Petrological and Structural Evolution of High-Grade Metamorphic Rocks from the Floor of the Alboran Sea Basin, Western Mediterranean

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The metamorphic basement beneath the Alboran Sea, a Neogene extensional basin drilled during Ocean Drilling Program (ODP) Leg 161 (Site 976), consists of high-grade pelitic schist overlying migmatitic pelitic gneiss. Inferred assemblages in the high-grade schist evolved from garnet + staurolite + biotite + muscovite + plagioclase (Assemblage 1) to biotite + sillimanite + K-feldspar + plagioclase ± garnet (Assemblage 2) to andalusite + biotite + K-feldspar + plagioclase (Assemblage 3). In the gneiss, which shows abundant migmatitic textures, the present assemblage is biotite + andalusite + sillimanite + muscovite + cordierite + K-feldspar + plagioclase ± garnet. Phase relations and thermobarometric calculations suggest that the high-grade schist experienced decompression accompanied by heating from ~500°C at 10±5 kbar through 600 ± 30°C at 6±7 kbar (Assemblage 1), and thence to 650–700°C at 3±4 kbar (Assemblage 2), followed by cooling through 500–600°C at 2 kbar or less (Assemblage 3). Peak temperature and melting occurred under low-pressure conditions. The widespread disequilibrium and overstepping of metamorphic reactions suggests that decompression was rapid. The P–T evolution is consistent with metamorphism in a late orogenic extensional basin, and suggests the existence of an external source of heat, probably indicating the complete removal of lithospheric mantle beneath the extending region.

KEY WORDS: decompression; extension; melting; P–T path; phase relations

INTRODUCTION

Late orogenic extension of previously thickened continental lithosphere is a widely documented phenomenon (e.g. Dewey, 1988; Malavieille, 1993; Ruppel, 1995), but it is unclear whether the thermal evolution of the lithosphere in this situation differs significantly from that associated with extension in non-orogenic settings. If extension is triggered by removal of part of the lower lithosphere (e.g. England & Houseman, 1989) there may be a detectable rise in temperature during exhumation of rocks (Platt & England, 1994), whereas extension caused by far-field forces is likely to produce isothermal decompression paths followed by cooling (Ruppel et al., 1988; Platt et al., 1999). Decompression accompanied by heating in extensional settings has been demonstrated in the Seward Peninsula in Alaska (Patrick & Lieberman, 1988), the Cordillera Darwin in the southern Andes (Kohn et al., 1993), the Tauern window in the Alps (Selverstone et al., 1984), the Aegean (Buick & Holland, 1989), the High Himalaya (Inger & Harris, 1992), and the Betic Cordillera (Van der Wal & Vissers, 1993). An interesting related question is the significance of crustal melting in extensional settings. In principle, partial melting can occur during decompression through dehydration-melting reactions, by increasing the heat supply through the crust, and by supplying water to rocks at temperatures above their H2O-saturated solidi (e.g.
Thompson & Connolly, 1995). Thermal modelling in extensional regimes, however, suggests that low-pressure granite rocks can only be formed during extension with the participation of an additional heat supply (Thompson & Ridley, 1987; Thompson & Connolly, 1995; Zen, 1995). The purpose of this paper is to document the $P\cdot T$ path followed by the high-grade metamorphic basement beneath a young, late-orogenic extensional basin: the Alboran Sea, in the western Mediterranean. The basement of the Alboran Sea was recently drilled by Ocean Drilling Program (ODP) Leg 161 at Site 976, from which some 250 m of high-grade metamorphic rock was recovered at a depth of nearly 2000 m below sea-level. The reconstruction of the $P\cdot T$ path of these metamorphic rocks is based on the detailed study of the growth zoning patterns of garnet, plagioclase, and staurolite porphyroblasts, and is characterized by a significant drop in pressure (from $>7$ to 3 kbar) accompanied by a rise in temperature (from below 500° to 650–700°C). Textural and mineral composition data suggest widespread disequilibrium conditions during the metamorphism, largely related to the rapid exhumation experienced by the basement during the formation of the Alboran Sea. We also document melting reactions during the decompression $P\cdot T$ path through the muscovite dehydration reaction. The decompression $P\cdot T$ path was followed most probably by a rapid cooling path resulting in the crystallization of residual granitic melts in the andalusite stability field and the preservation of textural and compositional disequilibrium relationships. The analyses of mineral zoning patterns and the complete re-compilation of textural observations and chemical data allow us to refine a previous interpretation of the $P\cdot T$ path experienced by this basement published by Platt et al. (1996).

**GEOLOGICAL SETTING**

The Alboran Sea in the western Mediterranean is one of a number of extensional basins of late Tertiary age that have formed in close association with the compressional orogens that form the Alpine system. These include the Aegean Sea, the Pannonian Basin, the Tyrrenhian Sea, and the Balearic Basin. Like these basins, the Alboran Sea is partly surrounded by an arcuate mountain chain within which contractional deformation continued during the formation of the basin. The Alboran Sea appears to have formed partly or wholly on the extended and thinned remains of a Late Mesozoic to Palaeogene orogen (known as the Alboran Domain) that forms the Internal Zones of the surrounding Betic and Rif mountain chains of the adjacent regions of Spain and Morocco (Fig. 1). During the Neogene evolution of the basin, contraction continued in the External Betic and Rif belts to form the present arcuate structure (Platt & Vissers, 1989; Comas et al., 1992; Garcia-Dueñas et al., 1992).

The formation of these arcuate orogens and associated extensional basins raises important geodynamic questions that remain unresolved. The main contending hypotheses at present are as follows:

1. Extension behind the arcuate thrust belt is induced by forces associated with a convergent plate boundary, for example, the negative buoyancy of the subducting slab of dense oceanic lithosphere (e.g. Royden, 1993; Lonergan & White, 1997).

2. Extension is purely an accidental consequence of the relative movement and rotation of microplates in the plate-boundary zone (e.g. Tapponnier, 1977; Avouac & Tapponnier, 1993).

3. Extension is a result of subcrustal forces induced by the delamination of lithospheric mantle (e.g. Channel & Mareschal, 1989).

4. Extension results from a substantial increase in potential energy caused by the convective removal of thickened lithosphere beneath a region of collisionally thickened crust (England & Houseman, 1989; Platt & England, 1994).

All of these hypotheses except (2) call on processes in the mantle to explain surface phenomena, and hence are difficult to test by direct observation. They do, however, predict significantly different thermal structures for the lithosphere during extension, and should therefore be testable by examining the $P\cdot T$ history of rocks within the lithospheric column. One of the goals of ODP Leg 161 in 1994 was to sample rocks from beneath the Alboran Basin, and to apply their structural and petrological history to test these various hypotheses. This paper reports on the results of post-cruise petrological and microstructural studies on samples from the drill-core obtained on that Leg.

**The Alboran basement at Site 976**

ODP Site 976 is located in the west Alboran Basin on a buried basement high bounded by normal faults (Fig. 1), probably active in Early to Middle Miocene time (Comas et al., 1992; Watts et al., 1993; Comas & Soto, 1999). The crust in the area is probably between 15 and 20 km thick (Hatzfeld et al., 1978). The results of the earlier Deep Sea Drilling Project (DSDP) Site 121 drill hole nearby had previously suggested the presence of a metasedimentary orogen (Kornprobst, 1973; Ryan et al., 1973). At Site 976 the metamorphic basement is covered by a 670 m thick, discontinuous sequence of sediments of Serravallian (Middle Miocene) to Pleistocene age (Shipboard Scientific Party, 1996). The basal contact of the sediments is sharp and characterized by the presence of a variety of basement pebbles, and appears to
be an unconformity, although the presence of breccias down to 40 m below the contact and the significant difference in the depth of this contact between two adjacent holes indicates active faulting during early to middle Miocene time (Shipboard Scientific Party, 1996). The overlying sedimentary sequence therefore places a younger age limit of ~13 Ma on the exhumation of the metamorphic rocks.

The top 124 m of cored basement at Hole 976B (1108 m, water depth) consists of dark grey, quartz–biotite–sillimanite–plagioclase schist, with visible porphyroblasts of garnet and andalusite. This metapelitic sequence has interlayers of calcite and dolomite marble and associated calc-silicate rocks as reaction bands that developed along the metapelitic–metacarbonate contacts. Metacarbonate and associated rocks are particularly abundant in the upper part of the cored section (Fig. 2). This metapelitic sequence overlies banded pelitic gneiss with coarse-grained K-feldspar, cordierite, and andalusite. Much of the gneiss is migmatitic, with centimetre-thick veins and segregations of weakly foliated or unfoliated felsic material (leucosome with a peraluminous granite composition) containing large crystals of cordierite, as well as biotite, andalusite, and sillimanite. The most common textural type for these migmatites is stromatic migmatites [terminology of Mehnert (1968) and McLellan (1983)]. In some instances, the granitic leucosome also occurs as veins cutting across the foliation and in pull-aparts, which locally gives the gneiss a brecciated structure (agmatitic texture). Leucosome geometries largely following pre-existing planes of anisotropy, the low volume of leucosome veins (<20%) with respect to the whole-rock volume, and the common occurrence of mafic selvedges (sillimanite-rich melanosomes) associated with them suggest that the low degree of melting was controlled by a strong, compositional foliation (see Brown
schist, granular cordierite-bearing gneiss, and marble intercalations, commonly cut by leucogranite dykes, characterizes those Alpujarride units underlying the Ronda ultramafic massif some 80 km north of the drill-site (Fig. 1). These rocks reach high-amphibolite to granulite grade conditions with evidence of migmatization and partial melting under relatively low-pressure conditions (<6 kbar) (Loomis, 1972; Westerhof, 1977; Lundeen, 1978; Sánchez-Gómez et al., 1995; Tubia et al., 1997). The Alpujarride Complex as a whole comprises a stack of units with different metamorphic evolutions and with metamorphic grade generally decreasing upward, which have been grouped in terms of their metamorphic and lithologic characteristics [for a classification of the Alpujarride units, see Aldaya et al. (1979), Tubia et al. (1992) and Azanén et al. (1994)]. The inferred stratigraphic sequence in all the Alpujarride units consists of (from bottom to top): graphite-rich schists and metapsammites of presumed pre-Permian age (up to 2000 m), fine-grained schists (or phyllites) of probable Permian–Triassic age (up to 600 m), and carbonate rocks (locally marbles; up to 800 m) of Middle and Late Triassic age (Kozur et al., 1974; Delgado et al., 1981). Compared with this sequence, the basement section on Site 976 resembles the lower metapelitic sequence of the Alpujarride units, of presumable Palaeozoic age.

Recovery 105R
Core 75R
80R
85R
90R
95R
100R

Site 976B Site 976E

Recovery Core Host lithology

12R
13R
14R
15R
16R
17R
18R
19R
20R
21R
22R
23R
24R
25R
26R
27R
28R

High-grade schist
Gneiss
Marble & calc-silicate rock
Leucogranite
Mid-Miocene sediments
Selected sample

Fig. 2. Schematic lithological columns of the metamorphic basement cored in ODP Hole 976B and Hole 976E [for more details, see Shipboard Scientific Party (1996)]. Recovered intervals are shown in black in the recovery column. ▲ location of samples discussed in the text.

et al., 1995; Vigneresse et al., 1996). Major and trace element contents of migmatite and pelite gneiss are similar to high-grade schist composition trends (Spada & Prosser, 1999) suggesting a local source and low mobility of melts during migmatite formation. Therefore, these leucosomes represent metatexite migmatites [following Brown (1973) and Sawyer (1996)].

Both rock units are cut by peraluminous, leucogranitic dykes, as well as numerous zones of brittle fault gouge and breccia. The high-grade schist and the gneiss show distinctive differences in their petrological and structural evolution, and at Hole 976B are probably separated by a significant tectonic boundary, represented by a thick band (up to 20 m thick) of fault rocks (consolidated clay breccia, fault gouges, and cataclastic rocks) (Comas & Soto, 1999).

The lithological association and metamorphic characteristics at the site broadly resemble those within parts of the Alpujarride Complex of the adjacent Betic Cordillera [and its counterparts the Sebtides in the Rif]. A sequence formed by biotite + K-feldspar + sillimanite-rich pelitic schist, granular cordierite-bearing gneiss, and marble intercalations, commonly cut by leucogranite dykes, characterizes those Alpujarride units underlying the Ronda ultramafic massif some 80 km north of the drill-site (Fig. 1). These rocks reach high-amphibolite to granulite grade conditions with evidence of migmatization and partial melting under relatively low-pressure conditions (<6 kbar) (Loomis, 1972; Westerhof, 1977; Lundeen, 1978; Sánchez-Gómez et al., 1995; Tubia et al., 1997). The Alpujarride Complex as a whole comprises a stack of units with different metamorphic evolutions and with metamorphic grade generally decreasing upward, which have been grouped in terms of their metamorphic and lithologic characteristics [for a classification of the Alpujarride units, see Aldaya et al. (1979), Tubia et al. (1992) and Azanén et al. (1994)]. The inferred stratigraphic sequence in all the Alpujarride units consists of (from bottom to top): graphite-rich schists and metapsammites of presumed pre-Permian age (up to 2000 m), fine-grained schists (or phyllites) of probable Permian–Triassic age (up to 600 m), and carbonate rocks (locally marbles; up to 800 m) of Middle and Late Triassic age (Kozur et al., 1974; Delgado et al., 1981). Compared with this sequence, the basement section on Site 976 resembles the lower metapelitic sequence of the Alpujarride units, of presumable Palaeozoic age.

