The evolution of the footwall to the Ronda subcontinental mantle peridotites: insights from the Nieves Unit (western Betic Cordillera)

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Abstract: Strongly heterogeneous deformation and extreme metamorphic gradients characterize the dominantly carbonate Nieves Unit in the footwall to the Ronda mantle extrusion wedge in the western Betic Cordillera. A well-developed foliation and mineral lineation, together with isoclinal intrafolial folds, occur in silicate-bearing, calcite/dolomite marbles within a *c*. 1,5 km-thick metamorphic aureole underlying the peridotites. For the inferred maximum pressure of 300 MPa, petrological investigations allow to define temperature ranges for the main zones of the metamorphic aureole: forsterite zone (> 510 ºC; probably *c*. 700 ºC), diopside zone (510–430 ºC), tremolite zone (430–360 ºC), and phlogopite zone (360–330 ºC). Field structural analysis integrated with petrological, microstructural and EBSD textural data document large finite strains consistent with general shear within the metamorphic aureole, associated with NW-ward thrusting of the peridotites. On the other hand, post-kinematic silicate growth suggests that heat diffusion from the high-temperature peridotites continued after the final emplacement of the Ronda mantle extrusion wedge, leading to final zoning of the metamorphic aureole and to local partial annealing of calcite marble

39 textures, particularly in the highest-temperature zone of the thermally softened footwall

40 carbonates. Following substantial cooling, renewed crustal shortening affected the

41 whole Nieves Unit, resulting in widespread development of NE-trending meso-scale

42 folds.

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46 In the western Betic Cordillera (Fig. 1), high-temperature tectonic emplacement of the

47 Ronda mantle peridotites (Fig. 2) on top of crustal units (e.g. Lundeen, 1978) is marked

48 by the occurrence of a 'metamorphic sole' in the footwall rocks (e.g. Tubía *et al.*, 1997).

49 In the southeastern, inner peridotite massif (Sierra Alpujata), a high pressure-

50 temperature (P-T) shear zone records orogen-parallel (Tubía & Cuevas, 1986; Orozco

51 and Alonso-Chaves, 2012), sinistral shear between the peridotites and the Alpine

52 eclogites of the Ojen Unit. In contrast, in the outer peridotite massif of Sierra Bermeja

53 located to the NW (Fig. 2), a high-T dynamothermal metamorphic aureole records

54 orogen-perpendicular, foreland-ward thrusting of the peridotites on top of both low-P

55 metamorphic rocks of the Guadaiza Unit (Cuevas *et al.*, 2006; Esteban *et al.*, 2008) and

56 sedimentary successions of the Nieves Unit (Mazzoli & Martín-Algarra, 2011). These

57 features have been interpreted in terms of deformation partitioning associated with

58 oblique convergence during continental subduction and subsequent exhumation

59 involving the coeval activity of kinematically linked systems of reverse, strike-slip and

60 'normal-sense' shear zones (Mazzoli & Martín-Algarra, 2011). In this context, orogen-

61 parallel wrenching would have dominated the deeper, high-pressure portions of the

62 continental subduction system, whereas orogen-perpendicular thrusting would have

63 characterized the shallower, outer parts. Top-to-the-hinterland shear along the contact

64 between the Ronda peridotites and overlying crustal rocks is consistent with extrusion of

65 subcontinental mantle rocks of the overriding plate of the continental subduction system.

66 Rapid exhumation of this extruding mantle wedge would have been further aided by

67 strike-parallel extension and thinning of the crustal rocks overlying the peridotites (e.g.

68 Tubía, 1994; Balanyà *et al.*, 1997; Argles *et al.*, 1999; Platt *et al.*, 2003).

69 The crustal units lying in the footwall to the northwestern (i.e. outer) peridotite massif of

70 Sierra Bermeja (Fig. 2) provide a unique opportunity to analyse strongly heterogeneous

71 deformation associated with the 'hot' emplacement of a mantle wedge at shallow crustal

72 levels. In the interior of the ultramafic massif, the contact at the base of the Ronda

73   peridotites is exposed in the Guadaiza River tectonic window (Fig. 2). There, the contact

74   is marked by a dynamothermal metamorphic aureole formed at the expense of footwall

75   rocks (Cuevas *et al.*, 2006; Esteban *et al.*, 2008). High-T metamorphism is well

76   constrained by U/Pb SHRIMP dating of zircons from thin, syn-kinematic leucogranite

77   sheets that emanated from the aureole and intruded the overlying peridotites, which

78   provided ages of 18.8 ± 4.9 Ma (Sánchez-Rodríguez & Gebauer, 2000) and 21.8 ± 0.5

79   Ma (Esteban *et al.*, 2007). The heat required for partial melting of footwall crustal rocks

80   was provided by the fast emplacement of a *c.* 7 km thick, hot peridotite slab on top of the

81   Guadaiza Unit, within the general framework of NW-directed overthrusting. This

82   metamorphic aureole represents a low-viscosity, high-T (> 650 °C) shear zone

83   characterized by a marked strain gradient towards the contact with the overlying

84   ultramafic rocks (Esteban et al., 2008). On the other hand, the structure of footwall

85   carbonate units occurring to the NW of the present leading edge of the Ronda mantle

86   extrusion wedge has received much less consideration. The Nieves Unit rocks exposed

87   in this area are characterized by a strong inverse metamorphic gradient (Martín-Algarra,

88   1987; Mazzoli & Martín-Algarra, 2011). Within a *c.* 1.5 km thick zone, non-

89   metamorphic limestones/dolostones pass to high-grade marbles in contact with the

90   overlying peridotites. The metamorphic gradient is marked by silicate minerals

91   occurring in the marbles, defining a series of zones varying in thickness from tens to

92   several hundred metres and characterized by the occurrence of (moving towards the

93   peridotites; Figs. 3 and 4): (i) talc, (ii) phlogopite, (iii) tremolite, (iv) diopside, and (v)

94   forsterite. Extreme deformation gradients and changes in structural styles accompany the

95   metamorphic gradient and are the focus of this study, along with the texture of silicate–

96   bearing calcite marbles. Our results provide new insights into the metamorphic and

97   deformation processes taking place in carbonate units adjacent to the Ronda high-T

98   mantle wedge emplaced within the Betic orogen. In this (Behrmann, 1983) as well as in

99   similar carbonate-rich orogenic belts such as the Alps (Pfiffner, 1982; Heitzmann, 1987;

100 Burkhard, 1993) or the Apennines (Iannace *et al.*, 2005, 2007; Vitale & Mazzoli, 2005,

101 2009), marbles may record important deformation processes in the upper crust. Although

102 naturally and experimentally deformed carbonates have been the subject of intense

103 laboratory investigations aimed at a better understanding of the localization of

104 plastic/superplastic flow in these rocks (e.g. Schmid *et al.*, 1977, 1980, 1987; Rutter *et*

105 *al.*, 1994; Busch & Van der Pluijm, 1995; Badertscher & Burkhard, 2000; Bestmann *et*

106 *al.*, 2000; Pieri *et al.*, 2001a, 2001b; Badertscher *et al.*, 2002; Renner *et al.*, 2002; Ulrich

107 *et al.*, 2002; Barnhoorn *et al.*, 2004, 2005; Ebert et al. 2007a, b; Vitale *et al.*, 2007), the

108 peculiar tectonic setting of the Nieves Unit allowed us analysing in detail the variation of

109 mesoscopic and microscopic structural features and of textural fabrics of impure

110 calcite/dolomite marbles for a range of temperature regimes – spanning over several

111 hundreds of degrees Celsius – within the km-thick dynamothermal metamorphic aureole

112 in the footwall to the Ronda peridotite mantle extrusion wedge.

113

114 **Geological Setting**

115

116 The Betic Cordillera, representing the westernmost part of the Alpine orogen in the

117 Mediterranean region, resulted from the convergence between the African and Iberian

118 plates since Late Cretaceous times (Andrieux *et al.*, 1970; Durand-Delga & Fontboté,

119 1980; Martín Algarra & Vera, 2004). The Betic Cordillera, similarly to further Alpine

120 mountain belts in the Mediterranean area, is traditionally subdivided into Internal (i.e.

121 inner) and External (i.e. outer) Domains, with the Flysch Domain representing the

122 suture between the two (Fig. 1). The area of the present study forms part of the Internal

123 Domain of the western Betic Cordillera (Fig. 2). This domain is characterized by the

124 occurrence of low-angle extensional shear zones of regional extent, reworking the pre-

125 existing nappe structure (Balanyà *et al.*, 1993; Argles *et al.*, 1999). These extensional

126 contacts bound major allochthonous units characterized by varying tectonic and

127 metamorphic pre-Alpine and Alpine evolutions (Sánchez-Navas et al., 2012, and

128 references therein). The lower allochthon (Alpujarride Complex) records variable

129 Alpine P-T-t paths, generally characterized by high-pressure metamorphism followed

130 by decompression and final cooling at 18–16 Ma (Sánchez Rodríguez & Gebauer, 2000).

131 The upper allochthon (Malaguide Complex) generally lacks Alpine metamorphism

132 (Martín-Algarra *et al.*, 2009). A general lack of Alpine metamorphism characterizes

133 also the so-called 'Frontal Units' of the Internal Domain, one of which – i.e. the Nieves

134 Unit – is the subject of the present study; these units are thrust over the Flysch and

135 External Domains (Fig. 2). A series of unconformable marine basin successions

136 (Viñuela Group; Martín-Algarra, 1987) sit on top of the collapsed nappe pile.

