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

S. MAZZOLI1*, A. MARTÍN-ALGARRA2, S. M. REDDY3, V. LÓPEZ SÁNCHEZ-VIZCAÍNO4, L. FEDELE1 & A. NOVIELLO1

1 Dipartimento Scienze della Terra, Università di Napoli Federico II, Largo San Marcellino 10, 80138 Naples, Italy
2 Departamento de Estratigrafía y Paleontología, Universidad de Granada, 18071 Granada, Spain
3 The Institute for Geoscience Research & ARC COE for Core to Crust Fluid Systems, Department of Applied Geology, Curtin University of Technology, GPO Box U1987, Perth WA 6845, Australia
4 Departamento de Geología (Unidad Asociada al IACT-CSIC, Granada), Escuela Politécnica Superior de Linares, Universidad de Jaén, 23700 Linares, Spain

*Corresponding author (e-mail: stefano.mazzoli@unina.it)

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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.
textures, particularly in the highest-temperature zone of the thermally softened footwall carbonates. Following substantial cooling, renewed crustal shortening affected the whole Nieves Unit, resulting in widespread development of NE-trending meso-scale folds.

In the western Betic Cordillera (Fig. 1), high-temperature tectonic emplacement of the Ronda mantle peridotites (Fig. 2) on top of crustal units (e.g. Lundeen, 1978) is marked by the occurrence of a ‘metamorphic sole’ in the footwall rocks (e.g. Tubía et al., 1997). In the southeastern, inner peridotite massif (Sierra Alpujata), a high pressure-temperature (P-T) shear zone records orogen-parallel (Tubía & Cuevas, 1986; Orozco and Alonso-Chaves, 2012), sinistral shear between the peridotites and the Alpine eclogites of the Ojen Unit. In contrast, in the outer peridotite massif of Sierra Bermeja located to the NW (Fig. 2), a high-T dynamothermal metamorphic aureole records orogen-perpendicular, foreland-ward thrusting of the peridotites on top of both low-P metamorphic rocks of the Guadaiza Unit ( Cuevas et al., 2006; Esteban et al., 2008) and sedimentary successions of the Nieves Unit (Mazzoli & Martín-Algarra, 2011). These features have been interpreted in terms of deformation partitioning associated with oblique convergence during continental subduction and subsequent exhumation involving the coeval activity of kinematically linked systems of reverse, strike-slip and ‘normal-sense’ shear zones (Mazzoli & Martín-Algarra, 2011). In this context, orogen-parallel wrenching would have dominated the deeper, high-pressure portions of the continental subduction system, whereas orogen-perpendicular thrusting would have characterized the shallower, outer parts. Top-to-the-hinterland shear along the contact between the Ronda peridotites and overlying crustal rocks is consistent with extrusion of subcontinental mantle rocks of the overriding plate of the continental subduction system. Rapid exhumation of this extruding mantle wedge would have been further aided by strike-parallel extension and thinning of the crustal rocks overlying the peridotites (e.g. Tubía, 1994; Balanyà et al., 1997; Argles et al., 1999; Platt et al., 2003).

The crustal units lying in the footwall to the northwestern (i.e. outer) peridotite massif of Sierra Bermeja (Fig. 2) provide a unique opportunity to analyse strongly heterogeneous deformation associated with the ‘hot’ emplacement of a mantle wedge at shallow crustal levels. In the interior of the ultramafic massif, the contact at the base of the Ronda
peridotites is exposed in the Guadaiza River tectonic window (Fig. 2). There, the contact is marked by a dynamothermal metamorphic aureole formed at the expense of footwall rocks (Cuevas et al., 2006; Esteban et al., 2008). High-T metamorphism is well constrained by U/Pb SHRIMP dating of zircons from thin, syn-kinematic leucogranite sheets that emanated from the aureole and intruded the overlying peridotites, which provided ages of 18.8 ± 4.9 Ma (Sánchez-Rodríguez & Gebauer, 2000) and 21.8 ± 0.5 Ma (Esteban et al., 2007). The heat required for partial melting of footwall crustal rocks was provided by the fast emplacement of a c. 7 km thick, hot peridotite slab on top of the Guadaiza Unit, within the general framework of NW-directed overthrusting. This metamorphic aureole represents a low-viscosity, high-T (> 650 °C) shear zone characterized by a marked strain gradient towards the contact with the overlying ultramafic rocks (Esteban et al., 2008). On the other hand, the structure of footwall carbonate units occurring to the NW of the present leading edge of the Ronda mantle extrusion wedge has received much less consideration. The Nieves Unit rocks exposed in this area are characterized by a strong inverse metamorphic gradient (Martin-Algarra, 1987; Mazzoli & Martin-Algarra, 2011). Within a c. 1.5 km thick zone, non-metamorphic limestones/dolostones pass to high-grade marbles in contact with the overlying peridotites. The metamorphic gradient is marked by silicate minerals occurring in the marbles, defining a series of zones varying in thickness from tens to several hundred metres and characterized by the occurrence of (moving towards the peridotites; Figs. 3 and 4): (i) talc, (ii) phlogopite, (iii) tremolite, (iv) diopside, and (v) forsterite. Extreme deformation gradients and changes in structural styles accompany the metamorphic gradient and are the focus of this study, along with the texture of silicate-bearing calcite marbles. Our results provide new insights into the metamorphic and deformation processes taking place in carbonate units adjacent to the Ronda high-T mantle wedge emplaced within the Betic orogen. In this (Behrmann, 1983) as well as in similar carbonate-rich orogenic belts such as the Alps (Pfiffner, 1982; Heitzmann, 1987; Burkhard, 1993) or the Apennines (Iannace et al., 2005, 2007; Vitale & Mazzoli, 2005, 2009), marbles may record important deformation processes in the upper crust. Although naturally and experimentally deformed carbonates have been the subject of intense laboratory investigations aimed at a better understanding of the localization of plastic/superplastic flow in these rocks (e.g. Schmid et al., 1977, 1980, 1987; Rutter et al., 1994; Busch & Van der Pluijm, 1995; Badertscher & Burkhard, 2000; Bestmann et al., 2000; Pieri et al., 2001a, 2001b; Badertscher et al., 2002; Renner et al., 2002; Ulrich
et al., 2002; Barnhoorn et al., 2004, 2005; Ebert et al. 2007a, b; Vitale et al., 2007), the peculiar tectonic setting of the Nieves Unit allowed us analysing in detail the variation of mesoscopic and microscopic structural features and of textural fabrics of impure calcite/dolomite marbles for a range of temperature regimes – spanning over several hundreds of degrees Celsius – within the km-thick dynamothermal metamorphic aureole in the footwall to the Ronda peridotite mantle extrusion wedge.

