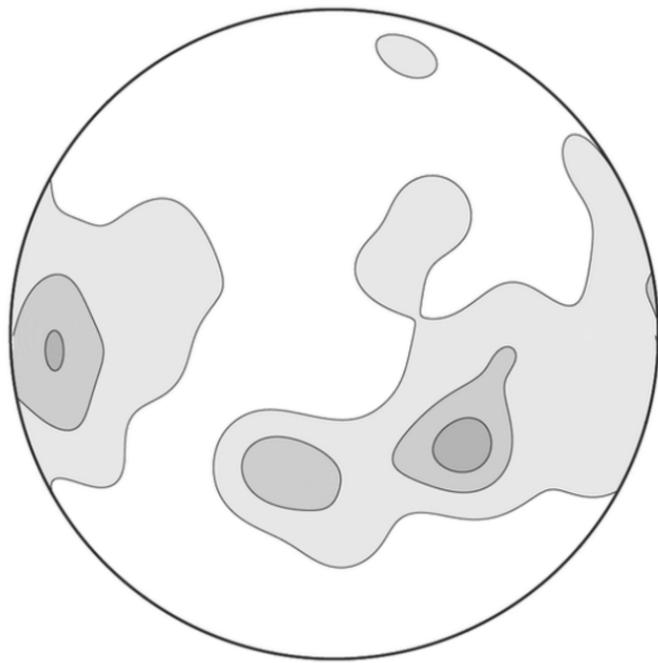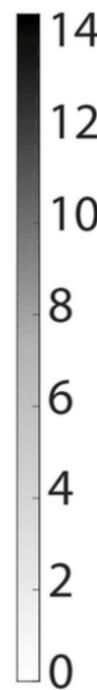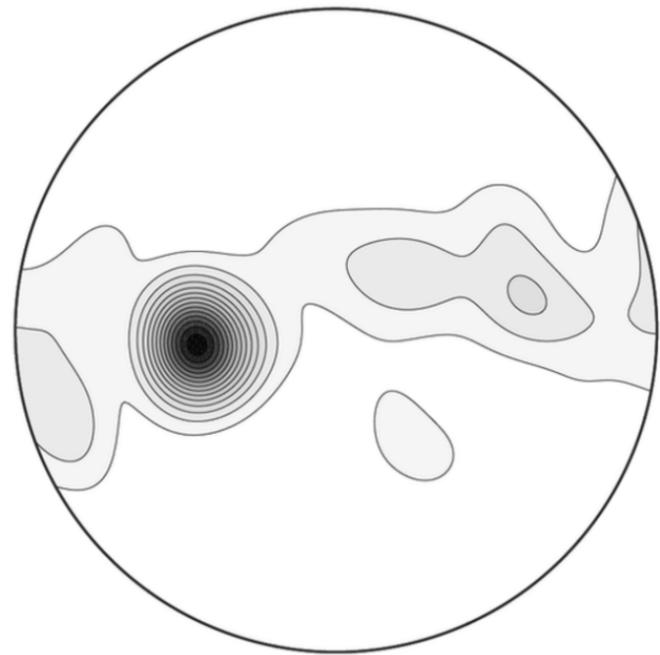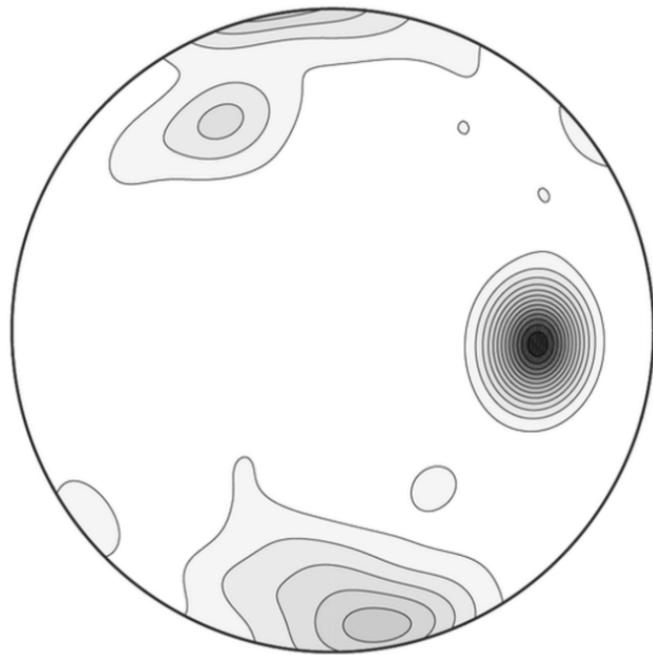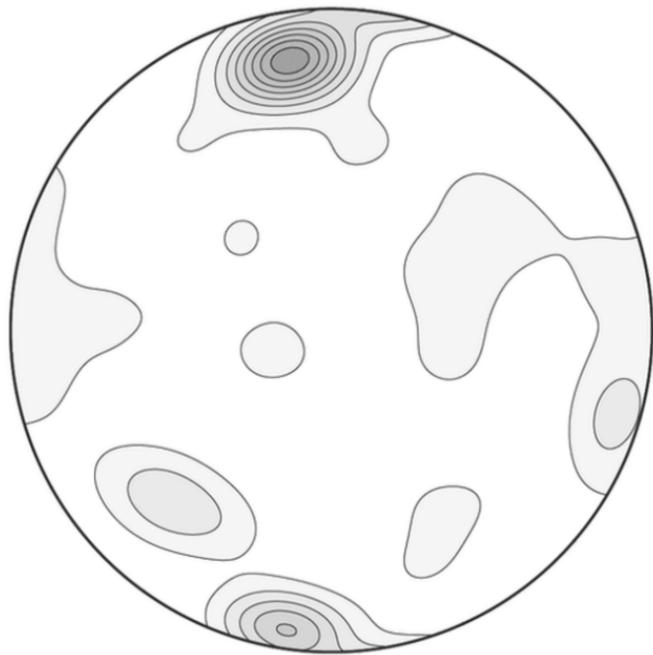
Table 1: Lithology, texture and mineral abundance for the Bear Creek samples

Sample	Lithology	Texture	Mineral abundance (%)*				
			Cpx	Opx	Olivine	Amphibole	Spinel
T4BC4A	Massive pyroxenite	Equigranular	95.66	0.74	2.98	0.61	0.01
T3BC4A	Massive pyroxenite	Equigranular	88.59	1.16	10.15	0.08	0.02
03T16A	Clinopyroxenitic intrusion	Pegmatitic	86.33	6.03	5.61	1.95	0.08
T4BC7B	Finely layered cumulates	Equigranular	89.2	0.56	6.1	4.12	0.02
03T46B	Finely layered cumulates	Equigranular	17.44	49.67	31.62	0.77	0.5
02T117A-coarse	Clinopyroxenitic intrusion	Pegmatitic	99.67	0.25	0	0.07	0.01
02T117A-cumulates	Finely layered cumulates	Equigranular	35.4	37.63	26.57	0.35	0.05
T3BC3A	Finely layered cumulates	Equigranular	58.44	26.61	12.56	2.37	0.02
01T13C	Finely layered cumulates	Equigranular	25.18	30.89	39.31	3.58	1.04
01T13B	Finely layered cumulates	Equigranular	81.13	6.01	8.87	3.88	0.11
01T13A	Finely layered cumulates	Equigranular	54.33	19.29	11.42	14.88	0.08
T3BC1A	Finely layered cumulates	Equigranular	39.66	21.52	34.92	3.83	0.07
T4BC8A	Finely layered cumulates	Equigranular	23.16	56.2	20.03	0.58	0.03
01T12A	Finely layered cumulates	Equigranular	26.14	44.3	25.58	3.53	0.45
T4BC2A	Lherzolite	Porphyroclastic	25.44	25.83	44.92	2.27	1.54
T4BC3A	Lherzolite	Porphyroclastic	5.59	30.69	63.56	0.02	0.14

* calculated on indexed EBSD data