137 Deposition of these Burdigalian clastic sediments predated backthrusting of the Flysch

138 Domain units above the Internal Domain (Fig. 2).

139 The Ronda peridotites form the lower portion of the traditionally recognized highest

140 Alpujarride nappe in the area. Above the mantle rocks, the highly attenuated Jubrique

141 crustal section – including granulitic gneisses, migmatites, high-to-low grade pre-
142 Mesozoic metapelites, and Triassic metasediments at the top – is characterized by
143 isograds lying roughly parallel to the lithological contacts and to the regional foliation
144 (Balanyà *et al.*, 1993, 1997; Tubía, 1994; Tubía *et al.*, 1997). Although the Ronda
145 peridotites and the Jubrique crustal section are traditionally grouped into the so-called
146 Los Reales Nappe (Didon *et al.*, 1973; Fig. 2), the contact between mantle and crustal
147 rocks consists of a high-T (T = 780 °C) major shear zone responsible for partial
148 exhumation of the peridotites (van Wees *et al.*, 1992; Balanyà *et al.*, 1993; Van der Wal
149 & Vissers, 1996; Argles *et al.*, 1999) and representing the top of the mantle extrusion
150 wedge according to Mazzoli & Martín-Algarra (2011). The Ronda peridotites are
151 mainly composed of plagioclase and spinel lherzolites (with subordinate garnet
152 lherzolites towards the top of the ultramafic body), mafic layers, and minor dunites and
153 harzburgites (Obata, 1980; Van der Wal & Vissers, 1996; Garrido *et al.*, 2011). They
154 form a series of massifs, the larger being those of Sierras Alpujata and Bermeja, where
155 ultramafic rocks are tectonically emplaced onto 'Blanca-type' Alpujarride units or onto
156 'Frontal Units' of the Internal Domain (Nieves Unit; Martín-Algarra, 1987; Fig. 2).
157 The 'Blanca-type' units (Ojen, Guadaiza, and Yunquera) share a typical Alpujarride
158 succession, with high-grade metapelites at the bottom, whose protoliths are of a
159 presumed Palaeozoic and Triassic age, and HT-marbles at the top, of probable Triassic,
160 and perhaps younger, sedimentary age (Hoeppener *et al.*, 1963, 1964). Stratigraphic
161 data (Martín-Algarra & Estévez, 1984; Martín-Algarra, 1987) and zircon geochronology
162 (Esteban *et al.*, 2011) demonstrate the Early Miocene age (c. 20–22 Ma) of the high-T
163 Alpine metamorphism that affects all 'Blanca-type' units, which was related to their
164 tectonic emplacement below the Ronda peridotites. However, the Ojen Unit records a
165 first stage of high-P metamorphic conditions (T = 790 °C, P ≥ 1.7 GPa; Tubía & Gil-
166 Ibarguchi, 1991) with a temperature peak dated at 19.9 ± 1.7 Ma (Sánchez-Rodríguez &
167 Gebauer, 2000) in eclogites derived from former Jurassic metabasites dated at 183 ± 3
168 Ma. Fast exhumation (at a rate of 6 km Ma^{-1}; Orozco and Alonso-Chaves, 2012, and
169 references therein) of these high-P rocks, rather than by ultra-rapid extension and
170 extreme thinning of an unrealistic (> 50 km thick) nappe stack, is likely to have
171 occurred within a subduction channel (Mazzoli and Martín-Algarra, 2011). On the
172 contrary, the Guadaiza Unit displays exclusively evidence of low-P metamorphic
173 conditions (T = 675–750 °C, P = 550–650 MPa; Esteban *et al.*, 2008). Finally, the
174 Yunquera Unit consists of an extremely thin, discontinuously exposed tectonic slice that

175   was possibly produced by tectonic delamination of the Ojen Unit in the footwall to the

176   Ronda peridotite wedge, as it also preserves relics of high-P metamorphism (Martín-

177   Algarra, 1987). In the area of the present study, the Yunquera Unit is locally exposed in

178   the footwall to the Ronda peridotites (Fig. 3). Both of these units record NW-directed

179   thrusting over the Nieves Unit (Esteban *et al.*, 2005; Mazzoli & Martín-Algarra, 2011).

180

181   **The Nieves Unit**

182

183   The Nieves Unit (Dürr, 1967) is formed by a non-metamorphic – apart from the aureole

184   at the contact with the Ronda peridotites – Triassic to Lower Miocene sedimentary

185   succession (Fig. 4). A few tens of metres thick sole of strongly sheared, brown scaly

186   clays of the Campo de Gibraltar Complex, including frequent quatzarenites and rare

187   micaceous sandstone phacoids, commonly occurs in its immediate footwall. This sole

188   constitutes a strongly thinned and stretched tectonic mélange interposed between the

189   Nieves Unit and the tectonically underlying Penibetic Units of the External Domain.

190   The hanging wall to the Nieves Unit is constituted by the Los Reales Nappe

191   (Alpujarride Complex), which in turn is overlain by the Malaguide Complex. Both

192   complexes thin rapidly northwards, forming a wedge-shaped, broad antiformal structure

193   whose core is occupied by the folded succession of the Nieves Unit (Fig. 4). The Ronda

194   peridotites at the base of the so-called Los Reales Nappe represent the thickest part of

195   the wedge but rapidly disappear towards the N due to the hanging-wall cutoff above the

196   Nieves Unit (Fig. 4). The Jubrique crustal succession overlying the peridotites is mostly

197   eroded above the main outcrop of the Nieves Unit; however, thin remnants are

198   preserved as small klippen on top of the youngest beds of the Nieves succession (Nava

199   Breccia Fm., see below), being in turn locally overlain by klippen of Malaguide

200   Palaeozoic rocks (Fig. 3).

201

202   *Stratigraphy*

203

204   The stratigraphic succession of the Nieves Unit includes a *c.* 1.5 km thick Meso-

205   Cenozoic succession (Fig. 4) mainly made of carbonate formations showing typical

206   Alpine-Mediterranean facies.

207   The lowest formation is made of very thick- to medium-bedded Norian dolostones

208   showing shallow marine facies and a cumulative thickness in excess of 1 km (unit 1 in

209   Fig. 4). Fine-grained terrigenous intercalations are locally present, especially in the

210   lowest part and towards the upper third part of the succession.

211   The top of the Triassic succession is made of up to 200 m thick, basinal, greyish and

212   locally yellowish marls alternating with hemipelagic micritic limestones (unit 2 in Fig. 4)

213   that gradually pass upwards (and laterally northwards) to black to dark-grey limestones

214   (unit 3 in Fig. 4). The Jurassic-Paleogene part of the succession (unit 4 in Fig. 4) starts

215   with cherty limestones (usually less than 200 m thick), locally including carbonate

216   turbidites rich in crinoidal bioclasts, mass-flow limestone breccias and red nodular

217   limestone horizons bearing lower Hettangian and Sinemurian ammonites. The

218   succession becomes marly towards the top of the Lower Jurassic section. Upwards, a

219   condensed pelagic succession develops (Martín-Algarra *et al.*, 1998; O'Dogherty *et al.*,

220   2001). It is made of Middle Jurassic red nodular and cherty limestones and radiolarian

221   cherts, Upper Jurassic and lowermost Cretaceous marly and cherty limestones with

222   ammonites, *Saccocoma*, *Aptychus* and *Calpionella* limestones (unit 5 in Fig. 4). The

223   latter are followed by varicoloured marls bearing Paleogene to Aquitanian planktonic

224   foraminifera and calcareous nannoplankton, with intercalated carbonate turbidites. The

225   succession of the Nieves Unit is topped by a Lower Miocene, rauhwacke-like,

226   continental carbonate breccia (Nava Breccia Fm; unit 6 in Fig. 4). This breccia is

227   cartographically unconformable onto the older formations of the Nieves Unit, although

228   being intensely folded together with the underlying part of the succession. The

229   palaeogeographic and palaeotectonic significance of the Nava Breccia Fm –

230   representing the youngest deposit involved in the deformation and, therefore, a key

231   stratigraphic unit of the study area – has been investigated by Martín-Algarra & Estévez

232   (1984) and Martín-Algarra (1987, 2004) to which the interested reader is referred to.

233   Towards the SE, the sedimentary succession described above is overthrusted by the

234   Ronda peridotites (Fig. 5A), intensely deformed and transformed into a metamorphic

235   succession essentially made of both calcite and dolomite marbles, locally rich in quartz

236   lenses and levels, with subordinate intercalations of calcschists, calcareous metapelites

237   and paramphibolites (Figs. 5B-F and 6A-E). This transformation can be laterally

238   followed in the field, bed by bed, from non-metamorphic lithotypes to high-grade rocks

239   (see below), especially in the Torrecilla area (Fig. 2).