**Geological Setting**

The Betic Cordillera, representing the westernmost part of the Alpine orogen in the Mediterranean region, resulted from the convergence between the African and Iberian plates since Late Cretaceous times (Andrieux et al., 1970; Durand-Delga & Fontboté, 1980; Martín Algarra & Vera, 2004). The Betic Cordillera, similarly to further Alpine mountain belts in the Mediterranean area, is traditionally subdivided into Internal (i.e. inner) and External (i.e. outer) Domains, with the Flysch Domain representing the suture between the two (Fig. 1). The area of the present study forms part of the Internal Domain of the western Betic Cordillera (Fig. 2). This domain is characterized by the occurrence of low-angle extensional shear zones of regional extent, reworking the pre-existing nappe structure (Balanyà et al., 1993; Argles et al., 1999). These extensional contacts bound major allochthonous units characterized by varying tectonic and metamorphic pre-Alpine and Alpine evolutions (Sánchez-Navas et al., 2012, and references therein). The lower allochthon (Alpujarride Complex) records variable Alpine P-T-t paths, generally characterized by high-pressure metamorphism followed by decompression and final cooling at 18–16 Ma (Sánchez Rodríguez & Gebauer, 2000). The upper allochthon (Malagueide Complex) generally lacks Alpine metamorphism (Martín-Algarra et al., 2009). A general lack of Alpine metamorphism characterizes also the so-called ‘Frontal Units’ of the Internal Domain, one of which – i.e. the Nieves Unit – is the subject of the present study; these units are thrust over the Flysch and External Domains (Fig. 2). A series of unconformable marine basin successions (Viñuela Group; Martín-Algarra, 1987) sit on top of the collapsed nappe pile. Deposition of these Burdigalian clastic sediments predated backthrusting of the Flysch Domain units above the Internal Domain (Fig. 2).

The Ronda peridotites form the lower portion of the traditionally recognized highest Alpujarride nappe in the area. Above the mantle rocks, the highly attenuated Jubrique
crustal section – including granulitic gneisses, migmatites, high-to-low grade pre-Mesozoic metapelites, and Triassic metasediments at the top – is characterized by isograds lying roughly parallel to the lithological contacts and to the regional foliation (Balanyà et al., 1993, 1997; Tubia, 1994; Tubia et al., 1997). Although the Ronda peridotites and the Jubrique crustal section are traditionally grouped into the so-called Los Reales Nappe (Didon et al., 1973; Fig. 2), the contact between mantle and crustal rocks consists of a high-T (T = 780 °C) major shear zone responsible for partial exhumation of the peridotites (van Wees et al., 1992; Balanyà et al., 1993; Van der Wal & Vissers, 1996; Argles et al., 1999) and representing the top of the mantle extrusion wedge according to Mazzoli & Martín-Algarra (2011). The Ronda peridotites are mainly composed of plagioclase and spinel lherzolites (with subordinate garnet lherzolites towards the top of the ultramafic body), mafic layers, and minor dunites and harzburgites (Obata, 1980; Van der Wal & Vissers, 1996; Garrido et al., 2011). They form a series of massifs, the larger being those of Sierras Alpujata and Bermeja, where ultramafic rocks are tectonically emplaced onto ‘Blanca-type’ Alpujarride units or onto ‘Frontal Units’ of the Internal Domain (Nieves Unit; Martín-Algarra, 1987; Fig. 2).

The ‘Blanca-type’ units (Ojen, Guadaiza, and Yunquera) share a typical Alpujarride succession, with high-grade metapelites at the bottom, whose protoliths are of a presumed Palaeozoic and Triassic age, and HT-marbles at the top, of probable Triassic, and perhaps younger, sedimentary age (Hoeppener et al., 1963, 1964). Stratigraphic data (Martín-Algarra & Estévez, 1984; Martín-Algarra, 1987) and zircon geochronology (Esteban et al., 2011) demonstrate the Early Miocene age (c. 20–22 Ma) of the high-T Alpine metamorphism that affects all ‘Blanca-type’ units, which was related to their tectonic emplacement below the Ronda peridotites. However, the Ojen Unit records a first stage of high-P metamorphic conditions (T = 790 °C, P ≥ 1.7 GPa; Tubía & Gil-Ibarguchi, 1991) with a temperature peak dated at 19.9 ± 1.7 Ma (Sánchez-Rodríguez & Gebauer, 2000) in eclogites derived from former Jurassic metabasites dated at 183 ± 3 Ma. Fast exhumation (at a rate of 6 km Ma^{-1}; Orozco and Alonso-Chaves, 2012, and references therein) of these high-P rocks, rather than by ultra-rapid extension and extreme thinning of an unrealistic (> 50 km thick) nappe stack, is likely to have occurred within a subduction channel (Mazzoli and Martín-Algarra, 2011). On the contrary, the Guadaiza Unit displays exclusively evidence of low-P metamorphic conditions (T = 675–750 °C, P = 550–650 MPa; Esteban et al., 2008). Finally, the Yunquera Unit consists of an extremely thin, discontinuously exposed tectonic slice that
was possibly produced by tectonic delamination of the Ojen Unit in the footwall to the 
Ronda peridotite wedge, as it also preserves relics of high-P metamorphism (Martín-
Algarra, 1987). In the area of the present study, the Yunquera Unit is locally exposed in 
the footwall to the Ronda peridotites (Fig. 3). Both of these units record NW-directed 
thrusting over the Nieves Unit (Esteban et al., 2005; Mazzoli & Martín-Algarra, 2011).

The Nieves Unit

The Nieves Unit (Dürr, 1967) is formed by a non-metamorphic – apart from the aureole 
at the contact with the Ronda peridotites – Triassic to Lower Miocene sedimentary 
succession (Fig. 4). A few tens of metres thick sole of strongly sheared, brown scaly 
clays of the Campo de Gibraltar Complex, including frequent quartzarenites and rare 
micaceous sandstone phacoids, commonly occurs in its immediate footwall. This sole 
constitutes a strongly thinned and stretched tectonic mélangé interposed between the 
Nieves Unit and the tectonically underlying Penibetic Units of the External Domain. 
The hanging wall to the Nieves Unit is constituted by the Los Reales Nappe 
(Alpujarride Complex), which in turn is overlain by the Malaguide Complex. Both 
complexes thin rapidly northwards, forming a wedge-shaped, broad antiformal structure 
whose core is occupied by the folded succession of the Nieves Unit (Fig. 4). The Ronda 
peridotites at the base of the so-called Los Reales Nappe represent the thickest part of 
the wedge but rapidly disappear towards the N due to the hanging-wall cutoff above the 
Nieves Unit (Fig. 4). The Jubrique crustal succession overlying the peridotites is mostly 
eroded above the main outcrop of the Nieves Unit; however, thin remnants are 
preserved as small klippen on top of the youngest beds of the Nieves succession (Nava 
Breccia Fm., see below), being in turn locally overlain by klippen of Malaguide 
Palaeozoic rocks (Fig. 3).

