Deciphering the tectono-metamorphic evolution of the Nevado-Filábride complex (Betic Cordillera, Spain) – A petrochronological study

Ángel Santamaría-López⁎, Pierre Lanari, Carlos Sanz de Galdeano

Departamento de Química Inorgánica, Cristalografía y Mineralogía, Facultad de Ciencias, Universidad de Málaga, 29071 Málaga, Spain
Institute of Geological Sciences, University of Bern, Baltzerstrasse 1 + 3, 3012 Bern, Switzerland
Instituto Andaluz de Ciencias de la Tierra, Consejo Superior de Investigaciones Científicas–Universidad de Granada, Facultad de Ciencias, 18071 Granada, Spain

ARTICLE INFO

Keywords:
Nevado-Filábride complex
Garnet thermobarometry
Allanite dating
XMapTools
GrtMod

ABSTRACT

The Nevado-Filábride metamorphic complex (Betic Cordillera, Spain) exhibits a succession of Paleozoic and Mesozoic metasediments, orthogneisses and metabasites. The complex has been divided into three tectono-metamorphic units, from bottom to top: the Ragua, Calar-Alto and Bédar-Macael units. The petrochronological evolution of the complex is not well constrained as P-T-t conditions differ between authors mainly because of the investigated rock types and the thermobarometric and geochronological techniques applied. Five garnet-bearing mica-schists from various areas of the complex were investigated for constraining and comparing the shape of the P-T paths recorded by the three units. Quantitative compositional mapping of garnet and K-white mica was combined with iterative thermodynamic models. The resulting P-T trajectories suggest that the units experienced similar clockwise P-T trajectories during Alpine metamorphism. The three units reached high-pressure and low-temperature conditions of ~2.0 GPa and ~520 °C in the Bédar-Macael unit; ~2.0–2.2 GPa and ~470–490 °C in the Calar-Alto unit; and ~2.2 GPa and ~480 °C in the Ragua unit. All samples recorded a temperature increase of ~130 °C during exhumation as shown by the successive stages of white mica re-equilibrations. Heating during exhumation was probably triggered by the hot hanging wall over the Nevado-Filábride complex. The garnet rims reflect a high-temperature and low-pressure stage which was dated at ~13 Ma using allanite U-Th-Pb geochemistry. Following the temperature peak, exhumation continued associated to cooling. Similarities in the shapes of P-T path throughout the units suggest a continuous metamorphic sequence rather a tectonically divided complex.

1. Introduction

Pressure-Temperature-time (P-T-t) path summarizes the changes in pressure and temperature conditions that have been recorded in a rock sample during an orogeny. In order to decipher mountain-building tectonic processes such as the subduction and exhumation of metamorphic complexes, P-T-t paths for metamorphic rocks of several units are required (e.g. Ernst, 1988; Guillot et al., 2009; Rolland et al., 2012). Quantification of the differences or similarities in the conditions recorded by several tectonic units is a great help in deciphering possible tectonic scenarios and in resolving tectonic and thermal histories (e.g. Thompson and England, 1984; Jamienson, 1991; Foster and Parrish, 2003; Brown, 2014).

The development of analytical and modelling techniques in metamorphic petrology has opened new possibilities for thermobarometry and improved our ability to reconstruct portions of P-T paths (e.g. Brown, 2007, 2014; Lanari et al., 2019; Lanari and Duesterhoeft, 2019). Thermobarometric methods include empirical thermobarometers (e.g. Hodges and Crowley, 1985), multi-equilibrium or average P-T approach (e.g. Holland and Powell, 1998; Berman, 1991) equilibrium phase diagrams (e.g. Holland and Powell, 1998, 2011), Raman spectrometry on carbonaceous material (e.g. Beyssac et al., 2002), etc. Generally, these techniques are based on the assumption of chemical equilibrium and have absolute uncertainties of ±50 °C (Kohn and Spear, 1991). By contrast, relative errors are smaller, allowing relative thermobarometry between units to be performed in a much precise way (Worley and Powell, 2000).

Thermobarometry requires the knowledge of the composition of minerals assumed to have been in chemical equilibrium at a given stage of their metamorphic history. Resilient porphyroblasts are excellent candidates as they can preserve chemical and textural record of metamorphic conditions affected the rock (e.g. Lanari et al., 2017; Brown, 2007, 2014; Lanari et al., 2019; Lanari and Duesterhoeft, 2019).
This is the case of garnet which is a common rock-forming mineral often preserving compositional zoning interpreted to reflect changes in equilibrium conditions (Gaidies et al., 2008; Caddick and Kohn, 2013; Cheng and Cao, 2015; Baxter et al., 2017; Lanari and Engi, 2017). For that purpose, the application of quantitative compositional mapping has proven to be of great help to image compositional zoning and relate these to P-T conditions of growth (Lanari et al., 2013, 2014, 2019). However, as garnet grows, it becomes isolated from the reactive part of the rock and the thermobarometric reconstruction based on bulk rock compositions are biased (Spear, 1988; Evans, 2004; Tinkham and Ghent, 2005). An alternative strategy to confront this problem is that followed by Lanari et al. (2017) which allows P-T conditions to be refined using an iterative optimization process based on the composition of each successive growth zones. Resorption and fractionation of garnet previously formed are optimized to assess the possible variations of the reactive bulk composition.

The collision of the African and Iberian plates led to the formation and staking of three metamorphic complexes included in the Internal Zone of the Betic Cordillera and named from bottom to top: the Nevado-Filábride, the Alpujárride and the Maláguide. In numerous studies, it has been proved that the Nevado-Filábride complex was affected by metamorphism during the Alpine orogeny (e.g. Augier et al., 2005b; Platt et al., 2005; Platt et al., 2006). Although several studies suggested that the Nevado-Filábride metamorphic history is characterized by an early Alpine HP/LT metamorphism followed by decompression at LP/HT, no consensus has been reached regarding the shape of the P-T-t trajectories recorded by each unit of this complex (Gómez-Pugnaire and Fernández-Soler, 1987; Bakker et al., 1985; Vissers et al., 1995; López-Sánchez-Vizcaíno et al., 2001; Puga et al., 2002; Augier et al., 2005a, 2005b; Behr and Platt, 2012; Ruiz-Cruz et al., 2015; Li and Massonne, 2018). The published P-T paths are scattered in particular for the decompression history. Differences strongly depend on the rock types and the thermobarometric methods applied, thus preventing the comparison to be made with a high degree of confidence. Another subject of debate is the division of the Nevado-Filábride complex into large-scale units (e.g.: Puga et al., 1974; Puga et al., 2002; Martínez-Martínez et al., 2002; Azañón and Goffé, 1997; Sánchez-Rodríguez and Gebauer, 2000; Booth-Rea et al., 2002; Janots et al., 2006) followed by a low pressure and locally high temperature stage (Platt et al., 2013). The lower complex, which is investigated in this study, is known as the Nevado-Filábride complex, and is formed by mica-schists (mainly), quartzites, marbles, orthogneisses and metabasites. The contact between both complexes is characterized by a ductile to brittle low-angle normal fault formed during the exhumation process (Gálindo-Zaldívar et al., 1989; Jabaloy et al., 1993; Augier et al., 2005c; Agard et al., 2011; Jabaloy et al., 2015).

According to Santamaria-López and Sanz de Galdeano (2018) the maximum age of deposition of the mica-schists protolith is ca. 349 Ma.