240

241   *Structure*

242

7

243 The map-scale structure of the Nieves Unit is dominated by a NW-verging, overturned
244 syncline (Figs. 3 and 4) that can be traced for several tens of kilometres along strike
245 (Fig. 2). Regional folding around a gently NE-plunging axis (Fig. 7A) is well defined
246 also statistically by the distribution of the poles to bedding (S_0) and to the main
247 composite planar fabric (S_0-S_1) measured from the metamorphic aureole (see below).
248 The stratigraphic succession, dominated by limestones and dolostones in the normal
249 limb of the syncline, gives way to calcite/dolomite marbles in the overturned fold limb,
250 which is characterized by the previously mentioned dramatic metamorphic gradient.
251 Each stratigraphic unit mapped in the normal fold limb has a corresponding, well-
252 defined high-T equivalent in the metamorphic aureole of the overturned limb (Fig 4).
253 Here the original bedding, marked by clearly distinguishable lithological changes (Fig
254 5B), is generally parallel to a foliation, defining a composite (S_0-S_1) planar fabric (Fig
255 5C-E). A dominantly SE-plunging (Fig. 7B) mineral lineation (L_1), defined by
256 elongated calcite/dolomite and locally tremolite crystals (Fig. 6A), occurs on the
257 composite (S_0-S_1) planar fabric. Isoclinal intrafolial folds (Fig. 6B) display variably
258 plunging hinge lines (mainly tending to lie close to L_1; Fig. 7C), and dominantly SSE
259 dipping axial surfaces (Fig. 7D). Field evidence for non-coaxial strain is rare; where
260 observed, it is represented by a sigmoidal foliation and S-C-C' structures indicating top-
261 to-the-NW sense of shear.

262 Minor NE-SW trending, open to close parasitic folds are associated with the regional
263 syncline in the non-metamorphic sedimentary succession of the Nieves Unit. They are
264 asymmetric (*s*-shaped looking towards NE) in the normal limb of the regional syncline
265 (Fig. 6D), and symmetric (*m*-shaped) in its core (hinge region). An axial plane cleavage
266 – or weakly convergent cleavage in competent beds – is associated with these folds.
267 This foliation is intensely developed in the core of the regional syncline, particularly in
268 Jurassic-Paleogene units.

269 Open to tight, post-metamorphic folds deform the main composite planar fabric (S_0-S_1)
270 in the metamorphosed overturned limb of the regional syncline (Fig. 6C), as well as
271 bedding (S_0) in its core and in the normal fold limb. These folds show NE-SW trending
272 hinge lines (Fig. 7E) and both SE- and NW-dipping axial surfaces (Fig. 7F). A variably
273 developed second-phase cleavage (S_2) is associated with these folds, being axial-planar
274 to them in less competent lithologies and forming convergent cleavage fans in more
275 competent beds. The lack of neo-formed metamorphic minerals on S_2 cleavage surfaces
276 suggests that this planar structure developed at very low-grade conditions. Within the

277   metamorphic aureole, the intersection between S_2 and the composite planar fabric (S_0-S_1)

278   defines a dominantly NE-trending linear structure parallel to the late fold hinges. A

279   similar NE-SW trend is also defined by calcite rods developed along buckled marble-

280   phyllite interfaces in the metamorphic units (the phyllites representing metamorphosed

281   terrigenous intercalations in the carbonates).

282   The youngest shortening-related structures within the Nieves Unit include scattered

283   kink bands (Fig. 6E), whereas high-angle, dominantly extensional faults dissect all of

284   the previous structures as well as the main nappe contacts.

285

286   **Petrology and Microstructural Analysis: Materials and Methods**

287

288   In the following we describe the most representative mineral assemblages and textures

289   of 25 selected silicate-bearing marbles sampled in the studied metamorphic aureole

290   (Figs. 3 and 4). Rocks were sampled along the A-397 road (km 17 to 21), near the Cerro

291   de los Cascajares peak, crossing the phlogopite, tremolite, diopside, and forsterite

292   metamorphic zones defined by Martín-Algarra (1987), and in further outcrops, within

293   the forsterite zone, close to the Cerro Alcojona peak (Table 1).

294   A first approximation to metamorphic conditions for the impure marbles of the Sierra de

295   las Nieves contact aureole was reached by means of appropriate petrogenetic grids. P-T

296   phase diagram projections (Fig. 9) were calculated using Perple X (Connolly, 1990,

297   2009) for the mixed-volatile K_2O-CaO-MgO-Al_2O_3-SiO_2-H_2O-CO_2 (KCMASHC)

298   system, following the method by Connolly & Trommsdorff (1991). Thermodynamic

299   mineral data and the equation of state for H_2O-CO_2 fluids were those of Holland &

300   Powell (1998, revised version 2002). The only solution model considered in the

301   calculations was that of the H_2O-CO_2 fluid (Trommsdorff & Connolly, 1991). In the

302   figures and following text, F represents the fluid composition expressed as X_{CO_2} [=

303    $CO_2/(CO_2+H_2O)$] and the numbers in parentheses are those of the pseudocompounds

304   (Connolly & Kerrick, 1987) used by Perple X to represent fluid composition. All other

305   phases were taken as pure phases and the following solid end-members were considered

306   (abbreviations after Whitney & Evans, 2010): calcite (Cal), clinochlore (Clc),

307   clinohumite (Chu), diopside (Di), dolomite (Dol), forsterite (Fo), microcline (Mc),

308   phlogopite (Phl), quartz (Qz), tremolite (Tr), and spinel (Spl).

309   This was a realistic simplification because most of the phases observed in the rocks

310   approach pure end-members compositions. In fact, X_{Mg} [= $Mg/(Mg+Fe)$] in all

311 ferromagnesian phases (amphibole, chlorite, clinohumite, clinopyroxene, olivine, and
312 spinel) commonly ranges between 0.90 and 0.95 (but in many cases it exceeds these
313 values) in the impure marble samples. Likewise, both calcite and dolomite display
314 compositions close to stoichiometry. Some exceptions are reported for phlogopite (with
315 up to 1.6 atoms per formula unit – a.p.f.u. – of total Al; 11 oxygens), chlorite (with up
316 to 2.6 a.p.f.u. of total Al; 14 oxygens) and clinohumite (very rich in fluorine: up to 1.7
317 a.p.f.u.; 13 cations). Pargasitic amphibole and clinopyroxene ($X_{Mg} \sim 0.60$) occurring in
318 some pelitic intercalations also differ from the general trend, but these assemblages
319 were not taken into account in the calculations. Uncertainties associated with ignoring
320 these solid solutions are discussed below.

321 Electron backscatter diffraction (EBSD) data collected by scanning electron microscope
322 (SEM) were used for textural analysis of seven calcite marble samples and one dolomite
323 marble sample from the metamorphic aureole found in the overturned limb of the
324 regional syncline affecting the Nieves Unit. Samples 09A-1 and 09A-3 are from the
325 forsterite zone; sample 09A-7 is from the diopside zone; samples 09A-12, 09A-15, 09A-
326 18 and 09A-19 are from the tremolite zone, and sample 09A-20 is from the phlogopite
327 zone. Thin sections of each sample were cut parallel to the mineral lineation (X) and
328 perpendicular to the composite planar fabric (S_0-S_1 = XY; see below). Thin sections
329 were mechanically polished using down to 1 μm diamond paste. To remove
330 mechanically-induced surface damage, a further 3 hours of polishing, with 0.06 μm
331 colloidal silica, was undertaken. Thin sections were coated with a thin layer of carbon to
332 prevent surface charging in the SEM. EBSD analysis was undertaken on a ZEISS-EVO
333 40XVP scanning electron microscope at Curtin University, Perth, Western Australia.
334 Collection, indexing and analysis of electron backscatter diffraction patterns (EBSP)
335 were carried out using the CHANNEL 5.10 software by Oxford Instruments. The EBSD
336 data were noise reduced using a 'wildspike' correction and a 7-neighbour zero solution
337 extrapolation following standard procedures (Reddy *et al.*, 2007). Orientation data were
338 plotted with respect to the mesoscopic sample lineation and foliation (X direction and
339 XY plane respectively), as lower hemisphere, equal area projections, using CHANNEL
340 Mambo software. Pole figure orientation data were plotted in terms of all of the
341 collected EBSD data ('all data') and as one point per grain data. For the latter, grain
342 boundaries were defined by misorientation boundaries in excess of 10°, and any twin
343 boundaries identified by a 78° rotation around <20-21> were ignored.

344 Finally, thin section microphotographs (acquired using a Leica DFC280 digital camera

345   mounted on a Leitz Laborlux 12 Pol polarizing microscope) were employed for modal

346   and carbonate grain size analyses of the same samples used for the textural analysis.

347   Modal analyses were aimed to the quantification of the volume abundances of non-

348   carbonate minerals (i.e. 'impurities', or secondary phases) and were performed using a

349   point counting routine of the Leica QWinPlus V3.2.1 image analysis software.

350   Carbonate grain size analyses were performed using the ImageJ 1.46r image analysis

351   software (further details are summarized in Table 2).

352

353   **Petrology and Microstructural Analysis: Results**

354

355   *Petrography of silicate–bearing marbles*

356

357   The studied rocks consist of impure marbles containing relatively low amounts (6.5% to

358   17.8% for the samples listed in Table 2) of non-carbonate minerals. Most of lower-T

359   dolomite marbles can be defined as phlogopite-tremolite marbles with the significant

360   local occurrence of chlorite or spinel (Table 1). They display a granoblastic texture with

361   a rough foliation defined by the preferred orientation of fine grained aggregates of

362   phlogopite, which may be also strongly folded (Figs. 8A-B). Phlogopite is also found as

363   randomly oriented, small, isolated flakes and as randomly oriented, hypidioblastic larger

364   flakes, commonly related to other silicates such as spinel and tremolite (*e.g.* sample

365   09A-11, Fig. 8D). Large, randomly oriented phlogopites may, however, display

366   distinctive undulose extinction. Colourless *amphibole* is commonly associated with

367   calcite grains or levels, and occurs as prismatic to acicular crystals rich in small

368   dolomite inclusions. Amphibole grains are isolated (Fig. 8F) or give place to palmed

369   aggregates, in many places with clear preferred orientation parallel to the main foliation

370   (Fig. 8C). Chlorite (sample 09A-16; Table 1) appears as large hypidioblastic flakes

371   having plenty of phlogopite and dolomite inclusions. Spinel (sample 09A-11; Table 1)

372   is found as xeno- to hypidioblastic, isolated grains that grew after – or related to –

373   recrystallized phlogopite flakes (Fig. 8D).

