Paleozoic Basement and Pre-Alpine History of the Betic Cordillera

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A. Martín-Algarra, A. García-Casco, M. T. Gómez-Pugnaire, A. Jabaloy-Sánchez, C. Laborda-López, V. López Sánchez-Vizcaíno, S. Mazzoli, P. Navas-Parejo, V. Perrone, R. Rodríguez-Cañero, and A. Sánchez-Navas

Abstract

Palaeozoic rocks in the Betic Cordillera are widespread in the Maláguide, Alpujárride and Nevado-Filábride complexes of its Internal Domain. The Malaguide stratigraphic successions record a deepening trend during the post-rift evolution of a divergent continental margin that was paleogeographically related to the Northern Paleotethys from the latest Ordovician to the Early Carboniferous. Since the Serpukhovian it evolved to a convergent margin with Culm-like synorogenic sedimentation. Also probably at that time most of the Nevado-Filábride and Alpujárride rocks were affected by a Variscan tectonometamorphic evolution that was followed by latest Variscan local granite emplacement at ca. 300–280 Ma.

9.1 Introduction: State of the Art and Main Controversies

A. Martín-Algarra

In the Betic Cordillera Cadomian basement rocks and Cambrian-Ordovician successions have not been recognized, and very limited evidence of Precambrian marine sediments has been found only in the Nevado-Filábride Complex.

Coordinator: Agustín Martín-Algarra.

A. Martín-Algarra (✉) · P. Navas-Parejo · R. Rodríguez-Cañero
Departamento de Estratigrafía y Paleontología, Facultad de Ciencias, Universidad de Granada, 18071 Granada, Spain
e-mail: agustin@ugr.es

P. Navas-Parejo
e-mail: pilarnpg@geologia.unam.mx

R. Rodríguez-Cañero
e-mail: charorc@ugr.es

A. Martín-Algarra · A. García-Casco · M. T. Gómez-Pugnaire · C. Laborda-López · V. López Sánchez-Vizcaíno · A. Sánchez-Navas
Departamento de Mineralogía y Petrología, Instituto Andaluz de Ciencias de la Tierra (IACT), Universidad de Granada (UGR)-Consejo Superior de Investigaciones Científicas (CSIC), Avenida de las Palmeras 4, 18100 Armilla, Granada, Spain
e-mail: agcasco@ugr.es

M. T. Gómez-Pugnaire
e-mail: teresa@ugr.es

C. Laborda-López
e-mail: claborda@ujaen.es

V. López Sánchez-Vizcaíno
e-mail: vlopez@ujaen.es

A. Sánchez-Navas
e-mail: asnavas@ugr.es

A. García-Casco · M. T. Gómez-Pugnaire · A. Sánchez-Navas
Departamento de Mineralogía y Petrología, Universidad de Granada, 18071 Granada, Spain

A. Jabaloy-Sánchez
Departamento de Geodinámica, Universidad de Granada, 18071 Granada, Spain
e-mail: jabaloy@ugr.es

V. López Sánchez-Vizcaíno
Departamento de Geología, Escuela Politécnica Superior, Universidad de Jaén, Alfonso X el Sabio 28, 23700 Linares, Spain

S. Mazzoli
Dipartimento di Scienze della Terra, dell'Ambiente e delle Risorse (DiSTAR), Università di Napoli Federico II, Complesso di Monte S. Angelo, Via Cinthia 21, 80126 Naples, Italy
e-mail: stefano.mazzoli@unina.it

Present Address:
P. Navas-Parejo
Estación Regional del Noroeste, Instituto de Geología, UNAM, 83240 Hermosillo, Mexico

V. Perrone
Dipto di Scienze Geologiche e Tecnologie Chimiche e Ambientali, Università di Urbino, Località Crocicchia, 61029 Urbino, Italy
e-mail: vincenzo.perrone@uniurb.it

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Nonetheless, geochronological data from detrital and magmatic zircon reworked in younger sedimentary, metamorphic and magmatic rocks from different complexes reveal that the Betic Paleozoic successions were originally related to crustal segments that were previously involved in the Cadomian subduction and followed by Cambrian-Ordovician magmatic events, as observed in many other North-Gondwanic regions. Solid although yet limited biostratigraphic data from the Maláguide Complex reveals, however, a well-developed marine basin since, at least, the latest Ordovician. The Silurian-Devonian Maláguide sediments are widespread, relatively diversified and systematically pelagic, although clasts of Devonian to Carboniferous shallow-water carbonates are also present in uppermost Devonian and Carboniferous deep-water clastics. The Maláguide successions record a deepening trend during the post-rift evolution of a divergent continental margin that was paleogeographically related to Paleotethyan margins preserved in other Alpine regions. After some limited basin unstability during the Frasnian-Famennian, the margin reached its mature post-drif stage, dominated by thermal subsidence, in Early Carboniferous time, with deposition of Tournaisian radiolarian cherts and Visean conodont limestones. Since then, an important tectonic event marked the onset of the synorogenic (Late) Variscan evolution in the Maláguide Complex. It was followed by deposition of thick clastic wedges, showing the classical Culm facies, which started to be deposited since the Serpukhovian or, more probably, after the beginning of the Bashkirian. Also probably during that time most of the Nevado-Filábride and Alpujárride rocks were affected by an intense Variscan tectonometamorphic evolution, which was followed by latest Variscan local granite emplacement at ca. 300–280 Ma. However, many aspects of the Betic pre-Mesozoic evolution remain controversial.

The Paleozoic successions and pre-Alpine geological history of the Betic Cordillera (Fig. 9.1) are poorly known. Paleozoic rocks are rarely present in the ***External Zones*** and the ***Campo de Gibraltar Flysch Complex*** and mostly as reworked sedimentary components in younger successions. On the contrary, they are widely present in the ***Internal Zones***, below Mesozoic successions (frequently metamorphosed) of many Alpine thrust nappes. The lithological monotony, scarcity of fossils, intensity and style of deformation, and metamorphism make difficult to recognize the original features of the different successions and, consequently, unravelling their Paleozoic history. Notwithstanding controversial aspects, recent studies make increasingly evident that (i) the intense Alpine evolution that overprints the Paleozoic features of Betic successions has not completely erased them, and that (ii) their pre-Alpine history can be reconstructed, at least partially.

Paleozoic rocks in the ***External Zones*** practically do not exist in outcrop. Only a few, decametre to hectometre sized fragments of putative Paleozoic rocks have been documented within the Antequera Subbetic Chaotic Complex, isolated within mélange-type rocks dominated by brecciated Triassic gypsum (Peyre 1974). In spite of lacking outcrops, a general consensus exists about the nature of the Paleozoic basement rocks below the *Prebetic* and *Subbetic* cover thrust sheets of the External Zones: seismic and some deep well data certainly show that they constitute the south-eastern prolongation of the Variscan Iberian Massif. This basement was thinned during the opening of the Mesozoic ***Southern Iberian Paleomargin***, which was the paleogeographic realm where the terrains of the External Zones were originally deposited (Vera 2004).

The Betic (and Rifian) Internal Zones constitute, at least in wide parts (see below), the outcropping crustal segment of the same lithosphere than that of the Alboran sea and, together with the basement of the Alboran Sea, they are included, since Andrieux et al. (1971), in the same tectonostratigraphic terrane, named ***Alboran Block*** or ***Domain*** (Balanyà and García-Dueñas 1987; Guerrera et al. 1993; Gómez-Pugnaire et al. 2012; Behr and Platt 2012; Augier et al. 2013). The Betic Internal Zones are subdivided in four Alpine thrusts sheet complexes named, from the deepest to the highest, ***Nevado-Filábride***, ***Alpujárride*** and ***Maláguide***, in addition to a group of ***Frontal Units*** located along the boundary between the Internal and External Zones (Vera 2004). The Nevado-Filábride, Alpujárride and Maláguide complexes constitute the main body of the Internal Zones forming an antiformal nappe stack of pre-Mesozoic and Meso-Cenozoic rocks, whereas the Frontal Units are exclusively made of post-Paleozoic cover rocks. The lithosphere of the Alboran Sea is made of a strongly thinned continental crust constituted by Alpujárride- and Maláguide-like terrains. In addition, the upper units of the Alpujárride Complex include the largest outcropping bodies in the World of subcontinental mantle rocks, the Ronda peridotites. The characterization and interpretation of the *pre-Alpine* processes that affected the Paleozoic successions associated with the peridotites play a crucial role for understanding of the *Alpine* geodynamic evolution of these mantle slices.

The pre-Mesozoic lithologic successions of the Betic Internal Zones are strongly deformed and affected by metamorphism of different grade (Vera 2004). Those of the Maláguide Complex yet preserve many original sedimentary features and fossils that allow accurate stratigraphic dating and environmental interpretation. Those of the Nevado-Filábride and Alpujárride complexes, however, are fully constituted by metamorphic rocks, mainly pelito-psammitic metasediments, with subordinate metaconglomerates, marbles and, occassionally, volcanic, subvolcanic and plutonic rocks, either acid and intermediate or mafic and ultramafic. The latter were transformed by the metamorphism to gneisses and amphibolites (or eclogites) and serpentinites,

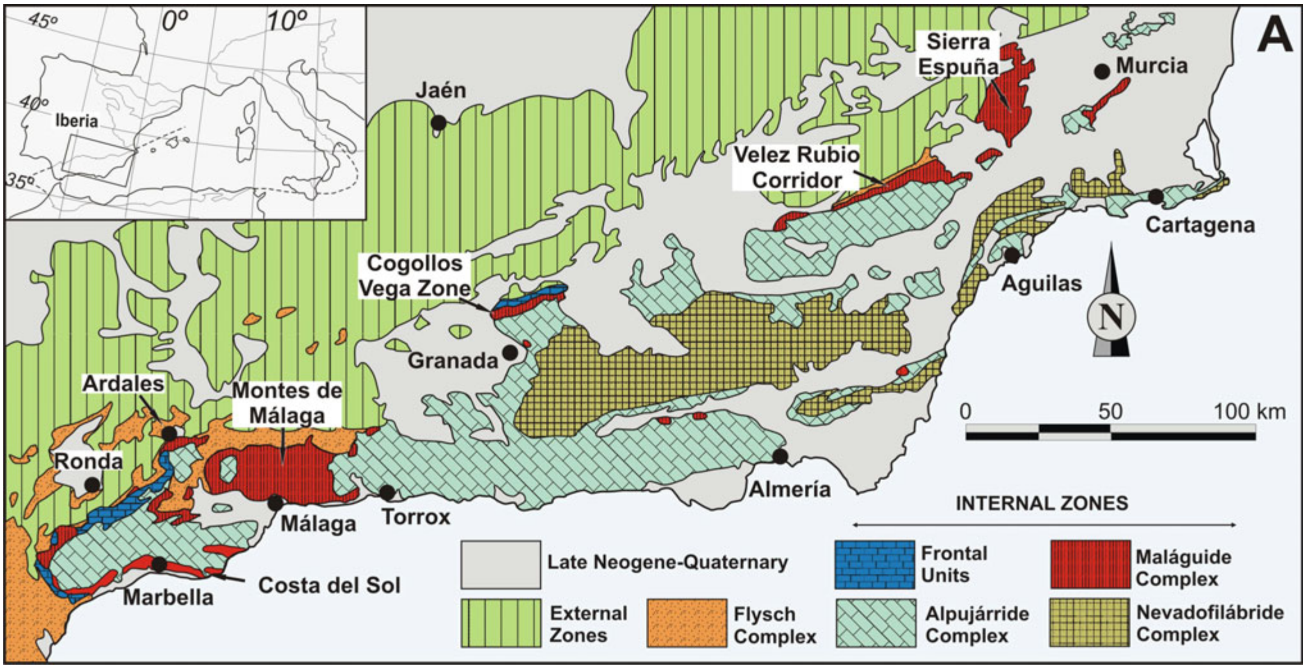

Fig. 9.1 Main divisions of the Internal Zones of the Betic Cordillera

respectively. These successions lack fossils, except in some localized outcrops. The combined stratigraphic-petrologic-structural analysis of the oldest metasedimentary successions of the three complexes and geochronologic dating from gneisses of diverse units, in particular by U-Pb zircon dating, make increasingly evident that they were affected by pre-Alpine (Variscan and maybe older) tectonic, metamorphic and magmatic events.

The main controversies on the Betic pre-Mesozoic successions deal with: (1) the stratigraphic interpretation and correlation of rock successions; (2) the distinction among petrographic and deformational features related to pre-Alpine and to Alpine events; (3) the significance of geochronological ages obtained from different rock types and by different methods; (4) the integration of available data in crustal/lithosphere scale geodynamic models compatible with the pre-Alpine and Alpine plate tectonic evolution.

It is yet unclear if the stratigraphy and paleogeographic evolution of the Betic Paleozoic terrains must be related to continental paleomargins and basins associated with the plate tectonic evolution of the Rheic Ocean, like those of the Iberian Massif, or to margins and basins associated with the evolution of the Prototethys and Paleotethys Oceans, as it is accepted for most Paleozoic terrains of the Alpine belt (Stampfli and Borel 2004; Stampfli et al. 2013 and references therein). Actually, most–but not all–Internal Betic Paleozoic successions show much stronger geological affinities with coeval succession of other Alpine areas (Alps, Maghrebides, Apennines, Carpathians…) than with those of the Variscan Iberian or Nord-African Paleozoic Massifs of the foreland of the Alpine-Mediterranean belts (Martín-Algarra et al. 2000, 2004; Rodríguez-Cañero et al. 2010; Navas-Parejo 2012; Rodríguez-Cañero and Martín-Algarra 2014).