2.2. Nevado-Filábride complex division

Several tectonic and tectono-metamorphic models have been proposed in the literature, but each of them strongly relies on how the units within the Nevado-Filábride complex are defined. Brouwer (1926) first distinguished two formations based on lithological criteria: the lowermost Crystalline of Sierra Nevada and the Mischungszone – mixed – zone. Since that, the division has remained a matter of debate depending on the criteria used by each author (i.e. petrological, structural and/or metamorphic). According to Puga et al. (1974) and Puga and Díaz de Federico (1976), the Nevado-Filábride complex is divided into the Veleta nappe (equivalent to the Crystalline of Sierra Nevada), and the overlying Mulhacén nappe (approximately equivalent to the Mischungszone). The Veleta nappe contains a monotonous sequence of dark schists (graphite-bearing) and quartzites, whereas the Mulhacén nappe shows a larger variety of lithologies such as light and dark schists, quartzites, marbles, gneisses, serpentinites and amphibolites. Puga et al. (2002) reclassified as “complexes” the “nappes” of Puga and Díaz de Federico (1976). Martínez-Martínez et al. (2002) renamed the Veleta nappe as Raguia unit (being not exactly, but approximately equivalent), and subdivided the Mulhacén nappe into the lowermost Calar-Alto unit and the overlying Bédar-Macael unit. However, authors...
Fig. 1. Geological map and transects of the southeast Iberian Peninsula. a General map of the Betic cordillera showing the Nevado-Filabride complex distribution. Sampling areas: Peñones de San Francisco and La Ragua, are shown as lines labeled as 1 and 2 respectively. b Geological cross sections of the Peñones de San Francisco and La Ragua areas. c Schematic lithological succession of the Nevado-Filabride Complex including the sample location.
as Galindo-Zaldívar (1993), Sanz de Galdeano and López-Garrido (2016) and Sanz de Galdeano et al. (2016) have rather suggested a continuous lithological sequence in the Nevado-Filábride complex. The location and even existence of the tectonic units remains imprecise nowadays despite the number of studies focused on the complex. In the present study the division used by Martínez-Martínez et al. (2002) has been adopted because its use in several recent studies of metamorphic conditions in the Nevado-Filábride complex (e.g. Platt et al., 2013; Kirchner et al., 2016; Li and Massonne, 2018). According to several authors (e.g. Martínez-Martínez et al., 2002; Augier et al., 2005a; Booth-Rea et al., 2015) the contact between the Ragua and Calar-Alto, and the Calar-Alto and Bédar-Macael units correspond to ductile shear zones.

### 2.3. Metamorphism in the Nevado-Filábride complex

Several lines of evidence suggest a two-stage metamorphic history with a HP/LT Alpine stage followed by a LP/HT stage during decompression. This latest stage largely overprinted the previous metamorphic record particularly, but not exclusively, in the metasediments. A few metamorphic studies were focused on metapelites (Augier et al., 2005a, 2005b; Ruiz-Cruz et al., 2015) and most of the constraints for the HP/LT stage are from the studies focused on mafic and felsic igneous rocks (e.g. Gómez-Pugnaire et al., 1994; López-Sánchez-Vizcaino et al., 2001, 2005; Puga et al., 2000, 2002; Padrón-Navarta et al., 2010; Ruiz-Cruz et al., 2015). However, the P-T estimates obtained in each lithology point toward a two-stage metamorphic history with HP/LT conditions and LP/HT conditions (Fig. 2). The resulting P-T paths however, strongly differ between authors, in particular for the decompression stage. Puga et al. (2000, 2002) and Behr and Platt (2012) proposed hairpin trajectories (Fig. 2a), which show coincidence in the P and T peaks; Augier et al. (2005a, 2005b), Ruiz-Cruz et al. (2015) and Laborda-López et al. (2018) suggested an isothermal decompression (Fig. 2b); Gómez-Pugnaire and Fernández-Soler (1987) proposed slight heating upon decompression (Fig. 2c); and Vissers (1981), Bakker et al. (1989), Booth-Rea et al. (2015) and Li and Massonne (2018) pointed to the existence of late reheating at lower pressure conditions (Fig. 2d). Discrepancies between authors have been attributed either to differences between thermobarometers, or variations of P-T trajectories throughout the cordillera (Platt et al., 2013).

### 2.4. Age constraints

Similarly to the internal division of the Nevado-Filábride complex and its P-T history, there is no consensus regarding the age of each metamorphic stage. Augier et al. (2005a) applied in situ laser $^{40}$Ar/$^{39}$Ar on K-white mica in metapelite samples from Calar-Alto and Bédar-Macael units. They concluded that the HP/LT stage took place at ca.
3. Methods

The petrochronology strategy used in this study involved the following steps (see Engi et al., 2017 for a detailed description): (1) classical thin section analysis for petrography, (2) quantitative compositional mapping combined with iterative thermodynamic modelling to retrieve P-T conditions and (3) in situ U-Th-Pb allotriomorphic dating. The analytical and computational methods employed are described in the following sections.

3.1. Petrography

Five mica-schist samples from two areas of the Nevado-Filábride complex were investigated in this study (samples 16.16, 16.8, RA-0 and RA-2, see Section 4). The studied samples were investigated first by optical microscopy and scanning electron microscopy (SEM) using an Environmental Scanning Electron Microscope (ESEM) FEI model Quanta 400, operating at 15–20 keV in Centro de Instrumentación Científica (CIC, Universidad de Granada, Spain). Allanite grains were identified in backscattered electron (BSE) images using a ZEISS EVO 50 scanning electron microscope at the Institute of Geological Sciences, University of Bern.

Table 1

<table>
<thead>
<tr>
<th>Sample</th>
<th>16.16</th>
<th>16.8</th>
<th>RA-0</th>
<th>10.1</th>
<th>RA-2</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>60.19</td>
<td>57.24</td>
<td>53.00</td>
<td>59.33</td>
<td>75.10</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.42</td>
<td>0.50</td>
<td>1.46</td>
<td>0.61</td>
<td>0.80</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>20.35</td>
<td>19.91</td>
<td>23.29</td>
<td>19.18</td>
<td>12.04</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>10.86</td>
<td>7.60</td>
<td>8.93</td>
<td>9.44</td>
<td>4.49</td>
</tr>
<tr>
<td>MnO</td>
<td>0.12</td>
<td>0.50</td>
<td>0.12</td>
<td>0.41</td>
<td>0.04</td>
</tr>
<tr>
<td>MgO</td>
<td>2.51</td>
<td>2.10</td>
<td>1.24</td>
<td>2.30</td>
<td>1.00</td>
</tr>
<tr>
<td>CaO</td>
<td>0.82</td>
<td>2.29</td>
<td>0.30</td>
<td>1.27</td>
<td>0.25</td>
</tr>
<tr>
<td>Na₂O</td>
<td>1.29</td>
<td>1.99</td>
<td>0.89</td>
<td>1.29</td>
<td>0.65</td>
</tr>
<tr>
<td>K₂O</td>
<td>2.55</td>
<td>2.14</td>
<td>3.29</td>
<td>2.24</td>
<td>1.81</td>
</tr>
</tbody>
</table>

3.2. Bulk-rock composition

X-ray fluorescence (XRF) analyses of major elements of whole rocks were carried out with a Philips PW1040/10 spectrometer in the Centro de Instrumentación (CIC, Universidad de Granada, Spain). The detection limit for major elements was 0.01 wt%, Loss on ignition (LOI) was determined with 0.5 g of powdered sample, first dried at 110°C and then heated to 1000°C for 1 h.

Representative bulk rock compositions were determined for samples 16.16, 16.8, 10.1 following the technique described in Laurent et al. (2018). In order to check the reliability of these estimates the same technique was applied for samples RA-0 and RA-2, and the modelling results were compared with those of obtained by XRF. The estimated bulk rock composition in samples RA-0 and RA-2 and the P-T conditions for garnet are similar within uncertainty. The bulk rock compositions used for modelling are reported in Table 1.

3.3. Quantitative compositional mapping

Two sets of compositional maps for major elements were obtained by EPMA. The first set of map was focused on garnet porphyroblasts and the second on micas in the mineral matrix. The analyses were carried out at the Institute of Geological Sciences, University of Bern, using a JEOL JXA-8200 superprobe instrument. The detailed analytical procedure is given in Lanari et al. (2013, 2014). The analytical conditions were 15 keV accelerating voltage, 100 nA specimen current, and variable dwell times. Dwell times for garnet maps were 100 ms (samples 10.1 and 16.16); 150 ms (sample RA-0); and 200 ms (samples 16.8 and RA-2). Dwell times for white-mica maps were 200 ns (samples RA-2, 10.1 and RA-0) and 160 ms (samples 16.16 and 16.8). The dwell time used in each analytical session was based on the analytical time available and the size of the mapped areas. The sizes of the garnet maps are: 948 × 1228 μm for sample 16.16; 2200 × 2200 μm for sample 16.8; 1800 × 1800 μm for sample RA-0; 770 × 900 μm for sample 10.1; and 800 × 800 μm for sample RA-2. The sizes of mica maps are: 400 × 600 μm for sample RA-2; 2010 × 900 μm for sample 10.1; 800 × 1200 μm for sample RA-0; 1050 × 2100 μm for sample 16.16; 1050 × 2100 μm for sample 16.8. For both datasets, nine elemental maps were acquired in two passes (Si, Mg, Na, Ca, K and Ti, Al, Fe, Mn). Data processing and calibration were performed using XMAPTOOLS 2.6.4 (Lanari et al., 2014, 2019). The analytical standardization was performed using the composition of high-quality spot analyses acquired on the same area (de Andrade et al., 2006) and the advanced procedure described in Lanari et al. (2019).