Stratigraphy

The stratigraphic succession of the Nieves Unit includes a c. 1.5 km thick Mesoe-
Cenozoic succession (Fig. 4) mainly made of carbonate formations showing typical 
Alpine-Mediterranean facies.
The lowest formation is made of very thick- to medium-bedded Norian dolostones 
showing shallow marine facies and a cumulative thickness in excess of 1 km (unit 1 in
Fine-grained terrigenous intercalations are locally present, especially in the lowest part and towards the upper third part of the succession. The top of the Triassic succession is made of up to 200 m thick, basinal, greyish and locally yellowish marls alternating with hemipelagic micritic limestones (unit 2 in Fig. 4) that gradually pass upwards (and laterally northwards) to black to dark-grey limestones (unit 3 in Fig. 4). The Jurassic-Paleogene part of the succession (unit 4 in Fig. 4) starts with cherty limestones (usually less than 200 m thick), locally including carbonate turbidites rich in crinoidal bioclasts, mass-flow limestone breccias and red nodular limestone horizons bearing lower Hettangian and Sinemurian ammonites. The succession becomes marly towards the top of the Lower Jurassic section. Upwards, a condensed pelagic succession develops (Martín-Algarra et al., 1998; O’Dogherty et al., 2001). It is made of Middle Jurassic red nodular and cherty limestones and radiolarian cherts, Upper Jurassic and lowermost Cretaceous marly and cherty limestones with ammonites, Saccocoma, Aptychus and Calpionella limestones (unit 5 in Fig. 4). The latter are followed by varicoloured marls bearing Paleogene to Aquitanian planktonic foraminifera and calcareous nanoplanckton, with intercalated carbonate turbidites. The succession of the Nieves Unit is topped by a Lower Miocene, rauhwacke-like, continental carbonate breccia (Nava Breccia Fm; unit 6 in Fig. 4). This breccia is cartographically unconformable onto the older formations of the Nieves Unit, although being intensely folded together with the underlying part of the succession. The palaeogeographic and palaeotectonic significance of the Nava Breccia Fm – representing the youngest deposit involved in the deformation and, therefore, a key stratigraphic unit of the study area – has been investigated by Martín-Algarra & Estévez (1984) and Martín-Algarra (1987, 2004) to which the interested reader is referred to. Towards the SE, the sedimentary succession described above is overthrust by the Ronda peridotites (Fig. 5A), intensely deformed and transformed into a metamorphic succession essentially made of both calcite and dolomite marbles, locally rich in quartz lenses and levels, with subordinate intercalations of calc schists, calcareous metapelites and paramphibolites (Figs. 5B-F and 6A-E). This transformation can be laterally followed in the field, bed by bed, from non-metamorphic lithotypes to high-grade rocks (see below), especially in the Torrecilla area (Fig. 2).

Structure
The map-scale structure of the Nieves Unit is dominated by a NW-verging, overturned syncline (Figs. 3 and 4) that can be traced for several tens of kilometres along strike (Fig. 2). Regional folding around a gently NE-plunging axis (Fig. 7A) is well defined also statistically by the distribution of the poles to bedding (S0) and to the main composite planar fabric (S0-S1) measured from the metamorphic aureole (see below). The stratigraphic succession, dominated by limestones and dolostones in the normal limb of the syncline, gives way to calcite/dolomite marbles in the overturned fold limb, which is characterized by the previously mentioned dramatic metamorphic gradient. Each stratigraphic unit mapped in the normal fold limb has a corresponding, well-defined high-T equivalent in the metamorphic aureole of the overturned limb (Fig 4).

Here the original bedding, marked by clearly distinguishable lithological changes (Fig 5B), is generally parallel to a foliation, defining a composite (S0-S1) planar fabric (Fig 5C-E). A dominantly SE-plunging (Fig. 7B) mineral lineation (L1), defined by elongated calcite/dolomite and locally tremolite crystals (Fig. 6A), occurs on the composite (S0-S1) planar fabric. Isoclinal intrafolial folds (Fig. 6B) display variably plunging hinge lines (mainly tending to lie close to L1; Fig. 7C), and dominantly SSE dipping axial surfaces (Fig. 7D). Field evidence for non-coaxial strain is rare; where observed, it is represented by a sigmoidal foliation and S-C-C’ structures indicating top-to-the-NW sense of shear.

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

Open to tight, post-metamorphic folds deform the main composite planar fabric (S0-S1) in the metamorphosed overturned limb of the regional syncline (Fig. 6C), as well as bedding (S0) in its core and in the normal fold limb. These folds show NE-SW trending hinge lines (Fig. 7E) and both SE- and NW-dipping axial surfaces (Fig. 7F). A variably developed second-phase cleavage (S2) is associated with these folds, being axial-planar to them in less competent lithologies and forming convergent cleavage fans in more competent beds. The lack of neo-formed metamorphic minerals on S2 cleavage surfaces suggests that this planar structure developed at very low-grade conditions. Within the
metamorphic aureole, the intersection between S2 and the composite planar fabric (S0-S1) defines a dominantly NE-trending linear structure parallel to the late fold hinges. A similar NE-SW trend is also defined by calcite rods developed along buckled marble-phyllite interfaces in the metamorphic units (the phyllites representing metamorphosed terrigenous intercalations in the carbonates).

The youngest shortening-related structures within the Nieves Unit include scattered kink bands (Fig. 6E), whereas high-angle, dominantly extensional faults dissect all of the previous structures as well as the main nappe contacts.

Petrology and Microstructural Analysis: Materials and Methods

In the following we describe the most representative mineral assemblages and textures of 25 selected silicate-bearing marbles sampled in the studied metamorphic aureole (Figs. 3 and 4). Rocks were sampled along the A-397 road (km 17 to 21), near the Cerro de los Cascajares peak, crossing the phlogopite, tremolite, diopside, and forsterite metamorphic zones defined by Martín-Algarra (1987), and in further outcrops, within the forsterite zone, close to the Cerro Alcojona peak (Table 1).

A first approximation to metamorphic conditions for the impure marbles of the Sierra de las Nieves contact aureole was reached by means of appropriate petrogenetic grids. P-T phase diagram projections (Fig. 9) were calculated using Perple X (Connolly, 1990, 2009) for the mixed-volatile K2O-CaO-MgO-Al2O3-SiO2-H2O-CO2 (KCMASHC) system, following the method by Connolly & Trommsdorff (1991). Thermodynamic mineral data and the equation of state for H2O-CO2 fluids were those of Holland & Powell (1998, revised version 2002). The only solution model considered in the calculations was that of the H2O-CO2 fluid (Trommsdorff & Connolly, 1991). In the figures and following text, F represents the fluid composition expressed as XCO2 [= CO2/(CO2+H2O)] and the numbers in parentheses are those of the pseudocompounds (Connolly & Kerrick, 1987) used by Perple X to represent fluid composition. All other phases were taken as pure phases and the following solid end-members were considered (abbreviations after Whitney & Evans, 2010): calcite (Cal), clinochlore (Clc), clinohumite (Chu), diopside (Di), dolomite (Dol), forsterite (Fo), microcline (Mc), phlogopite (Phl), quartz (Qz), tremolite (Tr), and spinel (Spl).