It is also unclear the timing and mode of exhumation of the Alpujárride subcontinental mantle rocks. The tectonic emplacement of the Ronda peridotites accounted in the frame of a subduction-related Alpine metamorphism of early Miocene age that also involved Mesozoic covers of the Alpujárride Complex and of some Frontal Units of the Internal Zones (Nieves Unit in particular: Mazzoli and Martín-Algarra 2011; Mazzoli et al. 2013). But this emplacement involved *extremely thinned* pre-Mesozoic Alpujárride crustal successions, in which recent geochronological data are demonstrating the existence of pre-Alpine HP and HT metamorphic and anatectic events (Sánchez-Navas et al. 2017 and references therein). Consequently, some authors interpret that the important thinning that affects the Alboran Domain crust could be at least partially *inherited* since (late) Paleozoic time and that it would not be *exclusively* related to Alpine extensional collapse of a previously thickened lithosphere as interpreted by most authors (e.g., Précigout et al. 2013; Gueydan et al. 2015; Frasca et al. 2017; Williams and Platt 2017, and references therein). The Paleozoic tectonometamorphic signatures of the Betic Paleozoic successions must have conditioned the Alpine evolution at crustal and lithospheric scales. The real problem arises from incomplete understanding of the role played by each orogeny in the configuration of the Internal Betic realms.

9.1.1 Controversies Related to the Stratigraphy and Structure of the Nevado-Filábride Complex

The Nevado-Filábride Complex is usually divided into two main groups of units that are grouped in the ***Veleta*** and ***Mulhacén*** tectonic ensembles (Puga et al. 1974, 2002). There is no agreement, however, neither on the cartographic distribution of these units nor on their lithological succession and stratigraphy, and even some authors reject this twofold subdivision and interpret the whole complex as a single tectonic unit (González-Casado et al. 1995; García-Dueñas et al. 1998a, b; Sanz de Galdeano and López-Garrido 2016a; Sanz de Galdeano et al. 2016; Santamaría-López and Sans de Galdeano 2018). The occurrence of Variscan magmatic rocks in the graphite bearing lower successions of the Mulhacén (Sierra Nevada), and in equivalent units of eastern areas, is well constrained, and the Paleozoic age of the host metasediments unanimously accepted. Moreover, some Veleta-type graphite-rich low-grade metapelites and black marble horizons have provided Paleozoic and older fossils (see below). Nevertheless, the interpretation of most upper Mulhacén successions, which are lithologically very heterogeneous, is controversial (see Sect. 2.1).

Different types of micaschists quartzites, calcschits and marbles constitute the predominant lithotypes in the Mulhacén successions (Puga 1977, 1990; Estévez and Pérez-Lorente 1974; Díaz de Federico et al. 1979; Gómez-Pugnaire 1981; Puga et al. 2002, 2011; Gómez-Pugnaire et al. 2012). The schists sometimes are dark-coloured, graphite bearing and/or biotite-rich, but also light-coloured, muscovite-rich and/or chlorite-rich. The marbles are locally massive, quite pure and several hundred of metres thick, but they are also, and mostly, impure and alternate with different types of metapelites, calcschists and amphibolites. Typically, these successions also include former magmatic rocks bodies, both felsic and mafic, with ultramafic lenses, all now systematically metamorphosed under high-P conditions. The felsic rocks are massive and layered orthogneisses, the latter frequently very rich in tourmaline (Nieto 1996; Nieto et al. 2000; Martínez-Martínez et al. 2010; Gómez-Pugnaire et al. 2012, and references therein). The felsic rocks have provided late Paleozoic magmatic U-Pb zircon ages (see Sect. 9.4). The massive orthogneisses are associated with graphite-rich metapelites unanimously interpreted as Paleozoic metasediments. Nevertheless, the layered gneissic lithotypes appear in the uppermost part of the Nevado-Filábride succession. They are commonly intercalated within light-coloured metasediments, including marble horizons in its upper part, which underlie a thick metasedimentary, mainly carbonatic succession, whose age is considered as, essentially, Permian although some authors have interpreted these layered lithotypes as Permian-Triassic vulcanoclastic rocks (Andriessen et al. 1991; Nieto 1996; Nieto et al. 2000). The mafic rocks are mainly amphibolites or amphibolitized eclogites with locally well preserved magmatic features evidencing that their protoliths were gabbros or basalts, sometimes pillowed (Morten and Puga 1984; Morten et al. 1987; Bodinier et al. 1987; Puga 1990; Gómez-Pugnaire and Muñoz 1991; Puga et al. 1989a, b, 1995, 1997, 1999, 2005, 2011, 2017). The ultramafic rocks are serpentinites and harzburgite metaperidotites (Burgos et al. 1980; Jabaloy-Sánchez et al. 2015 and references therein). The age of the mafic and ultramafic rocks and, especially, that of the upper Mulhacén metasedimentary successions is controversial. According to stratigraphic correlations with the Triassic Alpujárride succession, most authors interpret that the age of the metasediments (mainly marbles, sometimes bearing gypsum) above the layered gneisses and associated rocks reached the Permian-Triassic (Gómez-Pugnaire and Cámara 1990; De Jong and Bakker 1991; Gómez-Pugnaire et al. 1994). Other authors interpret part of the impure marbles and calcschists as Jurassic-Cretaceous metasediments originally deposited in deep marine environments deposited within an oceanic branch of the Western Tethys (Tendero et al. 1993; Puga et al. 2017), but others consider the same metasediments as part of the Paleozoic succession (Gómez-Pugnaire et al. 2012). Nevertheless, very recent studies on detrital zircon data on both sides of the Gibraltar Arc, coeval to the writing of this chapter, tend to accept the post-Paleozoic age (Triassic and younger) of at least part of the upper Mulhacén successions (Jabaloy-Sánchez et al. 2018; see also Azdimousa et al. 2019).

The geological meaning of the mafic and ultramafic Nevado-Filábride rocks is also another important unsolved topic because there is not agreement about the geodynamic and paleogeographic meaning of the original magmatism: for some authors it was only related to a continental rifting, for others the rifting progressed to seafloor spreading related to the Mesozoic opening of the Tethys and would be responsible for producing an ophiolitic succession (Puga et al. 2017, Lozano-Rodríguez 2018, and references therein). The upper Mulhacén mafic and ultramafic rocks and related metasediments would then constitute ophiolitic relics of a thinned *Mesozoic* lithosphere now tectonically sandwiched between continental crust successions belonging to opposite continental margins and Mesozoic plates, and transformed in a subduction-related tectonometamorphic mélange. In any case, there is increasing evidence supporting that most of the Nevado-Filábride Complex must be excluded from the Alboran Domain lithosphere, and that its lower part must be interpreted as a crustal section of the Iberian lithosphere now outcropping in a tectonic window in the very core of the Betic orogen.

9.1.2 Controversies Related to the Other Complexes of the Betic Internal Zones

One of the main controversies involving structurally higher Betic internal complexes deals with the distinction of pre-Alpine tectonometamorphic signatures in Alpujárride rocks that were strongly affected by Alpine metamorphism. Actually, different and sometimes contradictory PTt paths have been proposed for the metamorphic evolution of Alpujárride rocks. According to recent petrological research and geochronological dating, a polymetamorphic and poly-orogenic Alpine and pre-Alpine history, Variscan in particular, is becoming increasingly evident in the Alpujárride Paleozoic rocks (Zeck and Whitehouse 1999, 2002; Zeck and Williams 2001; Sánchez-Navas et al. 2012, 2014, 2016, 2017; Acosta-Vigil et al. 2014; Barich et al. 2014; Ruiz-Cruz and Sanz de Galdeano 2014a, b). Moreover, the importance of the Variscan Orogeny in the Maláguide Complex, the highest one in the Alpine tectonic stack, is clearly highlighted by the local preservation of pre-Alpine thrust sheets involving stratigraphically different Paleozoic successions (Martín-Algarra et al. 2009a, b), but this is usually ignored or underestimated (e.g. Williams and Platt 2017).

Another controversy deals with correlation of the Paleozoic (and post-Paleozoic) Internal Betic successions with other Iberian and Alpine terrains in North Africa, Apennines and Alps. If a Jurassic-Cretaceous ocean opened within the Nevado-Filábride Complex to the SE of Iberia before the onset of the proper Alpine Orogeny, the Paleozoic and younger terrains of the (true) Alboran Domain tectonic units would be originally located far towards the east, then in areas closer to the Alps and to the Apennines than to Iberia, and this putative ocean would have its continuation towards the Alps (compare Gómez-Pugnaire et al. 2000, 2004, 2012, with Puga et al. 2002, 2011, 2017). Consequently, the pre-Alpine (Variscan and older) evolution of these terrains (including upper Nevado-Filábride units and the whole Alpujárride and Maláguide complexes) should be different from that of the Iberian Variscan terranes and related units of the Nevado-Filábride Complex.

The paleogeography, geodynamics and tectonometamorphic evolution of the Alboran Domain is controversial during the Meso-Cenozoic (cf. Vol. 3, Ch. 7) but these aspects are practically unknown in pre-Mesozoic time, indeed. Actually, the proper existence of the Betic Palaeozoic terrains is neglected in widely difussed models devoted to the Palaeozoic evolution of the Alpine and Variscan regions in Southern Europe and North Africa, or misunderstood with that of other Iberian Variscan terranes (e.g. Von Raumer et al. 2002, 2003; Stampfli and Borel 2004; Sampfli and Hochard 2009; Stampfli et al. 2002, 2011, 2013; Torsvik and Cocks 2013). Only a few recent papers deal with this problem and they are mainly focused on the paleogeography of Maláguide sedimentary successions, whose stratigraphy and the understanding of their palaeogeographic and geodynamic evolution in relation to other Alpine Paleozoic terranes are rapidly improving (Rodríguez-Cañero et al. 2010, 2013; Navas-Parejo 2012; Navas-Parejo et al. 2008, 2009a, b, 2011, 2012a, b, 2015a, b; Somma et al. 2013; Rodríguez-Cañero and Martín-Algarra 2014).

In relation to the pre-Alpine tectonometamorphic evolution, recent studies are demonstrating that the effects of the pre-Alpine metamorphism and magmatism were very intense in the Alpujárride pre-Mesozoic successions. In addition, a probable pre-Alpine rise of the Ronda-Beni Bousera peridotites, and their emplacement in the shallow lithosphere in relation to a very important late Variscan crustal thinning, is becoming increasingly evident since some seminal papers (Reuber et al. 1982; Michard et al. 1997; see Sánchez-Navas et al. 2017 and references therein). In a different way, some authors have proposed that the Ronda peridotites were exhumed at surface in early Mesozoic time (Sanz de Galdeano and Ruiz-Cruz 2016; Sanz de Galdeano and López-Garrido 2016b) and that the Alpujárride continental crust rocks lying above them were previously affected by UHP metamorphism responsible for diamond and coesite within garnets (Ruiz-Cruz 2011b; Ruiz-Cruz and Sanz de Galdeano 2012b, 2013, 2014a, b). Accordingly, the Alpujárride pre-Alpine tectonometamorphic evolution and crustal thinning should be deeply re-evaluated. Moreover, the signatures related to the pre-Alpine emplacement of the peridotites at the base of the Alpujárride crustal succesions must be clearly distinguished from those related to the well documented Alpine thrust emplacement of the ultramafic slices within the upper Alpujárride crustal segment.

9.2 Pre-Mesozoic Successions of the Nevado-Filábride Complex

A. Martín-Algarra, M. T. Gómez-Pugnaire,
A. Jabaloy-Sánchez, R. Rodríguez-Cañero,
V. López Sánchez-Vizcaíno, P. Navas-Parejo,
C. Laborda-López

The Nevado-Filábride Complex is mainly exposed in the core of three main E–W trending antiforms and in separate outcrops toward the East (Fig. 9.1). Several tectonic subdivisions with diferent local names have been proposed. The most commonly used nomenclature distinguishes two tectonic ensembles (Puga and Díaz de Federico 1976; Puga et al. 1974, 2002): Veleta (footwall) and Mulhacén (hangingwall), which are subdivided in smaller tectonic units by other authors (Nijhuis 1964; Voet 1967; Bicker 1966; Egeler and Simon

1969; Puga et al. 1974, 2002; Kampschuur 1975; Vissers 1981; Bakker et al. 1989; De Jong and Bakker 1991; Martínez-Martínez et al. 2002; Gómez-Pugnaire et al. 2012, etc.). Both include thick and monotonous successions of graphite-bearing dark schists (locally including other subordinate rock types) of undisputable Paleozoic and older age. The dark schists of both the "*Veleta Succession*" and the "*Lower Mulhacén Succession*" are unanimously interpreted as pre-Alpine basements affected by Variscan metamorphism and overprinted by Alpine metamorphism. The "*Upper Mulhacén Successions*" are lithologically much more varied, structurally very complex and of controversial age and origin: they are dominated by light-coloured schists and marbles, and by magmatic rocks, both acid and mafic, also including some small ultramafic tectonic lenses. They have been classically interpreted as part of a Permo-Triassic and perhaps younger Alpine sedimentary cover (Egeler and Simon 1969; Puga et al. 2002), but other authors interpret that they are of exclusively Paleozoic age (Gómez-Pugnaire et al. 2000, 2004, 2012).