3.4. Phase equilibria modelling

Following the strategy described in Lanari et al. (2017), P-T conditions of garnet growth were obtained for each successive growth zone of 5 selected porphyroblasts. These grains were carefully selected based on BSE images and point analyses. The selection criteria applied in this study included the size of the porphyroblast, the preservation and absence of retrogression and their low inclusion content. Pixels showing mixing compositions, either along the grain boundaries or around mineral inclusions were filtered out using the BRC function of XMAPTOOLS (Lanari et al., 2019). A representative composition of each growth zone was obtained by averaging pixels from manually selected spatial domains. The corresponding compositions and their analytical uncertainties were exported and are reported in Table 2.

The composition of each successive growth zone was used to model the P-T-X conditions of garnet growth by using the program GeMoO 1.5.3 (Lanari et al., 2017). This program provides a numerical simulation of the garnet evolution based on the composition of the successive growth zones characterized in the compositional maps, combined with an iterative modelling strategy. This model relies on predictions made by Gibbs free energy minimization and the program...
Table 2
Representative mineral compositions of garnet.

<table>
<thead>
<tr>
<th>Sample</th>
<th>16.16</th>
<th>16.8</th>
<th>RA-0</th>
<th>10.1</th>
<th>RA-2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Grt</td>
<td>Gt₁</td>
<td>Gt₂</td>
<td>Gt₃</td>
<td>Gt₁</td>
<td>Gt₂</td>
</tr>
<tr>
<td>SiO₂</td>
<td>38.727</td>
<td>38.419</td>
<td>38.575</td>
<td>37.328</td>
<td>37.595</td>
</tr>
<tr>
<td>MnO</td>
<td>1.879</td>
<td>0.447</td>
<td>0.207</td>
<td>3.990</td>
<td>1.631</td>
</tr>
<tr>
<td>MgO</td>
<td>1.903</td>
<td>3.954</td>
<td>3.836</td>
<td>0.951</td>
<td>1.395</td>
</tr>
<tr>
<td>Na₂O</td>
<td>0.023</td>
<td>0.015</td>
<td>0.014</td>
<td>0.120</td>
<td>0.090</td>
</tr>
<tr>
<td>K₂O</td>
<td>0.003</td>
<td>0.002</td>
<td>0.003</td>
<td>0.015</td>
<td>0.008</td>
</tr>
</tbody>
</table>

Atom site distribution (12 anhydrous oxygen basis)

Sample RA-2 is a fine grained dark mica-schist (Fig. 3a) selected in

Á. Santamaría-López, et al.
Tectonophysics 767 (2019) 128158
metapelites attributed to the Ragua unit consisting of quartz, albite, K-white mica, paragonite, chlorite and garnet, and the accessory phases graphite, epidote, rutile, ilmenite, apatite, titanite and zircon. Most of the garnet porphyroblasts have been partially or completely replaced by chlorite. The preserved garnet grains are relatively small (< 1 mm), and they contain inclusions of quartz, ilmenite, epidote, zircon and titanite.

Sample 10.1 is a fine grained dark mica-schist (Fig. 3b, e) selected in the metaquartzites and metapelites alternations from the lower part of the Calar-Alto unit. The sample is made of quartz, K-white mica, paragonite, chlorite, chloritoid and garnet as main phases, and ilmenite, rutile, epidote and allanite as accessory phases. The size of the garnet porphyroblasts is up to 800 μm, and they show inclusions of quartz (mainly located in the core), graphite, ilmenite, epidote, allanite, zircon and titanite.

Sample RA-0 is a dark-grey mica-schist (Fig. 3c, f) which was selected in the mica-schists from the middle part of the Calar-Alto unit. This sample is formed by quartz, K-white mica, paragonite, chlorite, chloritoid and garnet, and the accessory phases apatite, zircon, rutile, tourmaline, ilmenite and monazite. Garnet porphyroblasts have a size of up to 1 mm and show fractures. The porphyroblasts are slightly deformed and contain numerous inclusions of quartz and in lower amount of graphite, rutile, zircon and apatite.

Sample 16.8 is a grey mica-schist (Fig. 3g) from the upper part of the Calar-Alto unit, showing major phases of quartz, albite, K-white mica, paragonite, chloritoid, chlorite and garnet and accessory phases of graphite, ilmenite, rutile, apatite, epidote, allanite, tourmaline and biotite. Garnet porphyroblasts have a size up to 3 mm. They are highly deformed and show numerous fractures. The inclusions in garnet are quartz, rutile, zircon, biotite and apatite.

Sample 16.16 is a dark grey mica-schist (Fig. 3d) selected in the uppermost levels of mica-schists in the Bédar-Macael unit, containing quartz, albite, K-white mica, paragonite, chloritoid, chlorite and garnet as major phases, and graphite, titanite, biotite, ilmenite, rutile, epidote, apatite and allanite as accessory phases. Garnet porphyroblasts have a size of up to 2 mm and inclusions of quartz, graphite, ilmenite, allanite, rutile and graphite. Snowball and atoll garnets are commonly observed.

The mica-schists pertaining to the Calar-Alto and Bédar-Macael units (samples 10.1, RA-0, 16.8 and 16.16), are comparatively light-coloured than those from the Ragua unit (sample RA-0). Allanite was found in samples 16.8 and 16.16, with comparable features. Allanite size ranges between 40 and 160 μm and the grains are generally located in the mineral matrix. Few of them were observed as inclusion in the outer rim of garnet.

4.2. Microstructures

The studied samples exhibit either two planar fabrics S1 and S2 (samples RA-0, 10.1, 16.8, and 16.16), or a single foliation Sp corresponding to S2 (sample RA-2). For more detailed studies of the main fabrics in the Nevado-Filábride complex the reader is referred to the papers of Booth-Rea et al. (2015), Jabaloy et al. (2015) and Vera (2004).

The first cleavage S1 is parallel to the compositional layering and it was folded during formation of a crenulation cleavage defined as S2. In the sample RA-2 pertaining to the Ragua unit, S1 is not preserved. In the Calar-Alto unit, S1 is preserved in sample 10.1, in which K-white mica and paragonite (up to 100 μm long) form aggregates intergrown by graphite lamellas; in sample RA-0 with K-white mica (up to 300 μm), paragonite (up to 200 μm) and graphite (up to 200 μm); and in sample 16.8 as shown by interbedded thin grains of K-white mica and paragonite (up to 300 μm). In the Bédar-Macael unit the sample 16.16 shows intergrowths of K-white mica (200 μm), paragonite (100 μm), chlorite (50 μm) and fine lamellas of graphite are observed. It has already been proposed that the S1 cleavage formed during the prograde history under greenschist facies conditions (e.g. Martínez-Martínez, 1986; de Jong and Bakker, 1991). Despite possible partial re-equilibrations at higher grade (e.g. Airaghi et al., 2017), local mineral assemblage described before for S1 (when it is visible) support the conclusion that it was formed under greenschist facies conditions.

In sample RA-2, S2 is underlined by thick aggregates of chlorite (up to 600 μm long), plagioclase (250 μm) and intergrowths of K-white mica and paragonite (200 μm). In the Calar-Alto unit, the following mineral phases are observed into S2: in sample 10.1, chlorite (500 μm), chloritoid (100 μm), K-white mica (400 μm) and paragonite (200 μm) showing graphite inclusions; in sample RA-0, K-white mica (up to
400 μm) included by graphite lamellas, paragonite (up to 200 μm), and chlorite (up to 300 μm); in sample 16.8 K-white mica (up to 800 μm), paragonite (up to 300 μm), plagioclase (up to 300 μm) and chlorite (up to 500 μm) which includes K-white mica. In the Bédar-Macael unit sample 16.16 S2 is marked by K-white mica (up to 400 μm), chlorite (up to 100 μm), and paragonite (up to 200 μm) include graphite lamellas. The crenulation-cleavage S2 has been proposed to have formed during exhumation and cooling (Martínez-Martínez, 1986; Platt and Behrmann, 1986; García-Dueñas et al., 1988; Augier et al., 2005b). The local mineral assemblages described above for S2 suggest that S2 developed under greenschist facies conditions in the three units.