MINERAL ASSEMBLAGES AND DUCTILE FABRICS
The bulk of the high-grade schist unit consists of dark graphitic quartz–biotite–sillimanite–plagioclase schist, but it contains interlayers a few millimetres to centimetres thick with differing compositions (Table 1), including garnet- and staurolite-rich layers (staurolite-bearing schist) with an Fe-rich bulk composition; layers with fibrolite [sillimanite crystals not exceeding 10 µm in minimum dimensions are considered to be fibrolite, following Kerrick & Speer (1988)] and biotite without plagioclase (biotite-rich schist); and corundum-bearing layers lacking quartz and staurolite, with a more aluminous and silica-poor bulk composition (corundum-bearing schist). Corundum-bearing schists are usually found near calc-silicate reaction bands. Quartz, plagioclase, biotite, and Al-silicates (andalusite and sillimanite) are always present as major components of the staurolite-bearing high-grade schist. Apatite, ilmenite, zircon, pyrite, and pyrrhotite (in late veins and as a transformation product of pyrite) are accessory minerals. Tourmaline is a major component in some of the corundum-bearing schists. Chlorite, muscovite, and haematite are retrogressive minerals. Kyanite has not been observed in any of the high-grade schist samples.
### Table 1: Distribution of metamorphic minerals in selected samples of the Alboran Basement

<table>
<thead>
<tr>
<th>Rock type</th>
<th>Sample</th>
<th>Quotation</th>
<th>Location: Leg-Hole-Core-Section, cm</th>
<th>Qtz</th>
<th>Bt</th>
<th>Ms</th>
<th>Chl</th>
<th>Sil</th>
<th>And</th>
<th>St</th>
<th>Grt</th>
<th>Crd</th>
<th>Kfs</th>
<th>Pl</th>
<th>Crn</th>
<th>Others</th>
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<tr>
<td>St-bearing HGS</td>
<td>21</td>
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<td>Ap, Tur, Py, Gr</td>
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<td>76B</td>
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<td>Ap, Ilm, Py, Gr</td>
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<tr>
<td>Crn-bearing HGS</td>
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<td>Ilm, Py, Po, Gr</td>
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<td>Ilm, Tur, Py, Gr</td>
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<tr>
<td>Bt-rich HGS</td>
<td>61</td>
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<td>Ilm, Gr</td>
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<td><strong>Gneiss</strong></td>
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<td>Grt-bearing gneiss</td>
<td>53</td>
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<td>54</td>
<td>161-976B-98R-1, 96-101 cm</td>
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<td>Pelitic gneiss</td>
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<td>161-976B-106R-1, 92-94 cm</td>
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<td>Tur, Grp</td>
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<td>Migmatite gneiss</td>
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<td>SOT</td>
<td>161-976B-95R-1, 69-71 cm</td>
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<td><strong>Granite</strong></td>
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<td>43</td>
<td>161-976B-82R-1, 93-98 cm</td>
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<td>77</td>
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<td>Ap, Tur</td>
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</table>

Mineral abbreviations from Kretz (1983). Textural symbols for minerals: relict (☐), primary (☐), and retrogressive (●). For garnet, the table also shows the occurrence of garnet I and garnet II porphyroblasts.
The pelitic gneiss consists of medium-grey biotite-bearing gneiss, with K-feldspar, plagioclase, sillimanite, andalusite, and cordierite porphyroblasts. Tourmaline, apatite, zircon, magnetite, Ti-magnetite, and ilmenite are accessory components. Muscovite is locally present, particularly in veins, migmatitic segregations, and as reaction products in andalusite and cordierite porphyroblasts. Idiomorphic garnet porphyroblasts (up to 500 μm) occur in some samples. Millimetre-long lenses of felsic material (with a peraluminous granite composition), as well as broader-scale compositional layering, mineral orientation, and abundant quartz-tourmaline and granite veins, define the main metamorphic foliation. Much of the gneiss is migmatitic, with centimetre-thick veins and segregations of weakly foliated or unfoliated leucosomes containing large crystals of cordierite, as well as biotite, andalusite, shreds and wisps of fibrolite, magnetite, and tourmaline. Zoned tourmaline porphyroblasts are always present as a minor component.

The AFM[Ti] assemblages (Table 1) in the high-grade schist and gneiss are [mineral abbreviations after Kretz, 1983]:

- Bt-rich high-grade schist: Bt + Gt + Fib (fibrolite) ± And + Kfs + Pl + Ilm + Qtz;
- St-bearing high-grade schist: St + Bt + Gt + Fib + And + Kfs + Pl + Ilm + Spl ± Rt + Qtz;
- Crn-bearing high-grade schist: Bt ± Gt + Fib + And + Crn + Kfs + Pl + Ilm;
- Gneiss: Bt ± Gt + Crd + Fib + And + Kfs + Pl + Ilm + Mag + Qtz.

As reaction textures among the AFM mineral phases, and in particular the Al-silicates, are always present in the high-grade schist, the above-mentioned mineral assemblages should be considered disequilibrium assemblages. Nevertheless, it is likely that the high-grade schist assemblage with St + Bt + Gt + Sil + And ± Kfs reflects lower temperature than the Bt + Crd + Sil + And + Kfs assemblage in the gneiss.

**Ductile fabrics**

Two systematically identifiable sets of ductile fabrics and structures can be distinguished in the high-grade schist, referred to as Di and D2. The earlier fabric, S1, is characterized by compositional layering, transposed quartz veins, 0.5–2 mm scale quartz-rich and biotite-rich laminae and oriented biotite, Garnet, staurolite, and early plagioclase porphyroblasts have grown over S1 and include it as graphite trails.

S1 has been intensively affected by 1–50 mm scale D2 folds. These are commonly tight and asymmetrical but may also be symmetrical. The compositional layering and quartz-biotite laminae that define S1 are considerably thickened in the hinges of these folds relative to the limbs. Biotite is recrystallized in the fold hinges, and tends to lie parallel to the axial surface, as do elongated mats of fibrolite, the two together defining a weak axial planar fabric, S2. In much of the high-grade schist D2 has been so strong that the main foliation is in fact a composite fabric formed by the transposition and modification of S1. D2 was followed by static growth of second-stage garnet, and subsequently by andalusite, plagioclase, and fine-grained K-feldspar intergrown with quartz.

We have observed no clear transition between the fabrics in the high-grade schist and that in the gneiss, so that the age of the gneissic foliation in relation to the deformational history of the schist is uncertain. On the basis of the textural relationships discussed below, we believe that it is a composite foliation, which may have formed initially at about the same time as D2 in the high-grade schist, but was subsequently heavily modified by compaction or further deformation during partial melting.

**MODEL P–T–XFe PHASE RELATIONS IN THE KFMASH SYSTEM**

The phase relations predicted for the simple system K2O–FeO–MgO–Al2O3–SiO2–H2O (KFMASH) along two contrasting decompression P–T paths are represented in the pseudobinary P–XFe sections of Fig. 3. The calculations were made using the Gibbs method (Spear et al., 1982; Spear, 1988) and the thermodynamic database of Berman (1988) modified by Spear & Cheney (1989). The two paths are those we consider the most probable alternative histories for the Alboran basement, considering the sequence of assemblages we observe and the thermobarometric results presented later in this paper. The two paths we investigate are (1) isothermal decompression at T = 675°C for the high-grade schist and at T = 725°C for the gneiss, followed by cooling at P < 3.5 kbar; and (2) a decompression path accompanied by a modest rise in temperature, reaching 675°C at 3.5 kbar for the high-grade schist, followed by cooling.

In view of the Fe-rich bulk-rock composition (see Fig. 4) of the high-grade schist and gneissic rocks of the Alboran basement (average XFe = 0.76; range 0.72–0.89), the P–T–XFe phase relations can be considered in terms of the relations between garnet, staurolite, biotite, Al-silicate (Als; sillimanite and andalusite), muscovite or K-feldspar, chlorite, quartz, and fluid H2O. Chlorite will be also considered tentatively in the rising T path because it may have participated in some of the reactions along this path.

The phase relations along both paths are described in terms of the following divariant reactions [which are labelled using the absent-phase notation of Zen (1966)]:
Fig. 3. Pseudobinary \( P-X_{Fe} \) sections and AFM diagrams (Thompson, 1957) showing the phase relations in the model KFMASH system, along two isothermal decompression \( P-T \) paths [at 675°C in (a), and 725°C in (b)] and along a decompression \( P-T \) path accompanied by a rise in \( T \), reaching 675°C at 3.5 kbar (c). These \( P-T \) paths are followed by cooling at low \( P \) calculated assuming a linear gradient [from 675°C and 4 kbar in (a), from 725°C and 3 kbar in (b), and from 660°C and 3 kbar in (c)]. The start of the cooling path is marked by wavy double lines. Projection points of AFM diagrams are pure components: quartz, muscovite (with the label: +), andalusite, garnet (with the label: +), and K-feldspar (with the label: + Kfs), and H2O. The calculations were done using the Gibbs method and the thermodynamic database of Berman (1988) modified by Spear & Cheney (1989). The starting conditions for the calculations are defined by the invariant point [Crd, Chl, Th] of Spear & Cheney (1989). Labelled regions, usually in shading, are the divariant assemblages described in the text, and horizontal lines correspond to KFMASH univariant reactions in Figs 19–21. Numbers in parentheses refer to reactions described in the text. Broken lines show the average bulk-rock \( X_{Fe} \) value (0.75) of the metapelites of the Alboran basement.

\[
\begin{align*}
\text{St} + \text{Ms} + \text{Qtz} & \Rightarrow \text{Grt} + \text{Als} + \text{H}_2\text{O} \quad \text{(1)} \\
\text{Grt} + \text{Als} + \text{H}_2\text{O} & \Rightarrow \text{[Bt]} + \text{St} + \text{Qtz} \quad \text{(2)} \\
\text{Grt} + \text{Ms} + \text{H}_2\text{O} & \Rightarrow \text{[Als]} + \text{St} + \text{Bt} + \text{Qtz} \quad \text{(3a)}
\end{align*}
\]

These reactions are operative up to the breakdown of staurolite by the univariant reaction [terminal reaction in the sense of Thompson (1976a)]

\[
\text{St} + \text{Ms} + \text{Qtz} = \text{Grt} + \text{Bt} + \text{Sil} + \text{H}_2\text{O} \quad \text{(4)}
\]

After this, the next reaction on the decompression path is the divariant reaction

\[
\text{Grt} + \text{Ms} + \text{H}_2\text{O} \Rightarrow \text{[Bt]} + \text{Sil} + \text{Qtz} \quad \text{(5a)}
\]

and its equivalent in which muscovite is replaced by K-feldspar,

\[
\text{Grt} + \text{Kfs} + \text{H}_2\text{O} \Rightarrow \text{[Bt]} + \text{Sil} + \text{Qtz} \quad \text{(5b)}
\]

after the muscovite-breakdown univariant reaction

\[
\text{Ms} + \text{Qtz} = \text{Sil} + \text{Kfs} + \text{H}_2\text{O} \quad \text{or L} \quad \text{(6a)}
\]

At 725°C the modelled decompression path intersects the cordierite-forming (non-terminal) reaction

\[
\text{Bt} + \text{Sil} + \text{Qtz} = \text{Grt} + \text{Crd} + \text{Kfs} + \text{H}_2\text{O} \quad \text{(7)}
\]

determines the generation at lower pressures [see fig 3B of Thompson (1976a)] of the following divariant equilibria:

\[
\begin{align*}
\text{Grt} + \text{Kfs} + \text{H}_2\text{O} & \Rightarrow \text{[Als]} + \text{Bt} + \text{Crd} + \text{Qtz} \quad \text{(8a)} \\
\text{Grt} + \text{Sil} + \text{Qtz} + \text{H}_2\text{O} & \Rightarrow \text{[Bt]} \quad \text{Crd.} \quad \text{(9)}
\end{align*}
\]

The \( P-T-X_{Fe} \) phase relations depicted in Fig. 3 reveal significant differences in composition and modal proportion of the phases according to the \( P-T \) path followed. For a fixed bulk \( X_{Fe} \) value of 0.70–0.75 (a typical value for metapelites and gneisses in the Alboran basement) these are as follows:

(1) The reaction history along the isothermal \( P-T \) path produces garnet consumption by triggering progressively the divariant reactions [Bt] (2) and [Als] (3a), garnet growth by the univariant reaction (4), and new episode...
of garnet consumption by the divariant equilibria [St, H$_2$O] (5a) and [St] (5b). This reaction history will produce progressively the AFM assemblages Grt + St + Ky, Grt + St + Bt and Grt + Bt + Sil + Ms, and Grt + Bt + Sil + Kfs, which correspond to the three-phase fields depicted in the AFM diagrams of Fig. 3a. Along this path, all the phases become more Fe rich during the decompression. In those rocks with slightly higher Fe contents ($X_F \sim 0.8$), a limited consumption of St could occur just before the St-breakdown reaction (4), by the divariant reaction [Bt] (2), resulting in the growth of fibrolite and garnet with decreasing $X_F$.

2) At 725°C the reaction history for an isothermal $P$-$T$ path differs only at low-$P$ conditions ($P < 4$ kbar) after the Crd-forming reaction (7) (Fig. 3b). This path is characterized by the consumption of garnet, K-feldspar, and sillimanite by the divariant reactions [Als] (8a) and [Bt] (9) producing an Fe enrichment in biotite and a rapid growth of cordierite with a constant $X_F$ ratio. Both divariant reactions (8a) and (9) involve rehydration.

3) Along the rising $T$ decompression path (Fig. 3c), the reaction history and chemical composition of the phases change abruptly when the Ky-Sil transition is crossed. At $P > 6.5$ kbar the phases become more Mg rich and the assemblage Grt + Ms is produced at expense of St + Bt by the divariant equilibria [Als] (3a). After the Ky-Sil transition, all the phases experience a small increase in the $X_F$ ratio and the reversal of the reaction [Als] (3a) determines a limited garnet resorption and a fairly constant composition of the phases garnet, staurolite, and biotite. The subsequent reaction history, from the St-breakdown reaction (4) towards lower $P$, the phase relations are similar to those depicted along the isothermal $P$-$T$ path. Finally, it is important to note that, because of the steep slope of the divariant reaction [St] (5b), if $T$ rises below the Ms-breakdown reaction (6a), garnet will grow and become more Mg rich (see Fig. 3a and c).

The main differences between the two $P$-$T$ paths therefore lie at pressures >5–6 kbar, where staurolite is stable, and also at pressures below the muscovite breakdown curve. $P$-$T$ estimates from thermobarometry, reaction histories deduced by textural analysis, and chemical variation of the phases will be critical for distinguishing between the two metamorphic evolutions.

**TEXTURES AND MINERAL CHEMISTRY IN THE HIGH-GRADE SCHIST**

To characterize the mineral chemistry and its variation with textural position among the different rock types described, we have selected 15 samples for microprobe analyses (see Fig. 2 for their location in the basement section; Table 1). The principal textures among the main mineral phases and their relationships with the different ductile fabrics in the high-grade schist are summarized in Fig. 5.