9.2.1 Veleta Succession

A. Jabaloy-Sánchez, A. Martín-Algarra,
R. Rodríguez-Cañero, P. Navas-Parejo, C. Laborda-López

The Veleta Succession is several kilometres thick, partly due to tectonic causes, and characterized by monotonous graphite-rich fine-grained black schists and phyllites, with increasingly abundant quartzites towards the top of the succession (Jabaloy-Sánchez 1993; Martínez-Martínez 1986; Puga et al. 2002, 2004). Ocasionally, the schists intercalate laterally discontinuous graphite-rich black marbles (Puga 1971; Puga and Díaz de Federico 1976; Díaz de Federico et al. 1979; Vissers 1981; Martínez-Martínez 1986; Álvarez-Lobato and Aldaya 1985), decimetres to a few metres thick bands of rhyolitic orthogneisses (Nieto et al. 2000) and sparse amphibolite lenses that, according to their compositional and isotopic features, correspond to former basaltic magmas generated in intracontinental mantle settings (Puga et al. 2002).

The black schists of the Veleta Succession in Sierra de Baza provided probable Neoproterozoic acritarchs (*Gloecapsomorpha* sp. and *Trematospahaeridium* sp.: Gómez-Pugnaire et al. 1982). The graphite–rich marbles are rare in the Veleta Succession of Sierra Nevada but in Sierra de Baza and in the Águilas-Cartagena region they are frequent and have provided pre-Mesozoic fossils. In the latter sector, the carbonate levels are locally thick (from 0.5 to 5 m). They concentrate in the middle part of the graphite-rich schists and have provided Devonian (Emsian-Eifelian) macrofossils: chaetetids (Lafuste and Pavillon 1976), crinoids and unclassifiable remnants of rugose corals, cephalopods, gastropods, brachiopods, trilobites and other fossils (Laborda-López et al. 2015a, b). In the Sierra the Baza, a low-grade metamorphic succession formed by fine-grained black schists very rich in graphite, including thin-bedded dark marbles and thick quartzites towards the top, underlies typical Veleta schists. The marbles have recently provided the first conodonts of the Nevado-Filábride Complex, of early Bashkirian age (*Declinognathodus inaequalis* Zone: Rodríguez-Cañero et al. 2018) and the quartzites contain well-preserved sedimentary structures (graded laminae, current ripples, parallel and cross bedding: Jabaloy-Sánchez 1993). These new data allow interpreting the Veleta Succession in the Aguilas-Cartagena area as deposited in marine (mainly terrigenous, subordinately carbonatic) shallow platform environments above storm base during the Devonian (Laborda-López et al. 2015b). In the Sierra de Baza the Late Carboniferous sedimentation accounted in deeper environments under calm and poorly oxygenated to anoxic bottom water conditions, although the seabottom was occassionally affected by slow bottom currents (Rodríguez-Cañero et al. 2018).

9.2.2 Lower Mulhacén Succession

A. Martín-Algarra, M. T. Gómez-Pugnaire,
A. Jabaloy-Sánchez

The Lower Mulhacén Succession is made, predominantly, of dark coloured graphite-rich schists with subordinate quartzites, quite similar to those of the Veleta Succession but without amphibolite lenses. The most significant difference is the local presence, in the Mulhacén black schists, of very large (cm-to dm-sized) and sometimes very well preserved pre-Alpine minerals (for details see Sect. 9.3.2) generated under static metamorphic conditions (Díaz de Federico and Puga 1976; Gómez-Pugnaire and Sassi 1983; Díaz de Federico et al. 1990). These porphyroblasts (Fig. 9.2) preserve microstructures defining two pre-Alpine phases of blastesis/deformation (Estévez and Pérez-Lorente 1974). Nonetheless, in most outcrops (in particular in the Sierra de Filabres) the porphyroblasts are strongly deformed, flattened, stretched and pseudomorphosed by micaceous lenses forming large white spots along rock surfaces defining the Alpine foliation (Puga et al. 1975; Díaz de Federico and Puga 1976; Puga et al. 2002, 2004). Hectometre sized ortogneissic (metagranite) bodies also appear locally included within these black schists (Nijhuis 1964; Bicker 1966). In the largest known (kilometre-sized) body, which is found in the eastern Sierra de Filabres (*Lubrín-Bedar Gneiss*), the original plutonic and pegmatoid magmatic structures are locally well preserved in spite of the intense Alpine deformation (see Sect. 9.4).

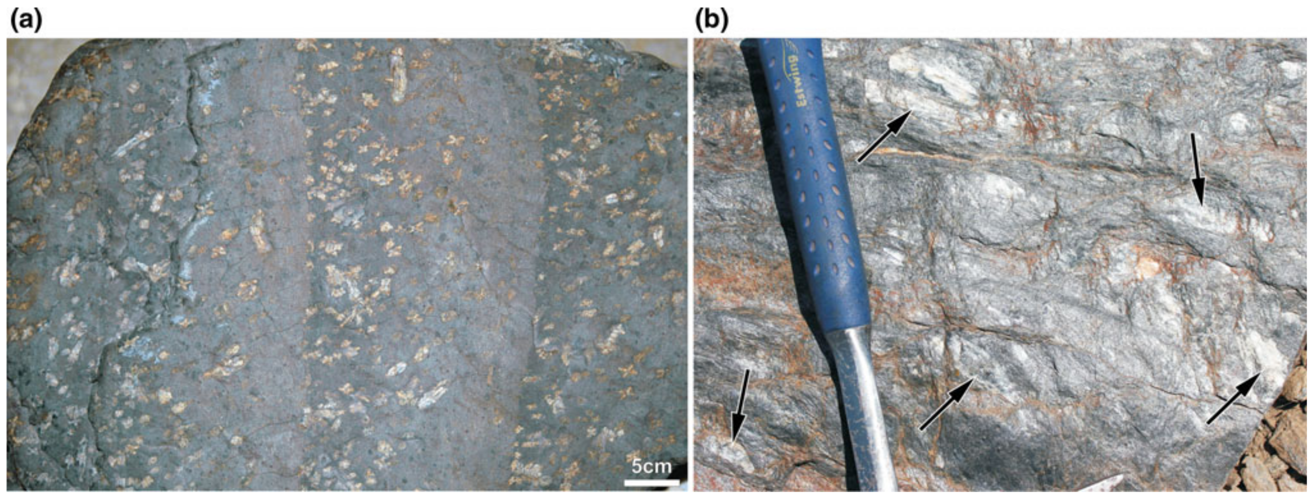

Fig. 9.2 Field views of pre-Alpine minerals of the Mulhacén Paleozoic succession: **a** grahite-rich schist with palm-like disoriented andalusite chiastolites (now fully transformed to Alpine kyanite), garnet, staurolite and chloritoid (smaller spots within the black matrix). **b** Alpine foliation surface showing white spots (arrows) corresponding to flattened and strongly sheared pre-Alpine andalusite crystals now transformed to fine-grained aggregates of Alpine muscovite and kyanite

9.2.3 Upper Mulhacén Succession

A. Martín-Algarra, M. T. Gómez-Pugnaire,
A. Jabaloy-Sánchez, V. López Sánchez-Vizcaíno

Above the black schists of the Lower Mulhacén Succession, a thick succession of light-coloured micaschists and quartzites constitutes the ***Tahal Schists*** (Nijhuis 1964; Kampschuur 1975; De Jong and Bakker 1991). Near to its base, this succession locally bears laterally discontinuous and *strongly deformed metaconglomerate lenses* (Fig. 9.3a), which probably constitute the base of a continental, probably alluvial succession unconformably deposited on their substratum of black schists (Gómez-Pugnaire et al. 2000). The widest outcrops of the proper Tahal Schists appear in the central part of the Sierra de Filabres (Vissers 1981). They constitute a thick (800–1000 m) monotonous sequence of silvery-gray muscovite schists with intercalated blue gray and white thick-to thin-layered quartzites, which are more abundant in the lower part of the succession. The quartzites frequently preserve cross lamination, cross bedding (Fig. 9.3b), ripples and bioturbation (De Jong and Bakker 1991; Gómez-Pugnaire et al. 2000). Silvery light gray metapelites with rare intercalations or marbles (>1 m thick) and calc-schists become predominant towards the top, although quartzites are always present. Some gneissic bodies with associated tourmaline-and garnet-skarns boudins and layers of garnet-tourmaline micaschists also appears in the upper part, as well as dikes of metabasites (Gómez-Pugnaire et al. 2012).

The Tahal Schists change upwards to a laterally discontinuous and usually *strongly brecciated calcite-dolomite marble succession, locally bearing gypsum and scapolite* (Gómez-Pugnaire et al. 1994; Puga et al. 1996), as well as metabasite lenses and dikes intruding different types of fine-grained schists and calc-schists. This complex lithologic ensemble has received many different local names (Nijhuis 1964; Voet 1967; Vissers 1981; De Jong and Bakker 1991; Puga et al. 1996; Martínez-Martínez et al. 2002; Puga et al. 2004 and references therein). This succession was deposited in evaporitic, probably coastal environments. The tectonic versus sedimentary origin of this succession and its age are controversial: (i) tectonic brecciation of evaporite-bearing Permian-Triassic rocks (Leine 1968; Vissers 1981); (ii) brecciation resulting from partial dissolution of rocks with high content of soluble salts (Duplaix and Fallot 1960; Jabaloy-Sánchez 1993; Gómez-Pugnaire et al. 1994); (iii) hybrid origins (Leine 1968; Bourgois 1979; Orozco et al. 1999); or (iv) deposition of an intra-orogenic volcano-sedimentary andesitic succession of Paleogene age that was subsequently tectonically brecciated (Puga et al. 1996). The presence of high-pressure minerals (kyanite-talc-phengite pseudomorphs) formed during the first high-pressure alpine metamorphic event in the very fine-grained metapelites does not fit well with the latter interpretation, however.

The uppermost part of the Upper Mulhacén succession is dominated by massive and pure calcitic and dolomitic marbles that overlie a heterogeneous succession of calcschists alternating with impure marbles, garnet micaschists (sometimes rich in graphite), chloritic and amphibolic micaschists, layered tourmaline-rich gneisses, metabasites and, occassionally, serpentinites. These lithologies show rapid lateral and vertical changes due to both stratigraphic and tectonic causes, the later resulting from intense syn- to postmetamorphic folding. Massive and pure calcitic and/or dolomitic marbles with white and/or banded white-gray colours are predominant in the uppermost part of the succession (***Macael Marbles***). Thinly to

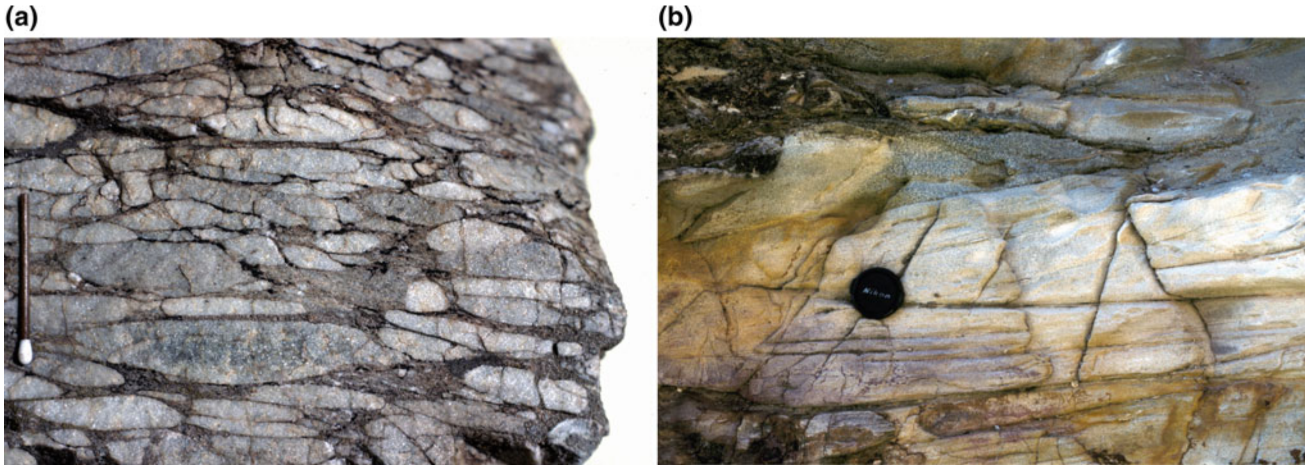

Fig. 9.3 Field views of the Tahal succession: **a** Stretched metaconglomerate of the lower part of the Tahal Schists succession (Sierra de Filabres). **b** Quartzites showing well preserved cross bedding (outcrop near Lubrin)

medium layered impure marbles lacking evaporites and rich in rounded detrital quartz grains, are essentially calcitic, usually dark coloured (grayish, brownish). They appear closely related to metabasites and serpentinites, and are associated with different types of yellowish-greenish calc-schists and micaschists with some quartzite horizons. They also, locally include a few meters thick bands of garnet-bearing, graphite-rich slightly calcareous dark schists. In the Cóbdar area the brown marbles include very thin layers with epidote, phengite, chlorite, amphibole, garnet, titanite and rutile, and some of them have unusually high Cr contents (López Sanchez-Vizcaíno et al. 1995) probably derived from detrital grains. In the same area, the impure marbles and calcschists intercalate a few metres thick bands of micaschists rich in graphite, and contain quartz nodules and bands (Fig. 9.4). These rocks bear rounded fine sand-sized detrital quartz grains and have provided ankeritic objects evoking marine pelagic microfossil remnants (Tendero et al. 1993): their size and shape in thin sections suggest mid-Cretaceous planktonic foraminifera, among them some fitting with *Helvetoglobotruncana helvetica* BOLLI, of Turonian age (Fig. 9.5). Nonetheless, this attribution is controversial, as Gómez-Pugnaire et al. (2000) interpret these remnants as benthic foraminifers of a wide geological time span.