Inclusion trails in garnet cores of samples 16.16 and 10.1 record the geometry of the S1 cleavage, whereas the porphyroblasts are deformed by S2 in sample 16.8. These observations suggest that garnet nucleated and grew significantly after S1 and before S2. In sample RA-0, garnet porphyroblasts are undeformed preventing the relative timing to be established, as it is for sample RA-2 in which garnet was replaced by chlorite.

5. Results

5.1. Compositional zoning of garnet

The compositional variability of garnet was characterized via quantitative compositional mapping. The maps of end-member fraction are shown in Fig. 5 for the sample 16.16 from the Bédar-Macael unit, Fig. 6 for samples 16.8, RA-0 and 10.1 from the Calar-Alto unit, and Fig. 7 for the sample RA-2 from Ragua unit. Several growth zones of garnet were defined for each sample based on the compositional zonations (delimited by dashed lines in Figs. 5–7). The limit of each growth zone was set using the grossular map as reference. Smoother transitions in almandine and pyrope maps suggest that the original composition may have been locally affected by diffusion. As these effects are restricted to ~10–40 μm around the limit (see below), it is assumed in the following that the central part of each growth zone was not significantly modified by diffusion allowing the growth composition to be preserved (other examples at similar conditions are given in Chapman et al., 2011). The average composition of each growth zone was determined by integrating the pixel compositions of a domain, whose size varies between 330 and 7200 μm² depending on the thickness of the growth zone. These garnet domains were carefully selected, avoiding mixing pixels, mineral inclusions and contact between zones that could have been affected by diffusion. For comparison among samples, these average compositions were also plotted in a ternary diagram (Fig. 4).

Garnet porphyroblasts in sample 16.16 exhibit a core with typical growth zoning: X_grs and X_sps decrease outward, whereas X_alm and X_prp increase outward from the core to the rim. The average compositions of these growth zones are plotted in a ternary diagram (Fig. 4).
Fig. 6. Quantitative compositional maps of end-member fractions for garnet porphyroblasts of the Calar-Alto unit: a sample 16.8, b sample RA-0, c sample 10.1. Dashed lines delimit the successive growth zones in each garnet. Circles show the location of the domains used for extracting the representative composition of each growth zone. Continuous lines show where compositional profiles reported in Supplementary Material S3 were obtained.
both increase (Fig. 5). The core is surrounded by two narrow rims, which are not thicker than 100 μm and show lower X$_{grs}$ and X$_{sps}$ and higher X$_{alm}$ and X$_{prp}$. A 50 μm thick fracture crosscuts the garnet core and is sealed by garnet with similar composition of the outermost rim (Fig. 5). Three garnet domains were selected in this sample. The first domain Grt1 corresponds to the core of the porphyroblast. The average composition was obtained far away from the fracture (Fig. 5). Two additional garnet domains were selected, Grt2 for the inner rim and Grt3 for the outer rim and the fracture (Fig. 5).

Garnet in sample 16.8 does not exhibit significant variations in Ca, (0.87 ± 0.03 apfu) as shown by the X$_{grs}$ map (Fig. 6a). By contrast, both X$_{alm}$ and X$_{sps}$ maps show lower values in the core (0.61 and 0.04) than in the rim (0.66 and 0.08). This chemical zonation is counterbalanced by a decrease of X$_{grs}$ from 0.09 in the core to 0.01 in the rim. Three garnet domains were selected for modelling: the core Grt1, the mantle Grt2 and the rim Grt3 (Fig. 6a).

Garnet porphyroblast in sample RA-0 shows a more complex zoning pattern including a core surrounded by three rings (Fig. 6b). The innermost ring is up to 300 μm thick, and the outermost is up to 150 μm. Garnet shows a oscillatory pattern in both X$_{alm}$ and X$_{grs}$ with the core and the second ring enriched in X$_{grs}$ and depleted in X$_{alm}$; the first and the outermost ring are both enriched in X$_{alm}$ and depleted in X$_{grs}$. By contrast, X$_{sps}$ gently increases from core (0.03) to rim (0.05), whereas X$_{grs}$ decreases (from 0.08 to 0.01). Based on the Ca map, four garnet domains were selected each corresponding to a single growth zone. Grt1 was defined in the core, Grt2 in the first ring, Grt3 in the second ring and Grt4 in the outermost rim (Fig. 6b). The sharp compositional transition between Grt3 and Grt4 in Ca, Fe and Mg support the assumption that post-growth intragranular diffusion was limited to a distance of < 10–40 μm.

Garnet porphyroblasts in sample 10.1 display three compositional domains delimited by roughly straight boundaries (Fig. 6c): a core surrounded by a mantle 50–80 μm thick, and a rim up to 160 μm thick. The core shows low X$_{alm}$ X$_{grs}$ and X$_{sps}$ (0.62, 0.22 and 0.05 respectively) and high X$_{grs}$ (0.09–0.12). By contrast, the two other domains are enriched in X$_{alm}$ X$_{grs}$ and X$_{sps}$ and depleted in X$_{grs}$. The Mn map shows small-scale diffusion between the core and the first rim, not larger than 15 μm (Fig. 6c). The slight manganese increase between the mantle and the rim from 0.02 to 0.05 suggests garnet resorption. This feature is only observed on the bottom side of the grain around a lobate structure generally interpreted as evidence of garnet resorption (e.g. Giuntoli et al., 2018). Two garnet domains were selected in the core for this sample, Grt1 near the centre and Grt3 in the outer part (Fig. 6c). Two additional garnet domains were chosen, Grt3 for the mantle and Grt4 for the rim (Fig. 6c).

Garnet porphyroblast from sample RA-2 are clustered of mineral inclusion making the zoning pattern more difficult to read and interpret. Nevertheless, several zones were identified including core surrounded by a thin rim (< 100 μm thick). The core shows a growth zoning where X$_{grs}$ and X$_{sps}$ decrease outward (from 0.03 to 0.02 and from 0.12 to 0.04 respectively), and X$_{alm}$ and X$_{sps}$ increase (from 0.59 to 0.69 and from 0.02 to 0.03). The rim is enriched in X$_{alm}$, X$_{grs}$ and X$_{sps}$ (0.67, 0.27, 0.03) and depleted in X$_{grs}$ (0.04). Two garnet compositions were defined for this grain: a core Grt1 and a rim Grt2 (Fig. 7).

As shown in Figs. 6 and 7, samples 16.8, RA-0, 10.1 and RA-2 yielded similar compositional zoning between core and rim. Only garnet from sample 16.16 shows a separate trend characterized a remarkable enrichment in Fe toward the rim (Fig. 5).

As a summary, garnet porphyroblasts of every sample show typical growth zoning in core and mantle with limited post-growth modification by diffusion. These growth zones were thus selected and are used in the following to model the P-T conditions using equilibrium thermodynamics. By contrast, the garnet rims in samples 16.16 and RA-0 show more complex compositional zoning with oscillatory zoning. Although these features suggest non-equilibrium processes an attempt to model the rims with equilibrium thermodynamics is made in the
Fig. 8. Quantitative compositional maps of the Si content (atoms per formula unit, apfu) and XMg (Mg/(Mg$^2^+$ + Fe$^{2+}$)) assuming no ferric iron, Fe$^{3+}$ = 0) in K-white mica from samples 16.16 (a, b), 16.8 (c, d), RA-0 (e, f), 10.1 (g, h) and RA-2 (i, j).

following. The goal of this test is to determine if equilibrium thermo-dynamics remain a reasonable assumption (or if it does not).

5.2. Compositional zoning of K-white mica

The compositional variability of white mica was also characterized via quantitative compositional mapping. The Fig. 8 shows the compositional maps of Si$^{4+}$ (unit apfu) and XMg for the five studied samples. As changes in Si$^{4+}$ content and XMg are symptomatic of changes in pressure and temperature conditions (e.g. Massonne and Schreyer, 1987), different white mica compositional groups (wm1, wm2, etc.; see Fig. 8) were defined. It is important to notice that wm1 systematically shows the higher Si$^{4+}$ content – interpreted to reflect the higher-pressure conditions. The representative composition of each group is reported in Table 3. The structural formulae of white mica were calculated assuming all the iron to be Fe$^{2+}$. Compositional variables other than XMg are in apfu.