374   Moving towards the peridotites, clinopyroxene is found in modal quartz-rich rocks

375   (with the exception of marble sample 09A-1; Table 1), usually consisting of pelitic or

376   quartz-feldspar-bearing intercalations within the calcite marbles. Clinopyroxene appears

377   as green pleochroic, isolated crystals or aggregates with largely variable grain size at the

378   contact between quartz-rich and carbonate-rich levels (Fig. 8E). Highly birefringent

379 scapolite poikiloblasts as well as minor amounts of alkali-feldspar, phlogopite and
380 tremolite (Fig. 8E) also appear in some of these rocks, (that will not be taken into
381 account in the phase diagram calculations below).

382 Within a few tens of metres to the peridotite, olivine-bearing calcite or calcite-dolomite
383 marbles contain variable silicate assemblages and textures (Table 1). The most
384 representative assemblages consist of Cal + Dol + Phl + Amp + Ol ± Chu ± Spl
385 (abbreviations after Whitney & Evans, 2010), while clinopyroxene and chlorite have
386 been found only in one sample within this zone. Silicate minerals typically display
387 textural characteristics indicative of static growth post-dating deformation (e.g. large
388 grain size, lack of preferred orientation, hypidioblastic shape; Fig. 8F). In further
389 instances silicates occur as non-deformed aggregates of mostly xenoblastic grains with
390 complex interrelationships reflecting metamorphic reactions (Fig. 8G). This is also
391 indicated by the occurrence of late poikiloblasts of dolomite surrounding these
392 aggregates.

393 In tremolite and higher-T mineral zones, centimetre-scale phlogopite-tremolite clots that
394 clearly displaced and overgrew the main composite planar fabric (S_0-S_1) are observed
395 also in outcrop (Fig. 5F). However, as mentioned previously, tremolite crystals also
396 occur on the main composite foliation.

397

398 *Metamorphic analysis*

399

400 A simplified petrogenetic grid for the KCMASHC system and a wide P-T range is
401 shown in Fig. 9 (A). Both fluid-present and fluid-absent univariant equilibria occur. The
402 most striking feature of this P-T projection is that most of the curves display very steep
403 and remarkably constant Clapeyron slopes, thus defining narrow zones in which the
404 stability of phase assemblages is strongly dependent on temperature and fluid
405 composition (e.g. Fig. 9B). Calculations for the K_2O-CaO-FeO-MgO-Al_2O_3-SiO_2-H_2O-
406 CO_2 system (KCFMASHC) that take into account the limited Fe-Mg solid solution (X_{Mg}
407 usually < 0.95) in the ferromagnesian minerals reveals a negligible shift towards higher
408 temperatures of the reactions of interest (not shown in the figures) and will not be
409 considered in the following. The effect of the solid solutions models for phlogopite
410 (between the phlogopite-eastonite end-members) and chlorite (between the clinochlore-
411 amesite end-members), applying for the Al amounts observed in these minerals, has
412 also been tested. As a general trend, for the reactions of interest, pseudounivariant

413   curves with Al-richer pseudocompounds shift toward higher temperatures. However,

414   within the range of the limited solid solutions observed in the studied minerals, this is a

415   minor effect when compared with that of the changing fluid phase composition and will

416   not be further considered. As a result, in the following analysis univariant reactions with

417   steep Clapeyron slopes (especially those involving phlogopite or chlorite) should be

418   regarded as indicators of the minimum temperature conditions at which observed

419   minerals or assemblages were stable. In contrast with these steep reactions, only a few

420   reactions display gentle or variable Clapeyron slopes. Among these, the reaction Cal +

421   Tr + Fo = H_2O + Di + Dol sets the highest pressure limit for the observed mineral

422   assemblage, in which calcite, amphibole and olivine coexist. Accordingly, all further P-

423   T projections (Fig. 9B to D) correspond to the expanded field between 50 and 500 MPa,

424   and between 300 and 700 ºC for the same KCMASHC system. This pressure range is

425   consistent with the previously described geological setting of the contact aureole. Based

426   on geological evidence pointing to a burial depth of *c.* 10 km (including the mantle

427   peridotites; Mazzoli & Martín-Algarra, 2011), maximum pressures may be further

428   constrained to around 300 MPa. These P conditions are roughly similar to those

429   suggested by Esteban et al. (2008) for the Guadaiza River tectonic window just south of

430   our study area (Fig. 2).

431   Metamorphic conditions for the occurrence of phlogopite, amphibole, spinel and

432   chlorite in the studied aureole can be discussed from selected representative equilibria

433   shown in Fig. 9 (B) and summarized in Fig. 9 (D). Minimum temperature conditions for

434   the first occurrence of phlogopite in the dolomite-bearing rocks are controlled by the

435   univariant reaction indicating the breakdown of detrital microcline (e.g. Rice, 1977) in

436   the presence of water-rich fluid: F(H_2O) + Dol + Mc = F(CO6.6) + Cal + Phl.

437   Depending on the latter parameter, the Phl-in reaction turns to be projected as

438   pseudounivariant equilibria (Connolly & Trommsdorff, 1991), represented as grey thin

439   dashed curves, in which only one (that of coexisting 6.6 and 9.9 pseudocompounds) is

440   labelled with the complete reaction and all the remaining ones are labelled with the

441   mean value of coexisting pseudocompounds (Fig. 9B). As a result, it can be deduced

442   that the first appearance of phlogopite can take place in a wide divariant field with

443   increasing temperature conditions for CO_2-richer fluids. For the estimated maximum

444   pressure of 300 MPa (thick grey dashed line in Fig. 9D), the minimum temperature for

445   the stability of phlogopite is 350 ºC.

446   A similar analysis can be carried out for defining the *tremolite*-in conditions. In

447   dolomite and quartz-bearing marbles, this takes place due to the univariant reaction:

448    $F(H_2O)$ + Qz + Dol = F(CO6.6) + Cal + Tr. The divariant field defined by

449   pseudounivariant equilibria (only that corresponding to pseudocompounds 16.4 and

450   19.7 is represented with a thin black dashed curve) is shaded in light grey in Fig. 9 (B).

451   In pure calcite marbles the univariant tremolite-in reaction is: $F(H_2O)$ + Qz + Cal + Phl

452   = F(CO6.6) + Tr + Mc; sample pseudounivariant equilibrium with pseudocompounds

453   13.1 and 16.4 is represented with a thin continuous black curve and the divariant field is

454   shaded in darker grey in Fig. 9 (B). Thus, at 300 MPa, the minimum temperature for the

455   appearance of tremolite ranges from 360 ºC in quartz-bearing dolomite marbles to 410

456   ºC in calcite marbles (Fig. 9D). In the studied selected samples the latter rock type is

457   more abundant than the former one (cf., Table 1).

458   The first appearance of *spinel* in the petrogenetic grids of Fig. 9 (B, D) is indicated by

459   the fluid-absent univariant reaction: Tr + Dol + Mc + Clc = Cal + Phl + Spl + Qz at an

460   almost constant temperature of 430 ºC. At higher T (500 ºC at 300 MPa), the reaction

461   Dol + Mc + Clc = Cal + Phl + Spl + Qz + F takes place. Both are clinochlore

462   breakdown reactions yielding the same solid products. The coexistence of Cal, Phl and

463   Spl is well documented (e.g. Fig. 8D), whereas Qz is always lacking. This can be easily

464   explained taking into account that the chlorite end-member phase employed in the

465   calculations (clinochlore) is richer in Si than real chlorites observed in the rocks (3 Si

466   a.p.f.u. versus 2.5 Si a.p.f.u.). Irrespective of composition, *chlorite* is stable in the

467   KCMASHC system, at Al-rich compositions, in the whole considered P-T range.

468   Selected relevant equilibria explaining the occurrence of diopside, olivine, and

469   clinohumite are displayed in Fig. 9 (C). Observed *diopside*-bearing rocks are lacking in

470   dolomite and very rich in quartz (Table 1). The first appearance of diopside in calcite

471   marbles takes place through the univariant reaction: $F(H_2O)$ + Qz + Cal + Tr = Di +

472   F(CO6.6). As in the previous cases, a divariant field (dark grey shading) towards higher

473   temperatures is defined by pseudounivariant reactions with increasing X_{CO_2} in the fluid

474   composition (reaching up to 0.58). The described Di-Cal-Qz divariant field (dark grey;

475   Fig. 9D) partially overlaps with that of the tremolite-in conditions (light grey). At 300

476   MPa, the minimum temperature of this divariant field exactly fits that of the first Spl-in

477   reaction (430 ºC). Taking into account that the sample with the product assemblage of

478   this reaction (09A-11; Table 1) is located only about 200 m away from the first

14

479   appearance of diopside, it could be suggested that Di-in temperatures must be somewhat

480   higher than those of Spl-in (Fig. 9D).