This was a realistic simplification because most of the phases observed in the rocks approach pure end-members compositions. In fact, XMg [= Mg/(Mg+Fe)] in all
ferromagnesian phases (amphibole, chlorite, clinohumite, clinopyroxene, olivine, and
spinel) commonly ranges between 0.90 and 0.95 (but in many cases it exceeds these
values) in the impure marble samples. Likewise, both calcite and dolomite display
compositions close to stoichiometry. Some exceptions are reported for phlogopite (with
up to 1.6 atoms per formula unit – a.p.f.u. – of total Al; 11 oxygens), chlorite (with up
to 2.6 a.p.f.u. of total Al; 14 oxygens) and clinohumite (very rich in fluorine: up to 1.7
a.p.f.u.; 13 cations). Pargasitic amphibole and clinopyroxene ($X_{Mg} \sim 0.60$) occurring in
some pelitic intercalations also differ from the general trend, but these assemblages
were not taken into account in the calculations. Uncertainties associated with ignoring
these solid solutions are discussed below.

Electron backscatter diffraction (EBSD) data collected by scanning electron microscope
(SEM) were used for textural analysis of seven calcite marble samples and one dolomite
marble sample from the metamorphic aureole found in the overturned limb of the
regional syncline affecting the Nieves Unit. Samples 09A-1 and 09A-3 are from the
forsterite zone; sample 09A-7 is from the diopside zone; samples 09A-12, 09A-15, 09A-
18 and 09A-19 are from the tremolite zone, and sample 09A-20 is from the phlogopite
zone. Thin sections of each sample were cut parallel to the mineral lineation (X) and
perpendicular to the composite planar fabric ($S_0-S_1 = XY$; see below). Thin sections
were mechanically polished using down to 1 µm diamond paste. To remove
mechanically-induced surface damage, a further 3 hours of polishing, with 0.06 µm
colloidal silica, was undertaken. Thin sections were coated with a thin layer of carbon to
prevent surface charging in the SEM. EBSD analysis was undertaken on a ZEISS-EVO
40XVP scanning electron microscope at Curtin University, Perth, Western Australia.
Collection, indexing and analysis of electron backscatter diffraction patterns (EBSP)
were carried out using the CHANNEL 5.10 software by Oxford Instruments. The EBSD
data were noise reduced using a ‘wildspike’ correction and a 7-neighbour zero solution
extrapolation following standard procedures (Reddy et al., 2007). Orientation data were
plotted with respect to the mesoscopic sample lineation and foliation (X direction and
XY plane respectively), as lower hemisphere, equal area projections, using CHANNEL
Mambo software. Pole figure orientation data were plotted in terms of all of the
collected EBSD data (‘all data’) and as one point per grain data. For the latter, grain
boundaries were defined by misorientation boundaries in excess of 10°, and any twin
boundaries identified by a 78° rotation around <20-21> were ignored.
Finally, thin section microphotographs (acquired using a Leica DFC280 digital camera
mounted on a Leitz Laborlux 12 Pol polarizing microscope) were employed for modal and carbonate grain size analyses of the same samples used for the textural analysis. Modal analyses were aimed to the quantification of the volume abundances of non-carbonate minerals (i.e. ‘impurities’, or secondary phases) and were performed using a point counting routine of the Leica QWinPlus V3.2.1 image analysis software. Carbonate grain size analyses were performed using the ImageJ 1.46r image analysis software (further details are summarized in Table 2).

Petrology and Microstructural Analysis: Results

Petrography of silicate–bearing marbles

The studied rocks consist of impure marbles containing relatively low amounts (6.5% to 17.8% for the samples listed in Table 2) of non-carbonate minerals. Most of lower-T dolomite marbles can be defined as phlogopite-tremolite marbles with the significant local occurrence of chlorite or spinel (Table 1). They display a granoblastic texture with a rough foliation defined by the preferred orientation of fine grained aggregates of phlogopite, which may be also strongly folded (Figs. 8A-B). Phlogopite is also found as randomly oriented, small, isolated flakes and as randomly oriented, hypidioblastic larger flakes, commonly related to other silicates such as spinel and tremolite (e.g. sample 09A-11, Fig. 8D). Large, randomly oriented phlogopites may, however, display distinctive undulose extinction. Colourless amphibole is commonly associated with calcite grains or levels, and occurs as prismatic to acicular crystals rich in small dolomite inclusions. Amphibole grains are isolated (Fig. 8F) or give place to palmed aggregates, in many places with clear preferred orientation parallel to the main foliation (Fig. 8C). Chlorite (sample 09A-16; Table 1) appears as large hypidioblastic flakes having plenty of phlogopite and dolomite inclusions. Spinel (sample 09A-11; Table 1) is found as xeno- to hypidioblastic, isolated grains that grew after – or related to – recrystallized phlogopite flakes (Fig. 8D).

Moving towards the peridotites, clinopyroxene is found in modal quartz-rich rocks (with the exception of marble sample 09A-1; Table 1), usually consisting of pelitic or quartz-feldspar-bearing intercalations within the calcite marbles. Clinopyroxene appears as green pleochroic, isolated crystals or aggregates with largely variable grain size at the contact between quartz-rich and carbonate-rich levels (Fig. 8E). Highly birefringent
scapolite poikiloblasts as well as minor amounts of alkali-feldspar, phlogopite and
tremolite (Fig. 8E) also appear in some of these rocks, (that will not be taken into
account in the phase diagram calculations below).

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

In tremolite and higher-T mineral zones, centimetre-scale phlogopite-tremolite clots that
clearly displaced and overgrew the main composite planar fabric (S₀-S₁) are observed
also in outcrop (Fig. 5F). However, as mentioned previously, tremolite crystals also
occur on the main composite foliation.

Metamorphic analysis

A simplified petrogenetic grid for the KCMASHC system and a wide P-T range is
shown in Fig. 9 (A). Both fluid-present and fluid-absent univariant equilibria occur. The
most striking feature of this P-T projection is that most of the curves display very steep
and remarkably constant Clapeyron slopes, thus defining narrow zones in which the
stability of phase assemblages is strongly dependent on temperature and fluid
composition (e.g. Fig. 9B). Calculations for the K₂O-CaO-FeO-MgO-Al₂O₃-SiO₂-H₂O-
CO₂ system (KCFMASHC) that take into account the limited Fe-Mg solid solution (X_Mg
usually < 0.95) in the ferromagnesian minerals reveals a negligible shift towards higher
temperatures of the reactions of interest (not shown in the figures) and will not be
considered in the following. The effect of the solid solutions models for phlogopite
(between the phlogopite-eastonite end-members) and chlorite (between the clinochlore-
amosite end-members), applying for the Al amounts observed in these minerals, has
also been tested. As a general trend, for the reactions of interest, pseudounivariant
curves with Al-richer pseudocompounds shift toward higher temperatures. However, within the range of the limited solid solutions observed in the studied minerals, this is a minor effect when compared with that of the changing fluid phase composition and will not be further considered. As a result, in the following analysis univariant reactions with steep Clapeyron slopes (especially those involving phlogopite or chlorite) should be regarded as indicators of the minimum temperature conditions at which observed minerals or assemblages were stable. In contrast with these steep reactions, only a few reactions display gentle or variable Clapeyron slopes. Among these, the reaction $\text{Ca} + \text{Tr} + \text{Fo} = \text{H}_2\text{O} + \text{Di} + \text{Dol}$ sets the highest pressure limit for the observed mineral assemblage, in which calcite, amphibole and olivine coexist. Accordingly, all further P-T projections (Fig. 9B to D) correspond to the expanded field between 50 and 500 MPa, and between 300 and 700 ºC for the same KCMASHC system. This pressure range is consistent with the previously described geological setting of the contact aureole. Based on geological evidence pointing to a burial depth of c. 10 km (including the mantle peridotites; Mazzoli & Martin-Algarra, 2011), maximum pressures may be further constrained to around 300 MPa. These P conditions are roughly similar to those suggested by Esteban et al. (2008) for the Guadaiza River tectonic window just south of our study area (Fig. 2).