Four white mica compositional groups were defined in sample 16.16 (Fig. 8a, b). The higher values for the Si-XMg pair correspond to wm1 (3.38 and 0.74), followed by wm2 (3.29 and 0.72), wm3 (3.18 and 0.53) and wm4 (3.17 and 0.50). The compositions of wm1 and wm2 are preserved in the core of thick flakes (up to 400 μm long) preserved both in S1 and S2 (Fig. 8a, b); whilst lower values correspond to wm4, fine lamellas. The Na content increases from 0.13 between wm1 and wm2 and up to 0.23 in wm4.

In sample 16.8 (Fig. 8c, d), the values for the Si-XMg pairs progressively decrease for the five white mica compositional groups from 3.43 and 0.70 (wm1), 3.31 and 0.65 (wm2), 3.25 and 0.58 (wm3), 3.18 and 0.48 (wm4), down to 3.16 and 0.36 (wm5). In both S1 and S2, the white mica flakes are up to 800 μm long and exhibits cores rich in Si$^{4+}$ and XMg (wm1 and wm2), with depletions (wm3 and wm4) and scarce smaller grains pertaining to wm5. The Na content increases between 0.13 and between wm1 and wm2 and up to 0.23 in wm4.

In sample RA-0 (Fig. 8e, f), high Si-content and XMg values were only observed in wm1 (3.40 and 0.53). Lower values (in particular for the Si$^{4+}$ content) were yielded by wm2 (3.17 and 0.52), wm3 (3.09 and 0.48) and wm4 (3.00 and 0.07), which is a muscovite. This sample mainly preserves wm2 and wm3 as long grains aligned along S2. Scarse wm1 (in grain cores) and wm4 (as thin overgrowths) were observed (Fig. 8e, f). The Na-content increases between wm1 and wm3 (from 0.07 to 0.22), followed by a significant decrease in wm4 (0.14).

Four white mica compositional groups were observed in sample 10.1 (Fig. 8g, h). In general both the Si-content and XMg values decrease from wm1 (3.43 and 0.64), wm2 (3.31 and 0.60), wm3 (3.14 and 0.48), up to wm4 (3.15 and 0.40). The groups of wm1, wm2 and wm3 form thick (150–200 μm) aggregates aligned along S2 (Fig. 8g, h). Few smaller grains of wm4 are observed. The Na-content varies between the four groups increasing from 0.04 (wm1) to 0.17 (wm3), and then decreasing to 0.14 (wm4).

The higher values of Si$^{4+}$ content and XMg in sample RA-2 (Fig. 8i, j) correspond to a few grains of wm1 (3.51 and 0.52). Remarkably lower values are recorded by wm2 (3.21 and 0.38), wm3 (3.11 and 0.35) and wm4 (3.12 and 0.18). The white mica wm2, wm3 and wm4 form remarkably long intergrowths aligned along S2, coating sparse wm1 grains (Fig. 8i, j). A notable increase in the Na content is observed between wm1 (0.06) and wm4 (0.24).

5.3. Phase equilibria

Optimal P-T conditions for garnet were determined using iterative forward models and the program GrtMoo (Lanari et al., 2017). A similar approach was followed for K-white mica using an in-house routine (see § 3.3). The results for all the studied samples are reported in Fig. 9. As the uncertainty envelopes are rarely symmetrical in both P and T for garnet, only the optimal solution is discussed in the following. Uncertainties are shown in Fig. 9 using the method proposed by Lanari et al. (2017) with a tolerance factor on C0 of 1.5.

Garnet nucleation is predicted at 518 °C and 1.8 GPa in sample 16.16 (Grt1 in Fig. 9a); 489 °C and 1.7 GPa in sample 16.8 (Grt1 in Fig. 9b); 511 °C and 0.8 GPa in sample RA-0 (Grt1 in Fig. 9c); 488 °C and 1.8 GPa in 10.1 (Grt1 in Fig. 8d); and 477 °C and 1.6 GPa in RA-2 (Grt1 in Fig. 8e).

The results of garnet mantle and rims of each sample occur at lower P and higher T conditions compared to the conditions of garnet nucleation. In sample 16.16, Grt2 (Fig. 9a) is predicted to have formed at 591 °C and 0.6 GPa, after 4.39 vol% of Grt1 was fractionated from the relictic part of the rock. In sample 16.8, garnet mantle (Grt2 in Fig. 9b) is predicted to have formed at 546 °C and 0.8 GPa (fractionation of 3.11 vol% of Grt1); whilst garnet rim (Grt2 in Fig. 9b) is modelled at 566 °C and 0.7 GPa (fractionation of 3.11 vol% and 3.66 vol% of Grt1 and Grt2 respectively). In sample RA-0 crystallization of Grt2 (Fig. 9c) is modelled at 521 °C and 0.7 GPa (fractionation of 0.57 vol% of Grt1). In sample 10.1, three additional domains of garnet were investigated: the outer core (Grt2 in Fig. 9d) is modelled at 512 °C and 1.5 GPa (fractionation of 0.012 vol% of Grt1) with large asymmetrical uncertainties in both temperature and pressure (Fig. 9d); the mantle (Grt1) at 543 °C and 0.7 GPa (fractionation of 0.01 vol% of Grt1) with large asymmetrical uncertainties in both temperature and pressure (Fig. 9d). The garnet rim in sample RA-2 (Grt2 in Fig. 9e) is modelled at 524 °C and 0.7 GPa (fractionation of 0.03 vol% of Grt1).

In samples 16.16 and RA-0 some of the rims were not satisfactory modelled. In sample 16.16, the outer rim Grt1 exhibits a complex oscillatory zoning (Fig. 5). The lower residual value of 0.082 (C0) was found at 585 °C and 0.6 GPa. In sample RA-0, the inner and outer rims Grt1 and Grt2 also show complex compositional zoning (Fig. 6b). The lower residual values of 0.088 and 0.078 were found at 527 °C and 0.7 GPa for Grt3 and 529 °C and 1.1 GPa for Grt4.

White mica P-T conditions were assigned to three different groups: (1) relict of HP K-white mica (blue ellipses in Fig. 9); (2) K-white mica equilibrated during exhumation and heating up to the temperature peak (green ellipses in Fig. 9); and (3) K-white mica re-equilibrated during cooling (red ellipses in Fig. 9).

In sample 16.16, K-white mica reflect HP conditions of 1.4–1.9 GPa, and 450–520 °C (wm1, wm2 and wm3, Fig. 9a); only a small fraction of K-white mica show lower pressure and higher pressure conditions of 0.7 GPa and ~580 °C (wm4, Fig. 9a).

Sample 16.8 preserves K-white mica relics predicted to be stable at 1.8–2.3 GPa and 450–510 °C (wm3 and wm4, Fig. 9b); Note that wm1 were not successfully modelled in the P-T window. A small fraction of K-white mica were equilibrated close to the temperature peak at 0.7–0.8 GPa and ~590 °C (wm4, Fig. 9b) upon cooling at 0.4–0.7 GPa and 475–555 °C (wm5, Fig. 9b).

Only a few K-white mica compositions suggest HP conditions in sample RA-0 (2.1–2.3 GPa and 450–490 °C; wm1 in Fig. 9c), they otherwise recorded lower pressure and higher temperature conditions between 1.2 and 1.7 GPa (wm2, Fig. 9c) and close to the temperature peak between 0.6 and 1.2 GPa (wm3, Fig. 9c). Post temperature peak K-white mica forms a cluster at 0.4–0.6 GPa and 475–525 °C.

K-white mica in sample 10.1 shows mostly high-pressure conditions of 1.6–2.3 GPa and 450–520 °C (wm1 and wm2 in Fig. 9d). A few compositions suggest equilibration during exhumation at 0.6–1.4 GPa and 505–556 °C (wm3, Fig. 9d), and during cooling at 0.4–0.8 GPa and 450–530 °C (wm4, Fig. 9d).
### Table 3

Representative mineral composition of K-white mica from the analysed samples. $\text{XMg} = \text{Mg}/(\text{Mg}^{2+} + \text{Fe}^{3+})$.