---

**Fig. 4.** Bulk-rock composition of high-grade schist (a, b) and gneissic rocks (c) projected in AFM diagrams [Thompson (1957); with $F = Fe + Mn$]. Projection points are pure components: in (a), quartz, muscovite, albite, anorthite, ilmenite, apatite, and H$_2$O. A corundum-bearing schist has also been projected for comparison, although it should be noted that this projection is actually invalid as the phases do not coexist with quartz. AFM diagrams show the calculated topologies [with the thermodynamic database of Berman (1988), modified by Spear & Cheney (1989)] in the model KFMASH system at intermediate-$P$ and low-$P$ conditions (b, c), which are representative of the $P$-$T$ conditions achieved by these rocks during their metamorphic evolution (a) represents Assemblage 1 in the high-grade schist; (b) Assemblage 2 at the end of $D_3$ in the high-grade schist, and (c) the main assemblage in the gneiss (see Fig. 1c).
Fig. 5. Selected textural relationships in the high-grade schist [mineral abbreviations after Kretz (1983); and Fib, fibrolite]. (a) $D_2$ fold in a biotite-rich high-grade schist. Sillimanite and biotite define $S_2$ parallel to the axial surface of the fold. Square shows the location of (b). Sample 161-976E-23R-2, 118–123 cm (biotite-rich high-grade schist). Photomicrograph, plane-polarized light. (b) Detail of a $D_2$ fold hinge with plagioclase porphyroblasts showing two stages of growth. Plagioclase has a graphite-rich core (plagioclase I) indicating that it grew statically over an early foliation ($S_1$), and was subsequently rotated during $D_2$. Plagioclase rims usually contain inclusions of biotite and sillimanite parallel to the external $S_2$ foliation, indicating late, post-$D_2$ growth (plagioclase II). Sample 161-976E-23R-2, 118–123 cm (biotite-rich high-grade schist). Photomicrograph, plane-polarized light. (c) Non-transformed, elongated staurolite porphyroblasts with straight inclusion trails of graphite, ilmenite, and rutile defining $S_1$. This foliation is oblique to the external $S_2$ fabric, defined by oriented biotite, plagioclase, and sillimanite. Sample 28 (161-976B-76R-1, 45–49 cm; staurolite-bearing high-grade schist). Photomicrograph, plane-polarized light. (d) Partially corroded garnet I porphyroblast core (Grt I) surrounded by a sub-idiomorphic overgrowth (Grt II) with graphite, ilmenite, and quartz inclusions trails parallel to the external foliation $S_2$. Arrow points to the irregular, sharp boundary between the two garnet zones. Sample 161-976E-14R-1, 45–50 cm (staurolite-bearing high-grade schist). Photomicrograph, plane-polarized light. (e) Plagioclase and K-feldspar intergrowths included in a garnet II porphyroblast (see location in Fig. 18b). Back-scattered electron image. Conditions for image acquisition: 512 × 512 pixels. Sample 28 (161-976B-76R-1, 45–49 cm; staurolite-bearing high-grade schist). (f) Relict staurolite preserved in a large andalusite porphyroblast. Staurolite is partially transformed to hessanite (Spl). Andalusite grows over the $S_2$ foliation in the matrix, which is defined by a fine intergrowth of fibrolite and K-feldspar. (Notice the minor occurrence of plagioclase surrounding the K-feldspar.) Back-scattered electron image. Conditions for image acquisition: 1024 × 1024 pixels. Sample 22 (161-976B-74X-1, 30–34 cm; staurolite-bearing high-grade schist). (g) Andalusite showing apparently resorbed outline, included in K-feldspar. Sample 161-976B-83R-1, 58–62 cm (staurolite-bearing high-grade schist). Photomicrograph, crossed polars.
The samples were analysed using the electron microprobe Cameca SX50 at the University of Granada. An acceleration voltage of 15 kV, beam current of 20 nA, and beam diameter of 8 μm were used. Standards were simple synthetic oxides (Al₂O₃, Fe₂O₃, Cr₂O₃, MnTiO₄, MgO), natural silicates (albite, orthoclase, and wollastonite), and natural sphalerite. Data were reduced using the PAP procedure supplied by the manufacturers (Pouchou & Pichoir, 1985). A complete dataset of analyses accompanying the paper can be found in the *Journal of Petrology* World Wide Web site.

**Andalusite and sillimanite**

Sillimanite occurs as elongated fibrolitic mts, and occasionally as larger prismatic porphyroblasts. Both textural types of sillimanite define the S₃ planar fabric. Fibrolite in the matrix forms aggregates with sub-microscopic intergrowths of K-feldspar and plagioclase, and these aggregates are commonly intergrown with biotite (Fig. 5b and f). Andalusite forms large randomly oriented porphyroblasts (up to 2–3 cm long) that grew post-kinetically with respect to S₂. It commonly contains oriented inclusions of ilmenite, biotite, and fibrolite parallel to S₂, and corroded staurolite porphyroblasts partly transformed to andalusite and minor spinel (Fig. 5f). A remarkable textural feature of andalusite is that it commonly shows irregular outlines interlayering with plagioclase and K-feldspar, and locally occurs as rounded or irregular blebs included in these minerals (Fig. 5g). Andalusite is also locally altered to randomly oriented white mica.

The content of ‘impurity’ cations (M = Fe³⁺, Ti, Cr, Mg, and Mn; X₀MAlSiO₅ is the sum of the ‘impurity’ cations) confined to the octahedral site in coexisting prismatic sillimanite and andalusite is significantly lower in the corundum-bearing high-grade schist than in the other metapelites and gneiss (staurolite-bearing schist: average X₀MAlSiO₅ = 0.013; corundum-bearing schist: average X₀MAlSiO₅ = 0.007; gneiss: average X₀MAlSiO₅ = 0.048). This probably indicates a larger P-T area of coexistence of the two polymorphs in the gneissic rocks (Kerrick & Spear, 1988), which may help explain some of the textural observations.

**Garnet**

We have identified two distinct families of garnet porphyroblasts (250 μm–5 mm in diameter) by considering their textural relationships with S₂ [see also Shipboard Scientific Party (1996)]. Garnet I usually has inclusion trails of graphite that define an internal foliation (S₃) oblique to the external S₂ foliation. This internal foliation is straight in the central part of the crystals, suggesting interkinematic growth between D₁ and D₂ (Fig. 5d). The porphyroblasts have inclusions of rutile, ilmenite, biotite, and early plagioclase (near the rims), and some are partly corroded to andalusite, plagioclase, K-feldspar, and biotite (Fig. 6), or to fine-grained aggregates of chlorite, muscovite, Na-rich plagioclase, and quartz.

Many garnet I porphyroblasts have an inclusion-poor outer zone with idiomorphic outlines (garnet II). This outermost rim usually has fibrolite inclusions that are continuous with the matrix. In some grains the boundary between the two zones is irregular and shows evidence of resorption (Fig. 5d). Garnet II grains without garnet I cores also occur in the matrix as large (up to 1 cm long), idiomorphic porphyroblasts with abundant inclusions parallel to the external S₂ foliation. Garnet II includes fibrolite, biotite, plagioclase and K-feldspar (Fig. 5e), tourmaline, and locally staurolite, chlorite, and corundum.

Garnet from corundum-bearing schist is characterized by increasing contents in almandine and pyrope and decreasing contents in spessartine and Fe/(Fe + Mg) from core to rims [e.g. in Fig. 7b: Alm₀·₄₈₋₀·₆₅, Pyr₀·₅₁₋₀·₆₀, Sp₀·₁₂₋₀·₁₅]. Grossular always decreases towards the rims (Gr₀·₂₀₋₀·₁₅). The two garnet types distinguished in the staurolite-bearing high-grade schist have a distinctive zoning pattern (Fig. 7a–d). Garnet I porphyroblasts have a normal growth zoning pattern indicating prograde growth: a strongly modulated bell-shaped spessartine profile, and increasing almandine and pyrope, from core to rim. There is also an overall gradual decrease in Fe/(Mg + Fe) from core (average 0·97) to rim (average 0·92). The inner cores of some of the larger garnet I porphyroblasts are depleted in the spessartine and grossular components, and the Fe/(Fe + Mg) ratio, and show a coupled increase in almandine and pyrope (Figs 7b and 8). This type of inner core could be the remains of an older relict garnet developed under prograde conditions in the presence of chlorite, which preferentially partitioned Mn (e.g. Spear, 1993). Incomplete or truncated zoning patterns similar to those observed in garnet I are present in grains included in andalusite porphyroblasts (Fig. 7g). Some of the garnet I rims show evidence of resorption, with a discontinuity in the Ca, Fe, and Mg zoning patterns (Figs 8 and 9).

Garnet II rims and grains show fairly constant composition (Fig. 10), similar to the rims of garnet I (Figs 7f and 8). Flat zoning patterns in this garnet may be a result of rapid growth (Holm & Selverstone, 1990), rather than diffusive homogenization of originally zoned garnets (e.g. Tracy et al., 1976; Yardley, 1977), because flat zoning is also observed in larger (>1·5 mm) porphyroblasts [intragranular diffusion is usually enhanced in garnets with radii <1 mm; Florence & Spear (1991); and Spear (1991)] and because, when it occurs, it affects only the outermost rims of garnet I (<100 μm, e.g. Fig. 7a). Some
small garnet II porphyroblasts show sharp increases in spessartine, Fe/(Fe + Mg) ratio, and grossular at the rims (Fig. 7c), but these increases are rare in the outermost rims of large garnet II porphyroblasts. This contrasting behaviour suggests that garnet grain size has a large effect on volume diffusion processes (Florence & Spear, 1991; Spear, 1991). There is a sharp inflection of the Fe/(Fe + Mg) and grossular contents from garnet I rims to garnet II overgrowths, suggesting that these two garnets have grown by different reactions or were separated by an inflection in the P-T path (e.g. Rumble & Finnerty, 1974; Karabinos, 1984; Spear et al., 1984; Whitney & Ghent, 1993).

### Plagioclase and K-feldspar

Plagioclase appears in the matrix as large ovoid porphyroblasts (up to 5–7 mm long), showing two main growth stages. The first, plagioclase I, is represented by cores densely crowded with included graphite, defining a straight internal foliation. This texture indicates interkinematic growth between D1 and D2, but locally plagioclase I has grown over D2 microfolds (early syn-D2 growth). Plagioclase II forms inclusion-free rims, as well as new clear porphyroblasts associated with K-feldspar and biotite (Fig. 5b). These may contain inclusions of fibrolite parallel to S2, indicating growth at a late stage (post-D2). In some samples plagioclase II contains corroded relics of staurolite and andalusite.

K-feldspar occurs in several textural settings: finely intergrown within fibrolite mats (Fig. 5f); interstitially with granular quartz in veins (cutting across the S2 foliation); as thin rims surrounding zoned plagioclase porphyroblasts; and as large irregular poikiloblasts. The last commonly contain abundant rounded inclusions of quartz and biotite, and may be intergrown with plagioclase. K-feldspar occasionally occurs as intergrowths with plagioclase within garnet II porphyroblasts (Fig. 5c).

The compositional range of plagioclase shows remarkable variations, both between rock types and within single samples, depending on textural position (Fig. 11a). Plagioclase in high-grade schist has a low to medium Ab content (staurolite-bearing schist X_{Ab} = 0.2–0.6; corundum-bearing schist X_{Ab} = 0.05–0.45). Plagioclase porphyroblasts in high-grade schist have variable non-symmetrical zoning patterns, with oscillatory (Fig. 12b and c) and normal zoning (i.e. more albite rich toward rims; Fig. 12a) as the most common types. Cores of plagioclase I, crowded with graphite and ilmenite inclusions, are characterized by low values of Fe^{3+} (Fig. 12a–d). Some plagioclase II porphyroblasts have thin rims of K-feldspar (total range X_{K} = 0.05–0.12) containing inclusions of fibrolite and biotite. The composition of plagioclase inclusions in garnet II porphyroblasts, similar to the composition range of matrix porphyroblasts, to X_{K} = 0.20–0.23 in the outer rim of garnet I grains. The highest Ab contents are found in plagioclase rims and plagioclase included with K-feldspar in fibrolite mats (X_{K} > 0.4).

K-feldspar porphyroblasts in high-grade schist are not zoned, and have low albite contents (X_{K} = 0.02–0.15) (Fig. 11b).
Fig. 7. Representative compositional profiles across garnet porphyroblasts in high-grade schist (HGS) [staurolite-bearing schist, (a–c) and (g); biotite-rich schist, (f); corundum-bearing schist, (h)]; and gneiss (i, j). All profiles are from rim to rim. Garnet profile in (c) corresponds to a garnet II porphyroblast adjacent to a garnet I grain [profile in (d)] (see location in Fig. 18b). The rim of this garnet II in contact with the matrix (grey area) has a remarkable increase in grossular, spessartine, and Fe/(Fe + Mg). Two-dimensional image maps of garnet profiles (a), (b), and (f) are shown in Figs 9, 8, and 10, respectively.
Oscillatory zoning in plagioclase may be the result of discontinuous growth, controlled by the local availability of reactant phases (e.g. garnet) and strain partitioning during deformation (Menard & Spear, 1996). The smooth variations in composition indicate that the zoning pattern was not controlled by any of the plagioclase compositional gaps (Smith, 1983). In the absence of any other recognizable synchronous Ca-bearing mineral phase, the normal and oscillatory zoning of plagioclase in the high-grade schist (i.e. growth of sodic plagioclase II rims after \(D_3\)) cannot be correlated with zoning of garnet I (i.e. growth of garnet I before \(D_3\)), which exhibits decreasing
Fig. 10. Back-scattered electron image (a) and X-ray composition maps (b–e) of a garnet II porphyroblast from sample 61 (161-976E-14R-1, 98–102 cm; biotite-rich high-grade schist). The line marks the location of the profile given in Fig. 7f. Conditions for image acquisition: 1024 × 1024 pixels, beam size 1 µm, 3 µm/pixel, Faraday cup current 25 nA, dwell time 30 ms/pixel.

Fig. 11. Compositional variation of single analysis of plagioclase (a) and K-feldspar (b) porphyroblasts in all the population of samples analysed.

grossular contents from core to rim. The normal zoning in plagioclase could, however, be explained as a result of breakdown of the paragonite component in the white mica. The textural evidence suggests that white mica breakdown occurred during $D_2$, however, whereas the growth of sodic plagioclase rims and porphyroblasts [plagioclase II] is post $D_2$. This is one line of evidence suggesting that the breakdown of white mica produced a small amount of melt, which took up the sodium, and that plagioclase II grew during crystallization of this melt phase. The oscillatory zoning may therefore result from short-term variations in $a_{H_2O}$ during crystallization, as is common in magmas crystallizing at low $P$ (e.g. Scaillet et al., 1995).