481   Concerning *forsterite* stability, first P-T constraints are defined by equilibrium Cal + Tr

482   + Fo = F + Dol + Di, with a strongly changing Clapeyron slope and increasing X_{CO_2}

483   composition of fluid with decreasing pressure (number labels of pseudoinvariant points

484   along the curve; Fig. 9D). This curve constrains the stability of the assemblage

485   Cal+Tr+Fo, observed in most rocks close to the contact with the peridotites (Table 1), to

486   maximum values of P=500 MPa (at 300 ºC) and T=575 ºC (at 200 MPa). Within this

487   field, the minimum temperature is further constrained by the reaction F(H_2O) + Tr +

488   Dol = F(CO6.6) + Cal + Fo and its related divariant field with coexisting Tr, Dol, Cal

489   and Fo and variable fluid composition (thin, grey, dashed pseudounivariant curves; Fig.

490   9C). At 300 MPa this field ranges from 510 to 560 ºC and partially overlaps with the

491   field of minimum temperatures for Di-in (Fig. 9D). This is consistent with the mineral

492   assemblage observed in rocks (sample 09A-1; Table 1) sampled close to the contact

493   with the peridotites. Additional information can be obtained from both chlorite and

494   clinohumite. The former has been observed at least in one Fo and Spl-bearing sample

495   (09A-4; Table 1). Both reactions Clc + Chu = H_2O + Fo + Spl and F(H_2O) + Dol + Clc

496   = F(CO6.6) + Cal + Spl + Fo are chlorite-breakdown reactions producing coexisting Fo

497   and Spl, consistent with the observed rock assemblages (Table 1, Fig. 8G), and

498   indicating minimum temperature conditions ranging from 525 to 600 ºC at 300 MPa

499   (Fig. 9D). Equilibria indicating the observed coexistence of clinohumite and olivine

500   (Table 1, Fig. 8G) are the already mentioned Clc + Chu = H_2O + Fo + Spl, and also H_2O

501   + Dol + Fo = Cal + Chu + F(CO6.6). The stability of clinohumite increases towards

502   higher temperatures with increasing fluorine contents (Rice, 1980). Taking this into

503   account and attending to petrogenetic grids in Fig. 9 (A to C), it could be suggested that

504   clinohumite-bearing assemblages sampled close to the contact with the peridotites

505   might have reached maximum temperatures above 700 ºC.

506   In conclusion, phase relations modelling in the KCMASHC system suggest the

507   following approximate temperature ranges for the observed metamorphic zones of the

508   Nieves Unit contact aureole at the maximum, geologically consistent, pressure of 300

509   MPa (Fig. 9D): forsterite (> 510 ºC; probably around 700 °C), diopside (510–430 ºC),

510   tremolite (430–360 ºC), phlogopite (360–330 ºC).

511

512   *Microstructural Analysis*

513

Calcite (dolomite poor) marbles occur as strongly deformed rocks, displaying a variety of conspicuous microstructures in calcite grains (Fig. 8A, F): elongated grains defining a main foliation, curved twins, undulose extinction or mortar textures. The presence of subgrains and new recrystallized grains is common. Irregular (lobate) grain boundaries are also observed, pointing to the occurrence of grain boundary migration recrystallization. Dolomite-bearing marbles, on the contrary, commonly show a granoblastic texture (Fig. 8B).

Besides a general – though irregular – trend of calcite grain size increase from lower to higher metamorphic grades (Table 2), a variation of marble microstructures with distance from the peridotites (and hence temperature) is observed. The more distant calcite marbles (phlogopite and tremolite zones) tend to preserve deformation features, such as S-C-C' structures, recording top-to-NW shearing (Fig. 8J) that are not observed in the strongly recrystallized high-T marbles from the diopside and forsterite zones. A secondary foliation, oblique with respect to the main composite (S_0-S_1) foliation and produced by grain boundary alignment of recrystallized grains, is observed in samples from the tremolite through the forsterite zones (Fig. 8K, H). All of these samples show evidences of dynamic recrystallization resulting from a combination of both subgrain rotation and grain boundary migration. However, marbles from the forsterite zone also show straight grain boundaries and calcite triple junctions that are indicative of variable degrees of static recrystallization (Fig. 8H, I). The latter process appears to have led in some cases to a partially annealed microstructure characterized by generally equant, polygonal calcite grains (Fig. 8I). However, in one of the forsterite zone samples (09A-1), finer-grained zones of recrystallized grains cut across the annealed, coarse-grained microstructural domains.

For samples from the diopside through the tremolite to the phlogopite zones (09A-7 to 09A-20), plots of all orientation data are very similar to the plots of 1 point per grain. This similarity reflects the equigranular nature of these samples. In contrast, orientation data from 09A-1 and 09A-3 (forsterite zone) show a significant difference between plots of all of the collected EBSD data (not shown) and 1 point per grain data because of the heterogeneous grain size and of the grain size bias associated with the larger grains. For forsterite zone samples we therefore only considered the one point per grain data.

EBSD data from the different metamorphic zones of the carbonate footwall show a number of similarities. In most cases (Fig. 10), the pole figures record a crystallographic

547   preferred orientation (CPO) with alignment of the calcite c-axis with the pole to the main

548   S_0-S_1 foliation (Z). This is most strongly pronounced in samples 09A-7 and 09A-20. In

549   09A-1 the single c-axis cluster lies oblique to Z and is rotated in the direction of the

550   macroscopically established top-to-NW sense of shear. In other samples a subordinate

551   component of the c-axis orientation also lies oblique to the Z direction, again synthetic

552   with respect to the direction of the macroscopically established top-to-NW sense of

553   shear (Fig. 10). The angular relationships between the different concentrations on the c-

554   axis pole figure (usually < 40°) are not consistent with these clusters being associated

555   with twinning in the samples. Comparison with high strain calcite deformation

556   experiments indicates that such a pattern in c-axis orientations is consistent with high

557   shear strains (Barnhoorn *et al.*, 2004) with varying degrees of recrystallization (Pieri *et*

558   *al.*, 2001a). Only in 09A-19, there is a less defined and more symmetric distribution of

559   c-axes (Fig. 10). This sample retains evidence for non-coaxial deformation in the form

560   of well-developed S-C fabrics with C-planes dominated by fibrous tremolite (Fig. 8J)

561   but again there is no significant asymmetry within the (0001) pole figure data. Along

562   with all other samples no variation in +<a> and –<a> was observed (Fig. 10). This

563   contrasts observation on non-coaxially deformed calcite marble where asymmetry in

564   +<a> and –<a> are recorded during both low- and high-T non-coaxial flow (Bestmann,

565   2000; Pieri *et al.*, 2001b) and may reflect a strong component of high-temperature

566   recrystallization and steady-state flow (Pieri *et al.*, 2001a), and/or a component of static

567   recrystallization (Barnhoorn *et al.*, 2005) that is consistent with the microstructure

568   observed in high-T marbles. The data presented here indicate that orthorhombic

569   symmetry of fabrics with respect to the principal (S_0-S_1) foliation plane cannot be used

570   to provide unambiguous constraints on the coaxial or non-coaxial nature of the

571   deformation at high temperature and high-strain rate conditions. However, where

572   asymmetry in the c-axis distribution relative to the foliation is recorded, then this

573   appears to be synthetic with respect to the rotational component of the non-coaxial strain

574   inferred from S-C fabrics and mesoscopic kinematic criteria.

575   Misorientation axes for calcite marble data record differences between samples. In

576   samples where significant low-angle boundaries are preserved (09A-1, 09A-7, and 09A-

577   20) there is a dominant misorientation axis concentration associated with the 'a' <11-

578   20> direction. This direction is still common in 09A-3 and 09A-19 which also have a

579   significant number of measurements (n > 1000), but other orientations are also apparent

580   associated with 'f' <02-21>, 'm'<10-10>. In samples where low-angle boundaries are

581 rare (09A-12 and 09A-18), misorientation axes are concentrated close to the <0001>

582 direction. The observed variation in the misorientation axes reflects the operation of

583 different slip systems in calcite with <11-20> corresponding to the common high

584 temperature (> 400 °C) slip systems r{10-14}<-2021> and f{-1012}<10-11>, whilst

585 'm'<10-10> is consistent with deformation by {0001}<-12-10> slip (De Bresser &

586 Spiers, 1997). In some of the samples (09A-1, 09A-18, 09A-19), misorientation axis

587 evidence for the operation of f{-1012}<0-2-21> is consistent with low temperature

588 deformation (De Bresser & Spiers, 1997) and this may reflect a lower temperature

589 overprint that may also have been responsible for the localized zones of reduced grain

590 size observed in the highest temperature sample (09A-1). Such a lower temperature

591 overprint could have led to a finer grain size of samples 09A-18 and 09A-19 (Table 2) as

592 a result of dynamic recrystallization dominated by subgrain rotation.

593 EBSD analysis has also been performed on a dolomite marble sample (09A-15) from the

594 tremolite zone (Fig. 10). The CPO in this sample, although clearly marked by the *c*-axis

595 distribution, is weaker than that generally recorded by calcite marbles and also shows

596 different misorientation axes from most of the calcite-dominated samples. This contrast

597 suggests the operation of a different set of slip systems, albeit including the common

598 (0001)<2-1-10> slip system (Wenk, 1985). The asymmetry of the *c*-axis distribution

599 with respect to the S_0-S_1 foliation plane is synthetic with the rotational component of

600 simple shear.