9.3 Pre-Alpine Metamorphism in the Nevado-Filábride Complex

M. T. Gómez-Pugnaire, A. Martín-Algarra, A. Jabaloy-Sánchez, V. López Sánchez-Vizcaíno

There is a general consensus about the existence of a pre-Alpine (Variscan) tectonometamorphic evolution in the Nevado-Filábride Complex. Nevertheless, some authors interpret that the Veleta Succession was exclusively affected by pre-Alpine metamorphism (Gómez-Pugnaire and Franz 1988) whereas others consider that the pre-Alpine event was also overprinted by Alpine metamorphic events (Puga et al. 2002) or interpret the whole evolution of the Veleta successions as related to, exclusively, an Alpine polyphasic evolution (Augier et al. 2005a; Behr and Platt 2012). The dark micaschists and quartzites of the Lower Mulhacén Succession show clear evidence of polyorogenic metamorphism that is unanimously accepted: these rocks underwent Variscan LP/IT metamorphism, which was overprinted by HP Alpine metamorphism (Puga and Díaz de Federico 1976; Gómez-Pugnaire and Sassi 1983; Puga et al. 2002; Martínez-Martínez et al. 2002).

9.3.1 Veleta Succession

M. T. Gómez-Pugnaire, V. López Sánchez-Vizcaíno

The Veleta rocks are dark, fine-grained micachists and quartzites, with garnet, chloritoid, plagioclase, white mica, green and brown-red chlorite, quartz, graphite, ilmenite, zircon, tourmaline and apatite; small amounts of biotite and stilpnomelane intergrowths are also locally present (Nijhuis 1964; Vissers 1981; Martínez-Martínez 1986). Phase relationships of the main minerals are sketched in Fig. 9.6. These rocks are affected by four deformational events. Dominant fabrics are related to D_2 and produced isoclinal folds and a penetrative schistosity (S_2) subparallel to the axial planes of folds and to lithological contacts (Fig. 9.6c). This S_2 is poorly preserved but it is locally visible mainly within porphyroblasts (Puga and Díaz de Federico 1976; Vissers 1981). The D_3 deformational episode produces asymmetrical folds and, locally, a less penetrative

Fig. 9.4 Marbles of the Upper Mulhacén successions at Cóbdar

crenulation foliation S_3 (Fig. 9.6d). The D_4 event generated open folds with an associated crenulation cleavage S_4.

Large and small chloritoid crystals are rich in inclusions of graphite and ilmenite defining the folded S_1. Undeformed (Fig. 9.6f), sigmoidal (Fig. 9.6e) or folded graphite and quartz inclusion (S_1, Fig. 9.6b) appear within plagioclase porphyroblasts. Garnet porphyroblasts show undeformed, folded or sigmoidal helicitic inclusion patterns of graphite and quartz (Fig. 9.6a), and, sometimes, inclusion-free rims. Some garnet including S_1 and apparently rotated, with deformational pressure shadows and flattening of the main schistosity, were interpreted as syn-S_2 to post-S_2 (Gómez-Pugnaire and Franz 1988).

Figures 9.7 and 9.8 show variations in chemical composition of garnet and chloritoid respectively. For comparison, they also include data of the same minerals, both pre-Alpine and Alpine, from Mulhacén rocks (Gómez-Pugnaire and Sassi 1983; Puga et al. 2002). Peak temperatures were crudely estimated around 400–450 °C for the main assemblage, without significant changes in the evolution of the metamorphic grade (Gómez-Pugnaire and Franz 1988). However, the oligoclase rim of the albite cores and the inclusion-free garnet rims would indicate a T-increase after recrystallization (Eeckhout and Konert 1983; Vissers 1981). It is noteworthy that similar chemical compositions are found in the inclusion-free garnet rims and in garnets recrystallized during Alpine metamorphism (see discussion below on controversies).

Low P conditions are generally accepted for the pre-Alpine metamorphism that affected the Veleta Succession (Puga and Díaz de Federico 1976; Gómez-Pugnaire and Sassi 1983; Gómez-Pugnaire and Franz 1988). Puga et al. (2002) estimated $P \approx 3$ kb from the low-Si phengite in white micas found in pressure shadows of porphyroblasts. This agrees with chemical composition of the mineral assemblage and allows distinguishing this metamorphism from the HP Alpine metamorphism found in the Mulhacén rocks. Actually, systematic high-Fe contents in some minerals such as chloritoid and garnet are chemically incompatible with the high-Mg content found in the same minerals when formed in relation to the Alpine metamorphism (Figs. 9.7a, b and 9.8).

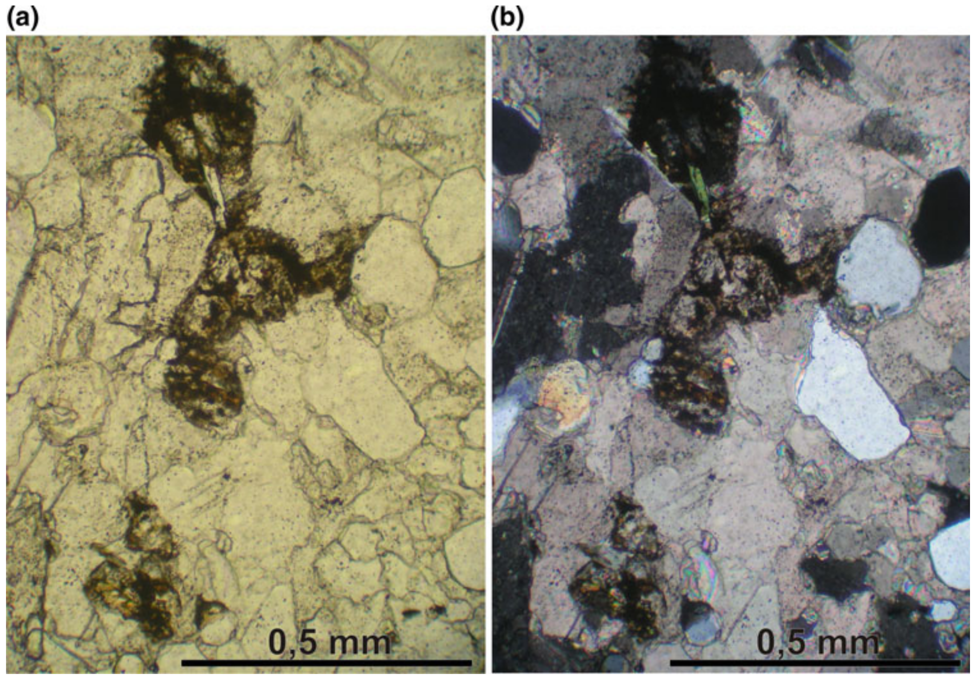

Fig. 9.5 Thin section of impure marbles near Cóbdar, with subrounded detrital quartz grains and one ankeritic object whose size and shape fits exactly with the axial section of the planktonic foraminifer *Helvetoglobotruncana helvetica* BOLLI, of Turonian age (Tendero et al. 1993)

Recent estimates using Raman spectrometry in carbonaceous materials of the Veleta Succession and local equilibria of chlorite-phengite pairs (Vidal and Parra 2000) yield P–T conditions of 450–530 °C at 12–14 kbar followed by heating and decompression reaching a maximum temperature of 530 °C at 8–10 kbar (Augier et al. 2005b). Behr and Platt (2012) also used Raman spectrometry in Sierra Alhamilla to quantify the temperature peak at the top of the Veleta succession, obtaining an average T = 493 °C. These authors attributed this metamorphic event in the Veleta rocks to the Alpine cycle. In addition, isopleth thermobarometry and pseudosections obtained from compositional zoning of garnet from Veleta schists (Aerden et al. 2013) indicate prograde metamorphic paths from ca. 5 kb/500 °C to 10 kb/550 °C, with P peak between 7 and 12 kbar. The pressure peak was calculated from garnet inclusion-free overgrowths that are interpreted to be Alpine. This would probably explain the overlap of some Veleta garnets in the field of the Alpine garnet of the Mulhacén rocks (Fig. 9.7b; full dots).

9.3.2 Lower Mulhacén Succession

M. T. Gómez-Pugnaire, A. Jabaloy-Sánchez, A. Martín-Algarra

The graphite-bearing micaschists of the lowest part of the Lower Mulhacén Succession include LP/IT mineral parageneses consisting of andalusite, chloritoid, staurolite, garnet, biotite, white micas, chlorite, quartz, magnetite and rutile. This assemblage, unanimously interpreted as Variscan, was overprinted by the HP/LT Alpine metamorphism (Puga and Díaz de Federico 1976; Gómez-Pugnaire and Sassi 1983; Martínez-Martínez 1986). The most striking feature of the Variscan mineral assemblage is the large size and the euhedral habit of porphyroblasts such as chloritoid, staurolite and andalusite (up to several cm long) and the very fine-grained matrix in the less overprinted samples (Fig. 9.9).

In the alpine overprint kyanite replaces andalusite; pre-Alpine sericite is found in rim and cracks of large porphyroblasts of andalusite and staurolite, and is replaced by Mg-chloritoid + garnet; garnet crystallized after biotite, and a pyrope-rich corona occurs overgrowing almandine-rich euhedral core of porphyroblastic garnets. The chemical composition of pre-Alpine garnet and choritoid is very similar to that described in the Veleta Succession. Both assemblages are typically rich in FeO compared with the Mg-rich Alpine minerals (Figs. 9.8a, b and 9.9). Peak pressures and temperatures recorded by phase relationships and the conventional geobarometric method are in the range of 2–4 kb and 500–600 °C (Puga and Díaz de Federico 1976; Gómez-Pugnaire and Sassi 1983).

9.3.3 Variscan Pre-Magmatic UHP Metamorphism in the Nevado-Filábride Complex?

M. T. Gómez-Pugnaire

Ruiz-Cruz et al. (2015, 2016) and Ruiz-Cruz and Sanz de Galdeano (2017) have reported diamond inclusions in

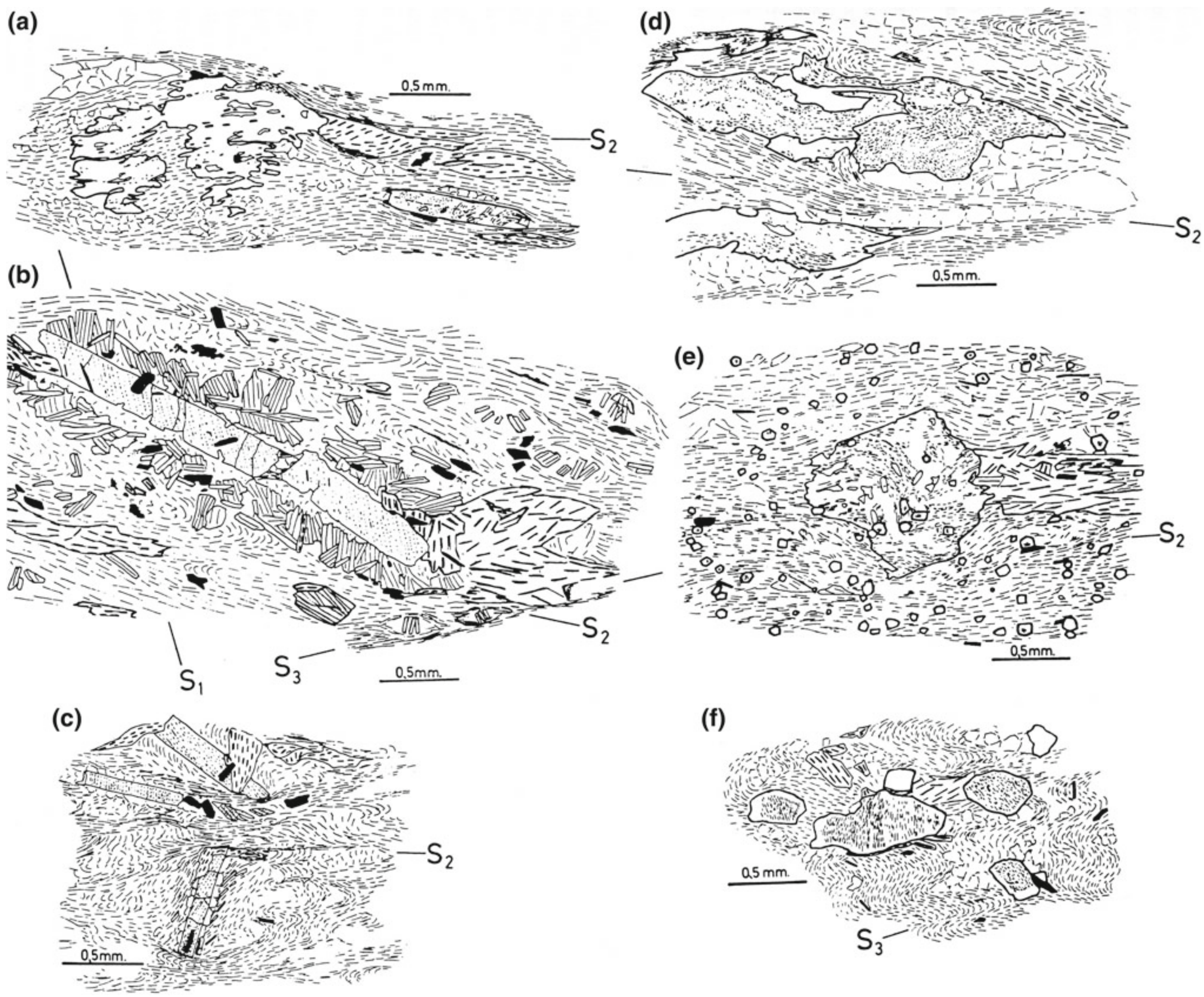

Fig. 9.6 Drawings illustrating phase relationships of porphyroblasts in the Veleta Succession. **a** Garnet (left) with S-inclusion pattern, chlorite (thick dashed) and chloritoid (slightly folded and irregular inclusion pattern), quartz and white mica. **b** Cracked chloritoid with undeformed inclusion pattern defining S_1, surrounded by small chloritoid crystals (thin dashed) and partly replaced by oxychlorite (thick dashed). **c** Irregular distribution of chloritoid porphyroblasts with partly continuous, partly discontinuous internal inclusion pattern. **d** Albite blast with irregular shape growing parallel to S_2 and a mostly continuous inclusion pattern of an early crenulation. **e** S-shaped inclusion pattern in albite with small garnets and chlorite to the right of the blast. **f** Albite blasts in a mica rich matrix, affected by S_3 crenulation, and with a narrow inclusion-free rim (from: Gómez--Pugnaire and Franz 1988)

apatite from different Nevado-Filábride rocks in Sierra Nevada, which they interpret as related to a pre-Alpine UHP metamorphism in the Nevado-Filábride Complex. The studied apatite comes from biotitic gneisses dated by $^{238}U/^{206}Pb$ in zircon as 285 ± 3 Ma. Based on petrographic and SEM analyses, and on geochemical data from biotite gneisses, these authors distinguished several generations of minerals that they attributed respectively to pre-magmatic, magmatic (Hercynian) and metamorphic (Alpine) events. Apatite concentrated in biotite-rich layers is closely associated with poorly preserved garnet interpreted as pre-magmatic on the base of garnet zoning pattern and isopleth thermobarometry. The coexistence of garnet and apatite suggests their common pre-magmatic origin, and the presence of diamond inclusions in apatite would indicate that UHP metamorphism might have affected source rocks before melting and latter gneissic recrystallization. The interpretations by Ruiz-Cruz and co-workers have important objections that are discussed by the authors themselves. One is related to distinguishing true diamond inclusions from artifacts produced by contamination during sample preparation, although the authors ruled out this possibility. The second objection relates to the easy metamorphic recrystallization of apatite below 500 °C, and to the possible dissolution of