**Biotite and muscovite**

Dark brown biotite porphyroblasts occur in very different textural positions: biotite is locally oriented parallel to $S_1$;
it is commonly well oriented parallel to $S_2$, particularly in the hinge areas of $D_2$ folds; it occurs as randomly oriented aggregates, associated with plagioclase II and andalusite porphyroblasts; it occurs as inclusions in plagioclase II and garnet II porphyroblasts, and as reaction products of garnet I and staurolite. Biotite porphyroblasts defining $S_2$ are commonly closely associated or intergrown with fibrolite. In the corundum-bearing schist, large porphyroblasts of biotite with pyrite inclusions partly transformed to pyrrhotite are common. All of these textural relationships indicate that biotite was stable during almost the entire metamorphic evolution of these rocks. Mimetic alteration of biotite to chlorite has occurred at a very late stage.

Representative biotite analyses are illustrated in Fig. 13. We have rejected all the chemical analyses in high-grade schist with total oxide amounts <94.5 wt %, and selected those analyses with a sum of $[\text{cations}] >1.6$ ions p.f.u. This is to avoid possible chloride intergrowths at small scale, mass deficiencies, and analytical uncertainties in the determination of alkali elements in micas (García-Casco et al., 1993).

Biotite in the high-grade schist is in general Fe rich and exceptionally aluminous ($[\text{Al}]^{[6]}$ ranges between 0.75 and 1.25 ions p.f.u.). Matrix biotite is homogeneous in individual samples, but varies with rock type, suggesting that the main control is bulk-rock composition. In staurolite-bearing schist, for example, $\text{Fe}/(\text{Fe} + \text{Mg})$ is ~0.8 (range 0.67–0.85) and Ti varies from 0.3 to 0.5 ions p.f.u., whereas in corundum-bearing schist $\text{Fe}/(\text{Fe} + \text{Mg})$ is ~0.7 (range 0.5–0.8) and Ti ~0.2 ions p.f.u. (Fig. 13). Biotite included in garnet porphyroblasts (important for thermobarometric calculations) shows similar $\text{Fe}/(\text{Fe} + \text{Mg})$ and $[\text{Al}]^{[6]}$ to matrix biotite, but shows bell-shaped and linear or crescent-shaped patterns of zoning in $\text{Fe}$, $[\text{Al}]^{[6]}$, and Ti (Fig. 14). The bell-shaped zoning pattern suggests diffusional interchange with the garnet host, and both zoning patterns contradict the common assumption of infinitely fast diffusion in biotite, which produces rapid homogenization of the crystals in this
context (e.g. Spear & Florence, 1992; Spear & Parrish, 1996), and might suggest a rapid cooling history for these rocks.

Muscovite in the high-grade schist is always retrogressive (e.g. replacing andalusite), and has low Ti and Si contents (usually <0.05 and 3.05–3.3 ions p.f.u., respectively), consistent with crystallization at relatively low P and T (Guidotti, 1984; Massone & Schreyer, 1987; Guidotti et al., 1988).

**Staurolite**

This mineral is present in two different textural positions

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**Fig. 13.** Biotite composition per sample, expressed in terms of Al\(^{vi}\), Fe/(Fe\(^{3+}\) + Mg), and Ti contents. Mica analyses were normalized to 20 oxygens, 4 OH\(^-\), and Fe\(^{total}\) = Fe\(^{2+}\). Chemical analyses with total oxide amounts <94.5 wt% in the high-grade schist and <93.5 wt% in the gneiss and granite, and with a sum of [six] cations <1.6 ions p.f.u., have been rejected for this study.

**Fig. 14.** Compositional profiles of biotite included in a large garnet II porphyroblast (for location of these profiles, see Fig. 18b). Sample 28 (161-976B-76R-1, 45–49 cm). Both profiles illustrate the two extreme zoning patterns observed in biotite in this context: (a) a bell-shaped zoning pattern; (b) a linear, crescent-shaped zoning pattern.
that can coexist in the same rock sample. Small staurolite crystals (<200 μm) are preserved as corroded inclusions in garnet II, andalusite, and plagioclase II porphyroblasts. Inside andalusite, staurolite is locally transformed into an Fe-rich spinel (hercynite) (Fig. 5f). In some samples, it occurs as unaltered crystals in the matrix with an internal foliation defined by straight inclusions of graphite and ilmenite (Fig. 5c). Both textural positions suggest that staurolite grew inter-kinematically between $D_1$ and $D_2$.

Staurolite shows little chemical variation in Fe (Fig. 15a), but unaltered porphyroblasts in the matrix are zoned with increasing Fe/(Fe + Mg) toward the rims (0.925–0.946) (Fig. 16b). Mg and Zn contents are always very low (between 0.2–0.4 and 0.05–0.17 ions p.f.u., respectively). Relict inclusions of staurolite in garnet II and andalusite have lower Fe/(Fe + Mg) than matrix porphyroblasts (average 0.93 and 0.88, respectively), and may be derived from the cores of the original grains.

**Corundum**

Corundum forms elliptical porphyroblasts up to several millimetres in diameter in quartz-poor assemblages. The crystals have inclusion trails that are generally concordant with $S_2$, but showing some evidence of continued flattening of $S_1$ around the crystals. They are probably synkinematic with $S_2$. In some rocks there is a distinct halo of K-feldspar around the corundum, suggesting that it formed at the expense of muscovite (see also Shipboard Scientific Party, 1996).

**Ti–Fe oxides**

Ilmenite is the most abundant Ti–Fe oxide in all of the different types of high-grade schist. It is present as a matrix component parallel to $S_3$, or as inclusions in garnets I and II, plagioclase, andalusite, corundum, staurolite, and K-feldspar porphyroblasts. Inside andalusite, worm-like crystals of ilmenite with a radial pattern are common. Rutile is only present as needle-shaped crystals preserved in large ilmenite crystals, suggesting that rutile was transformed into ilmenite before $D_2$. Another Fe-rich oxide present as a minor component is spinel (hercynite), which has significant contents in Mg and Mn, and negligible Zn contents (average composition Fe$_{0.85}$Mg$_{0.09}$Mn$_{0.14}$Zn$_{0.07}$). Hercynite commonly occurs as small grains in poikiloblastic andalusite surrounding relict inclusions of staurolite (Fig. 5f).

**METAMORPHISM AND MELTING IN THE GNEISS**

We summarize here the main petrological features of the gneiss, with emphasis on those features that constrain melting conditions. An accompanying file, placed in the *Journal of Petrology* World Wide Web site, contains the detailed description of the mineral chemistry and textures of the gneiss. Additional and more detailed information on the reaction history and the P-T path followed by these rocks have also been given by Soto et al. (1999).

The main assemblage in the gneiss is biotite + muscovite + sillimanite + andalusite + cordierite + plagioclase + K-feldspar + ilmenite ± garnet ± quartz. This is clearly not an equilibrium assemblage, but the textural relationships do not allow an unequivocal paragenetic sequence to be established. Rather, the gneiss is best interpreted as a disequilibrium assemblage resulting from a series of strongly overstepped reactions that did not continue to completion (Soto et al., 1999).

Textural evidence suggests that melting occurred in these rocks as a result of the muscovite and biotite
Fig. 16. Selected textural relationships in the gneiss and granite rocks [mineral abbreviations after Kretz (1983); and Fib, fibrolite]. (a) Principal foliation defined by intensely oriented harmonious fibrolite, with augen magnetite and porphyroblasts of quartz and K-feldspar. Sample 51 (161-976B-95R-2, 24–28 cm; pelitic domain, migmatite gneiss). Photomicrograph, plane-polarized light. (b) Zoned plagioclase phenocryst surrounded by K-feldspar. Dark areas in the K-feldspar are small patches of Ab-rich plagioclase. Both feldspars are replaced by a fine intergrowth of white mica and sericite, with needle-like quartz (in black). The line marks the location of the profile given in Fig. 12c. Back-scattered electron image. Conditions for image acquisition: 512 × 512 pixels. Sample 43 (161-976B-82R-1, 93–98 cm; granite). (c) Andalusite with inclusions of fibrolite and biotite parallel to the external foliation. The andalusite porphyroblast shows apparently resorbed outlines with respect to surrounding plagioclase. Sample 161-976B-93R-1, 52–57 cm (pelitic gneiss). Photomicrograph, plane-polarized light. (d) Andalusite with augen texture. Opaque oxide inclusions define an included fabric at high angle to external foliation defined by biotite and fibrolite. Sample 161-976B-93R-1, 52–57 cm (pelitic gneiss). Photomicrograph, plane-polarized light. (e) Andalusite with inclusion trails of opaque oxide defining a folded internal fabric. External foliation is defined by oriented aggregates of fibrolite, clearly oblique to the internal fabric inside the andalusite. Sample 161-976B-96R-1, 30–34 cm (pelitic gneiss). Photomicrograph, plane-polarized light. (f) Idiomorphic andalusite phenocrysts included in a K-feldspar + quartz matrix. Sample 161-976B-96R-1, 46–48 cm (pelitic gneiss). Photomicrograph, plane-polarized light.
breakdown reactions. Muscovite dehydration melting reaction (6a) occurs at pressures <4–5 kbar and temperatures >700°C. The intersection between the Ms-out reaction (6a) and the granite melting curve defines an invariant point placed at ~725°C and 5 kbar. Consequently, to form the granitic assemblage Qtz + Bt + Sil + Kfs + Pl, the gneissic rocks must surpass the Ms-out melting curve at temperatures between this invariant point and those at the reactions involving melting of biotite to form pyroxene

\[ \text{Bt + Pl + Qtz} = \text{Opx + Kfs + L} \]  
(10)

i.e. at \( T = 700–750°C \). \( P \) conditions for this assemblage are constrained by the presence of cordierite.

Textural observations support a final growth of cordierite and K-feldspar including andalusite, probably representing the final crystallization of granitic melts. Taking all the textural evidence together, we suggest a continuous growth of cordierite accompanying the melting reactions. Cordierite was probably formed at \( P < 4–5 \text{kbar} \) (Soto et al., 1999) by the univariant reaction (7) and by the reaction

\[ \text{Al-rich Bt + Sil + Qtz} = \text{Crd + Kfs + L} \]  
(11)

(Vielzeuf & Holloway, 1988). Previous garnet porphyroblasts (Garnet I) were consumed by this reaction, whereas new garnet porphyroblasts (Garnet II) could have been locally produced by

\[ \text{Bt + Sil = Grt + Spl + Mag + L} \]  
(12)

(Clarke et al., 1989) or by the cordierite-forming univariant reaction (7). The composition of garnet II, with higher Fe/(Fe + Mg) ratio than in garnet I, confirms the phase relations deduced for the reaction (7) in the model KFMASH system (Fig. 3b).

Andalusite shows what appear to be contradictory textural relations with the main foliation in the gneiss. The porphyroblasts contain oriented inclusion trails of opaque oxide that are commonly oblique to the external foliation (Fig. 16d), and in many cases outline microfolds (Fig. 16c). The main foliation wraps around the andalusite porphyroblasts, truncating the included fabric, and creating a typical augen structure (Fig. 16d). This evidence suggests that the andalusite pre-dates the main foliation, and includes an earlier fabric and microfolds. These relationships led Platt et al. (1996) to conclude that the gneiss had passed from the andalusite stability field into the sillimanite field. Nevertheless, andalusite also includes and appears to post-date fibrolite (Fig. 16c), and in places either includes or is intergrown with prismatic sillimanite parallel to the main foliation. The simplest resolution of this paradox is that the andalusite post-dates the sillimanite foliation, but that deformation continued during and after andalusite growth, and resulted in folding of the foliation, rotation of the crystals, and creation of the augen structure.

Other textural evidence suggests that andalusite probably crystallized from a melt phase: andalusite porphyroblasts in the gneiss are commonly broken up into numerous small and partly resorbed fragments (Fig. 16c), which form rounded blebs surrounded by a granitic matrix; in places andalusite is included in and partly replaced by cordierite; and in leucosome domains andalusite occurs as large coherent euhedral crystals, included in phenocrysts of K-feldspar (Fig. 16f). These textures suggest that andalusite crystallized together with a granitic melt (e.g. Clarke et al., 1976; Price, 1983; Kerrick & Speer, 1988; Vernon et al., 1990).

The coexistence of andalusite and a granitic melt can be explained by considering the expansion of the \( P-T \) melt field because of the addition of \( \text{B}_2\text{O}_3 \) (Pichavant & Manning, 1984), F (Manning, 1981), and excess \( \text{Al}_2\text{O}_3 \) all the textural evidence together, we suggest a continuous (Holtz et al., 1989) or by the cordierite-forming univariant reaction (7) and by the granite melting curve de®nes an invariant point placed at ~725°C and 5 kbar. This proposition is also congruent with the high B contents measured by Acosta et al. (1997) (B ≤ 600 ppm) in comparable leucogranite dykes and migmatites which crop out in the vicinity of the Ronda ultramafic massif.

Andalusite was partly replaced by coarse-grained muscovite at a relatively late stage, indicating that the final stage in the cooling evolution of the gneiss was growth of muscovite in the presence of andalusite, perhaps as a result of the reaction with a residual hydrous fluid which could arise from the melt crystallization:

\[ \text{Crd + Kfs + And + H}_2\text{O} = \text{phengitic-Ms + Qtz} \]  
(13a)

\[ \text{Kfs + And + H}_2\text{O} = \text{Ms + Qtz} \]  
(13b)

(Vielzeuf & Holloway, 1988).