601

602 **Discussion**

603

604 The Nieves Unit is characterized by significant changes in styles and intensity of

605 deformation, with strongly heterogeneous structural development being closely linked

606 to an inverse metamorphic gradient that is controlled by distance from the hanging-wall

607 Ronda peridotites. This variation is expressed by different deformation features at

608 different structural position within the main regional syncline involving the studied unit.

609 This major structure may be interpreted as a result of footwall deformation within the

610 general framework of bulk non-coaxial strain associated with top-to-the-NW shearing.

611 The normal limb and core region of the regional syncline are characterized by the

612 occurrence of NE-SW trending, meso-scale parasitic folds, whereas in the metamorphic

613 aureole of the inverted limb, isoclinal, intrafolial folds occur. The hinge lines of these

614 latter folds tend to lie close to the stretching direction and are interpreted to have been

615   strongly rotated towards this direction during progressive deformation. The occurrence

616   of a well-developed foliation and associated mineral lineation, together with isoclinal

617   intrafolial folds in the marbles, indicates that these rocks record a significant amount of

618   finite strain, although shear-sense indicators are rarely observed in the field. These

619   features are consistent with progressive strain localisation in the inverted limb of the

620   major syncline. As a matter of fact, the overturned fold limb of the regional syncline

621   appears to constitute a SE dipping, thermally softened carbonate rock panel that

622   underwent intense deformation and metamorphism associated with the emplacement of

623   the Ronda peridotites.

624   Both field and thin section evidence indicate a limited syn-kinematic growth of silicate

625   minerals (locally testified by the preferred orientation of early generations of tremolite

626   and phlogopite, Fig. 8B, C). However, most silicate grain growth appears to have

627   occurred statically, following the development of the composite planar fabric (S_0-S_1) and

628   mainly post-dating the main internal deformation in the marbles (Fig. 8F). This suggests

629   that high-T footwall deformation associated with peridotite emplacement occurred at a

630   much faster rate with respect to heat diffusion, which continued well after peridotite

631   emplacement. Estimated maximum temperatures in the metamorphic aureole range from

632   values in excess of 510 °C (probably around 700 ºC) close to the contact with the

633   peridotites to < 350º C in the more distant, less thermally overprinted rocks, for the

634   inferred maximum pressure of 300 MPa. Higher-pressure values are rather unlikely due

635   to the complete lack of regional metamorphism of the Nieves Unit away from the

636   aureole at the contact with the peridotites. Had the overlying tectonic units been

637   significantly thicker than *c.* 10 km, the whole footwall (Nieves Unit) would have

638   undergone greenschist facies metamorphism (which is not observed). Therefore, it can

639   be realistically considered that the present-day geometry (taking erosion into account;

640   see geological sections in Figs. 2 and 4) is representative of the original (Miocene)

641   tectonic load on top of the Nieves Unit (Mazzoli and Martín-Algarra, 2011, their fig. 6).

642   This is consistent with syn-orogenic extension (Argles *et al.*, 1999) of the Jubrique

643   crustal rocks originally overlying the Ronda peridotites. The crustal succession was

644   being tectonically attenuated as the Ronda peridotites were exhuming, so that the Los

645   Reales Unit was already significantly thinned (essentially to its present-day thickness)

646   when final emplacement on top of the Nieves Unit occurred (Mazzoli and Martín-

647   Algarra, 2011, their fig. 6). A maximum temperature of *c.* 700 ºC is consistent with a

648   peridotite emplacement postdating the latest metamorphic event recorded within the

649   ultramafic body, which occurred at conditions of P = 1 GPa and T = 800–900 ºC

650   (obtained for the aluminous mafic rocks alternating with peridotites and sampled near

651   the marbles section studied in this work; Morishita *et al.*, 2001). Such high P-T

652   conditions are interpreted as corresponding to a sustained synkinematic decompression

653   event (Garrido *et al.*, 2011) that took place well before the final peridotite emplacement

654   described here. The general – though irregular – trend of increasing calcite grain size

655   with metamorphic grade observed in the studied footwall carbonates confirms the

656   fundamental control exerted by temperature on calcite recrystallization processes,

657   although the variable amount of secondary phases (Table 2) present in our impure

658   marbles is likely to have played a major role, resulting in complex calcite grain growth

659   patterns (Ebert *et al.* 2007a, b). As a matter of fact, the irregular distribution of calcite

660   grain size shown in Table 2 is likely to result from a complex interplay of different

661   factors including original grain size, amount of impurities, differential stress and thermal

662   regime controlling the dominant deformation mechanisms. The microstructure of calcite

663   marbles, characterized by the occurrence of undulose extinction, subgrains and new

664   recrystallized grains, is indicative of deformation by slip and climb of dislocations

665   (dislocation creep). The analysed calcite marbles show a well-developed CPO at all

666   metamorphic grades, for a wide range of grain sizes and different – though generally

667   moderate – amounts of secondary phases (Table 2). This is consistent with the

668   dominance of dislocation creep at all temperatures characterizing the studied

669   metamorphic aureole. Both microstructures and CPOs show no indications for a switch

670   in deformation mechanisms from dislocation to diffusion creep or to grain-size sensitive

671   mechanisms even at the highest temperature conditions (forsterite zone). Such an

672   interpretation is in general agreement with the experimental findings of Barnhoorn *et al.*

673   (2004). The associated occurrence of grain boundary migration recrystallization

674   (testified by the common presence of lobate grain boundaries) is also consistent with

675   large strains and relatively high temperatures in calcite marbles (e.g. Barnhoorn *et al.*,

676   2004). The generally symmetric CPO patterns suggest high-T plastic deformation,

677   dynamic recrystallization and steady-state flow, probably followed by variable degrees

678   of static recrystallization that may have weakened the CPO (Barnhoorn *et al.*, 2005). In

679   the highest-grade forsterite zone, the presence of a carbonate melt may have also

680   contributed to a weakening of the CPO. In fact, the highest temperatures close to the

681   peridotite contact were potentially higher than the eutectic of marbles (e.g. Boettcher &

682   Wyllie, 1969). Therefore, at least part of the microfabric suggesting static

683   recrystallization associated with thermal diffusion – and the related weakened CPO –

684   could actually have resulted from grain growth following partial melting. Melt fractions

685   – no matter how small – may additionally have controlled deformation processes in the

686   highest-grade marbles by enhancing grain boundary sliding and migration, as well as

687   rotation.

688   The lack of strong asymmetries in +<a> and –<a> distribution in marbles from all

689   metamorphic zones, suggesting the occurrence of a coaxial strain component (Bestmann,

690   2000; Pieri *et al.*, 2001b), is consistent with bulk deformation by general shear and/or

691   partitioning of the deformation into zones dominated by either non-coaxial or coaxial

692   strain. Furthermore, the occurrence of minor zones of strain localization and grain size

693   reduction (associated with subgrain rotation recrystallization) overprinting previously

694   annealed calcite textures (e.g. sample 09A-1) suggests a complex interplay between

695   thermal effects and deformation events. Notwithstanding this, a final static overprint

696   appears to be dominant in carbonates throughout the metamorphic aureole. This is

697   consistent with the interpretation that most of the petrologic features characterizing the

698   metamorphosed rock panel (inverted limb of the Nieves Unit regional syncline) located

699   immediately in the footwall to the Ronda peridotites were acquired following the

700   emplacement of the mantle extrusion wedge. Continued heat transfer from the

701   peridotites to the footwall succession led to post-kinematic silicate growth and locally –

702   especially in the highest-T forsterite zone – to partial annealing of calcite marble

703   textures. Following substantial cooling (Esteban *et al.*, 2004), horizontal crustal

704   shortening was resumed – as testified by meso-scale refolding affecting both

705   metamorphic and non-metamorphic parts of the Nieves Unit – probably in relation with

706   NW-ward thrusting of the 'Frontal Units' of the Internal Domain over the Flysch and

707   External Domains (Figs. 2 to 4). Such a late, SE-NW oriented shortening of the Nieves

708   Unit occurred at much lower temperatures, as testified by the lack of metamorphic

709   mineral assemblages along the axial plane foliation (S_2) associated with the second-

710   phase, NE-SW trending mesoscopic folds.

711

712   **Conclusions**

713

714   Structural analysis of the dominantly carbonate Nieves Unit located in the footwall to

715   the Ronda mantle extrusion wedge unravelled strongly heterogeneous deformation

716   accompanying a dramatic metamorphic gradient. Marbles within a several hundreds of

717 meters-thick metamorphic aureole record high strain in the form of a well-developed
718 foliation and associated mineral lineation, accompanied by widespread isoclinal
719 intrafolial folds. Field evidence for non-coaxial strain, although clearly consistent with
720 top-to-the-NW shearing, appears to be subordinate. On the other hand, the
721 microstructures of calcite marbles – particularly the lower-T phlogopite-tremolite ones –
722 effectively record top-to-the-NW kinematics, whereas their CPOs suggest the possible
723 occurrence of a coaxial strain component and therefore of an overall deformation by
724 general shear. Where asymmetry in the c-axis distribution is observed with respect to
725 the main composite foliation (sub-parallel to the shear plane, due to very likely high-
726 strain values), this appears to be synthetic with respect to the rotational component of
727 the non-coaxial strain inferred from mesoscopic kinematic criteria, this being consistent
728 with large shear strains and high-T conditions. Our results indicate that large strains and
729 attenuation of a thermally softened carbonate rock panel, probably under a general shear
730 deformation regime, dominantly controlled structural development and marble textural
731 fabrics in the metamorphic aureole in the footwall to the Ronda peridotites.
732 Petrological characteristics and mineral assemblages within the zoned metamorphic
733 aureole appear to have mostly developed after the final emplacement of the Ronda
734 mantle extrusion wedge. Post-kinematic, continued heat transfer from the hanging-wall
735 peridotites led to static growth of silicate assemblages and local partial annealing of
736 calcite marble textures, especially in the highest-T zone in the immediate footwall.
737 Following substantial cooling, renewed crustal shortening affected the Nieves Unit,
738 probably in relation with foreland-directed thrusting.
739

751

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973

Figure Captions

975

976 **Fig. 1.** Tectonic sketch map of the Betic Cordillera, showing the Ronda peridotites
977 outcrop area (boxed SW-most sector).