Metamorphic conditions for the occurrence of phlogopite, amphibole, spinel and chlorite in the studied aureole can be discussed from selected representative equilibria shown in Fig. 9 (B) and summarized in Fig. 9 (D). Minimum temperature conditions for the first occurrence of phlogopite in the dolomite-bearing rocks are controlled by the univariant reaction indicating the breakdown of detrital microcline (e.g. Rice, 1977) in the presence of water-rich fluid: $\text{F}(\text{H}_2\text{O}) + \text{Dol} + \text{Mc} = \text{F}(\text{CO}_{6.6}) + \text{Cal} + \text{Phl}$. Depending on the latter parameter, the Phl-in reaction turns to be projected as pseudounivariant equilibria (Connolly & Trommsdorff, 1991), represented as grey thin dashed curves, in which only one (that of coexisting 6.6 and 9.9 pseudocompounds) is labelled with the complete reaction and all the remaining ones are labelled with the mean value of coexisting pseudocompounds (Fig. 9B). As a result, it can be deduced that the first appearance of phlogopite can take place in a wide divariant field with increasing temperature conditions for CO$_2$-richer fluids. For the estimated maximum pressure of 300 MPa (thick grey dashed line in Fig. 9D), the minimum temperature for the stability of phlogopite is 350 ºC.
A similar analysis can be carried out for defining the tremolite-in conditions. In dolomite and quartz-bearing marbles, this takes place due to the univariant reaction:

\[ \text{F(H}_2\text{O)} + \text{Qz} + \text{Dol} = \text{F(CO}_6\text{.6)} + \text{Cal} + \text{Tr}. \]

The divariant field defined by pseudounivariant equilibria (only that corresponding to pseudocompounds 16.4 and 19.7 is represented with a thin black dashed curve) is shaded in light grey in Fig. 9 (B). In pure calcite marbles the univariant tremolite-in reaction is:

\[ \text{F(H}_2\text{O)} + \text{Qz} + \text{Cal} + \text{Phl} = \text{F(CO}_6\text{.6)} + \text{Tr} + \text{Mc}; \]

sample pseudounivariant equilibrium with pseudocompounds 13.1 and 16.4 is represented with a thin continuous black curve and the divariant field is shaded in darker grey in Fig. 9 (B). Thus, at 300 MPa, the minimum temperature for the appearance of tremolite ranges from 360 °C in quartz-bearing dolomite marbles to 410 °C in calcite marbles (Fig. 9D). In the studied selected samples the latter rock type is more abundant than the former one (cf., Table 1).

The first appearance of spinel in the petrogenetic grids of Fig. 9 (B, D) is indicated by the fluid-absent univariant reaction:

\[ \text{Tr} + \text{Dol} + \text{Mc} + \text{Clc} = \text{Cal} + \text{Phl} + \text{Spl} + \text{Qz} \]

at an almost constant temperature of 430 °C. At higher T (500 °C at 300 MPa), the reaction \[ \text{Dol} + \text{Mc} + \text{Clc} = \text{Cal} + \text{Phl} + \text{Spl} + \text{Qz} + \text{F} \]

takes place. Both are clinochlore breakdown reactions yielding the same solid products. The coexistence of Cal, Phl and Spl is well documented (e.g. Fig. 8D), whereas Qz is always lacking. This can be easily explained taking into account that the chlorite end-member phase employed in the calculations (clinochlore) is richer in Si than real chlorites observed in the rocks (3 Si a.p.f.u. versus 2.5 Si a.p.f.u.). Irrespective of composition, chlorite is stable in the KCMASHC system, at Al-rich compositions, in the whole considered P-T range.

Selected relevant equilibria explaining the occurrence of diopside, olivine, and clinohumite are displayed in Fig. 9 (C). Observed diopside-bearing rocks are lacking in dolomite and very rich in quartz (Table 1). The first appearance of diopside in calcite marbles takes place through the univariant reaction:

\[ \text{F(H}_2\text{O)} + \text{Qz} + \text{Cal} + \text{Tr} = \text{Di} + \text{F(CO}_6\text{.6)} \]

As in the previous cases, a divariant field (dark grey shading) towards higher temperatures is defined by pseudounivariant reactions with increasing \(X_{CO2}\) in the fluid composition (reaching up to 0.58). The described Di-Cal-Qz divariant field (dark grey; Fig. 9D) partially overlaps with that of the tremolite-in conditions (light grey). At 300 MPa, the minimum temperature of this divariant field exactly fits that of the first Spl-in reaction (430 °C). Taking into account that the sample with the product assemblage of this reaction (09A-11; Table 1) is located only about 200 m away from the first
appearance of diopside, it could be suggested that Di-in temperatures must be somewhat higher than those of Spl-in (Fig. 9D).

Concerning forsterite stability, first P-T constraints are defined by equilibrium Cal + Tr + Fo = F + Dol + Di, with a strongly changing Clapeyron slope and increasing $X_{CO_2}$ composition of fluid with decreasing pressure (number labels of pseudoinvariant points along the curve; Fig. 9D). This curve constrains the stability of the assemblage Cal+Tr+Fo, observed in most rocks close to the contact with the peridotites (Table 1), to maximum values of P=500 MPa (at 300 ºC) and T=575 ºC (at 200 MPa). Within this field, the minimum temperature is further constrained by the reaction $F(H_2O) + Tr + Dol = F(CO_6.6) + Cal + Fo$ and its related divariant field with coexisting Tr, Dol, Cal and Fo and variable fluid composition (thin, grey, dashed pseudounivariant curves; Fig. 9C). At 300 MPa this field ranges from 510 to 560 ºC and partially overlaps with the field of minimum temperatures for Di-in (Fig. 9D). This is consistent with the mineral assemblage observed in rocks (sample 09A-1; Table 1) sampled close to the contact with the peridotites. Additional information can be obtained from both chlorite and clinohumite. The former has been observed at least in one Fo and Spl-bearing sample (09A-4; Table 1). Both reactions Clc + Chu = H_2O + Fo + Spl and $F(H_2O) + Dol + Clc = F(CO_6.6) + Cal + Spl + Fo$ are chlorite-breakdown reactions producing coexisting Fo and Spl, consistent with the observed rock assemblages (Table 1, Fig. 8G), and indicating minimum temperature conditions ranging from 525 to 600 ºC at 300 MPa (Fig. 9D). Equilibria indicating the observed coexistence of clinohumite and olivine (Table 1, Fig. 8G) are the already mentioned Clc + Chu = H_2O + Fo + Spl, and also $H_2O + Dol + Fo = Cal + Chu + F(CO_6.6)$. The stability of clinohumite increases towards higher temperatures with increasing fluorine contents (Rice, 1980). Taking this into account and attending to petrogenetic grids in Fig. 9 (A to C), it could be suggested that clinohumite-bearing assemblages sampled close to the contact with the peridotites might have reached maximum temperatures above 700 ºC.