**Fe–Mg PARTITIONING IN COEXISTING MINERALS**

Compositions of the coexisting garnet, staurolite, and biotite in the high-grade schist, and of garnet, biotite, and cordierite in the gneissic rocks, are plotted in a semi-logarithmic plot and AFM projections in Fig. 17. In general, the Fe-Mg partitioning in coexisting phases follows the sequence \( \text{Fe/Mg}^{\text{solidus}} > (\text{Fe/Mg})^{\text{granite}} > (\text{Fe/Mg})^{\text{plagioclase}} \) in the staurolite-bearing schist, and \( (\text{Fe/Mg})^{\text{granite}} > (\text{Fe/Mg})^{\text{staurolite}} > (\text{Fe/Mg})^{\text{solidus}} \) in those gneisses in which garnet is present. This normal Fe–Mg partitioning occurs in Fe-rich systems and is consistent with natural data and with
predictions based on model systems (e.g. Thompson, 1976a, 1976b; Spear & Cheney, 1989). A reverse Fe–Mg partitioning occurs, however, for staurolite grains included in garnet II porphyroblasts \([\text{Fe/Mg}_{\text{Grt II}}} > \text{Fe/Mg}_{\text{St}}\) \], which indicates that textural disequilibrium is associated with chemical disequilibrium (e.g. Garcia-Casco & Torres-Roldán, 1996), or is probably the result of the non-ideal behaviour of the Fe–Mg solid solution in staurolite (Holdaway et al., 1988; Ballevère et al., 1989). Garnet shows a wider variation in Fe/Mg ratio than the other ferromagnesian phases. The Fe/Mg ratio decreases from core to rim and tends to be higher in the schists (>10) than in the gneisses (<12); but there are significant differences among the different types of staurolite in garnet I rims of garnet I porphyroblasts in St-bearing schist Fe/Mg is 10–23; in Crn-bearing it is ~7; and in Bt-rich schist it is 7–10, which could be related to different bulk-rock compositions (see Fig. 4). There are also differences in the Fe/Mg ratio of garnet rims within samples of the same rock type (e.g. compare samples of St-bearing schist), which probably reflect subsequent resorption of garnet. A reverse Fe–Mg zonation characterizes some of the garnet porphyroblasts in the gneissic rocks \([\text{Fe/Mg}_{\text{Grt II}}} < \text{Fe/Mg}_{\text{St}}\) \], which could be due either to modification of their composition by diffusion processes (see above sections) or to crystallization within the cordierite stability field. The latter interpretation is based on the theoretical Fe–Mg partitioning in the KFMASH system, which predicts a significant increase of the Fe/Mg ratio in garnet when it coexists with cordierite and K-feldspar [e.g. see fig. 3B of Thompson (1976a)].

Staurolite has a variable Fe/Mg ratio (7–16), but always maintains a constant difference from the composition of garnet I rims. As the theoretical Fe–Mg partitioning in the KFMASH system predicts a progressive Mg enrichment of staurolite with rising temperature (e.g. Spear & Cheney, 1989) and Fe enrichment with decreasing pressure (see below), the inverse zonation in Fe/Mg ratio
from core to rim suggests that the staurolite grew during decompression.

Matrix biotite in all rock types is generally homogeneous in terms of Fe/Mg ratio (see also Fig. 13), but varies significantly according to rock type (2.5–4.5 for St-bearing and Bt-rich schist; 1.3–2.5 for Crn-bearing schist; 1.5–2.5 for gneissic rocks). Where variations in the Fe/Mg ratio occur within single samples, there is no correlation with the textural position of the biotite, or with the Fe/Mg of garnet in the same rock. The largely homogeneous composition of matrix biotite within single samples together with the lack of correlation with the chemical composition of garnet suggest that biotite composition is probably not in equilibrium with garnet, and that biotite grains of different ages have been largely homogenized with respect to Fe/Mg ratio.

**METAMORPHIC EVOLUTION OF THE HIGH-GRADE SCHIST**

**Reaction history**

In the high-grade schist, we can identify an early (post-$S_1$, pre-$S_2$) mineral assemblage that includes garnet I + staurolite + biotite + plagioclase + Ti-Fe oxide + quartz (Assemblage 1). Given the bulk composition of the schist, muscovite was almost certainly present, but is not now preserved. In the KFMASH system Assemblage 1 corresponds to a divariant assemblage Grt + St + Bt + Ms [Als] (3a), and it is therefore unlikely that an Al-silicate was present in equilibrium with this assemblage (Fig. 4a).

The normal zoning pattern of garnet I suggests prograde metamorphism with rising temperature and possibly decreasing pressure, but the timing of this garnet growth is not clear from textural evidence, although thermobarometric calculations (see below) suggest that their rims grew near the Ky–Sil transition. These observations, together with the decrease in the Fe/ (Fe + Mg) ratio in garnet I towards the rim, suggest that the growth of this mineral may have occurred by the divariant reaction [Als] (3a) in the Ky stability field (see Fig. 3c). Inner garnet cores characterized by a decreasing Fe/(Fe + Mg) ratio might be relics of previous garnet which may have grown in greenschist facies conditions from chlorite or chloritoid. Subsequent resorption of garnet I, before the growth of garnet II rims, may have occurred either as a result of the prograde univariant reaction

\[ \text{Grt} + \text{Chl} + \text{Ms} = \text{St} + \text{Bt} + \text{Qtz} + \text{H}_2\text{O} \quad (14) \]

or during decompression by the net transfer reaction [Als] (3a).

During $D_3$, the stable assemblage in the high-grade schist was probably biotite + sillimanite + K-feldspar + plagioclase + ilmenite + quartz ± garnet (Assemblage 2). Staurolite became unstable when the conditions surpassed the univariant reaction (4). Muscovite breakdown occurred probably as a result of the dehydration melting reaction (6a) or its equivalent:

\[ \text{Ms} + \text{Pl} + \text{Qtz} = \text{Sil} + \text{Kfs} + \text{L} \quad (6b) \]

(Thompson & Algor, 1977; Thompson & Tracy, 1979; Thompson, 1982). Reaction [Als] (3a), which probably occurred before (5b), also contributed to the consumption of muscovite. Textural evidence that suggests the presence of minor amounts of melts formed by this reaction are the occurrence of K-feldspar + plagioclase + quartz intergrowths included in the new garnet II porphyroblasts, and also K-feldspar on rims of plagioclase, and interstitial K-feldspar together with quartz in veins. Additionally, the apparent simultaneous crystallization of K-feldspar, plagioclase, and andalusite, to form Assemblage 3, is interpreted as resulting from the later crystallization of interstitial melts. The development of the assemblage corundum + biotite + K-feldspar + plagioclase in quartz-depleted rocks (corundum-bearing schist), represents the reaction

\[ \text{Ms} = \text{Crn} + \text{Kfs} + \text{H}_2\text{O} \quad (15) \]

(Huang et al., 1973; Huang & Wyllie, 1974).

Therefore, Assemblage 2 results from the change of the divariant assemblage Grt + Bt + Sil + Ms to form Grt + Bt + Sil + Kfs. Taken altogether, this sequence of assemblages indicates decompression, accompanied by constant or rising $T$.

Textural evidence suggests that garnet II grew after $D_3$, and therefore at lower $P$ and/or higher $T$ than the muscovite breakdown reaction. Nevertheless, barometric studies on inclusion suites inside these garnet porphyroblasts (see below) suggest that garnet grew under decreasing-$P$ conditions (from the Ky–Sil transition to $P \sim 3$–4 kbar). Garnet growth at this stage probably involved staurolite breakdown (4), as suggested by the fact that some garnet II grains include corroded staurolite relics. Another garnet-forming reaction that could have been involved is the Crd-forming reaction (7), which is ruled out because of the absence of cordierite in the high-grade schist. The range of $P$ estimations on this garnet, together with the flat-zoning pattern in the Fe/(Fe + Mg) ratio, can only be explained by a rising-$T$ decompression $P-T$ path (Fig. 3c). Along this path limited growth of garnet could have occurred by reaction (5b), but given the abundant evidence of disequilibrium in these rocks (in particular, the preservation of staurolite in the matrix), we suggest that garnet II grew also by overstepping of reaction (4) under low-$P$ conditions.

The final assemblage we can establish in these rocks is biotite + andalusite + K-feldspar + plagioclase + ilmenite + quartz (Assemblage 3). The common
occurrence of staurolite relics included in andalusite suggests that the breakdown of staurolite was considerably overstepped, and probably contributed to substantial growth of an Fe-rich biotite during this final stage. Retrogressive breakdown of garnet occurred during the cooling path by the divariant reaction \([\text{St}]\) (5b), as is revealed by the occurrence of andalusite including garnet, K-feldspar, and fibrolite (Fig. 6).

**Quantitative \(P-T\) determinations**

We have applied several geothermometers and geobarometers to help constrain the \(P-T\) conditions achieved by the basement rocks. Table 2 shows selected \(P-T\) estimates for representative analyses per sample and in different textural positions, and illustrates the diversity of results obtained within a single thermometer or barometer depending on the type of calibration. The absence of suitable parageneses for thermobarometry in the gneiss together with the spurious and scattered results obtained with several geothermometers and geobarometers (e.g. garnet + cordierite geobarometer) give us additional data to suggest that many of the mineral assemblages in the gneissic rocks are not in equilibrium.

Among the different calibrations of the garnet + biotite (GARB) thermometer, the calibrations which include corrections for the biotite composition (in particular, for the Ti and Al\(^{[\text{vi}]}\) contents; e.g. Indares & Martignole, 1985; Patiño Douce et al., 1993; Holdaway et al., 1997) and for the non-ideality in both the garnet and biotite (e.g. Bhattacharya et al., 1992; Kleemann & Reinhardt, 1994; Holdaway et al., 1997) seem to be the most reasonable calibrations for the studied rocks, on the basis of biotite and garnet compositions (i.e. biotite with a significant Ti content and garnet relatively rich in Ca and Mn). The results obtained with the calibration of Holdaway et al. (1997) have a smaller range in temperature estimates in the high-grade schist (620–780°C), although scattered results are always obtained in the gneissic rocks (600–950°C). In general, the calibration of Holdaway et al. (1997) yields temperature estimates close to those obtained with the calibration of Kleeman & Reinhardt (1994) (maximum differences of \(+15\)–20°C) and Perchuk & Lavrent’eva (1983) (maximum differences of \(-20\)–40°C), whereas the calibration of Patiño Douce et al. (1993) always gives extremely high temperatures, with deviations of up to 175–180°C.

We have applied the GARB thermometer to garnet–biotite grains in contact, either in the matrix or as inclusions in large garnet II porphyroblasts (see Figs 18 and 19a). The temperatures obtained vary from 620 to 770°C [henceforth values with the calibration of Holdaway et al. (1997); always for an estimated pressure of 4 kbar]. The calculated temperatures obtained from single garnet II porphyroblasts crowded with biotite inclusions vary from 637 to 693°C (662 \(\pm\) 14.5°C).

The wide span in the calculated temperature within single samples (some incompatible with the stability field of the observed mineral assemblage) gives further evidence of disequilibrium in these rocks. Additionally, the temperatures calculated with GARB should be used with circumspection because of the following contributing factors:

1. Matrix biotites display variable Fe/(Fe + Mg) contents depending on the textural position and reaction site.
2. Biotite in the matrix has probably undergone a late-stage homogenization towards more Fe-rich compositions (resulting in an overestimate of the temperature with GARB), by the breakdown of garnet at lower \(P\) to form andalusite + biotite by the net transfer reaction \([\text{St}]\) (5b) (Fig. 3a and c).
3. Some of the biotite inclusions in garnet II porphyroblasts have also probably undergone this late-stage enrichment in Fe by diffusion along fractures in garnet (Whitney, 1996), which are particularly abundant in garnet II porphyroblasts (see Figs 8 and 18). The occurrence of Fe-rich chlorite inclusions in some of these porphyroblasts (Fig. 18) provides further evidence for retrogression, probably induced by diffusion along these fractures.
4. Dissolution of garnet exposed different rim compositions [in general with higher Fe/(Fe + Mg) ratios than the absolute rim] not in equilibrium with matrix biotites. This could explain some of the lower temperatures obtained with rims of garnet I porphyroblasts (Figs 18b and 19a).

All of these factors are related to the widespread disequilibrium relationships between garnet and biotite in these high-grade rocks, and suggest caution in the use of this geothermometer to estimate peak-temperature conditions (e.g. Ghent et al., 1982; Selverstone & Chamberlain, 1990; Spear & Florence, 1992; Spear & Parrish, 1996).

Application of the garnet + sillimanite + quartz + plagioclase (GASP) barometer (calibrations of Newton & Haselton, 1981; Hodges & Spear, 1982; Ganguly & Saxena, 1984; Hodges & Crowley, 1985; Koziol & Newton, 1988; Koziol, 1989) to plagioclase inclusions in garnet from the staurolite-bearing high-grade schist consistently gives pressures between 6 and 7 kbar in garnet I rims \(\left[X_{\text{Bi}}\right] = 0.20\)–0.25 and assuming kyanite as the coexisting phase of Holdaway et al. (1997).
Table 2: Representative garnet, biotite, and plagioclase analyses with thermobarometric calculations