978

979 **Fig. 2.** Tectonic sketch map of the western Betic Cordillera and regional geological
980 cross-section X-X' (after Mazzoli and Martín-Algarra, 2011, modified). Box shows
981 location of area mapped in detail (Fig. 3).

982

983 **Fig. 3.** Geological map of field study area (showing sample location) and trace of cross-
984 section reproduced in Fig. 4.

985

986 **Fig. 4.** Geological cross-section (roughly parallel to the Ronda-San Pedro de Alcantara
987 road) through the leading edge of the Ronda peridotites at Sierra Bermeja (after Mazzoli
988 and Martín-Algarra, 2011, modified), showing projected location of the studied samples.
989 Zoning of the metamorphic aureole (shaded) in the Nieves Unit is shown by means of
990 the following index minerals: talc (tc), phlogopite (phl), tremolite (tr), diopside (di), and
991 forsterite (fo). Key for stratigraphic units: (1) Norian dolostones and (1') dolomite
992 marbles; (2) Rhaetian marls and (2') calcschists; (3) Rhaetian limestones and (3') calcite
993 marbles; (4) Lower Jurassic cherty limestones and marls and (4') marbles including
994 quartz lenses and calcschists; (5) Middle Jurassic-Paleogene (mainly marly) condensed
995 succession and (5´) calc-schists; (6) Nava Breccia Fm.

996

997 **Fig. 5.** Field examples of structural features in the Nieves Unit in the footwall to the
998 Ronda peridotites. (A) View of the tectonic contact with the overlying Ronda peridotites
999 (Cerro Corona area). (B) View of the hinge zone of the regional syncline, including
1000 second-order overturned folds (Fuenfría area). (C) Composite planar fabric (S_0-S_1) in
1001 dolomite marble (km 19.5 of the A-473 road from Ronda to San Pedro de Alcantara). (D)
1002 Close-up of the boxed area in (C), showing minor folds in the hinge zone of a larger
1003 syncline. (E) Composite planar fabric (S_0-S_1; horizontal) in metamorphosed Jurassic
1004 cherty limestones (Torrecilla area). Note symmetric pressure shadows around framboidal
1005 pyrite grains (arrowed). (F) Composite planar fabric (S_0-S_1) displaced and overgrown by
1006 phlogopite-tremolite clot in metamorphosed Norian levels (dolomite marbles of the
1007 Cerro Alcojona area).

1008

1009   **Fig. 6.** Outcrop examples of structural features in the Nieves Unit. (A) Mineral lineation

1010   (L_1) defined by elongated tremolite (arrowed) and dolomite crystals lying on the main

1011   foliation in metamorphosed Norian levels (dolomite marbles at km 19.5 of the A-473

1012   road from Ronda to San Pedro de Alcantara). (B) Early intrafolial folds in dolomite

1013   marbles (Cerro Alcojona area). (C) Late (post-metamorphic) folds affecting the main

1014   composite planar fabric (S_0-S_1) in dolomite marbles (km 19.5 of the A-473 road from

1015   Ronda to San Pedro de Alcantara). (D) Asymmetric (*z*-shaped looking towards SW)

1016   parasitic folds affecting bedding (S_0) in the normal limb of the regional syncline

1017   (Jurassic cherty limestones of the Torrecilla area). (E) Late kink bands affecting the

1018   main composite planar fabric (S_0-S_1) along the San Pedro de Alcantara road.

1019

1020   **Fig. 7.** Orientation data (lower hemisphere, equal-area projections) for structures in the

1021   Nieves Unit. (A) Poles to bedding (S_0)/composite planar fabric (S_0-S_1; metamorphic

1022   aureole); pole to the best-fit great circle (or π-girdle, plunging 19° toward 54°N)

1023   provides a statistical axis consistent with NE-SW trending folding. (B) Mineral lineation

1024   (L_1) in silicate-bearing marbles. (C) Hinges of isoclinal, intrafolial folds. (D) Poles to

1025   axial surfaces of isoclinal, intrafolial folds. (E) Hinge lines of late, open to tight folds

1026   within metamorphic aureole. (F) Poles to axial surfaces of late, open to tight folds within

1027   metamorphic aureole.

1028

1029   **Fig. 8.** Thin section microphotographs (crossed polars) showing representative

1030   petrological and microstructural features of the studied marble samples (mineral name

1031   abbreviations after Whitney and Evans, 2010). (A) Intensely deformed, high-T calcite

1032   marble (sample 09A-02; forsterite zone). (B) Granoblastic, mainly dolomite marble

1033   (with scarce, "dusty" calcite grains) with a thin folded layer of phlogopite (sample 09A-

1034   15; tremolite zone). (C) Palmed aggregate of amphibole subparallel to the main rock

1035   foliation as indicated by the white dashed line (sample 09A-16; tremolite zone). (D)

1036   Xenoblastic spinel grains associated with calcite and large phlogopite flakes (sample

1037   09A-11; tremolite zone). (E) Typical mineral assemblage of clinopyroxene, calcite and

1038   quartz-bearing rock (sample 09A-08; diopside zone). (F) Hypidioblastic clinopyroxene

1039   and amphibole grains overgrowing strongly deformed calcite grains (sample 09A-01;

1040   forsterite zone). (G) Typical high-grade mineral assemblage of olivine-clinohumite-

1041   spinel marble close to the contact with peridotite (sample 09A-24; forsterite zone). (H)

1042 High-T calcite marble (sample 09A-3; forsterite zone) showing lobate grain boundaries

1043 (red arrows), straight grain boundaries and triple junctions (yellow arrows) and

1044 secondary foliation (sf) defined by recrystallized grains oblique to the main composite

1045 foliation (horizontal; NW is to the right of the picture). (I) High-T marble with equant

1046 calcite grains showing both lobate and straight boundaries (sample 09A-25; forsterite

1047 zone). (J) Lower-T calcite marble (sample 09A-19; tremolite zone) displaying S-C-C'

1048 fabrics (NW is to the right of the picture). (K) Lower-T calcite marble (sample 09A-20;

1049 phlogopite zone) showing secondary foliation (sf) defined by recrystallized grains

1050 oblique to the main composite foliation (horizontal; NW is to the right of the picture).

1051

1052 **Fig. 9.** Calculated P-T projections for the K_2O-CaO-MgO-Al_2O_3-SiO_2-H_2O-CO_2 system

1053 with the following minerals (abbreviations after Whitney and Evans, 2010): calcite

1054 (Cal), clinochlore (Clc), clinohumite (Chu), diopside (Di), dolomite (Dol), forsterite

1055 (Fo), microcline (Mc), phlogopite (Phl), quartz (Qz), spinel (Spl) and tremolite (Tr).

1056 Abbreviation for the fluid solution is F. All curves in (A) correspond to univariant

1057 equilibria. Field b-d in (A) is enlarged in diagrams (B), (C) and (D). Note that minimum

1058 pressure in these cases widens towards 50 MPa. Thick and thin dashed curves

1059 correspond to univariant and pseudounivariant equilibria, respectively (see text for

1060 details). Big circles and small filled circles along univariant curves correspond to

1061 invariant and pseudoinvariant points, respectively. Different grey areas represent

1062 divariant fields in which pseudounivariant equilibria are stable with the same solid

1063 phases and changing fluid composition (X_{CO2} always increases with temperature). In

1064 (D), thick dashed line at a constant pressure of 300 MPa acts as reference for obtaining

1065 the temperature ranges for the relevant phases observed in the studied rocks (upper part

1066 of the diagram).

1067

1068 **Fig. 10.** CPO data for calcite marble samples from: (i) forsterite zone (09A-1, 09A-3);

1069 (ii) diopside zone (09A-7); (iii) tremolite zone (09A-18, 09A-19); and (iv) phlogopite

1070 zone (09A-20). Dolomite marble sample (09A-15) is from the tremolite zone. All

1071 projections are lower hemisphere and equal area. Contour lines are at intervals of 0.5

1072 Multiples of Uniform Density (MUD) and the maximum MUD is given to two

1073 significant figures. For forsterite zone samples, heterogeneous grain size gives rise to a

1074 grain-size dependent bias in standard pole figures. For 09A-1 and 09A-3 data are

1075 therefore shown as 1 point per grain. For other samples, pole figure data (columns 1-5)

1076   represents all of the collected EBSD data. For comparison, 1 point per grain data for all

1077   samples is given in column 6, along with the MUD and the total number of analysed

1078   grains. Misorientation axes data are calculated for minimum angular misorientations

1079   between adjacent EBSD analyses of 2-10° and are given with MUD and number of

1080   misorientation axes. Orientation of oblique foliation is indicated in column 1 for

1081   samples (09A-3, 19, 20) whose microstructure is shown in Fig. 8.