In conclusion, phase relations modelling in the KCMASHC system suggest the following approximate temperature ranges for the observed metamorphic zones of the Nieves Unit contact aureole at the maximum, geologically consistent, pressure of 300 MPa (Fig. 9D): forsterite (> 510 ºC; probably around 700 ºC), diopside (510–430 ºC), tremolite (430–360 ºC), phlogopite (360–330 ºC).

Microstructural Analysis
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 (S₀-S₁) 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
preferred orientation (CPO) with alignment of the calcite c-axis with the pole to the main S₀-S₁ foliation (Z). This is most strongly pronounced in samples 09A-7 and 09A-20. In 09A-1 the single c-axis cluster lies oblique to Z and is rotated in the direction of the macroscopically established top-to-NW sense of shear. In other samples a subordinate component of the c-axis orientation also lies oblique to the Z direction, again synthetic with respect to the direction of the macroscopically established top-to-NW sense of shear (Fig. 10). The angular relationships between the different concentrations on the c-axis pole figure (usually < 40°) are not consistent with these clusters being associated with twinning in the samples. Comparison with high strain calcite deformation experiments indicates that such a pattern in c-axis orientations is consistent with high shear strains (Barnhoorn et al., 2004) with varying degrees of recrystallization (Pieri et al., 2001a). Only in 09A-19, there is a less defined and more symmetric distribution of c-axes (Fig. 10). This sample retains evidence for non-coaxial deformation in the form of well-developed S-C fabrics with C-planes dominated by fibrous tremolite (Fig. 8J) but again there is no significant asymmetry within the (0001) pole figure data. Along with all other samples no variation in +<a> and −<a> was observed (Fig. 10). This contrasts observation on non-coaxially deformed calcite marble where asymmetry in +<a> and −<a> are recorded during both low- and high-T non-coaxial flow (Bestmann, 2000; Pieri et al., 2001b) and may reflect a strong component of high-temperature recrystallization and steady-state flow (Pieri et al., 2001a), and/or a component of static recrystallization (Barnhoorn et al., 2005) that is consistent with the microstructure observed in high-T marbles. The data presented here indicate that orthorhombic symmetry of fabrics with respect to the principal (S₀-S₁) foliation plane cannot be used to provide unambiguous constraints on the coaxial or non-coaxial nature of the deformation at high temperature and high-strain rate conditions. However, where asymmetry in the c-axis distribution relative to the foliation is recorded, then this appears to be synthetic with respect to the rotational component of the non-coaxial strain inferred from S-C fabrics and mesoscopic kinematic criteria.

Misorientation axes for calcite marble data record differences between samples. In samples where significant low-angle boundaries are preserved (09A-1, 09A-7, and 09A-20) there is a dominant misorientation axis concentration associated with the ‘a’ <11-20> direction. This direction is still common in 09A-3 and 09A-19 which also have a significant number of measurements (n > 1000), but other orientations are also apparent associated with ‘f’ <02-21>, ‘m’<10-10>. In samples where low-angle boundaries are
rare (09A-12 and 09A-18), misorientation axes are concentrated close to the <0001> direction. The observed variation in the misorientation axes reflects the operation of different slip systems in calcite with <11-20> corresponding to the common high temperature (> 400 °C) slip systems r{10-14}<20-21> and f{-1012}<10-11>, whilst ‘m’<10-10> is consistent with deformation by {0001}<12-10> slip (De Bresser & Spiers, 1997). In some of the samples (09A-1, 09A-18, 09A-19), misorientation axis evidence for the operation of f{-1012}<0-2-21> is consistent with low temperature deformation (De Bresser & Spiers, 1997) and this may reflect a lower temperature overprint that may also have been responsible for the localized zones of reduced grain size observed in the highest temperature sample (09A-1). Such a lower temperature overprint could have led to a finer grain size of samples 09A-18 and 09A-19 (Table 2) as a result of dynamic recrystallization dominated by subgrain rotation.

EBSD analysis has also been performed on a dolomite marble sample (09A-15) from the tremolite zone (Fig. 10). The CPO in this sample, although clearly marked by the c-axis distribution, is weaker than that generally recorded by calcite marbles and also shows different misorientation axes from most of the calcite-dominated samples. This contrast suggests the operation of a different set of slip systems, albeit including the common (0001)<2-1-10> slip system (Wenk, 1985). The asymmetry of the c-axis distribution with respect to the S0-S1 foliation plane is synthetic with the rotational component of simple shear.

Discussion

The Nieves Unit is characterized by significant changes in styles and intensity of deformation, with strongly heterogeneous structural development being closely linked to an inverse metamorphic gradient that is controlled by distance from the hanging-wall Ronda peridotites. This variation is expressed by different deformation features at different structural position within the main regional syncline involving the studied unit. This major structure may be interpreted as a result of footwall deformation within the general framework of bulk non-coaxial strain associated with top-to-the-NW shearing.

The normal limb and core region of the regional syncline are characterized by the occurrence of NE-SW trending, meso-scale parasitic folds, whereas in the metamorphic aureole of the inverted limb, isoclinal, intrafolial folds occur. The hinge lines of these latter folds tend to lie close to the stretching direction and are interpreted to have been
strongly rotated towards this direction during progressive deformation. The occurrence of a well-developed foliation and associated mineral lineation, together with isoclinal intrafolial folds in the marbles, indicates that these rocks record a significant amount of finite strain, although shear-sense indicators are rarely observed in the field. These features are consistent with progressive strain localisation in the inverted limb of the major syncline. As a matter of fact, the overturned fold limb of the regional syncline appears to constitute a SE dipping, thermally softened carbonate rock panel that underwent intense deformation and metamorphism associated with the emplacement of the Ronda peridotites.