<table>
<thead>
<tr>
<th>Sample</th>
<th>RA1</th>
<th>RA2</th>
</tr>
</thead>
<tbody>
<tr>
<td>19.16</td>
<td>16.16</td>
<td>16.8</td>
</tr>
<tr>
<td>RA-0</td>
<td>50.54</td>
<td>48.56</td>
</tr>
<tr>
<td>RA2</td>
<td>52.49</td>
<td>50.99</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>SiO$_2$</th>
<th>TiO$_2$</th>
<th>Al$_2$O$_3$</th>
<th>FeO</th>
<th>MgO</th>
<th>CaO</th>
<th>Na$_2$O</th>
<th>K$_2$O</th>
</tr>
</thead>
<tbody>
<tr>
<td>RA1</td>
<td>50.973</td>
<td>0.632</td>
<td>29.688</td>
<td>1.809</td>
<td>2.112</td>
<td>0.012</td>
<td>1.035</td>
<td>9.094</td>
</tr>
<tr>
<td>RA2</td>
<td>49.126</td>
<td>0.416</td>
<td>30.983</td>
<td>1.464</td>
<td>2.087</td>
<td>0.014</td>
<td>1.028</td>
<td>9.436</td>
</tr>
<tr>
<td>RA-0</td>
<td>47.056</td>
<td>0.378</td>
<td>32.666</td>
<td>2.126</td>
<td>1.392</td>
<td>0.014</td>
<td>1.172</td>
<td>9.277</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>Al (T2)</th>
<th>Si (T2)</th>
<th>MnO</th>
<th>Fe (M1)</th>
<th>Mg (M1)</th>
<th>Al (M2 M3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>RA1</td>
<td>3.387</td>
<td>1.384</td>
<td>0</td>
<td>0.015</td>
<td>0.015</td>
<td>1.699</td>
</tr>
<tr>
<td>RA2</td>
<td>3.287</td>
<td>1.287</td>
<td>0</td>
<td>0.006</td>
<td>0.006</td>
<td>1.731</td>
</tr>
<tr>
<td>RA-0</td>
<td>3.174</td>
<td>1.187</td>
<td>0</td>
<td>0.011</td>
<td>0.011</td>
<td>1.779</td>
</tr>
</tbody>
</table>

| | XMg | XFe | | | |
|-----|-----|-----|-----|-----|
| RA1 | 0.739 | 0.261 | | | |
| RA2 | 0.717 | 0.283 | | | |
| RA-0| 0.702 | 0.298 | | | |

<table>
<thead>
<tr>
<th></th>
<th>Si$_4$</th>
<th>Al (T2)</th>
<th>Si (T2)</th>
<th>V (M1)</th>
<th>Fe (M1)</th>
<th>Mg (M1)</th>
<th>Al (M2 M3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>RA1</td>
<td>3.387</td>
<td>1.384</td>
<td>1.384</td>
<td>0.992</td>
<td>0.006</td>
<td>0.015</td>
<td>1.699</td>
</tr>
<tr>
<td>RA2</td>
<td>3.287</td>
<td>1.287</td>
<td>1.287</td>
<td>0.982</td>
<td>0.006</td>
<td>0.006</td>
<td>1.731</td>
</tr>
<tr>
<td>RA-0</td>
<td>3.174</td>
<td>1.187</td>
<td>1.187</td>
<td>0.972</td>
<td>0.011</td>
<td>0.011</td>
<td>1.779</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>Sum</th>
<th>Xcel</th>
<th>Xms</th>
<th>Xprl</th>
<th>Xpg</th>
</tr>
</thead>
<tbody>
<tr>
<td>RA1</td>
<td>0.903</td>
<td>0.298</td>
<td>0.476</td>
<td>0.093</td>
<td>0.133</td>
</tr>
<tr>
<td>RA2</td>
<td>0.939</td>
<td>0.259</td>
<td>0.555</td>
<td>0.058</td>
<td>0.128</td>
</tr>
<tr>
<td>RA-0</td>
<td>0.959</td>
<td>0.201</td>
<td>0.61</td>
<td>0.046</td>
<td>0.143</td>
</tr>
</tbody>
</table>

In sample RA-2 the high-pressure K-white mica shows conditions of 2.2–2.4 GPa, 400–495 °C (wm1 in Fig. 9e) and 1.5–1.8 GPa, 450–490 °C (wm2). A large fraction of K-white mica compositions show lower pressure conditions between 0.6 and 1.4 GPa, and higher temperatures of 450–520 °C. Finally, a remarkable population of K-white mica equilibrated during cooling at conditions of 0.4–0.7 GPa and 455–550 °C.

As a summary, in most of the samples (except for RA-0), garnet nucleation is modelled at high-pressure conditions, above ~1.6 GPa and temperatures ranging between 480 and 525 °C (Fig. 9). This HP-LT stage is also largely recorded by K-white mica (blue ellipses in Fig. 9). Garnet in sample RA-0 apparently recorded nucleation at lower pressure conditions below 0.8 GPa at 511 °C (Fig. 9c). By contrast all the following growth stages of garnet, involving mantle and rims, yield lower pressure and slightly higher temperature conditions characteristic of a LP/HT stage. The behaviour of white mica upon exhumation and heating is more complex and sample dependent, but the results are systematically in line with the garnet conditions. Finally it was not possible to model with reasonable residuals the outer garnet rim(s) of samples 16.16 and RA-0.

5.4. Allanite dating

The dating results of allanite for samples 16.16 and 16.8 are summarized in Fig. 10 (see methods in § 3.5, and Burn et al., 2017). The data were plotted in (1) uncorrected Terra-Wasserburg diagrams to estimate the $^{207}\text{Pb}^{206}\text{Pb}$ common lead composition; (2) Th-isochron
diagrams, in which the uncertainty on the common lead fraction ($f_{206}$) was propagated through the age calculation procedure using a Monte-Carlo technique (Burn et al., 2017); and as (3) common lead corrected U ages using the common lead compositions from (1) and (2).

In sample 16.16, the U- and Th- isochrons defined by 21 allanite analyses yield the intercept ages of 12.87 ± 3.31 Ma (2σ, Fig. 10a) and 12.40 ± 4.19 Ma (2σ, Fig. 10b). The corresponding weighted mean U–Pb age is of 12.91 ± 1.10 Ma (2σ). Allanite core and rim analyses return the same age within uncertainty (MSWD = 0.75, see Fig. 10c).

In sample 16.8, the U- and Th- isochrons defined by 20 allanite analyses yield the intercept ages of 14.61 ± 4.28 Ma (2σ, Fig. 10d), and 18.08 ± 5.83 Ma (2σ, Fig. 10e). The corresponding weighted mean U–Pb age is of 14.51 ± 2.01 Ma (2σ, Fig. 10f).
6. Discussion

6.1. Phase equilibria modelling

In this study, garnet and K-white mica were modelled using Gibbs free energy minimization and an iterative approach. The predicted mineral assemblages for the HP stage were not considered to refine the P-T conditions which only rely on the garnet and white mica compositions.

6.1.1. Garnet

This technique applied to model garnet cores (Grt 1) is similar in essence to isopleth thermometry without considering the coexisting assemblage. The main advantages of Grt-Mo are (1) to enable garnet fractionation and (2) resorption in order to compute a suitable reactive bulk composition at each stage and (3) the weighting procedure included in the inversion (Lanari et al., 2017), allowing “optimal” P-T conditions to be derived in a quantitative way (see Fig. 11 in Lanari and Duesterhoeft, 2019 for a comparison between the weighted and unweighted procedure).

The robustness of the model predictions, such as the garnet compositions and the volume fractions, was tested along a profile (the positions are shown in Figs. 5–7). The results are given in Supplementary Material S3. The chemical composition of garnet was extracted using XMAPTools. The distances between each pixel was transformed into a material position and the volume fractions, was tested along a profile (the positions and the volume fractions). The modelled garnet zoning profiles are consistent with the modelled using Gibbs free energy minimizations relying on the assumption of chemical equilibrium.

As presented in Section 5.3, the outer rims of garnet in samples 16.16 and RA-0 yielded high residual values. The observed garnet compositions cannot be modelled with our model based on equilibrium thermodynamics; these are interpreted to reflect disequilibrium processes such as transport-controlled growth and they are excluded from the comparison shown above.

6.1.2. K-white mica

Mineral fractionation and resorption were not considered for white mica modelling as this is known to have little effects on the position of the Si4+ and X-Mg isopleths (Airaghi et al., 2017). K-white mica records a similar trend in all the studied samples with the preservation of a HP generation and partial re-equilibration during exhumation and heating, eventually followed by new growth upon cooling (Fig. 9). The HP generation is observed as cores (e.g. samples 16.16, RA-0 and 10.1) or single-grains (e.g. samples 16.8 and RA2) that are preserved in S2 and S1 (if present, see Fig. 8). This observation shows that white mica re-equilibration occurs by dissolution-precipitation and that the composition cannot be related to D1 and D2 deformations stages. Similar conclusions were reached for amphibolite facies metapelites in the Longmen Shan (Airaghi et al., 2017; Airaghi et al., 2019). The degree of re-equilibration is likely controlled by the availability of fluids driving pseudomorphic replacement (Putnis, 2009; Airaghi et al., 2017; Berger et al., 2017). In the present study, white mica from samples RA-0 and RA-2 were strongly re-equilibrated compared to white mica in samples 16.16, 16.8 and 10.1 that preserve a higher fraction of HP mica. This study shows that the interpretation of white mica thermobarometry in link with deformation can be erroneous if not assisted by quantitative compositional maps.