Table 1: Mineral assemblages of the studied marbles. Names abbreviations after Whitney and Evans (2010)

Samp. Area*	Met. Zone**	Sample	Cal	Dol	Qz	Mc	Chl	Phl	Amp	Cpx	Scp	Ol	Chu	Spl	Sph	Mag
CC	forsterite	09A-1	x		x			x	x	x						
CC		09A-2	x					x				x				
CC		09A-3	x					x				x				x
CC		09A-4	x	x			x	x	x			x		x		x
CA		09A-21	x	x				x	x			x	x	x		x
CA		09A-22	x	x				x	x			x		x		x
CA		09A-23	x	x				x	x							x
CA		09A-24	x	x				x	x			x	x	x		x
CA		09A-25	x					x	x							x
CC	diopside	09A-5	x		x			x		x						x
CC		09A-6	x		x			x	x	x	x				x	x
CC		09A-7	x		x			x								x
CC		09A-8	x		x	x		x		x	x				x	x
CC		09A-9	x		x			x	x	x					x	x
CC	tremolite	09A-10	x	x				x								x
CC		09A-11	x	x				x	x					x		x
CC		09A-12	x	x	x			x	x							x
CC		09A-13	x	x				x								x
CC		09A-14	x	x				x								x
CC		09A-15	x	x				x								x
CC		09A-16	x	x			x	x	x							x
CC		09A-17	x	x				x	x							x
CC		09A-18	x	x				x	x							x
CC		09A-19	x					x								x
CC	**phlogopite**	09A-20	x		x			x								x

* Sampling area. CC: Cerro de los Cascajares; CA: Cerro Alcojona

** Metamorphic zone, after Martín-Algarra (1987)

Table 2: Modal analysis and carbonate grain size analysis of the studied marbles.

Sampling Area*	Metamorphic Zone**	Sample	Impurities	Dmin	Dmax	Dmedian
CC	forsterite	09A-1	6.5 n = 4700	9	1329	178 (±153) n = 713
CC	forsterite	09A-3	6.8 n = 6000	68	2083	230 (±136) n = 512
CC	diopside	09A-7	12.0 n = 8000	35	429	111 (±52) n = 705
CC	tremolite	09A-12	10.1 n = 10000	51	1586	354 (±244) n = 561
CC	tremolite	09A-15	7.2 n = 5000	55	666	143 (±59) n = 1218
CC	tremolite	09A-18	17.8 n = 8000	22	394	70 (±31) n = 740
CC	tremolite	09A-19	16.6 n = 10000	12	96	30 (±12) n = 619
CC	phlogopite	09A-20	11.0 n = 10000	26	437	112 (±55) n = 614

* CC: Cerro de los Cascajares.
** After Martín-Algarra (1987).
Impurities = abundance of the non-carbonate minerals (in vol. %; n = number of counted points); Dmin = minimum size (in μm) of the carbonate minerals; Dmax = maximum size (in μm) of the carbonate minerals; Dmedian = median size (in μm) of the carbonate minerals (standard deviation values reported in brackets; n = number of measured grains).

Foreland

Late Neogene-Quaternary

External Domain

Flysch Domain

Internal/External Domain Boundary

Frontal Units

Maláguide Complex

Alpujárride Complex

Nevado-Filábride Complex

INTERNAL DOMAIN

IBERIAN MASSIF

Guadalquivir Basin

Córdoba

Jaén

Granada

Ronda

Málaga

Almería

Murcia

Alicante

N

0   50   100 km

Iberia

Betic Cordillera

W. Mediterranean Sea

Africa

Rif

Tell

Alps

Apennines

Pyrenees

Iberian Range

Calabrian-Peloritanian Arc

45°

40°

35°

5°   0°   5°   10°   15°

Map labels (top map):

5°10'W 5°00'W 4°50'W 4°40'W

36°50'N

X

Ronda Yunquera

Nieves

Torrecilla

36°40'N

Fig. 3

Guadaiza Blanca

Ojén

Los Reales

36°30'N

Marbella

Estepona

X' N

0 5 10 15km

Cross section labels:

X
NW
Ronda

2000m
0m

Oligocene-Aquitanian of the
Malaguide Complex and Frontal
Units of the Internal Domain

Frontal
Units
(eroded)

Sierra de
las Nieves

Viñuela Group (eroded)

Backthrusted Campo de Gibraltar
Complex onto the Internal Domain (eroded)

X'
SE

Malaguide
Complex

Torrecilla

Sierra
Blanca

Sierra
Alpujata

(eroded)

0m

Nieves Unit
(yellow: Nava Breccia Fm)

Campo de Gibraltar
Complex
(Flysch Domain)

Penibetic
(External Domain)

Yunquera Unit Guadaiza Unit Ojen Unit

Blanca-type Units

Ronda peridotites Crustal succession

Los Reales Nappe

Alpujarride Complex

Legend:

Internal Domain

Postorogenic Deposits
(Late Miocene-Holocene)

External Domain

Flysch Domain

Torrecilla Syncline

Top of the Ronda peridotite

Thrusts Footwall Hangingwall

Dextral strike-slip

Malaguide Complex

Cover
(Triassic-Aquitanian)

Basement
(Paleozoic, and older?)

Frontal Units

Nava Breccia Fm
(Lower Miocene)

Carbonate rocks
(Triassic-Aquitanian)

Viñuela Group

Late orogenic
deposits (Burdigalian)

Metapelites Los Reales
Nappe

Peridotites

Marbles Blanca-type
Units

Metapelites

Alpujarride Complex

Geologic map legend:

Quaternary — **Penibetic** — **Campo de Gibraltar Complex**

Nieves Unit
- Lower Miocene Nava Breccia
- Middle Jurassic-Paleogene condensed succession
- Lower Jurassic cherty limestones
- Rhaetian limestones and calcitic marbles
- Rhaetian marls and calcschists
- Norian dolostones
- Norian dolomitic marbles with tremolite
- Impure dolomitic marbles very rich in calc-silicates
- Dolomitic marbles with high-grade calc-silicates

Los Reales Nappe
- Crustal succession (Jubrique Unit)
- Subcrustal detachment
- Ronda peridotites (Bermeja Unit)

Yunquera Unit (Blanca-type)
- Marbles
- Gneisses

09A-1 to 25: studied samples

A ——— B Location of the cross section

Unconformable base of the Nava Breccia Fm.: — — — normal, — — — reversed

- Thrust
- Normal fault
- Undifferentiated fault
- Trace of bedding
- Horizontal bedding
- 80 Normal bedding
- Vertical bedding
- 80 Reversed bedding
- Subhorizontal reversed bedding
- Water courses
- Roads and paths

Map labels: Igualeja, Cascajares, Fuenfria, Cerro Alcojona, Cerro Abanto, Cerro Alcojona

Sample labels: 09A-20, -19, 09A-18, -15, 09A-14, -13, -12, -11, 09A-10, -7, -3, 09A-1, 09A-21 to 09A25

1km

Nieves Unit

Jurassic-Paleogene / Rhaetian / Norian

Lower Miocene Nava Breccia (6)

Condensed limestones, marls and cherts (5), and calcschists with quatz nodules and talc (5')

Cherty limestones and marls (4), and calcitic marbles with quatz nodules and talc (4')

Limestones (3) and calcitic marbles with talc (3')

Phlogopite-in

Marls (2) and calcschists with talc and phlogopite (2')

Dolostones (1) and fine-grained dolomitic marbles (1' with talc (tc)

Dolomitic marbles with phlogopite (phl)

Tremolite-in

Dolomitic marbles (1') with tremolite (tr)

Diopside-in

Impure dolomitic marbles (1') with diopside (di)

Forsterite-in

Dolomitic marbles (1') with forsterite (fo)

Los Reales Nappe

Crustal succession

Ronda peridotite

Yunquera Unit (Blanca-type)

Marbles

Gneisses

Campo de Gibraltar Complex

Penibetic

Road from Ronda to San Pedro de Alcantara, with location of the studied samples (projected)

Alpujarride Complex

A — WNW / ESE — Nieves Unit — Ronda peridotites

B — SE / NW

C — NW / SE — $(S_0 + S_1)$

D

E

F — 10 cm — www.sgu.se

A — NW ... SE (with coin for scale, red arrow and white arrow)
B — SE (www.sgu.se scale card)
C — NW ... SE ($S_0 + S_1$)
D — SE ... NW, S_0
E — SE ... NW

A N
n = 339

B N
n = 26

C N
n = 29

D N
n = 29

E N
n = 40

F N
n = 40

Pole Figures

Misorientation Axes

Z / Y / X orientation key

c {0001} -a {11$\bar{2}$0} +a {$\bar{1}\bar{1}$20} f {01$\bar{1}$2} r {10$\bar{1}$4} c {0001}

2-5° 5-10°

09A-1
mud=5.72
n=2009
5.72
n=2009
0001
01$\bar{1}$0
10$\bar{1}$0
5.72
n=6905

09A-3
mud=2.78
n=498
2.78
n=498
2.14
n=2409

09A-7
mud=3.67
3.51
n=5076
1.56
n=22562

09A-12
mud=6.97
4.90
n=642
5.33
n=464

09A-18
mud=3.87
2.64
n=3399
2.29
n=479

09A-19
mud=2.34
2.10
n=2977
1.79
n=1316

09A-20
mud=3.97
3.68
n=3760
1.69
n=20965

09A-15
mud=2.51
2.64
n=1114
1.94
n=1322