Both field and thin section evidence indicate a limited syn-kinematic growth of silicate minerals (locally testified by the preferred orientation of early generations of tremolite and phlogopite, Fig. 8B, C). However, most silicate grain growth appears to have occurred statically, following the development of the composite planar fabric (S₀-S₁) and mainly post-dating the main internal deformation in the marbles (Fig. 8F). This suggests that high-T footwall deformation associated with peridotite emplacement occurred at a much faster rate with respect to heat diffusion, which continued well after peridotite emplacement. Estimated maximum temperatures in the metamorphic aureole range from values in excess of 510 °C (probably around 700 ºC) close to the contact with the peridotites to < 350º C in the more distant, less thermally overprinted rocks, for the inferred maximum pressure of 300 MPa. Higher-pressure values are rather unlikely due to the complete lack of regional metamorphism of the Nieves Unit away from the aureole at the contact with the peridotites. Had the overlying tectonic units been significantly thicker than c. 10 km, the whole footwall (Nieves Unit) would have undergone greenschist facies metamorphism (which is not observed). Therefore, it can be realistically considered that the present-day geometry (taking erosion into account; see geological sections in Figs. 2 and 4) is representative of the original (Miocene) tectonic load on top of the Nieves Unit (Mazzoli and Martín-Algarra, 2011, their fig. 6). This is consistent with syn-orogenic extension (Argles et al., 1999) of the Jubrique crustal rocks originally overlying the Ronda peridotites. The crustal succession was being tectonically attenuated as the Ronda peridotites were exhuming, so that the Los Reales Unit was already significantly thinned (essentially to its present-day thickness) when final emplacement on top of the Nieves Unit occurred (Mazzoli and Martín-Algarra, 2011, their fig. 6). A maximum temperature of c. 700 °C is consistent with a peridotite emplacement postdating the latest metamorphic event recorded within the
ultramafic body, which occurred at conditions of $P = 1$ GPa and $T = 800–900$ °C (obtained for the aluminous mafic rocks alternating with peridotites and sampled near the marbles section studied in this work; Morishita et al., 2001). Such high P-T conditions are interpreted as corresponding to a sustained synkinematic decompression event (Garrido et al., 2011) that took place well before the final peridotite emplacement described here. The general – though irregular – trend of increasing calcite grain size with metamorphic grade observed in the studied footwall carbonates confirms the fundamental control exerted by temperature on calcite recrystallization processes, although the variable amount of secondary phases (Table 2) present in our impure marbles is likely to have played a major role, resulting in complex calcite grain growth patterns (Ebert et al. 2007a, b). As a matter of fact, the irregular distribution of calcite grain size shown in Table 2 is likely to result from a complex interplay of different factors including original grain size, amount of impurities, differential stress and thermal regime controlling the dominant deformation mechanisms. The microstructure of calcite marbles, characterized by the occurrence of undulose extinction, subgrains and new recrystallized grains, is indicative of deformation by slip and climb of dislocations (dislocation creep). The analysed calcite marbles show a well-developed CPO at all metamorphic grades, for a wide range of grain sizes and different – though generally moderate – amounts of secondary phases (Table 2). This is consistent with the dominance of dislocation creep at all temperatures characterizing the studied metamorphic aureole. Both microstructures and CPOs show no indications for a switch in deformation mechanisms from dislocation to diffusion creep or to grain-size sensitive mechanisms even at the highest temperature conditions (forsterite zone). Such an interpretation is in general agreement with the experimental findings of Barnhoorn et al. (2004). The associated occurrence of grain boundary migration recrystallization (testified by the common presence of lobate grain boundaries) is also consistent with large strains and relatively high temperatures in calcite marbles (e.g. Barnhoorn et al., 2004). The generally symmetric CPO patterns suggest high-T plastic deformation, dynamic recrystallization and steady-state flow, probably followed by variable degrees of static recrystallization that may have weakened the CPO (Barnhoorn et al., 2005). In the highest-grade forsterite zone, the presence of a carbonate melt may have also contributed to a weakening of the CPO. In fact, the highest temperatures close to the peridotite contact were potentially higher than the eutectic of marbles (e.g. Boettcher & Wyllie, 1969). Therefore, at least part of the microfabric suggesting static
recrystallization associated with thermal diffusion – and the related weakened CPO –
could actually have resulted from grain growth following partial melting. Melt fractions
– no matter how small – may additionally have controlled deformation processes in the
highest-grade marbles by enhancing grain boundary sliding and migration, as well as
rotation.
The lack of strong asymmetries in +<\textit{a}> and –<\textit{a}> distribution in marbles from all
metamorphic zones, suggesting the occurrence of a coaxial strain component (Bestmann,
2000; Pieri \textit{et al.}, 2001b), is consistent with bulk deformation by general shear and/or
partitioning of the deformation into zones dominated by either non-coaxial or coaxial
strain. Furthermore, the occurrence of minor zones of strain localization and grain size
reduction (associated with subgrain rotation recrystallization) overprinting previously
annealed calcite textures (e.g. sample 09A-1) suggests a complex interplay between
thermal effects and deformation events. Notwithstanding this, a final static overprint
appears to be dominant in carbonates throughout the metamorphic aureole. This is
consistent with the interpretation that most of the petrologic features characterizing the
metamorphosed rock panel (inverted limb of the Nieves Unit regional syncline) located
immediately in the footwall to the Ronda peridotites were acquired following the
emplacement of the mantle extrusion wedge. Continued heat transfer from the
peridotites to the footwall succession led to post-kinematic silicate growth and locally –
especially in the highest-T forsterite zone – to partial annealing of calcite marble
textures. Following substantial cooling (Esteban \textit{et al.}, 2004), horizontal crustal
shortening was resumed – as testified by meso-scale refolding affecting both
metamorphic and non-metamorphic parts of the Nieves Unit – probably in relation with
NW-ward thrusting of the ‘Frontal Units’ of the Internal Domain over the Flysch and
External Domains (Figs. 2 to 4). Such a late, SE-NW oriented shortening of the Nieves
Unit occurred at much lower temperatures, as testified by the lack of metamorphic
mineral assemblages along the axial plane foliation (S\textsubscript{2}) associated with the second-
phase, NE-SW trending mesoscopic folds.

\textbf{Conclusions}

Structural analysis of the dominantly carbonate Nieves Unit located in the footwall to
the Ronda mantle extrusion wedge unravelled strongly heterogeneous deformation
accompanying a dramatic metamorphic gradient. Marbles within a several hundreds of
meters-thick metamorphic aureole record high strain in the form of a well-developed foliation and associated mineral lineation, accompanied by widespread isoclinal intrafolial folds. Field evidence for non-coaxial strain, although clearly consistent with top-to-the-NW shearing, appears to be subordinate. On the other hand, the microstructures of calcite marbles – particularly the lower-T phlogopite-tremolite ones – effectively record top-to-the-NW kinematics, whereas their CPOs suggest the possible occurrence of a coaxial strain component and therefore of an overall deformation by general shear. Where asymmetry in the c-axis distribution is observed with respect to the main composite foliation (sub-parallel to the shear plane, due to very likely high-strain values), this appears to be synthetic with respect to the rotational component of the non-coaxial strain inferred from mesoscopic kinematic criteria, this being consistent with large shear strains and high-T conditions. Our results indicate that large strains and attenuation of a thermally softened carbonate rock panel, probably under a general shear deformation regime, dominantly controlled structural development and marble textural fabrics in the metamorphic aureole in the footwall to the Ronda peridotites.

Petrological characteristics and mineral assemblages within the zoned metamorphic aureole appear to have mostly developed after the final emplacement of the Ronda mantle extrusion wedge. Post-kinematic, continued heat transfer from the hanging-wall peridotites led to static growth of silicate assemblages and local partial annealing of calcite marble textures, especially in the highest-T zone in the immediate footwall. Following substantial cooling, renewed crustal shortening affected the Nieves Unit, probably in relation with foreland-directed thrusting.