6.2. Comparison between units

The combination of thermobarometric results obtained for garnet and K-white mica allows the P-T paths of each sample to be reconstructed (Fig. 11). The absolute uncertainty on any P-T estimate is commonly assumed to be close to ± 50 °C and ± 0.2 GPa. However, as the same modelling technique and thermodynamic properties are used, the relative uncertainty is assumed to be much smaller (between 10 and 25 °C, see the uncertainty envelopes in Fig. 9) permitting to perform relative thermobarometry (e.g. Worley and Powell, 2000).

The studied samples exhibit clockwise P-T paths characterized by a HP/LT stage followed by heating during decompression toward a LP/HT stage. The burial conditions however cannot be reconstructed as no relic was found. For the HP/LT stage, a small difference of 0.2 GPa is observed between samples 16.8, RA-0 (Calar-Alto unit) and RA-2 (Ragua unit) (~2.3 GPa), and samples 16.16 (Bédar-Macael unit) and 10.1 (Calar-Alto unit) (~2 GPa). The samples reached mostly the same temperature during the HP/LT stage (~465 °C in 16.8, RA-0 and RA-2; ~455 °C in 16.16; ~475 °C in 10.1). These differences can be related to the uncertainty in the thermodynamic model to predict the conditions of the Si4+ rich cores. Minor differences of < 30 °C are observed between the P-T path during the exhumation and heating stage. Samples 16.16 and 16.8 recorded a higher peak temperature of ~585 °C; this is not recorded by samples RA-0, 10.1 and RA-2 (~535–550 °C). Finally, following the LP/HT stage all samples show a similar cooling trajectory.

It is interesting to note that, despite stark differences in the thermal peak conditions between samples 16.16 (Bédar-Macael unit), 16.8, RA-0 and 10.1 (Calar-Alto unit), the pre-peak P-T trajectories are similar, including for the HP stage. Only sample RA-2 (Ragua unit) evolved at slightly lower temperature during exhumation, but this difference cannot be interpreted as statistically different given the range of uncertainty (see above).

The similarities in the P-T conditions during HP/LT stage and during decompression suggest that the units underwent a similar heating and exhumation history. By contrast, during the LP/HT stage, the temperature was higher in the upper part of the complex (samples 16.16
and 16.8) compared to the lower part (samples 10.1, RA-0 and RA-2).
This result suggests that this inverted metamorphism observed in the
Nevado-Filábride complex was recorded during collision. Other studies
found higher P-T conditions for this stage in the upper part of the
complex than in the lower one (e.g. Augier et al., 2005a, 2005b; Puga
et al., 2000; Behr and Platt, 2012; Booth-Rea et al., 2015) but such
features were not visible in our dataset.

6.3. Petrochronology

Allanite dating can provide age constraints of single metamorphic
stage(s), provided that growth episodes can be correlated to major
phases (Rubatto et al., 2011; Cenki-Tok et al., 2014; Loury et al., 2016;
Engi, 2017). In this study, allanite from samples 16.16 and 16.8, per-
taining to the Bédar-Macael and Calar-Alto units respectively, crystal-
lized during Miocene at 12.91 ± 1.10 Ma and 14.51 ± 2.01 Ma re-
spectively (Fig. 10c, f). In these two samples, allanite is present in the
matrix (Fig. 12) and as inclusion in the outer rim of garnet (Fig. 13). No
age difference was found between these two locations or between the
REE-rich core and the REE-depleted rim of single grains (Fig. 12c). In
both samples, allanite inclusions were found in the garnet rim (Fig. 13a)
as well as along late fractures cross-cutting the previous growth zones.
In Sample 16.16, allanite was found only in the outer rim (Grt 3)
showing oscillatory zoning. These observations show that allanite
formed before or syn Grt 3 and allanite dating provides a maximum age
for the formation of the garnet rim in these samples. These rims cor-
respond to the garnet domains that were not modelled using a chemical
equilibrium model, i.e. Grt 3 in sample 16.16 (see § 5.3). The textural
relationships and the potential presence of reactive fluids to form the
last garnet rims in these samples suggest that the allanite formed during
the LP/HT stage at ca. 13 Ma. Accordingly, the HP/LT stage should take
place earlier as proposed by Augier et al. (2005a). Li and Massonne
(2018) recognized two monazite populations formed at 40.2 ± 1.7 and
24.1 ± 0.8 Ma. According to these authors, a first HP event took place
before the first monazite population formed, whereas the exhumation
and a second P-T loop coincide with the age of the younger population.
These stages are all older than the LP/HT stage recorded by allanite.
The allanite dating results are also in line with Monié et al. (1991) that
suggested an early Miocene HP stage. Other studies reported HP me-
tamorphism occurring between 18 and 13 Ma (López-Sánchez-Vizcaíno
et al., 2001; Platt et al., 2006; Gómez-Pugnaire et al., 2012; Kirchner
et al., 2016) previous or nearly coincident with the allanite data for the
LP/HT stage. The results of de Jong (2003), ~14 and ~17 Ma for the
early stages of exhumation are previous or nearly coincident with the
allanite data.

6.4. Comparison with literature data

The P-T trajectories obtained in this study are compared in the
following sections with the P-T paths proposed in previous studies
(Fig. 14).

6.4.1. Ragua unit

Augier et al. (2005b) proposed a clockwise P-T path for the Ragua
unit (Fig. 14a). These conditions are similar to sample RA-2 except for
wm 1 that indicates higher-pressure conditions (Fig. 9e). As only a few

Fig. 12. BSE images (a, c, e) and optical microscopy microphotographs (b, d, f) of allanite grains from samples 16.16 and 16.8 matrix. Abbreviations: ab (albite), aln (allanite), grt (garnet), pg (paragonite), Kwm (K-white mica), qtz (quartz), rt (rutile).
grains aligned along S2 recorded these conditions, they are easy to miss
without the assistance of compositional maps. For the LP/HT stage, the
temperature reached by sample RA-2 is slightly higher (~25 °C) than
proposed by Augier et al. (2005b) but this remain within the absolute
uncertainty of each method. Li and Massonne (2018) proposed a dif-
f erent P-T path consisting of two successive hairpin P-T loops (path 2 in
Fig. 14a); such pressure cycling is not recorded by white mica in our
dataset. The authors interpreted the second loop as a consequence of
overlying the hot Alpujárride complex. In all the samples investigated
using quantitative compositional mapping, neither garnet, nor white
mica recorded such complex P-T path involving two burial stages (see
Rubatto et al., 2011 and Regis et al., 2014 for other examples of white
mica zoning recording pressure cycling).

6.4.2. Calar-Alto unit

The results of Augier et al. (2005b) for the Calar-Alto unit show an
isothermal decompression at ~550 °C (Fig. 14b). Our white mica data
from samples RA-0, 10.1 and 16.8 rather suggest a non-isothermal
decompression between ca. 2 and 0.6 GPa associated with a tempera-
ture rise of ~100 °C. The LP/HT stage of sample 16.8 is predicted to
occur at temperatures 50 °C higher than Augier et al., (2005b) as al-
ready proposed by Li and Massonne (2018). Such reheating event of
60 °C as proposed by Li and Massonne (2018) is only observed in
sample 16.8 and not associated to a second burial stage (Fig. 14b).

6.4.3. Bédar-Macael unit

In contrast with the two other units, the Bédar-Macael unit appar-
ently recorded a large variety of P-T paths spreading over the entire P-T
range. Bakker et al. (1989), Vissers et al. (1995) and Gómez-Pugnaire
and Fernández-Soler (1987) (Fig. 14c) yielded values for the HP/LT
between 0.5 and 0.7 GPa, significantly lower than the estimates from
our garnet and white mica modelling in sample 16.16. However the
continuous heating during exhumation observed in these P-T paths is in
line with our results. The P-T estimates of Augier et al. (2005a) and
Ruiz-Cruz et al. (2015) (Fig. 14c) have similar pressure conditions for
the HP/LT stage (~1.9 GPa) but at significant higher temperature
conditions (~600 °C). Behr and Platt (2012) and Puga et al. (2002)
(Fig. 14c) proposed hairpin P-T trajectories with peak conditions of
550 °C, 1.4 GPa and 650 °C, 2.1 GPa respectively. In both cases de-
compression occurs together with cooling. Finally, higher temperature
conditions were obtained by López-Sánchez-Vizcaíno et al. (2001) from
ultramafic rocks (Fig. 14c). Such high-temperatures conditions are not
compatible with the absence of partial melting in the metasediments of
this area.