<table>
<thead>
<tr>
<th>Sample</th>
<th>St-bearing schist</th>
<th>Crn-schist</th>
<th>Bt-rich schist</th>
<th>Gneiss</th>
<th>Gnt-bearing gneiss</th>
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<tr>
<td></td>
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<tr>
<td>Garnet porphyroblasts (rim of garnet I, [garnet I near inclusion], and [garnet II])</td>
<td></td>
<td></td>
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<td></td>
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<tr>
<td>Analysis</td>
<td>C-15 (C-67)</td>
<td>d-25</td>
<td>D-43 [B-177]</td>
<td>(G-27)</td>
<td>F-15</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>49-6</td>
</tr>
<tr>
<td>Anh</td>
<td>0.737 (0.701)</td>
<td>0.787</td>
<td>0.831 [0-790]</td>
<td>(0-702)</td>
<td>0-629</td>
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<tr>
<td>XGr</td>
<td>0.081 (0.095)</td>
<td>0.094</td>
<td>0.045 [0-063]</td>
<td>(0-167)</td>
<td>0-177</td>
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<tr>
<td>Xgr</td>
<td>0.068 (0.052)</td>
<td>0.043</td>
<td>0.046 [0-063]</td>
<td>(0-025)</td>
<td>0-088</td>
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<tr>
<td>Xsps</td>
<td>0.114 (0.152)</td>
<td>0.076</td>
<td>0.079 [0-083]</td>
<td>(0-106)</td>
<td>0-106</td>
</tr>
<tr>
<td>Fe-no.</td>
<td>0.916 (0.930)</td>
<td>0.948</td>
<td>0.947 [0-926]</td>
<td>(0-965)</td>
<td>0-877</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td>0.928</td>
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<tr>
<td>Biotite (matrix and [inclusion in garnet II porphyroblast])</td>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Analysis</td>
<td>C-29</td>
<td>d-7</td>
<td>D-45 [B-120]</td>
<td>(G-28)</td>
<td>F-49</td>
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<td>9-51</td>
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<tr>
<td>Anh</td>
<td>0.167</td>
<td>0.157</td>
<td>0.179 [0-147]</td>
<td>(0-147)</td>
<td>0-149</td>
</tr>
<tr>
<td>XTi</td>
<td>0.072</td>
<td>0.083</td>
<td>0.070 [0-070]</td>
<td>(0-070)</td>
<td>0-027</td>
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<tr>
<td>Fe-no.</td>
<td>0.713</td>
<td>0.834</td>
<td>0.822 [0-747]</td>
<td>(0-747)</td>
<td>0-572</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.744</td>
</tr>
<tr>
<td>Plagioclase (rim of porphyroblast in the matrix, [inclusion in garnet I porphyroblast], and [inclusion in garnet II porphyroblast])</td>
<td></td>
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<td></td>
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<tr>
<td>Analysis</td>
<td>C-76 (C-35)</td>
<td>d-2</td>
<td>D-47 [B-178]</td>
<td>(G-28)</td>
<td>F-49</td>
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<td>81</td>
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<tr>
<td>Anh</td>
<td>0.845 (0.770)</td>
<td>0.549</td>
<td>0.434 [0-526]</td>
<td>(0-776)</td>
<td>0-805</td>
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<tr>
<td>XAb</td>
<td>0.153 (0.230)</td>
<td>0.441</td>
<td>0.556 [0-466]</td>
<td>(0-215)</td>
<td>0-124</td>
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<tr>
<td></td>
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<td>0.527 (0.497)</td>
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<tr>
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</tr>
<tr>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>T1&amp;5</td>
<td>670</td>
<td>750</td>
<td>743 [685]</td>
<td>(685)</td>
<td>607</td>
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<tr>
<td>T2&amp;5</td>
<td>704</td>
<td>799</td>
<td>762 [711]</td>
<td>(711)</td>
<td>675</td>
</tr>
<tr>
<td>T5 &amp; 1</td>
<td>727</td>
<td>853</td>
<td>831 [756]</td>
<td>(756)</td>
<td>643</td>
</tr>
<tr>
<td>T5 &amp; 2</td>
<td>773</td>
<td>933</td>
<td>864 [795]</td>
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<td>711</td>
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<tr>
<td>T6 &amp; L</td>
<td>627</td>
<td>665</td>
<td>661 [634]</td>
<td>(634)</td>
<td>604</td>
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<tr>
<td>T6 &amp; M1</td>
<td>621</td>
<td>663</td>
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<td>691</td>
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<tr>
<td>T6 &amp; M2</td>
<td>580</td>
<td>640</td>
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<td>(637)</td>
<td>626</td>
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<tr>
<td>T7</td>
<td>707</td>
<td>802</td>
<td>771 [723]</td>
<td>(723)</td>
<td>676</td>
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<tr>
<td>T8</td>
<td>534</td>
<td>558</td>
<td>508 [606]</td>
<td>(606)</td>
<td>624</td>
</tr>
<tr>
<td>T10</td>
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<td>650</td>
<td>629 [620]</td>
<td>(620)</td>
<td>595</td>
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<tr>
<td>T11</td>
<td>562</td>
<td>598</td>
<td>586 [584]</td>
<td>(584)</td>
<td>573</td>
</tr>
<tr>
<td>T12</td>
<td>562</td>
<td>776</td>
<td>733 [700]</td>
<td>(700)</td>
<td>709</td>
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<td>627</td>
<td>670</td>
<td>641 [646]</td>
<td>(646)</td>
<td>629</td>
</tr>
<tr>
<td>T14</td>
<td>627</td>
<td>668</td>
<td>652 [656]</td>
<td>(656)</td>
<td>653</td>
</tr>
<tr>
<td>T15 &amp; 1</td>
<td>638</td>
<td>663</td>
<td>647 [648]</td>
<td>(648)</td>
<td>643</td>
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<td>Garnet + Al-silicate + quartz + plagioclase (GASP) barometer</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>P[Al2O3]</td>
<td>5812 (5062)</td>
<td>7211</td>
<td>4534 [5077]</td>
<td>(5077)</td>
<td>6822</td>
</tr>
<tr>
<td>P[Al2O3]</td>
<td>2900 (3259)</td>
<td>5586</td>
<td>3412 [3716]</td>
<td>(3716)</td>
<td>4976</td>
</tr>
<tr>
<td>P[Al2O3]</td>
<td>3732 (3935)</td>
<td>5722</td>
<td>2237 [3272]</td>
<td>(3272)</td>
<td>6516</td>
</tr>
<tr>
<td>P[Al2O3]</td>
<td>4030 (4183)</td>
<td>6621</td>
<td>4774 [4484]</td>
<td>(4484)</td>
<td>6196</td>
</tr>
<tr>
<td>P[Al2O3]</td>
<td>6319 (5708)</td>
<td>7316</td>
<td>4243 [5037]</td>
<td>(5037)</td>
<td>7464</td>
</tr>
<tr>
<td>Garnet + plagioclase + biotite + quartz (GPBQ) barometer</td>
<td></td>
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<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>P[SiO2]</td>
<td>5560</td>
<td>7257</td>
<td>4485 [4945]</td>
<td>(4945)</td>
<td>9261</td>
</tr>
<tr>
<td>P[SiO2]</td>
<td>5595</td>
<td>6834</td>
<td>4261 [4810]</td>
<td>(4810)</td>
<td>9423</td>
</tr>
</tbody>
</table>
Al-silicate; henceforth quoted pressure estimates are those obtained with the calibration of Hodges & Crowley (1985), close to the kyanite–sillimanite transition, for an assumed temperature of 600°C (Table 2 and Fig. 19a). These estimates with GASP should be considered as maximum pressures because of the previously reported absence of an Al-silicate in equilibrium with garnet I (Assemblage 1). Using rim compositions for garnet I in contact with rims of matrix plagioclase \( (X_{Al} = 0.45–0.60) \), and assuming that both minerals are in equilibrium with sillimanite, gives a wide range of lower pressures, 4–6.7 kbar at 675°C (Table 2, Fig. 19a). Plagioclase inclusions

Footnote to Table 2

\( P \) (in bar) obtained for an estimated temperature of 675°C, except for garnet I estimates, where 600°C has been used. \( T \) (in °C) obtained for an estimated pressure of 4 kbar; with the exception of high-grade schist 32 (8 kbar) and gneiss 53 (6 kbar). Inside square brackets are shown \( P-T \) estimates from inclusions in garnet II porphyroblasts. Inside braces are shown \( P-T \) estimates from inclusions in rims of garnet I porphyroblasts. n.d., not determined; n.e., out of range estimation because of clear disequilibrium relationships.

\* \( X_{Alm} = \frac{Fe}{Fe + Ca + Mg + Mn} \); \( X_{Grs} = \frac{Ca}{Fe + Ca + Mg + Mn} \); \( X_{Pyr} = \frac{Mg}{Fe + Ca + Mg + Mn} \); \( X_{SpS} = \frac{Mn}{Fe + Ca + Mg + Mn} \); \( X_{Fe-No} = \frac{Fe}{Fe + Mg} \).

\^ \( X_{Al} = \frac{Al}{\text{Sum}} \); \( X_{Ti} = \frac{Ti}{\text{Sum}} \); \( \text{Sum} = Al + Ti + Cr + Mg + Fe + Mn + Zn \).

\& GARB calibrations: \( T_{F&S} \), Ferry & Spear (1978); \( T_{G&S} \), Ganguly & Saxena (1984), symmetrical mixing model for garnet; \( T_{G&S-1} \), Ganguly & Saxena (1984), asymmetrical mixing model for garnet (both calibrations have been calculated with a value of the Mn-Margules mixing parameter \( \Delta W_{Mn} = 2500 \text{ cal} \) (Ganguly & Saxena, 1985; Holdaway et al., 1989)); \( T_{P&E} \), Perchuk & Lavrent’eva (1983); \( T_{G&S-2} \), Ganguly & Saxena (1984) with garnet activity model from Ganguly & Saxena (1984); \( T_{I&M} \), Indares & Martignole (1985) with garnet activity model from Ganguly & Saxena (1984); \( T_{Bh} \), Bhattacharya et al. (1992) with garnet activity model from Berman (1990); \( T_{D} \), Dasgupta et al. (1991); \( T_{S} \), Sass & Bose (1992); \( T_{S} \), Bhattacharya et al. (1992) with garnet activity model from Ganguly & Saxena (1984); \( T_{R} \), Roy (2013) with a non-ideal mixing model for biotite and garnet, assuming all Fe in biotite is Fe\(^{2+}\), and activity coefficients of garnet calculated with the model of Berman (1990); \( T_{K&R} \), Kleemann & Reinhardt (1994) with a biotite activity model assuming non-ideal mixing of (Mg, Fe)–Al and (Mg, Fe)–Ti and ideal Mg–Fe mixing, garnet activity model from Berman (1990); \( T_{M} \), Holdaway et al. (1997) adopting new Margules parameters for garnet from Mukhopadhyay et al. (1997), assuming that total Fe is Fe\(^{2+}\) in garnet and biotite \( (T_{M}=Fe^{2+}) \) or that garnet contains 3% Fe\(^{3+}\) replacing Al in octahedral sites and biotite has 11.6% Fe\(^{3+}\) \( (T_{M}=Fe^{3+}) \).

\$ GASP calibrations: \( P_{M&H} \), Newton & Haselton (1981); \( P_{G&S} \), Ganguly & Saxena (1984); \( P_{H&S} \), Hodges & Spear (1982); \( P_{B} \), Berman (1990); \( P_{K&S} \), Koziol & Newton (1988) and Koziol (1989). In all cases the Al-silicate is sillimanite, with the exception of estimates from inclusions in garnet I porphyroblasts, where kyanite is assumed to be in equilibrium.

in garnet II porphyroblasts, which tends to be more Ab rich towards the garnet rim (plagioclase inclusions in the garnet II core: $X_{Al} = 0.33-0.45$; inclusions near the garnet II rim: $X_{Al} = 0.48-0.38$) also record decreasing pressures during their growth, from 7.3-6.2 kbar at the garnet core to 3.5-3 kbar at the garnet rim (at 675°C and coexisting with sillimanite; Fig. 18b). In view of the strong zonation of plagioclase in the matrix, and the evidence of late-stage re-equilibration of feldspars, the GASP results with rims of porphyroblasts should be treated with caution.

Additional $P$ estimations have been obtained by using the garnet + plagioclase + biotite + quartz (GPBQ) barometer calibrated by Hoisch (1990, 1991), which gives pressures consistent with GASP in the range 8.6-7.6 kbar (for an assumed $T$ of 675°C) for plagioclase included in garnet II porphyroblasts, and 5.3-4.3 kbar for rims in contact with matrix plagioclase (Table 2, Fig. 19a).

Pressure estimates for garnet I porphyroblasts have been performed by applying the garnet + rutile + Al-silicate + ilmenite + quartz (GRAIL) barometer (Bohlen et al., 1983) to rutile-ilmenite intergrowths occasionally preserved in these garnet porphyroblasts. Pressures calculated, assuming pure ilmenite, range from 10.5 to 8 kbar (at 500 and 600°C) for inclusions in garnet I core and rim, respectively (Fig. 19a). Considering the above-mentioned presumable absence of an Al-silicate in equilibrium with garnet I, the GRAIL estimates can be
Thermobarometry of the High-grade schist
GASP, GPBQ, GRAIL, and GARB estimates

Thermobarometry on the cooling P-T path
Ternary-Feldspar calibration (Elkins & Grove, 1990)

[Diagrams showing P-T relationships and inclusions in minerals]
considered as tentative minimum pressures (see Bohlen et al., 1983, p. 1055).

Keeping in mind the need for caution in interpreting the thermobarometric results, we suggest that the results of the GASP and GBPQ barometers indicate a significant pressure drop during $D_h$ from 6-7 kbar to 3-4 kbar. Approximate temperature conditions at the end of $D_h$ are likely to have been between 650-700°C in the metapelites and 700-750°C in the gneiss. Approximate prograde $P$-$T$ conditions in these rocks can only be estimated using the GRAIL barometer estimates, indicating a pressure of 10-5-8 kbar (for a presumed temperature of 500 and 600°C) for core and rim of garnet I porphyroblasts, respectively (see Fig. 19a).

**P–T conditions along the cooling path**

The decompression $P$-$T$ path was followed by a cooling path ($P < 3$ kbar) into the andalusite stability field ($T \sim 600°C$). During this portion of the path, rapid crystallization of residual melts and a large overstepping of reactions occurred, allowing complex textural relationships involving andalusite to form. In the metapelites these include relic staurolite preserved inside large andalusite porphyroblasts, and intergrowths of andalusite and either plagioclase or K-feldspar. In the gneiss, andalusite is found either included in or intergrown with cordierite, plagioclase, and K-feldspar.

To constrain the $P$-$T$ conditions experienced by the rocks during the cooling history, we have used the two-feldspar geothermometer, based on the common interpretation that this thermometer could yield estimates of solidus temperatures for granitic systems (Anderson, 1996). We have used the binary-feldspar calibrations of Stormer (1975), Powell & Powell (1977), Whitney & Stormer (1977), and Haselton et al. (1983), together with the ternary-feldspar calibrations of Ghiorsor (1984), Green & Usdansky (1986), Fuhrman & Lindsley (1988), and Elkins & Grove (1990). Mole fraction information of solidus temperatures for granitic systems (Anderson, & Usdansky (1986), Fuhrman & Lindsley (1988), and Joyce & Voigt, 1994; Johannes & Holtz, 1996) (see Figs. 19a). Key reactions from the KFMASH grid of Spear & Cheney (1989) are also shown for reference. Muscovite-out reactions ([6a] and [6b]) are from Vielzeuf & Clemens (1992). Long-dashed lines in the hornblende stability field represent the effect of the addition of $F$ (4 wt %) and $B_2O_3$ (4 wt %) on the granite solidus. [Note the large degree of equilibrium during crystallization of residual melts and a large overstepping of reactions occurred, allowing complex textural relationships involving andalusite to form. In the metapelites these include relic staurolite preserved inside large andalusite porphyroblasts, and intergrowths of andalusite and either plagioclase or K-feldspar. In the gneiss, andalusite is found either included in or intergrown with cordierite, plagioclase, and K-feldspar.]