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Figure Captions

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

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

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

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

Fig. 5. Field examples of structural features in the Nieves Unit in the footwall to the Ronda peridotites. (A) View of the tectonic contact with the overlying Ronda peridotites (Cerro Corona area). (B) View of the hinge zone of the regional syncline, including second-order overturned folds (Fuenfría area). (C) Composite planar fabric ($S_0$-$S_1$) in dolomite marble (km 19.5 of the A-473 road from Ronda to San Pedro de Alcantara). (D) Close-up of the boxed area in (C), showing minor folds in the hinge zone of a larger syncline. (E) Composite planar fabric ($S_0$-$S_1$; horizontal) in metamorphosed Jurassic cherty limestones (Torrecilla area). Note symmetric pressure shadows around frambooidal pyrite grains (arrowed). (F) Composite planar fabric ($S_0$-$S_1$) displaced and overgrown by phlogopite-tremolite clot in metamorphosed Norian levels (dolomite marbles of the Cerro Alcojona area).
Fig. 6. Outcrop examples of structural features in the Nieves Unit. (A) Mineral lineation (L1) defined by elongated tremolite (arrowed) and dolomite crystals lying on the main foliation in metamorphosed Norian levels (dolomite marbles at km 19.5 of the A-473 road from Ronda to San Pedro de Alcantara). (B) Early intrafolial folds in dolomite marbles (Cerro Alcojona area). (C) Late (post-metamorphic) folds affecting the main composite planar fabric (S0-S1) in dolomite marbles (km 19.5 of the A-473 road from Ronda to San Pedro de Alcantara). (D) Asymmetric (z-shaped looking towards SW) parasitic folds affecting bedding (S0) in the normal limb of the regional syncline (Jurassic cherty limestones of the Torrecilla area). (E) Late kink bands affecting the main composite planar fabric (S0-S1) along the San Pedro de Alcantara road.

Fig. 7. Orientation data (lower hemisphere, equal-area projections) for structures in the Nieves Unit. (A) Poles to bedding (S0)/composite planar fabric (S0-S1; metamorphic aureole); pole to the best-fit great circle (or π-girdle, plunging 19° toward 54°N) provides a statistical axis consistent with NE-SW trending folding. (B) Mineral lineation (L1) in silicate-bearing marbles. (C) Hinges of isoclinal, intrafolial folds. (D) Poles to axial surfaces of isoclinal, intrafolial folds. (E) Hinge lines of late, open to tight folds within metamorphic aureole. (F) Poles to axial surfaces of late, open to tight folds within metamorphic aureole.

Fig. 8. Thin section microphotographs (crossed polars) showing representative petrological and microstructural features of the studied marble samples (mineral name abbreviations after Whitney and Evans, 2010). (A) Intensely deformed, high-T calcite marble (sample 09A-02; forsterite zone). (B) Granoblastic, mainly dolomite marble (with scarce, “dusty” calcite grains) with a thin folded layer of phlogopite (sample 09A-15; tremolite zone). (C) Palmed aggregate of amphibole subparallel to the main rock foliation as indicated by the white dashed line (sample 09A-16; tremolite zone). (D) Xenoblastic spinel grains associated with calcite and large phlogopite flakes (sample 09A-11; tremolite zone). (E) Typical mineral assemblage of clinopyroxene, calcite and quartz-bearing rock (sample 09A-08; diopside zone). (F) Hypidioblastic clinopyroxene and amphibole grains overgrowing strongly deformed calcite grains (sample 09A-01; forsterite zone). (G) Typical high-grade mineral assemblage of olivine-clinohumite-spinel marble close to the contact with peridotite (sample 09A-24; forsterite zone). (H)
High-T calcite marble (sample 09A-3; forsterite zone) showing lobate grain boundaries (red arrows), straight grain boundaries and triple junctions (yellow arrows) and secondary foliation (sf) defined by recrystallized grains oblique to the main composite foliation (horizontal; NW is to the right of the picture). (I) High-T marble with equant calcite grains showing both lobate and straight boundaries (sample 09A-25; forsterite zone). (J) Lower-T calcite marble (sample 09A-19; tremolite zone) displaying S-C-C’ fabrics (NW is to the right of the picture). (K) Lower-T calcite marble (sample 09A-20; phlogopite zone) showing secondary foliation (sf) defined by recrystallized grains oblique to the main composite foliation (horizontal; NW is to the right of the picture).

**Fig. 9.** Calculated P-T projections for the K₂O-CaO-MgO-Al₂O₃-SiO₂-H₂O-CO₂ system with the following minerals (abbreviations after Whitney and Evans, 2010): calcite (Cal), clinochlore (Clc), clinohumite (Chu), diopside (Di), dolomite (Dol), forsterite (Fo), microcline (Mc), phlogopite (Phl), quartz (Qz), spinel (Spl) and tremolite (Tr). Abbreviation for the fluid solution is F. All curves in (A) correspond to univariant equilibria. Field b-d in (A) is enlarged in diagrams (B), (C) and (D). Note that minimum pressure in these cases widens towards 50 MPa. Thick and thin dashed curves correspond to univariant and pseudounivariant equilibria, respectively (see text for details). Big circles and small filled circles along univariant curves correspond to invariant and pseudoinvariant points, respectively. Different grey areas represent divariant fields in which pseudounivariant equilibria are stable with the same solid phases and changing fluid composition (X_{CO₂} always increases with temperature). In (D), thick dashed line at a constant pressure of 300 MPa acts as reference for obtaining the temperature ranges for the relevant phases observed in the studied rocks (upper part of the diagram).

**Fig. 10.** CPO data for calcite marble samples from: (i) forsterite zone (09A-1, 09A-3); (ii) diopside zone (09A-7); (iii) tremolite zone (09A-18, 09A-19); and (iv) phlogopite zone (09A-20). Dolomite marble sample (09A-15) is from the tremolite zone. All projections are lower hemisphere and equal area. Contour lines are at intervals of 0.5 Multiples of Uniform Density (MUD) and the maximum MUD is given to two significant figures. For forsterite zone samples, heterogeneous grain size gives rise to a grain-size dependent bias in standard pole figures. For 09A-1 and 09A-3 data are therefore shown as 1 point per grain. For other samples, pole figure data (columns 1-5)
represents all of the collected EBSD data. For comparison, 1 point per grain data for all samples is given in column 6, along with the MUD and the total number of analysed grains. Misorientation axes data are calculated for minimum angular misorientations between adjacent EBSD analyses of 2-10° and are given with MUD and number of misorientation axes. Orientation of oblique foliation is indicated in column 1 for 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)

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* 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.

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* 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).
**Nieves Unit**

- Lower Miocene Nava Breccia (6)
- Condensed limestones, marls and cherts (5), and calcshists with quartz nodules and talc (5’)
- Cherty limestones and marls (4), and calcitic marbles with quartz nodules and talc (4’)
- Limestones (3) and calcitic marbles with talc (3’)

**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 Alcántara, with location of the studied samples (projected)