Despite the variety of P-T paths reported in the literature, our P-T
data for both white mica and garnet in sample 16.16 indicate conditions
similar to the other units (Fig. 11). Even though a higher internal de-
gree of complexity cannot be excluded for the Bédar-Macael unit, our
new dataset suggests a very similar metamorphic history for both the
HP/LT and LP/HT stages.

Fig. 13. BSE images of allanite inclusions in garnet porphyroblasts from samples 16.16 (a) and 16.8 (b). Abbreviations: aln (allanite), ap (apatite), grt (garnet), ilm (ilmene), qtz (quartz), rt (rutile), zr (zircon).
Following the LP/HT stage, the retrograde stage follows a linear cooling trajectory. Other proposed P-T paths enter the field of andalusite and/or sillimanite (Augier et al., 2005b, Fig. 14a, b; Bakker et al., 1989; Vissers et al., 1995; Augier et al., 2005a, Fig. 14c) during decompression. However, kyanite is the only Alpine-age aluminosilicate reported in the Nevado-Filábride complex (e.g. Behr and Platt, 2012). With the exception of the sample 16.16, the trajectories presented in this study remain inside the kyanite stability field (Fig. 11).

6.4.4. Summary

The present study brings a new dataset for the Ragua unit, confirming the HP metamorphism proposed by Augier et al. (2005b) and Li and Massonne (2018). However, our data support similar peak-pressure conditions in the three units. The studied samples recorded a similar exhumation path associated to heating (Fig. 11). The thermal peak was recorded at similar depths of ~20 km but with a thermal gradient of 50 °C. This later stage was dated at 13–15 Ma. A higher degree of complexity in the Bédar-Macael unit – maybe involving several subunits with contrasted tectono-metamorphic histories – cannot be excluded from the literature data (see above).

6.5. Tectonic implications

The results of this study suggest that the three units followed similar exhumation trajectories. These results question the tectonic model of two independent complexes proposed by Puga et al. (2002). Despite the similarities in the P-T shapes, the uppermost rocks of the Nevado-Filábride record higher temperature conditions that the lowermost ones during the LP/HT stage. This difference has been attributed to various factors including: the existence of different tectono-metamorphic units or complexes (e.g. Puga et al., 2002); the proximity of the top to a hot hanging wall (e.g. Behr and Platt, 2012); a late-stage thermal overprint (Bakker et al., 1989) involving either shear heating or hot fluid flows.

Fig. 14. Comparison between the P-T paths obtained in the present study and those available in the literature for: a Ragua unit, b Calar-Alto unit, c Bédar-Macael unit.

Fig. 15. Simplified model showing the evolution of the Nevado-Filábride complex during subduction and exhumation. According to the allanite data, exhumation should have started before ~13 Ma.
The upwelling of hot mantle material followed by magmatism (de Jong, 2003). The increase of the maximum temperature conditions toward the top of the series can be explained by the proximity with the mantle wedge (Fig. 15). The linear cooling after the HP/HT stage (Fig. 11) probably coincides with the final emplacement of the Nevado-Filábride complex underneath the Alpujárride complex.

Regarding the age of metamorphism, allanite dating from samples 16.16 (12.91 ± 1.10 Ma) and 16.8 (14.51 ± 2.01 Ma) allowed us to constrain the timing of the LP/HT stage at ca. 13–15 Ma. Cooling rate for retrograde metamorphisms estimated at ~120 °C/Ma, in line with the results of Johnson et al. (1997) of 105–200 °C/Ma. The peak-metamorphism age obtained from allanite is compatible with the existing time frame for the HP stage of 20–13 Ma (e.g. López-Sánchez-Vizcaíno et al., 2001; Gómez-Pugnaire et al., 2012; Platt et al., 2006; Kirchner et al., 2016), and 48–30 Ma (Monié et al., 1991; Augier et al., 2005a; Li and Massonne, 2018). A Miocene age for the HP/HT stage implies a rapid exhumation of the Nevado-Filábride complex as previously proposed (e.g., López-Sánchez-Vizcaíno et al., 2001; de Jong, 2003; Platt et al., 2006; Behr and Platt, 2012; Kirchner et al., 2016), and rapid exhumation rates of ~12 km/My. By contrast, the scenario of an Eocene HP/HT stage involves exhumation rate of 2.4–4.0 km/My. Rapid exhumation and cooling rates have also been proposed for the Western and Central Mediterranean region such as 20–24 km/My (Gebauer et al., 1997) and 36 km/My in Dora Maira Massif (Rubatto and Hermann, 2001); 6–8 km/My for the Edough massif (Northern Africa, Carminati et al., 1998); and 25 km/My for the Beigua Unit (Western Alps, Rubatto and Scambelluri, 2003). In addition, such rapid cooling rates have also been proposed for the overlying Alpujárride complex (e.g. Platt et al., 2005; Platt et al., 2013) together with high cooling rates (350 °C/My, Lonergan and Johnson, 1998; 400 °C/My, Esteban et al., 2004; ~500 °C/My, Zeck, 2004).

Finally, our results question the existence of several tectono-metamorphic units and major discontinuities in the Nevado-Filábride complex. It cannot be excluded that the ductile to ductile-brittle shear zones observed throughout the complex are local discontinuities rather than crustal scale shear zones with km-scale displacements, which would have resulted in the superposition of rocks with different metamorphic histories. In this sense, the similarity of the P-T paths is interpreted here as a similar tectono-metamorphic history even though the units could have been subducted and exhumed sequentially. The similar age range of the LP/HT stage in the Bédar-Macael (sample 16.16) and the upper part of the Calar-Alto (sample 16.8) units also support this conclusion of a similar P-T-t history.

7. Conclusions

Five garnet-bearing mica-schists have been studied to obtain detailed P-T paths for the Alpine metamorphism followed by the Nevado-Filábride complex. The most notable findings in this study are:

- This study confirms the Nevado-Filábride complex recorded an apparent inverted metamorphism, characterized by higher P-T conditions in the upper part.
- The samples pertaining to the previously proposed Ragua, Calar-Alto and Bédar-Macael units followed clockwise P-T trajectories, both characterized by a first HP/HT stage, followed by heating during decomposition leading to the LP/HT stage. The final part of the exhumation took place along a nearly linear cooling following the LP/HT stage.
- During the HP/HT stage, the Bédar-Macael unit reached ~2 GPa and ~520 °C. The temperature difference increased during the decompression up to ~70 °C during the LP/HT stage.
- The HP/HT stage in the Calar-Alto unit reached ~2.0–2.2 GPa and ~470–490 °C. The temperature difference increased during the decompression up to ~50–100 °C during the LP/HT stage.
- The HP/HT event in the Ragua unit was recorded at ~2.2 GPa and ~480 °C. The LP/HT stage reached a temperature of ~535 °C.
- It is remarkable that similarities in the shapes of the P-T paths from the proposed units could imply the existence of a continuous sequence showing an inverted metamorphism, rather than a complex tectonically divided into several tectono-metamorphic units.
- Allanite dating of two samples one from the Bédar-Macael and one from the Calar-Alto units allowed us to interpret the date range of 13–15 Ma as the possible age of the LP/HT stage.

Supplementary data to this article can be found online at https://doi.org/10.1016/j.tecto.2019.06.028.

Acknowledgement

The authors like to express their gratitude to Hans-Joachim Massonne and Romain Augier for their stimulating comments and suggestions and Philippe Agard for his editorial handling. We acknowledge Daniela Rubatto and Jörg Hermann (Institute of Geological Sciences, University of Bern), and Jesús Galindo Zaldívar and Manuel Martínez Martos (Facultad de Ciencias, Universidad de Granada) for their comments and discussion on this work. This study was supported by the Ministerio de Economía y Competitividad (Spain) (Project CGL 2012-31872), and by the project DAMAGE (AIE/FEDER CGL 2016-80687-R).

References