![Fig. 19. Thermobarometric constraints on the P-T evolution of the Alboran basement. (a) Thermobarometric estimates in the high-grade schist (samples used for the plot: 22, 28, 32, 61, and 76B; see Table 2) with GARB thermometer (Holdaway et al., 1997; assuming that garnet contains 3% $Fe^{3+}$ and biotite has 11.6% $Fe^{3+}$), and GASP (Hodges & Crowley, 1985), GBPQ (Hoisch, 1990, 1991), and GRAIL (Bohlen et al., 1983) barometers. Key reactions from the KFMASH grid of Spear & Cheney (1989) are also shown for reference. Muscovite-out reactions ([6a] and [6b]) are from Huang et al. (1973), Huang & Wyllie (1974), Thompson & Algor (1977), and Thompson (1982). The lightly shaded box represents the average temperatures and pressures ($\pm 1σ$) for the rims in contact of matrix porphyroblasts of garnet, biotite, and plagioclase. [Note the large degree of equilibrium during crystallization of residual melts and a large overstepping of reactions occurred, allowing complex textural relationships involving andalusite to form. In the metapelites these include relic staurolite preserved inside large andalusite porphyroblasts, and intergrowths of andalusite and either plagioclase or K-feldspar. In the gneiss, andalusite is found either included in or intergrown with cordierite, plagioclase, and K-feldspar.]

(b) Thermobarometric estimates in the high-grade schist (dotted lines) and granite + gneissic rocks (continuous lines) on plagiochriste + K-feldspar intergrowths (selected results from samples: 22, 28, 43, 76, and 76B; see Table 3), performed with the ternary-feldspar calibration of the two-feldspar thermometer [calibration of Elkins & Grove (1990)]. These triangular areas represent the $P$-$T$ conditions achieved by these rocks during the cooling path. The lightly shaded area corresponds to one of the most accurate $P$-$T$ estimates (sample SOT) obtained with this thermometer (centred in this area is placed at 500°C and 3-2 kbar). Data sources for the different granite solidus have been summarized by Johannes & Holtz (1996) [H$_2$O-saturated granite solidus is from Le Breton & Thompson (1988); Al$_2$O$_3$-saturated granite solidus is from Holtz et al. (1992) and Joyce & Voigt (1994). Dehydration melting reactions ([6a] and [6b]) are from Vliezeuf & Clemens (1992). Long-dashed lines in the And-stability field represent the effects of the addition of $F$ (4 wt %) and $B_2O_3$ (4 wt %) on the granite solidus. [Note the large degree of equilibrium during crystallization of residual melts and a large overstepping of reactions occurred, allowing complex textural relationships involving andalusite to form. In the metapelites these include relic staurolite preserved inside large andalusite porphyroblasts, and intergrowths of andalusite and either plagioclase or K-feldspar. In the gneiss, andalusite is found either included in or intergrown with cordierite, plagioclase, and K-feldspar.]

The proposed prograde and decomposition $P$-$T$ path for the high-grade schist have also been plotted for comparison. Dashed portion of the path shows approximate $P$-$T$ conditions along the prograde trajectory (growth of garnet I). (b) Thermobarometric estimates in the high-grade schist (dotted lines) and granite + gneissic rocks (continuous lines) on plagiochriste + K-feldspar intergrowths (selected results from samples: 22, 28, 43, 76, and 76B; see Table 3), performed with the ternary-feldspar calibration of the two-feldspar thermometer [calibration of Elkins & Grove (1990)]. These triangular areas represent the $P$-$T$ conditions achieved by these rocks during the cooling path. The lightly shaded area corresponds to one of the most accurate $P$-$T$ estimates (sample SOT) obtained with this thermometer (centred in this area is placed at 500°C and 3-2 kbar). Data sources for the different granite solidus have been summarized by Johannes & Holtz (1996) [H$_2$O-saturated granite solidus is from Le Breton & Thompson (1988); Al$_2$O$_3$-saturated granite solidus is from Holtz et al. (1992) and Joyce & Voigt (1994). Dehydration melting reactions ([6a] and [6b]) are from Vliezeuf & Clemens (1992). Long-dashed lines in the And-stability field represent the effects of the addition of $F$ (4 wt %) and $B_2O_3$ (4 wt %) on the granite solidus. [Note the large degree of equilibrium during crystallization of residual melts and a large overstepping of reactions occurred, allowing complex textural relationships involving andalusite to form. In the metapelites these include relic staurolite preserved inside large andalusite porphyroblasts, and intergrowths of andalusite and either plagioclase or K-feldspar. In the gneiss, andalusite is found either included in or intergrown with cordierite, plagioclase, and K-feldspar.]

The proposed cooling $P$-$T$ path for the Alboran basement has also been plotted for comparison. Numbers in parentheses refer to reactions described in the text. Al-silicate triple point from Berman (1988).
Table 3: Selected plagioclase and K-feldspar analyses with two-feldspar thermometric calculations

<table>
<thead>
<tr>
<th>Sample</th>
<th>Textures</th>
<th>Sample</th>
<th>Textures</th>
<th>Sample</th>
<th>Textures</th>
</tr>
</thead>
<tbody>
<tr>
<td>22</td>
<td>mtrx-fksil</td>
<td>28</td>
<td>mtrx</td>
<td>31</td>
<td>Z-Pl</td>
</tr>
<tr>
<td>54</td>
<td>mtrx</td>
<td>61</td>
<td>D-5</td>
<td>76</td>
<td>mtrx</td>
</tr>
<tr>
<td>51</td>
<td>mtrx</td>
<td>53</td>
<td>D-5</td>
<td>43</td>
<td>mtrx</td>
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<td>77</td>
<td>mtrx</td>
<td>43</td>
<td>D-5</td>
<td>77</td>
<td>mtrx</td>
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Plagioclase

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<th>Texture</th>
<th>XAn</th>
<th>XAb</th>
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<tbody>
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<td>A22-inc-2</td>
<td>A-12</td>
<td>0.541</td>
<td>0.447</td>
</tr>
<tr>
<td>B28-inc-1</td>
<td>D-57</td>
<td>0.415</td>
<td>0.572</td>
</tr>
<tr>
<td>C-15</td>
<td>1-per 4</td>
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K-feldspar

<table>
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<tr>
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<th>Texture</th>
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<th>XAb</th>
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<tbody>
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<td>A22-inc-1</td>
<td>A-10</td>
<td>0.001</td>
<td>0.100</td>
</tr>
<tr>
<td>B28-inc-1</td>
<td>D-56</td>
<td>0.002</td>
<td>0.116</td>
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<td>C-42</td>
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</table>

Binary-feldspar calibrations

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<th>Temperature (°C)</th>
<th>Precision</th>
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<tbody>
<tr>
<td>TS</td>
<td>571</td>
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<tr>
<td>TP</td>
<td>541</td>
</tr>
<tr>
<td>TW</td>
<td>504</td>
</tr>
<tr>
<td>TH</td>
<td>566</td>
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</tbody>
</table>

Ternary-feldspar calibrations

<table>
<thead>
<tr>
<th>Temperature (°C)</th>
<th>Precision</th>
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</thead>
<tbody>
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<td>TGi</td>
<td>622</td>
</tr>
<tr>
<td>TG&amp;U</td>
<td>527</td>
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<tr>
<td>TF&amp;L</td>
<td>523</td>
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<tr>
<td>TE&amp;G</td>
<td>566</td>
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</tbody>
</table>

$T$ (in °C) has been calculated for an estimated pressure of 3 kbar. In the ternary-feldspar calibrations, the average $T$ between the three end-member solutions ($T_{\text{G}}, T_{\text{U}}, and T_{\text{L}}$) is shown. For the gneissic rocks the centroid of intersection points in the results of Elkins & Grove (1990) is located at 3 ± 0.2 kbar. Therefore, there is an independent test on the reliability of the pressure value used with this geothermometer. Textural abbreviations: mtrx, rims of porphyroblast in the matrix; fksil, Pl + Kfs intergrowths within fibrolite mats; Grt II, inclusions within garnet II porphyroblasts; Z-Pl, zoned plagioclase porphyroblast with an outer rim of K-feldspar. n.e., out of range estimation, probably representing disequilibrium conditions during crystallization (see Green & Usdansky, 1986). Mineral abbreviations from Kretz (1983).

Modelling of the $P$-$T$ path in the KNaCaFMASH system

Table 4 lists the mineral compositions for Assemblage 2 biotite + sillimanite + K-feldspar + plagioclase ± garnet II + quartz, which is assumed to represent equilibrium conditions immediately after $D_3$ in the high-grade schist. From the thermobarometric studies we have assigned $P$-$T$ conditions of 675°C and 4 kbar. To overcome the observed disequilibrium relationships, we have back-calculated the modal abundance of the phases under these $P$-$T$ conditions by using bulk-rock chemistry data. The results are also included in the Table 4, and constitute the starting $P$-$T$-$X$-$M$ point for the modelling. We have computed in the KNaCaFMASH system the isopleths of mineral composition and mineral abundance.
Table 4: P–T–X–M values for the peak metamorphic assemblage in the high-grade schist of the Alboran basement, and P–T–X–M values calculated with the Gibbs method in the KNaCaFMASH system

<table>
<thead>
<tr>
<th>Assemblage</th>
<th>P (bar)</th>
<th>T (°C)</th>
<th>XAlm</th>
<th>XPhl</th>
<th>XAn</th>
<th>XMg</th>
<th>XKfs</th>
<th>XMds</th>
<th>XMts</th>
<th>XPyr</th>
<th>fe-no.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Grt + Bt + Sil + Kfs + Pl + Qtz</td>
<td>4000</td>
<td>675</td>
<td>0.82</td>
<td>0.25</td>
<td>0.62</td>
<td>0.06</td>
<td>0.40</td>
<td>0.09</td>
<td>0.75</td>
<td>0.07</td>
<td>0.92</td>
</tr>
<tr>
<td>Grt + Bt + Sil + Ms + Pl + Qtz</td>
<td>5000</td>
<td>675</td>
<td>0.38</td>
<td>0.20</td>
<td>0.09</td>
<td>0.18</td>
<td>0.38</td>
<td>0.85</td>
<td>0.08</td>
<td>0.11</td>
<td>0.91</td>
</tr>
<tr>
<td>Grt + Bt + Crd + Kfs + Pl + Qtz</td>
<td>3000</td>
<td>675</td>
<td>0.01</td>
<td>0.25</td>
<td>0.11</td>
<td>0.18</td>
<td>0.77</td>
<td>0.06</td>
<td>0.06</td>
<td>0.93</td>
<td></td>
</tr>
</tbody>
</table>

Mineral abundance is expressed as molar proportion of the phase (M), fe-no. = Fe/(Fe + Mg), XAlm = Fe/(Fe + Mg + Ca), XPhl = Mg/(Fe + Mg + Ca), XAn = Ca/(Ca + Na), XMs = Ms/(Ms + Ms + Ms), XMt = Mt/(Mt + Mt + Mt), and XKfs = Kfs/(Kfs + Kfs + Kfs).

Calculations have been done from the reference conditions at 675°C, 4 kbar, with the Gibbs method (Spear et al., 1982) and the thermodynamic database of Berman (1988) modified by Spear & Cheney (1989), using the programs provided by Spear et al. (1991). Starting conditions are those deduced from sample 28 (see Table 1) for Assemblage 2; i.e. garnet II (rim of garnet II porphyroblast; analysis B-104), biotite (included in garnet II porphyroblast; analysis B-120), plagioclase (rim of plagioclase II porphyroblast; analysis D-16), K-feldspar (pure), sillimanite, quartz, and H2O. Starting biotite composition corresponds to an inclusion in a large garnet II porphyroblast instead of a matrix porphyroblast, because biotite in the matrix was largely homogenized during retrogression, achieving high Fe contents (see text for discussion). Mineral abundances have been calculated by using bulk-rock chemistry data of this rock (which is considered a representative sample of the St-bearing high-grade schist; see text for discussion). Garnet (2%), biotite (35%), plagioclase (10%), K-feldspar (10%), sillimanite (20%), quartz (23%), and H2O (0%). For modelling purposes the hydrous cordierite (XH2O = 0.7) model of this database has been used. Mineral abbreviations from Kretz (1983).

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variant fields [Al$_3$] (8a), [St] (5b), [St, H$_2$O] (5a), and [Al$_3$] (5a), respectively. The boundary reactions of these trivariant fields are now divariant in the new system (for the calculation details, see Table 4), and correspond to the previously reported univariant reactions: Crd-in (7), Ms-out (6b), and St-out (4). As we reported for the KFMASH system, all the calculations have been done with the Gibbs method, using the programs provided by Spear et al. (1991). Because the assemblage at the starting point of the modelling is not in full equilibrium, however, the detailed results of the modelling can only be used to give a qualitative indication of the $P$-$T$ path.

The diagrams in Fig. 20 show the $P$-$T$ evolution which best reconciles the reaction history in the high-grade schist with the $P$-$T$-$X$-$M$ phase relations in the model KFMASH system. The depicted $P$-$T$ evolution has also been constructed following the more reliable results of the thermobarometric calculations. This path has three domains well constrained by textural, chemical, and thermobarometric analyses:

1. Prograde growth of garnet I porphyroblasts from the chlorite-in to the staurolite-in stability fields. This portion of the $P$-$T$ path occurs from 10–10.5 kbar to 6–8 kbar, near the Ky-Sil transition, as is revealed by GRAIL and GASP results. Along this path, $T$ probably rose from 500°C to 600°C to account for the observed garnet I zoning patterns.

2. Growth of garnet II porphyroblasts along a path with decreasing $P$ and rising $T$. This path occurs, as is documented by GARB, GASP, and GPBQ results, from the lower-$P$ portion of the staurolite-in stability field (6–7 kbar) to low-$P$ conditions (4 kbar at 650–700°C), once the muscovite dehydration reactions [6a], [6b], and (15) have been surpassed.

3. In the gneiss, we have evidence of higher-$T$ conditions during the decompression ($T > 700°C$) which resulted in a long-lived history of melting reactions, the possible participation of some biotite-breakdown reactions [e.g. (11) and (12)], and the crossing of the cordierite-forming reaction (7). From the present study, we can rule out the occurrence of this reaction in the high-grade schist.

4. The decompression was followed by near-isoharic cooling ($P < 3$ kbar), producing rapid crystallization of residual melts in the andalusite stability field ($T < 600°C$), and a late growth of muscovite in coexistence with andalusite.