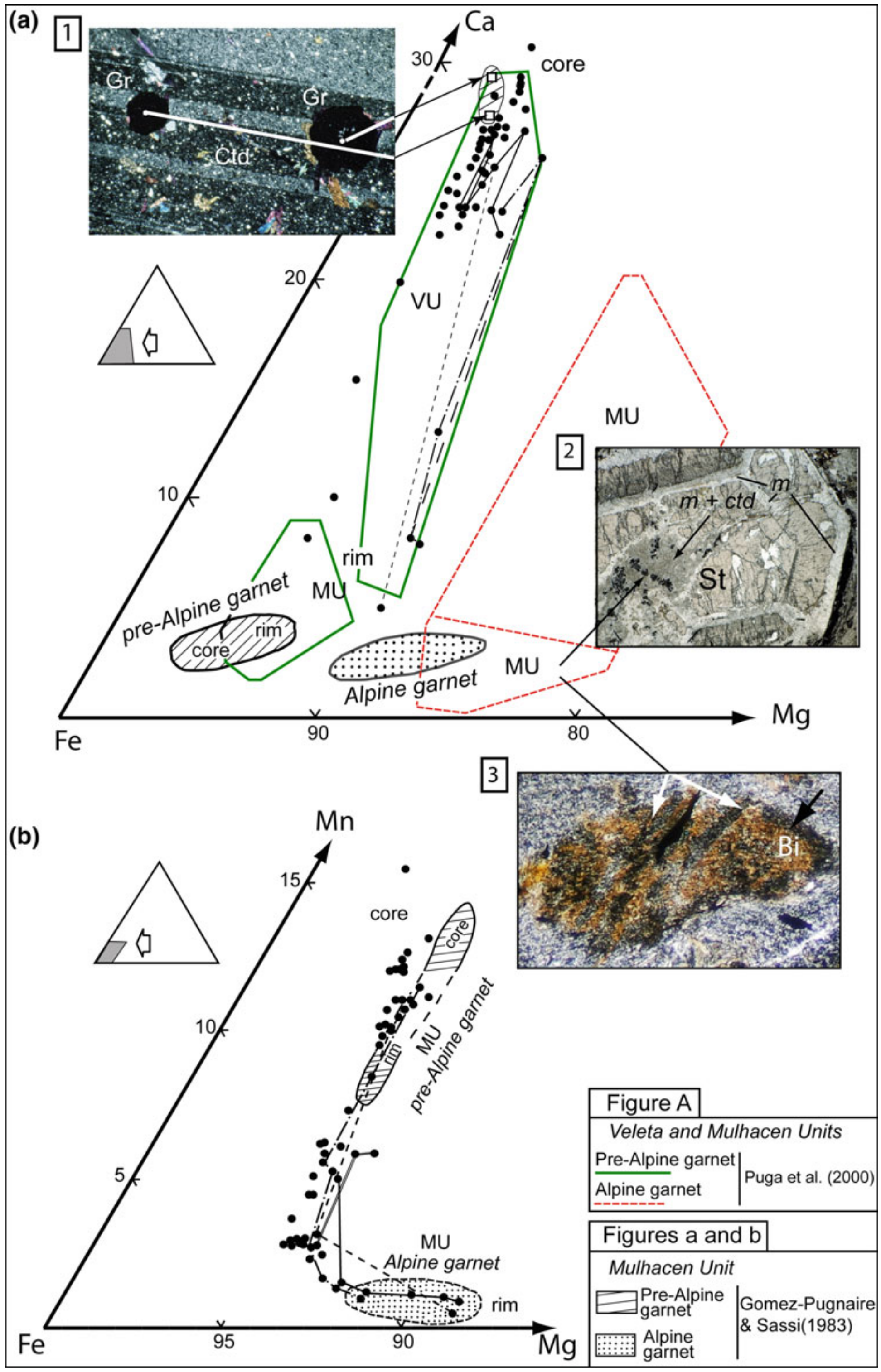

Fig. 9.7 Chemical composition of garnets: **a** Ca–Fe–Mg; **b** Mn–Fe–Mg of the Veleta and Mulhacén schists (VS and full dots, and MS, respectively). Lines connect point analyses along compositional profiles. Empty squares represent chemical compositions of euhedral garnet (cores) included within large pre-Alpine chloritoid poikiloblasts (insert picture 1). Mineral abbreviations according to Whitney and Evans (2010). Modified from Gómez-Pugnaire and Franz (1988), with analyses of the Mulhacén Alpine and Pre-Alpine garnets from other studies (insert in the lower right corner). See text for details

apatite during partial melting (Hammerli et al. 2014; Bea and Montero 1999). Nonetheless, Ruiz-Cruz et al. (2016) report several occurrences of apatite that persisted melting in eclogites and under crustal anatexis conditions (e.g. Puchelt and Emmermann 1976; Bingi et al. 1996). To avoid possible misinterpretations, geochemical data of well-characterized apatite domains must be used to testify the preservation or re-equilibration of the pre-magmatic apatite during anatexis. In any case, the existence of UHP conditions in the Nevado-Filábride Complex, if confirmed, would have very

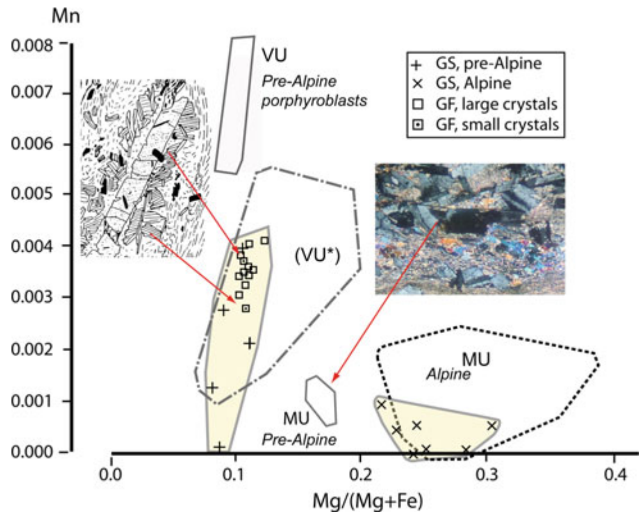

Fig. 9.8 Chemical composition of chloritoid (modified from Gómez-Pugnaire and Franz 1988; yellow shaded areas; GF) compared with data from Gómez-Pugnaire and Sassi (1983; GS) and Puga et al. (2002). VU: Veleta; MU; Mulhacén; VU*: interpreted as Alpine garnet by Puga et al. (2002)

important consequences not only for understanding the geological evolution of the Nevado-Filábride Complex but also for that of the Betic Cordillera as a whole, and must be carefully checked in future research.

9.3.4 Controversial Aspects on the Metamorphism Affecting the Veleta Successions

M. T. Gómez-Pugnaire, V. López Sánchez-Vizcaíno

There is a general agreement on the polymetamorphic character of the Lower Mulhacén Succession: the main low-pressure assemblage has been overprinted by other metamorphic parageneses chemically similar to those related to the Alpine metamorphism. In the Veleta Succession, however, most if not all the metamorphism is considered to be of Alpine age by some authors, despite of the metamorphic regime is similar to that of pre-Alpine metamorphism in the Lower Mulhacén Succession (Gómez-Pugnaire and Franz 1988). Nevertheless, the lower metamorphic degree, the lack of index minerals and strong retrogression do not allow reaching conclusive estimates for the P-T conditions that affected the Veleta Succession. For this reason, the consistence of some data addressed to identify HP Alpine metamorphic conditions in the Veleta Succession, as proposed by some recent studies, is discussed below.

Puga and Díaz de Federico (1976) and Vissers (1981) proposed a polymetamorphic history of the Veleta Succession from folded inclusion patterns found in large chloritoid porphyroblasts. They attributed an Alpine age to the main deformational event (D_2) and assumed a pre-Alpine age for the porphyroblasts. Since folded inclusions are similar in chloritoids of the matrix, they can also be interpreted as resulting from porphyroblasts growth while pre-Alpine S_1 was developing. Alpine HP conditions in the Veleta schists have been calculated on the base of the Si content (a.p.f.u) in phengite (Puga et al. 2002; Augier et al 2005b). Phengite component of muscovite is a function not only of pressure but also of the composition and this barometer can only be used in metapelites of specific compositional systems: KMASH (phengite + quartz + K-feldspar + biotite + H_2O) and KFASH (phengite + almandine + kyanite + quartz + H_2O; Massonne and Szpurka 1997). Assemblages containing feldspar, biotite and kyanite do not coexist with phengite in the Veleta schists, however. Additionally, the pyrophyllite substitution in muscovite should be considered because this substitution can be so large that geobarometry solely based on Si content can result in pressure overestimation (Essene 1989; Agard et al. 2011). By applying the chlorite-phengite multiequilibria thermobarometric method recent studies have concluded that the low-grade Veleta micaschists were affected by HP Alpine metamorphism (Augier et al. 2005b; Booth-Rea et al. 2005). Nevertheless, this method requires ascertaining whether adjacent mineral phases are in chemical equilibrium, which is extremely difficult to prove, in practice, in low-grade rocks, especially when they are strongly retrogressed like the Veleta schists. Recent attempts to reproduce these results from Veleta rocks of Sierra Alhamilla have actually failed (Behr and Platt 2012).

Scattered K/Ar and $^{40}Ar/^{39}Ar$ ages obtained from the Veleta Succession, ranging from the Early Jurassic to the Early Miocene, have been interpreted as related to Alpine metamorphic events (De Jong 1991, 1993a, b, 2003; Puga et al. 2004; Behr and Platt 2012). These ages are difficult to reconcile with much more robust results recently obtained from multiple geochronometers that constrain the peak of metamorphism in the Nevado-Filábride Complex to the early-middle Miocene (15.0 ± 0.6 Ma: López Sanchez-Vizcaíno et al. 2001; 18.2 ± 0.8 Ma and 16.8 ± 0.3 Ma: Platt et al. 2006; 17.3 ± 0.4 Ma: Gómez-Pugnaire et al. 2012; 20.1 ± 1.1 Ma: Kirchner et al. 2016).

RAMAN thermometry on carbonaceous material in the graphite schists shows that the shear zone at the Veleta-Mulhacén contact superposes the higher-T Lower Mulhacén Succession onto lower-T Veleta Succession (Augier et al. 2005b; Platt et al. 2006; Behr and Platt 2012; Booth-Rea et al. 2015). The inverted metamorphic gradient was classically interpreted as a result of a late-stage thermal overprint before Alpine nappe emplacement (De Roever and Nijhuis 1964; Vissers 1981; Bakker et al. 1989) and, recently, as formed during subduction of the cold Nevado-Filábride Complex beneath the hot thin Alboran

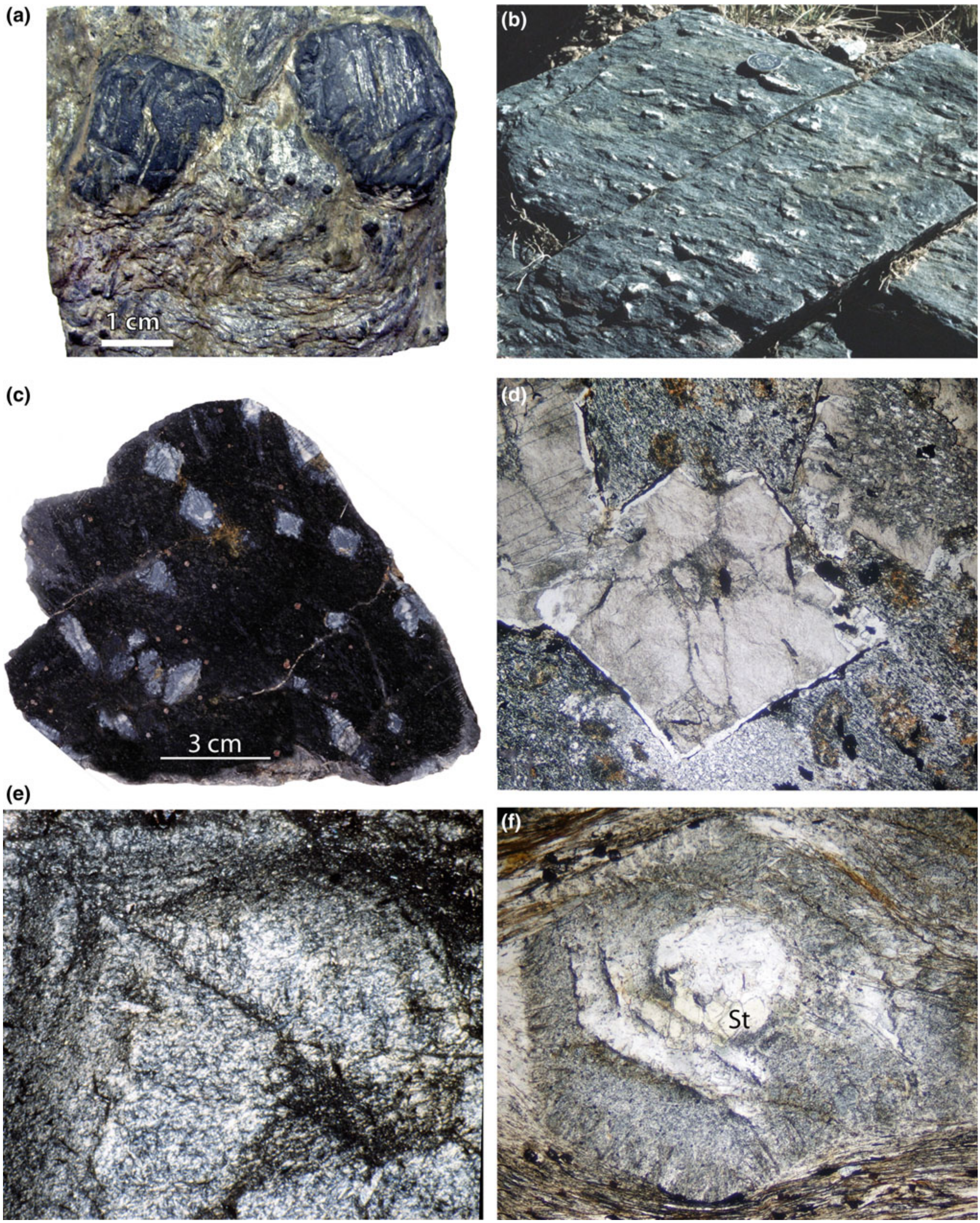

Fig. 9.9 Pre-Alpine minerals of the pre-Mesozoic Mulhacén dark schists. **a** Large chloritoid porphyroblasts with garnet inclusions. **b** Field view of elongated staurolite crystals. **c** Pseudomorphs of euhedral andalusite crystals in fine-grained schists with euhedral garnet. **d** Microphotography of the same sample as C showing euhedral chiastolite pseudomorphs and biotite flakes (plane-polarized light). **e** Cross-polarized photography of chiastolitic andalusite fully replaced by tiny Alpine kyanite crystals. **f** Petrographical image of a large staurolite crystal almost completely replaced by Alpine chloritoid and muscovite (plane-polarized light). Mineral abbreviations after Whitney and Evans (2010)

lithosphere (Behr and Platt 2012). In any case, the upper part of the Veleta Succession must have been affected, at least locally, by the metamorphic evolution related to the latest Alpine deformational events responsible for the ductile shear zone that separates it from the overlying Lower Mulhacén Succession: thin oligoclase rims of plagioclase and overgrowths of garnet could be the metamorphic products of this process. The temperature calculated from the rocks involved in the shear zone ranges between 493 and 510ºC (Augier et al. 2005b; Behr and Platt 2012), which is in the same range of the temperature calculated for the main event of the regional metamorphism of the Veleta Succession.

In summary, inaccuracies associated with the methods used lead to conclude that, for the moment, the estimated P-T conditions and age (pre-Alpine vs. Alpine) for the metamorphism underwent by the Veleta rocks should be taken with caution. This brings into question the widespread attribution of the Veleta metamorphism to the sole Alpine orogeny. If the main metamorphic processes would be related mainly to the Variscan Orogeny, the interpretation of the Veleta successions as a pre-Alpine basement involved in the Alpine orogeny only at a relatively late stage (Gómez-Pugnaire and Franz 1988) would be reinforced. Consequently, the rocks of the Veleta Succession might correspond to the prograde lower temperature segment of the pre-Alpine PT path while those of the Lower Mulhacén Succession would represent the peak PT conditions of the Variscan metamorphism.

9.4 Paleozoic Magmatism of the Nevado-Filábride Complex

M. T Gómez-Pugnaire, A. Jabaloy-Sánchez, V. López Sánchez-Vizcaíno

In addition to the above-mentioned metasediments, the Mulhacén Successions include different types of magmatic rocks strongly affected by the Alpine metamorphism (orthogneisses, metabasites and serpentinites). This section deals with acidic orthogneisses, whose latest Paleozoic magmatic age is demonstrated by geochronological data. Metamorphosed acidic igneous rock bodies of variable dimensions are commonly found throughout the Nevado-Filábride successions (Nijhuis 1964; Puga et al. 2002; Nieto 1996; Gómez-Pugnaire et al. 2012; Martínez-Martínez et al. 2010). In the Veleta Succession, Puga et al. (2002) mention the local presence of dm-to m-thick lenses of undated rhyolitic orthogneisses. In the Lower and Upper Mulhacén Successions, Gómez Pugnaire et al. (2012) and Ruiz-Cruz et al. (2016) distinguished gneisses with two different compositions: (1) peraluminous leucocratic gneisses, with textures that ranged from metagranites (Fig. 9.10a), augen gneisses and even-grained gneisses depending of the different degree of deformation and metamorphism, which affected the granites, and (2) strongly peraluminous biotite-rich gneisses. The largest outcrop of the type 1 gneisses is exposed in the Sierra de los Filabres near to El Chive and Bédar (Fig. 9.10), where evidences of contact metamorphism and metasomatism in the country metasediments are common.

Nieto et al. (2000) and Puga et al. (2002) distinguished among syncollisional and postcollisional peraluminous leucocratic gneisses. The latter type consists of tourmaline-rich layered gneisses made of alternating centimeter-to metre-thick leucocratic layers, with Na- and K-feldspar, and quartz, and melanocratic lithotypes with phengite, green biotite, epidote and tourmaline (Díaz de Federico et al. 1990; Nieto 1996; Nieto et al. 2000; Torres-Ruiz et al. 2003; Gómez-Pugnaire et al. 2012). They were interpreted as metarhyolites and meta-volcanoclastic rocks (Nieto et al. 2000; Puga et al. 2002), and have been dated by K/Ar as Triassic in age (Andriessen et al. 1991; Nieto 1996; Nieto et al. 2000). The same rocks, however, have provided Paleozoic zircon ages (Martínez-Martínez et al. 2010; Gómez-Pugnaire et al. 2012) and the alternation of dark and light layers can be also explained by segregation during deformation by ductile shear zones. Most compositional data of the gneisses protoliths plot in the *anatectic* (syn- and post-orogenic) granitoid field in the R1–R2 multicationic diagram of Batchelor and Bowden (1985), whereas some of the biotite gneiss samples and one leucocratic gneiss display very high R1 and plot in the compositional field of the Nevado-Filábride metapelites (Gómez-Pugnaire et al. 2012; Ruiz-Cruz et al. 2016). All geochemical data suggest a derivation of gneisses by melting of a crustal parent. Additionally, the presence of detrital zircon cores with different zoning patterns in all dated gneisses (Gómez-Pugnaire et al. 2012) indicates metasedimentary rather than igneous sources. The high A/CNK ratio, high SiO_2, low CaO, high LILE, and REE patterns, are comparable to other peraluminous, post-collisional granitic bodies in the European Variscan orogen. The sequence can be interpreted within the context of the late Variscan extension that generates magmatic activity between 295 and 240 Ma (Gómez-Pugnaire et al. 2012; Ruiz-Cruz et al. 2016).

The age of the igneous rocks indicates the minimum age of the host rocks if relationships between them prove to be original. Rb/Sr whole rock dating and U-Pb SHRIMP dating of magmatic zircons from gneisses from the Upper Mulhacén Successions near the marbles yield and age of 247 ± 11 and 301 ± 7 Ma respectively (Table 9.1: Gómez-Pugnaire et al. 2000, 2004). Martínez-Martínez et al. (2010) reported U-Pb LA-ICPMS ages of 314 ± 7 Ma and 304 ± 23 Ma in gneisses from one outcrop studied by Gómez-Pugnaire et al. (2004) in Sierra Nevada, but disagree with the interpretation of field relationships between the gneisses and the country

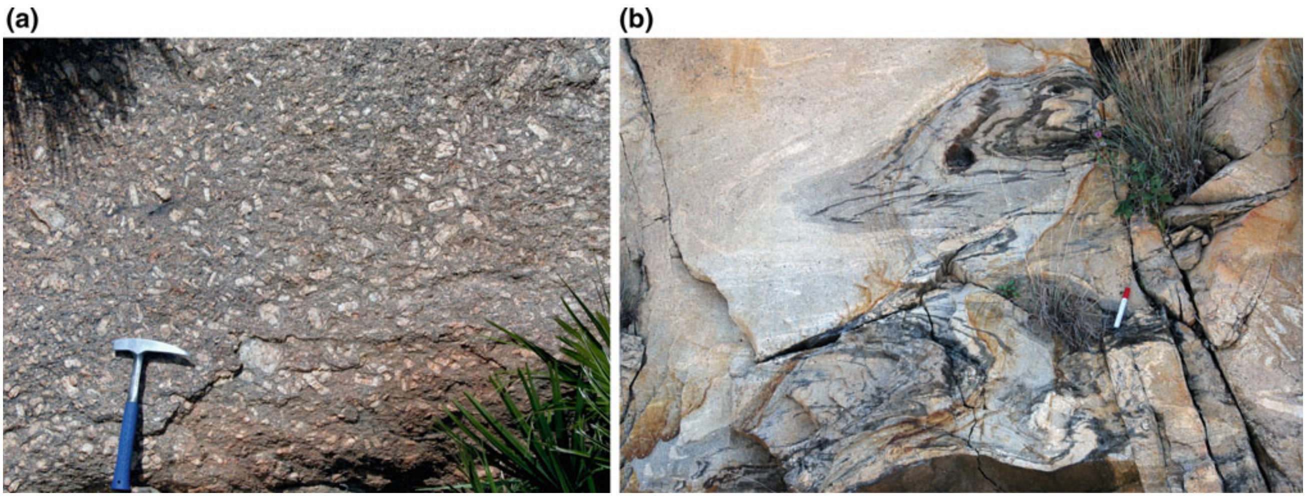

Fig. 9.10 **a** Undeformed metagranite from the Monda outcrop. **b** Strongly deformed gneiss from the Bédar outcrop

rocks (upper marbles) made by the previous authors. The intense thinning of the Upper Mulhacén Successions makes difficult assessing the relationships of this gneiss body and marbles in this locality. Nevertheless, thick marble layers from the Sierra de los Filabres underlie, with obvious stratigraphic relationships, a tourmaline gneiss complex yielding and age of 291 ± 3 Ma to Gómez-Pugnaire et al. (2012, their Fig. 9.3). The same authors also reported U-Pb SHRIMP dating of zircons from metagranites and gneisses intruded in the Lower Mulhacén Succession and associated with metasediments of the Upper Mulhacén Successions (Fig. 9.11) up to the base of the marbles that characterize their highest part (Table 9.1). Ruiz-Cruz and Sanz de Galdeano et al. (2017) have reported slightly younger ages (286 ± 3 Ma) in equivalent rocks in Sierra Nevada.

9.5 Pre-Mesozoic Successions and Pre-Alpine Evolution of the Alpujárride Complex

A. Sánchez-Navas, A. Martín-Algarra, A. García-Casco, S. Mazzoli

The stratigraphic successions of the Alpujárride Complex (Fig. 9.11) yet preserve abundant Triassic fossils, in spite of having been affected by Alpine metamorphism (cf. Vol. 3, Ch. 7), but they have not provided any pre-Mesozoic fossil. So, the Paleozoic and/or older stratigraphic age of many Alpujárride rocks is primarily based on lithostratigraphic correlation. The mineral paragenesis and metamorphic conditions of the Alpujárride rocks change depending on their stratigraphy (pre-Mesozoic vs. Mesozoic), lithology (pelitic-psammitic *vs.* carbonate), tectonic position within the Alpujárride nappe stack (lower, intermediate, upper), or regional location (eastern, central, western). In general, the metamorphic grade in the Lower Alpujárrides is lower than in the Intermediate and Upper Alpujárrides, and it is the highest in the western sectors where the Upper Alpujárride pre-Mesozoic rocks directly overlie the Ronda ultramafic massifs. The main lithological and petrological features of the pre-Mesozoic Alpujárride successions are quite constant, with only minor changes from the Lower to the Upper Alpujárrides, although they are much better developped and thicker in the latter. Actually, the Upper Alpujárrides are dominantly made of pre-Mesozoic rocks (Fig. 9.12) because Triassic and eventually younger rocks above them are exclusively present close to the Internal-External Zone boundary, along a narrow and strongly imbricated belt below the Maláguide Complex (Didon et al. 1973; Martín-Algarra et al. 1995; Sanz de Galdeano et al. 1995a, b, 2001).