The prograde portion of the $P$-$T$ path is shown tentatively as it is only supported by the zoning pattern and GRAIL pressure estimates on garnet I. Growth of inner zones of garnet I could have occurred from chlorite and biotite on a trajectory from 500°C, 10 kbar to 560°C, 8–5 kbar. These conditions satisfy GRAIL data and yield large amounts of garnet zoned down to an Fe/Fe + Mg ratio of ~0.92. After the intersection with the staurolite-

...zoning patterns we have described (Fig. 12a and b). Consequently, the Ab-rich composition of plagioclase considered at the starting point (Table 4) generates an isopleth useless for the consideration of phase relations.

Fig. 20. (a–b) P–T diagrams showing composition (expressed as mole fractions) and abundance (expressed as molar proportions of the phases) isopleth contours corresponding to the assemblages St + Ms + Grt + Ms + Grt + Bt + Pl + Qtz + H2O [Als] (3b), Ms + Als + Grt + Bt + Qtz [St, H2O] (5d), Kfs + Als + Grt + Bt + Pt + Qtz + H2O [St] (5c), and Crd + Kfs + Grt + Bt + Pt + Qtz + H2O [Als] (6b), in the KNa-CaFMASH system. The isopleths for the Crd-bearing assemblage have been omitted in most of the diagrams [with the exception of (e) and (g)] for simplicity and because of the lack of cordierite in the high-grade schist. Contours corresponding to the assemblage Chl + Ms + Grt + Bt + H2O are shown only for reference in (d) and (h) for the molar abundance and Fe/(Fe + Mg) composition of garnet. (i) P–T diagram showing the isopleths ofmonds, expressed as wt % of garnet, for the reactions (4) St + Ms + Qtz + Pl = Grt + Bt + Als + H2O and (7) Bt + Als + Qtz + Pl = Grt + Crd + Kfs + H2O. The proposed P–T path for the high-grade schist has also been plotted in the diagrams for discussion of predicted P–T–X–M phase relations (see text). Dashed portion of the path shows approximate P–T conditions along the prograde trajectory of the high-grade schist based on zoning patterns of garnet I porphyroblasts. Starting P–T–X–M conditions at 675°C and 4 kbar are shown by a white circle. Calculated intersections with divariant reaction (7), (6b), and (4) are marked with white squares, and their P–T–X–M conditions are listed in Table 4. Approximate composition of chlorite at the intersection with the divariant staurolite-in reaction (14) has been calculated from the invariant point [Crd, Als, Te] of Spear & Cheney (1989) at 589°C and 9.5 kbar, Xc, in chlorite 0.333, chlorite abundance 13.5 wt %). Calculations have been done using the Gibbs method of Spear et al. (1982) and Spear (1988, 1998), and the thermodynamic database of Berman (1988) modified by Spear & Cheney (1989). Numbers in parentheses refer to reactions described in the text. Al-silicate triple point from Berman (1988).
(5) With respect to phase relations under the lower-P
conditions of the P-T evolution of the gneissic rocks, we
can support with the modelling a large growth of cordi-erite with a fairly constant Xc, composition (0.30 ± 0.37)
(Fig. 20c and g). It is important to note that contours of
the molar abundance of cordierite allow for an increasing
and continuous growth of this mineral along the cooling
path in the gneiss (from T > 700°C and P ~ 3 kbar; 
Fig. 20g). If we explore the phase relations of the cor-
dierite-forming divariant reaction [equivalent to (7)] we
obtain a marked positive slope of the isopleths of moles
of garnet, which diverge with temperature (Fig. 20e).
Furthermore, we can deduce qualitatively that a large
amount of garnet could be produced under low-T con-
ditions (> 600°C) if this reaction is crossed (the same
conclusions could be depicted for moles of cordierite).
Consequently, some amount of garnet could be produced
in the gneiss by this reaction, and we reinforce the
assumption of relatively higher T during the de-
compression in the gneiss (T > 700°C) because reaction
(7) results probably in the production of Grt + Crd under
higher P than the reaction depicted by Spear & Cheney
(1989). In conclusion, the main assemblage in the gneissic
rocks requires T > 700°C at P > 4 kbar. An upper
temperature limit of 750–800°C is indicated by the
absence of orthopyroxene (Fig. 21). Hence we suggest
that the melting event in the gneissic rocks occurred
probably in the range 700–750°C and 4–6 kbar (Soto et
al., 1999).

Finally, during the cooling P-T path, Assemblage 3 was
formed in the high-grade schist including andalusite + K-
feldspar + plagioclase + quartz. We interpret the tex-
tural relations between the andalusite and the feldspars
as indicating crystallization of small amounts of granitic
melt in the andalusite stability field. Assuming the avail-
ability of water during crystallization, the lack of mus-
covite suggests that conditions still lay above the
muscovite breakdown curve. These considerations sug-
ject P-T conditions of T ~600°C and P < 2 kbar for
this assemblage. Cooling continued through 450–550°C
as indicated by the results obtained with the two-feldspar
thermobarometer. Very similar textures involving an-
dalusite also developed in the gneissic rocks, with the
difference that a much larger proportion of melt was
involved, and cordierite also crystallized together with
andalusite and the granitic assemblage. The P-T con-
ditions for crystallization of this late assemblage must
have been similar to those in the high-grade schist.

DISCUSSION AND CONCLUSIONS
The conclusion that the high-grade schist underwent
a significant rise in temperature during decompression
(Fig. 21) has profound tectonic implications, as it imposes
important constraints on the thermal structure and evol-
ution of the lithosphere during the exhumation of the
Alboran Sea basement. Although the uncertainties in our
P-T determinations for the various stages in the evolution
of these rocks are large, we believe that this conclusion
is robust. Overall, a rise in temperature of between 50°C
and 100°C has occurred during decompression from >8
kbar to <4 kbar. In view of the tectonic setting of these
rocks, the decompression is likely to reflect the extensional
tectonics that created the Alboran Sea.

The rise in temperature of the rocks during de-
compression requires that there was a substantial heat
source during this process [see discussion by Inger (1994)].
One possible heat source could be provided by magmatic
activity. In the Betic Cordillera intrusive rocks have a
limited areal extent, and consist of leucogranite and
basaltic dykes, both of Early Miocene age [18–20 Ma
for the leucogranite (Zeck et al., 1992); 22–23 Ma for the
basaltic rocks (Torres-Roldán et al., 1986)]. In the Alboran
Sea volcanism is particularly important in the East Al-
boran Basin (>50 km from Site 976) and is of middle to
late Miocene age [see summary by Comas et al. (1996)].
At Site 976 the relative volume of the leucogranite dykes
to the overall metapelite volume is, in addition, too small
to have acted as an additional heat source. In conclusion,
there is little evidence for magmatic activity of the right
age and scale near Site 976 to produce the observed
heating during the decompression history.

Thermal modelling by Platt et al. (1999) to account for
the P-T conditions presented here leads to the conclusion
that the only way to explain a continuous rise in tem-
perature during the decompression path is a combination
of a high radiogenic heat production in the initial
thickened crust and a complete removal of the litho-
spheric mantle. The most important boundary conditions
controlling this result are a time lapse (pause) between
crustal thickening and the start of the extension of the
order of at least 10 my, removal of the lithospheric
melt at <70 km depth, and subsequent lithospheric
stretching by a factor of three in ~6 my. Platt et al. (1999)
have also modelled the P-T paths that would characterize
other tectonic scenarios developed to explain the origin
of the Alboran Sea: namely (1) no removal of the litho-
sphere after thickening, followed by uniform extension
of the thickened region; (2) convective removal of the
lithospheric mantle below 125 km; and (3) oceanic-type
subduction–accretion with no lithospheric thickening and
without removal of the lithosphere. None of these models
can explain the metamorphic evolution of the Alboran
Sea basement, because they predict either isothermal
decompression followed by cooling, or decompression
P-T paths reaching peak temperatures at higher pressure
conditions (P > 6 kbar) than those reported here. The
almost complete removal of the lithospheric mantle, in
conclusion, seems to provide the most plausible additional
heat source to explain the $P\!-\!T$ path followed by the high-grade schist of the Alboran Sea basement. This interpretation of the $P\!-\!T$ path appears to favour models involving the delamination of lithospheric mantle [following Bird (1979)] as a factor contributing to the change from compressional to extensional tectonics in the region (Platt et al., 1999), and provides an additional and independent argument to other workers who have also suggested a similar origin for the Alboran Sea (Comas et al., 1992; García-Dueñas et al., 1992; Docherty & Banda, 1994; Seber et al., 1996).

The decompression $P\!-\!T$ path was followed by cooling through 600°C at ≤2 kbar (Fig. 21) forming Assemblage 3 ($\text{And} + \text{Kfs} + \text{Pl} + \text{Qtz}$). Ar/Ar dating of coexisting muscovites forming part of Assemblage 3 gives mean ages of 20.0 ± 0.2 Ma ($\pm 2\sigma$) for muscovite and 19.2 ± 0.7 Ma ($\pm 2\sigma$) for biotite Kelley & Platt, 1999). Apatite fission-track analyses on the same rocks give a mean central age of 18.3 ± 1.0 Ma ($\pm 2\sigma$) for the time of cooling below the apatite partial annealing zone (60°C) (Hurford et al., 1999). These data suggest that the Alboran Sea basement cooled from 426 ± 22°C (the estimated closure temperature of muscovite for Ar in these rocks) to 60°C in a period of between 0.5 and 2.9 my, during the early Miocene, giving a minimum average cooling rate of 126°C/my (Platt et al., 1999). Thermal modelling of this final cooling, taking both the petrological and radiometric data into account, indicates that it requires an average denudation rate of 4 km/my (Platt et al., 1999). A similar rapid cooling history seems to characterize many of the Alpujarride units of the Betic Cordillera (estimated cooling rates range between 100 and 350°C/my) (Zeck et al., 1992; Monié et al., 1994; Andreasen & Zeck, 1996; Sossou et al., 1998).

Considering both the results of the thermal modelling of the exhumation history as a whole and the absolute timing of the cooling portion of the $P\!-\!T$ path, Platt et al. (1999) estimated that extension started at ~27 Ma (Oligocene), and that exhumation of the Alboran basement to the surface took ~9 my ending at 18 Ma (upper Early Miocene).

Several important textural observations also indicate that the decompression $P\!-\!T$ path was rapid; namely, the preservation of widespread disequilibrium relationships, the occurrence of overstepped reactions, and the crystallization of andalusite in coexistence with a melt phase.
The preservation at low pressure of early minerals far outside their stability fields means that the common assumption of equilibrium among members of an assemblage of coexisting minerals is invalid. This completely rules out the application of integrated thermobarometric calculations based on internally consistent datasets, and makes the application of individual thermobarometers such as GARB and GASP difficult. Further, reactions among the minerals were commonly strongly over-stepped, and therefore did not proceed according to the theoretical sequence determined from petrogenetic grids. The breakdown of staurolite in the high-grade schist, for example, occurred partly outside the stability field of muscovite, and therefore involved a reaction different from the normal terminal reaction of this mineral, in which muscovite is consumed. Its breakdown probably led to the growth of garnet II, at a time when garnet growth would not normally be predicted. A third petrologically important effect of the preservation of phases outside their stability fields and the strong overstepping of reactions is conflicting textural evidence on relative timing of mineral growth. An example is the formation of deformed and partially resorbed augen of andalusite surrounded by foliated sillimanite in the gneiss. These textural relationships reflect deformation within the andalusite stability field of gneiss that contained both Al-silicates, concurrent with overstepped crystallization from melt of feldspars and cordierite causing consumption of andalusite.

An important additional conclusion arising from the petrological study of the Alboran basement is that melting was a direct consequence of the combination of rapid exhumation and heating. Considering the positive slope of the several muscovite dehydration melting reactions [see Fig. 21], such a decompressional \( P-T \) path enhances the generation of leucogranite magmas during exhumation, and confirms that the generation of migmatites at low-\( P \) conditions requires the contribution of an additional heat supply (De Yoreo et al., 1991).

Apart from the decompressional history of the high-grade schist, we have also reconstructed qualitatively the high-pressure part of the \( P-T \) path (based on phase relations of Assemblage 1 of the high-grade schist and the zoning pattern of garnet I), up to \( \sim 500^\circ \text{C} \) at 10–5 kbar. This portion of the path suggests an initial low geothermal gradient of 13–15°C/km (assuming an average crustal density of 2700 kg/m\(^3\)). Comparable low gradients for the initial compressional orogen have been proposed in some of the Alpujarride units of the Betic Cordillera, on the basis of carpholite + chloritoid + kyanite-bearing assemblages (\( \leq 17^\circ \text{C/km} \)) (Azañon & Goffè, 1997). We have found no evidence to constrain the increasing pressure part of the \( P-T \) path, nor do we know at what crustal level these rocks lay before the compressional event. It is clear, however, that these were not originally lower-crustal rocks.

The comparison between the \( P-T \) evolution presented here and those reconstructed in the Betic and Rif Cordilleras gives us evidence for the existence of a distinct thermal structure beneath the Alboran Sea region. The onshore Alpujarride and Nevado-Filabride units show thermal evolutions comparable from high-pressure conditions, preserved as relic eclogite and carpholite-bearing assemblages (e.g. Bakker et al., 1989; Puga et al., 1989; Tubia & Gil Ibarguchi, 1991; Azañon & Goffè, 1997), to low- and medium-pressure conditions. The decompressional \( P-T \) paths are characterized by a conspicuous pressure drop (with an estimated \( -\Delta P > 10 \) kbar) under isothermal conditions (e.g. Bakker et al., 1989; García-Casco et al., 1993; Monié et al., 1994; García-Casco & Torres-Roldán, 1996; Azañon et al., 1997, 1998; Balanya et al., 1997; Tubia et al., 1997). Similar isothermal decompression \( P-T \) paths characterize also the various units of the Sebídes in the Rif [see compilation by Michard et al. (1997)]. The parallelism among the published metamorphic evolutions for the various metamorphic units in the Betic and Rif, and the dissimilarity with the \( P-T \) path presented here, suggest some differences in the thermal evolution of the lithosphere during extension between the Alboran Sea basin and the surrounding regions. On the basis of the modelling results presented by Platt et al. (1999) and summarized above, we suggest a lesser amount of removal of the lithospheric mantle under the surrounding mountain chain areas. In addition, and in the light of the evidence discussed above for extensive metamorphic disequilibrium, future work on the Betic Cordillera and the Rif is required to define more precisely the metamorphic \( P-T \) evolution.

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