The pre-Mesozoic Alpujárride successions are predominantly made of dark brown and gray metapelites and metapsammites, usually rich in graphite (Gr; mineral abbreviations after Whitney and Evans 2010) but not as much as in their Nevado-Filábride counterparts. The metapelites are fine-grained schists in the upper part, and become progressively coarser-grained downwards. In their middle and lower part, the schists include mm to occasionally cm-sized porfiroblasts, especially of garnet (Grt), chloritoid (Cld), plagioclase (Pl) and andalusite (And), which mainly concentrate in discrete, pelitic horizons and in quartz (Qz)-rich veins (Fig. 9.12e, f). Deeper levels also contain staurolite (St), kyanite (Ky), fibrolite (Fi) and, occasionally, prismatic sillimanite (Sil). In some Intermediate and especially Upper Alpujárride units, the coarsest-grained schists gradually change downwards to gneisses and migmatites (Fig. 9.12a–d), with local evidence of partial melting. Consequently, the pre-Mesozoic Alpujárride rocks were affected by a metamorphism of gradually increasing grade

Table 9.1 Field occurrence, petrography and age of some gneiss samples (modified from Gómez-Pugnaire et al. 2012)

Area	Locality	Sample	Country rocks	Field occurrence	Magmatic ages (U-Pb) zircon
Eastern Sierra de Filabres	*El Chive*	**CHIVE-2**	**Bottom of gneiss**: light schists, marbles, metabasites, metaevaporites and serpentinites **Top of gneiss**: dark- and light-coloured schists, serpentinites and marbles. Skarn with eclogitic assemblages in the border	*Chive:* porphyritic gneiss with feldspar megacrystals *Bédar:* tectonite	291 ± 3 Ma
	Bédar	**BED-3**			283 ± 4 Ma
Western Sierra de Filabres	*El Pocico*	**PC-3**	Uppermost light schists and marbles. Small dikes of metabasites	Stretching lineation and parallel folds	283 ± 4 Ma
		PC-4	Light micaschists		
	Charches	CHA-1(1)	Replacement tourmalinites, dark and light micaschists and marbles	Small and strongly sheared even-grained layers and lenses augen textured, abundant tourmaline nodules	247 ± 11 Ma(1) (Rb/Sr whole rock)
Central Sierra de Filabres	*Líjar River*	**LR-3**	Marble layers (1 m thick) and dark and light micaschists. Skarn boudins (~40 cm thick)	Small banded body, 4–5 m thick	295 ± 3 Ma
Eastern Sierra Nevada	*Cerro Blanco*	**CB-4**	Uppermost marbles and metaserpentinites. Skarn rocks in the contact	75 m thick body. Banded structure. Well-developed mylonitic foliation	279 ± 5 Ma
Western Sierra Nevada	*Sierra Nevada road*	SAB-00(2)	Host tourmalinites, dark, light micaschists and marble. Skarn like rocks in the original intrusive contacts	Augen and even-grained layers	301 ± 7 Ma(2)
	Prado del Cebollar	**CEB-3**	Graphite-bearing schists. Dismembered skarns transformed to eclogites	Lens (10 × 40 m) of alternating leucocratic and melanocratic layers. Augen texture and even-grained layers, parallel to the foliation	285 ± 3 Ma

downwards, with highest-grade conditions locally associated with anatexis in the deepest structural levels.

The Alpujárride pre-Mesozoic rocks show a polymetamorphic and polyorogenic pre-Alpine and Alpine evolution (Sánchez-Navas et al. 2017 and references therein). In the Upper Alpujárrides of the central sector (Fig. 9.1) the lowest known pre-Mesozoic rocks make part of the *Torrox Gneissic Complex* (TGC: Fig. 9.12). The TGC is made of several types of gneissic rocks bearing muscovite (Ms) biotite (Bt), K-feldspar (Kfs), Grt, Al-silicates (Als) and, subordinately, cordierite (Crd). The ortho- or paraderivated origin of the TGC, the Alpine or pre-Alpine age of the high-T metamorphism that affect it, and the precise metamorphic conditions reached during both evolutions are controversial (Cuevas et al. 1989; García-Casco 1993; García-Casco et al. 1993; Zeck and Williams 2001; Zeck and Whitehouse 2002). Traditionally considered of fully Alpine metamorphic age, recent detailed petrologic and geochronologic studies show, however, that the TGC evolved from pre-Mesozoic granitic rocks (Fig. 9.12a, b) that were affected by a pre-Alpine high T metamorphism responsible for migmatization and for intrusion of granitic pegmatite dikes bearing And at 283 ± 16 Ma, according to SHRIMP zircon dating (Sánchez-Navas et al. 2014; Fig. 9.12c).

The TGC and its schistose envelope were strongly deformed and metamorphosed during the Alpine Orogeny under medium- to high-grade metamorphic conditions at around 20–22 Ma (Zeck et al. 1989; García-Casco et al. 1993; García-Casco and Torres-Roldán 1996, 1999; Sánchez-Navas 1999). Ruiz-Cruz and Sanz de Galdeano (2012b, 2013, 2014a, b) affirm that the Torrox gneisses and other Grt-bearing high-grade gneisses in Ceuta and immediately above the Ronda peridotites would have been affected by ultrahigh P metamorphism producing coesite and microdiamond, although this has been considered doubtfull in the Ronda area by Massonne (2014).

At outcrop scale, the main tectonic fabric of the Alpujárride pre-Mesozoic schists above the TGC, and of equivalent rocks in other Alpujárride units, is a metamorphic foliation, which is generally sub-parallel to the lithologic contacts. This foliation includes one older foliation and is strongly folded and partially or totally transposed by at least one (probably two) new foliation(s). These foliations are usually named S_1 (oldest), S_2 (main) and S_3 (youngest) and

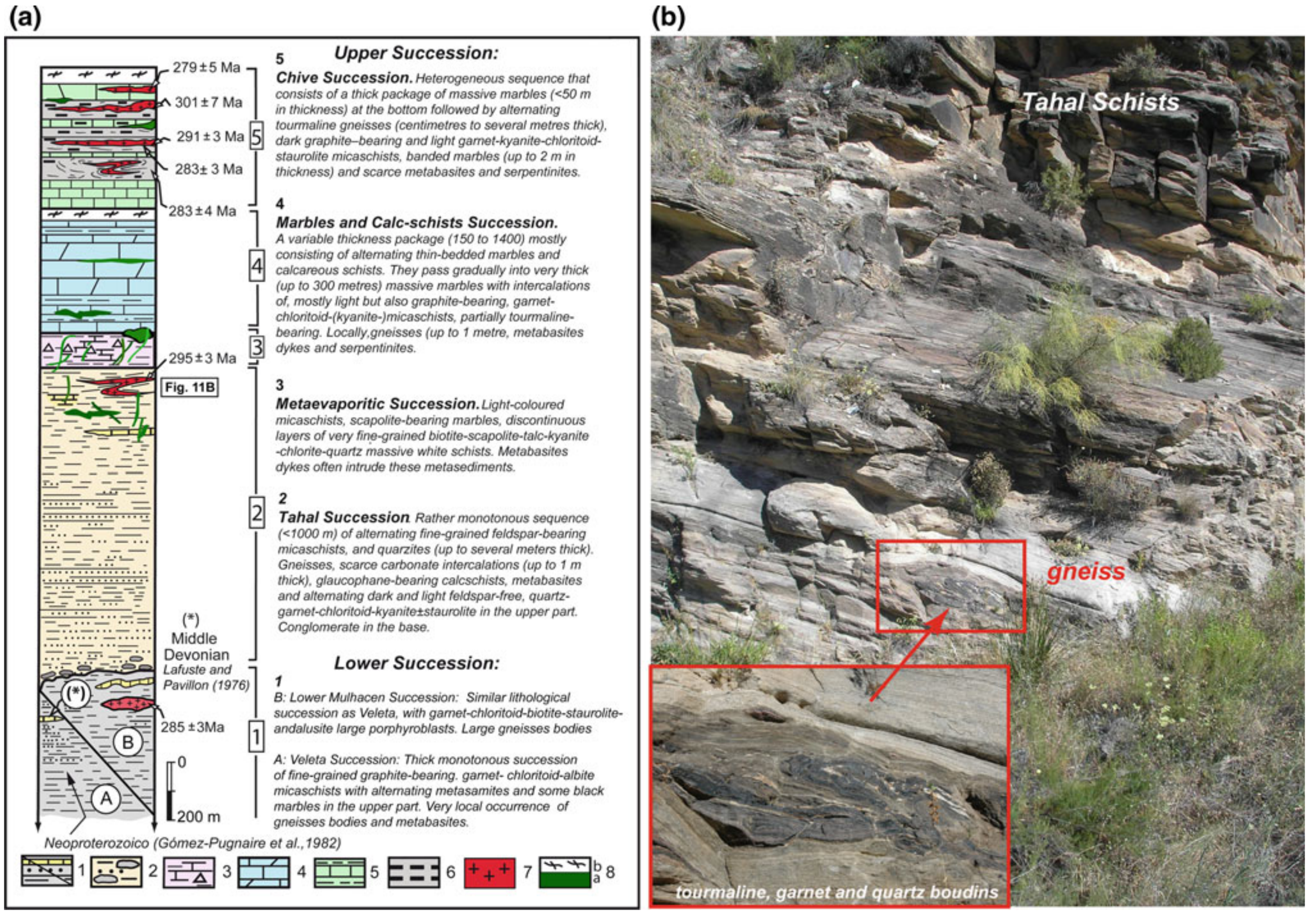

Fig. 9.11 **a** Lihological successions of the Nevado-Filábride Complex and location of the dated gneisses bodies. **b** A gneiss body embedded in the marbles and schists of the Upper Mulhacen Successions (see location in Fig. 9.11a)

were formed during prograde evolutions from low- to medium- to high-grade metamorphic stages during a polycyclic, polyphasic and plurifacial tectonometamorphic evolution, ending with retrograde evolutions. The S_1 and S_2 foliations and the associated minerals (see below) are also found in pelitic enclaves included within the most evidently granitic lithotypes of the TGC, dated as Early Permian (Sánchez-Navas et al. 2017). Consequently, these foliations must be related to a prograde pre-Alpine tectonometamorphic cycle that culminated with crustal anatexis and granite emplacement in latest Variscan (Hercynian) time. The S_3 foliation crosscuts all gneissic lithotypes of the TGC including, in particular, the latest Variscan pegmatite dikes and their discordant intrusive contacts with the enclosing rocks (Sánchez-Navas et al. 2014). Consequently, this S_3 foliation is usually considered as Alpine. Hereafter, these pre-Alpine (S_1, S_2) and Alpine (S_3) foliations will be named S_{1V} S_{2V} and S_A, respectively.

The oldest pre-Alpine foliation of the pre-Mesozoic schists (S_{1V}) is rarely visible at outcrop scale but it is commonly found under the microscope, delineated by Gr, opaque minerals and fine-grained micas, in lensoid domains bounded by later foliations, or as oriented inclusions within relict minerals that grew synkinematically and/or postkinematically with respect to S_{1V}: Grt, Ky, St, and (only in the upper levels of the succession) Cld (Fig. 9.13a–d). These relict minerals usually appear strongly dissolved and deformed by the later foliations (Sánchez-Navas et al. 2012, 2014, 2016, 2017).

The S_{2V} was associated with the synkinematic blastesis of the largest Bt and Ms crystals visible within these rocks. It was clearly overprinted by static blastesis of the largest minerals visible at the naked eye (mm to cm sized), And in particular (Figs. 9.12f, 9.13d-e-f). It is also statically overprinted by disoriented Ms and, especially, by red Bt crystals, Pl and, in the deepest parts of the succession affected by the highest metamorphic grade, by Cdr (Fig. 9.13b), which is intimately associated with And and usually appears, under the microscope, strongly pinnitized (Sánchez-Navas et al. 2012, 2014).

In most outcrops where the pre-Mesozoic Alpujárride schists are coarse-grained the And quite evidently grew

Fig. 9.12 Field views of Upper Alpujárride pre-Mesozoic rocks from the Torrox Gneissic Complex (**a**–**d**) and surrounding high-grade metapelites (**f**), and of the Benamocarra Unit (**e**): **a** Porphyritic orthogneiss yet preserving the magmatic texture of the granite protolith, which is intruded by a pegmatite dike, the whole affected by intense folding. **b** Well preserved granite texture, with large undeformed porphyroblasts of Kfs. **c** Late Variscan granitic dike dated by zircon SHRIMP at 283 ± 16 Ma, intruding intensely deformed gneisses in the outer part of the TGC; SHRIMP data of this dike, which is affected by a later sillimanite-bearing foliation, also record an Alpine event at ca 22 Ma. **d** Intensely folded leucocratic and melanocratic bands of the TGC close to the contact with surrounding metapelites. **e** Disoriented pre-Alpine Ky crystals partially transformed to pre-Alpine pink And cristals within a small Qz seggregation in dark schists of the Benamocarra Unit). **f** Close-up of dark (graphite-rich) pelitic bands lying directly on the gneisses shown in (**d**), full of large Variscan pink And crystals that are strongly deformed along the Alpine foliation, and partially transformed to a whitish aggregate of Alpine Ky and Ms

statically on the best visible foliation planes S_{2V}, forming palm-like aggregates and, in the coarsest-grained lithotypes, also forming chiastolites up to several cm long and, sometimes, nearly 1 cm thick (Figs. 9.12f and 9.13f). Nevertheless, a careful examination of the outcrops and thin sections reveal that the post-S_{2V} And as well as all the other above mentioned minerals in relation to the S_{1V} and S_{2V} foliations, are systematically deformed and sheared along the Alpine foliation S_A (Fig. 9.12f), which is frequently the most evident at outcrop scale, although it can be locally difficult to distinguish it from the S_{2V}.

A typical feature of the pre-Mesozoic schists, especially in the Upper Alpujárride units, is the presence of Qz plus albite (Ab)-rich veins. These are discordant with respect to a previous foliation (probably S_{2V}) but are usually strongly deformed and parallelized to the most evident foliation at oucrop scale (S_A). Along most of the succession-but not in its highest and finest-grained levels-these veins frequently contain pink And, sometimes forming well-preserved rosettes of disoriented crystals. In some veins, but much less frequently, there also appear large blue Ky crystals (Fig. 9.12e) that are replaced by pink And, which evidently postdate the Ky (Fig. 9.13d), during a prograde Variscan metamorphic evolution (Sánchez-Navas et al. 2016). The formation of blue Ky and its partial or total replacement by pink And within the veins has been tentatively correlated with the prograde evolution (Fig. 9.14) that affects their schistose matrix and that was responsible for the pre-Alpine syn- to postkinematic blastesis (with respect to S_{1V}) of variscan Ky (Ky_V), followed by S_{2V}-related deformation and, finally, by the post-S_{2V} blastesis of And, which replaced partially or totally the syn- to post-S_{1V} previous Ky (Ky_V). The same reaction has been found in pelitic enclaves included within ortogneisses of the TGC, which evidence a pre-Alpine evolution that produced a high-T metamorphism of pre-Permian metasediments (Sánchez-Navas et al. 2017). In the deepest parts of the Upper Alpujárride successions this tectonometamorphic evolution was followed by anatexis under moderate to high P, followed by decompression and granite emplacement at the end of the Variscan Orogeny (near to the Carboniferous-Permian boundary) whereas lower P and T are recorded in the upper structural levels represented by the Benamocarra Unit and the Maláguide Paleozoic successions (Fig. 9.14). In any case, these veins, and the pink And within them, as well as the And within both the schistose matrix and the enclaves included in the TGC orthogneisses, have been also strongly affected by the S_A-related deformation: the And within these rocks appears stretched, sheared and crushed to form granular aggregates of small sized cataclastic And and fine-grained Ms, and partially transformed to fine-grained Alpine Ky (Sánchez-Navas et al. 2012, 2017).

The development of S_A is associated with a generalised grain size reduction that affects most previous minerals, especially micas, and that produces Alpine Ky (Ky_A) plus Grt, and Fi in deeper structural levels. This foliation is mainly defined by fine-grained phengitic Ms and Bt plus strongly stretched Qz bands that are locally associated with neoformed Als from larger minerals (Ky and/or Fi, which forms, in particular, from the post-S_{2V} And crystals and from the pelitic matrix), and by retrograde chlorite Chl (Sánchez-Navas et al. 2012, 2014, 2016). The Ky_A is of much lesser size than the pre-Alpine Ky_V and, together with Fi, it clearly postdates the post-S_{2V} pre-Alpine And. In the Upper Alpujárrides of the central sector (Torrox area) the S_A foliation is associated with dominantly top-to-the-W shear zones, and records an evolution from kyanite-grade ductile shear zones, to Chl-bearing low-grade S-C mylonites, to brittle duplexes produced by west-directed imbrication (Sánchez-Navas et al. 2017).

At present, it is unclear if staurolite also formed synkinematically with kyanite and garnet in relation to the S_A-related Alpine tectonometamorphic event, or not (see, however, Williams and Platt 2017). It is also yet unclear if Alpine Ky and Fi formed coevaly and in relation to the *same* foliation, or not. Actually, Fi formation from And could also develop during the latest stages of a *prograde* pre-Alpine evolution at low P, as a consequence of the development of and extensional foliation under high-T conditions in relation to the widespread and well-documented latest Variscan extensional collapse, which accounted, essentially and everywhere, during the early Permian. This hypothesis does not fit, however, with higher P mineral asssemblages predating pre-Alpine And in the enclaves within the TGC nor in the surrounding schists (Sánchez-Navas et al. 2017, and Fig. 9.14). In the pre-Mesozoic Alpujárride schists, St usually appears strongly corroded and associated with the previous Ky (syn-S_{1V} to post S_{1V} above-mentioned), being both included, together with Grt and with Cld (when present), within post-S_{2V} And. In relation to Grt, the picture is also complicated: the compositional mapping of zoned Grt from pre-Mesozoic schists belonging to Upper Alpujárride units (García-Casco 1993; see also Ruiz-Cruz 2011a, b and Sánchez-Navas et al. 2012) and also to Intermediate Alpujárride units (Manjón-Cabeza et al. 2014) show complex compositional patterns, including relatively sharp boundary zones between cores and overgrowths, and chemical reversals at the boundaries, which clearly record two superimposed metamorphic events: pre-Alpine (cores) and Alpine (rims). On the contrary, garnets from Permo-Triassic schists overlying the pre-Mesozoic black schists in Intermediate Alpujárride units are unzoned and they only record, obviously, Alpine events (Manjón-Cabeza et al. 2014). The blastesis of the pre-Alpine Grt cores accounted under high-gradient

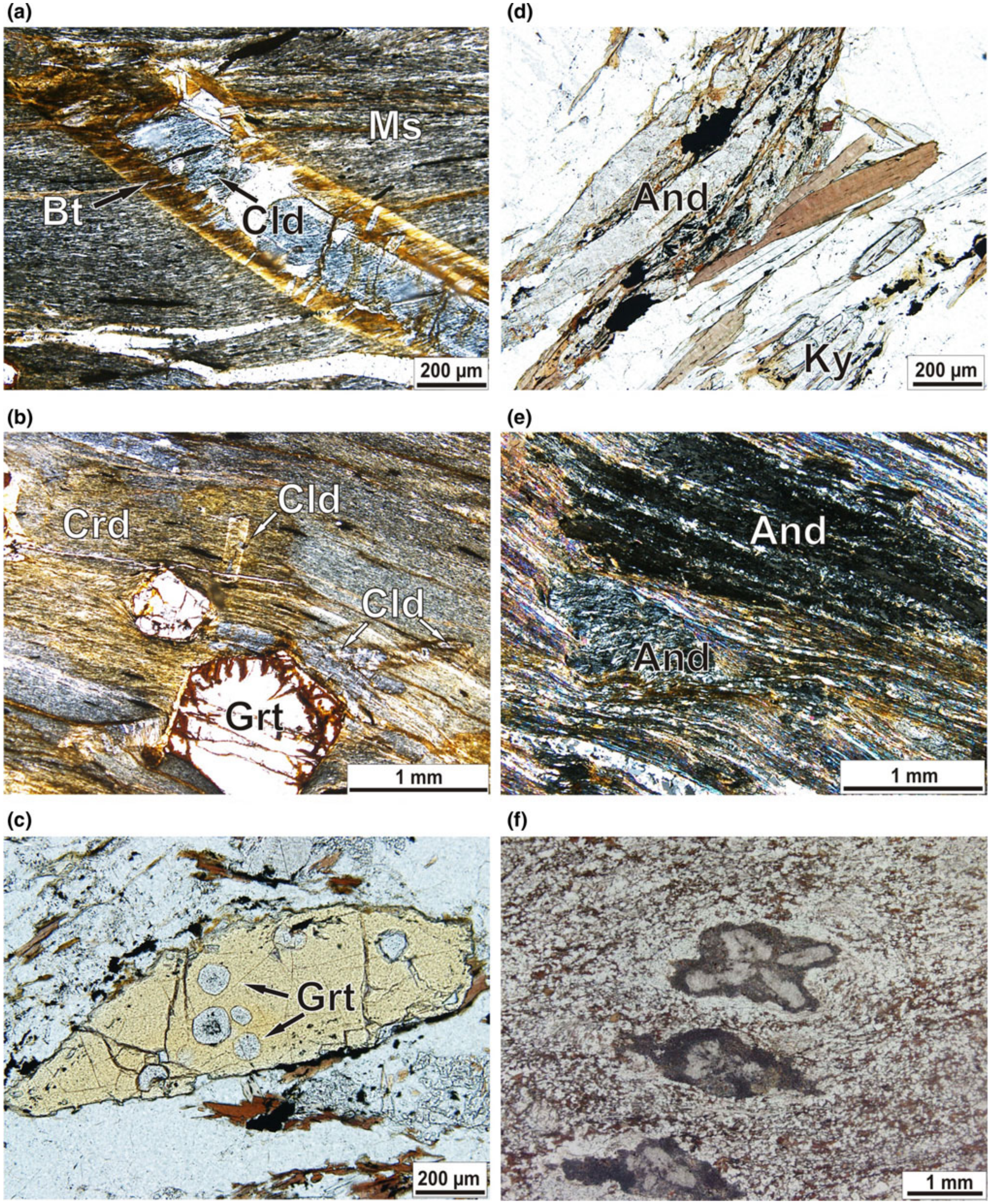

Fig. 9.13 Optical image with parallel nicols (**a**–**d** and **f**) and with crossed nicols (**e**) of black Gr-rich Upper Alpujárride metapelites (**a**-**b**-**e**: Benamocarra Unit, upper part of the succession; (**c**-**d**-**f**): Torrox Unit, lower part of the succession, immediately above the TGC). **a** Slightly deformed pre-Alpine Cld prism, partially transformed to Alpine Bt, Ms and opaque phases, within a foliated fine-grained matrix (S_{1V} foliation). **b** Pre-Alpine Cld and Grt porphyroblasts enclosed by altered pre-Alpine Crd. **c** Pre-Alpine St porphyroblast including small rounded Grt. **d** Pre-Alpine And pseudomorphs after Ky and relict Ky (lower part of the image). **e** Gr-rich domain corresponding to sheared And ghost, with a pre-Alpine chiastolite remnant on the left, wrapped by syn-S_A Qz, Bt and Ms. **f** Post S_{2V} pre-Alpine And chiastolites transformed to Alpine Ky at their rims, affected by the S_A foliation and surrounded by a mantle composed by Alpine Ms (lighter regions)

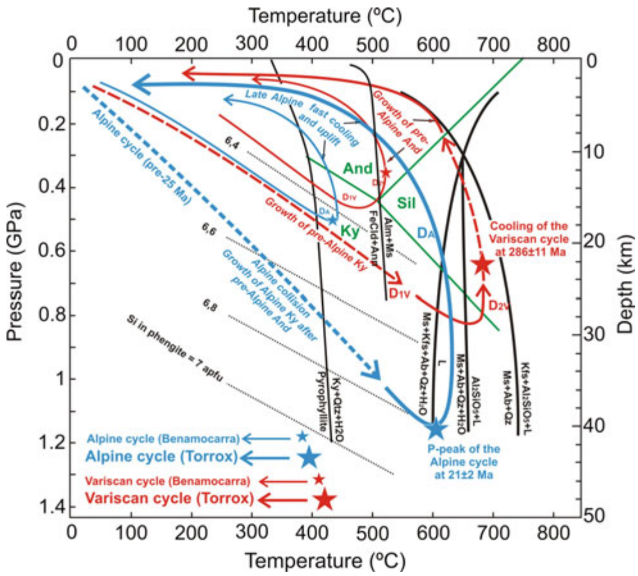

Fig. 9.14 Tentative pressure-temperature-time (P-T-t) Variscan (red) and Alpine (blue) paths of the lowest and upper parts of the pre-Mesozoic successions in the central Upper Alpujárrides, corresponding, respectively, to the Torrox and Benamocarra units (Sánchez-Navas et al. 2016, 2017), based on relevant reactions for partial melting and granitic melt crystallization in the symplified NKASH system (White et al. 2001), Si isopleths in muscovite-phengite solid solution in the KMASH system (Massonne and Schreyer 1987) and the phase diagram for Al_2SiO_5 polymorphs (Holdaway 1971). The garnet-in reaction is defined by the Fe end-member reaction: Fe-Cld + Ann = Alm + Ms (Spear and Cheney 1989). Dashed lines correspond to poorly defined parts of the P-T-t paths. The main pre-Alpine (D_{1V}, D_{2V}) and Alpine (D_A) deformation phases and growth episodes of kyanite (Ky) and andalusite (And) are also indicated

conditions, sometimes producing layeritic growth. After a sharp stop in its growth, the rims made of Alpine Grt formed, starting with sharp Ca-enrichments related to a marked P-increase, followed by a gradual P-decrease associated with a T-increase that reached maximum values during the final growth of the rims, probably during the Early Miocene.

The uppermost, finest-grained part of the pre-Mesozoic Alpujárride schists does not contain visible minerals at the naked eye or these minerals are strongly retromorphosed: they are represented by Gr-rich black spots in which the original mineralogy can be recognized with great difficulty and only occasionally under the microscope. These rocks, either when spotted or finest-grained, are indistinguishable from those that constitute the lowest part of the Maláguide Paleozoic succession. Consequently, in wide areas of the Montes de Malaga (especially in their eastern part: Fig. 9.1) and of the Costa del Sol, delineating the contact among typical Upper Alpujárride schists in lower tectonic position (*Benamocarra Unit:* Sánchez-Navas et al. 2016) and typical Maláguide Paleozoic rocks (*Piar Group*, see below) becomes an extremely difficult or simply impossible task. Actually, in these areas, the pre-Mesozoic Alpujárride schists seem to change, gradually upwards, to the lowest siliciclastic formations of the Maláguide Paleozoic, which are only slightly affected by metamorphism (*Morales Formation*). Nonetheless, a detailed study of the Benamocarra schists allows recognizing an equivalent evolution to that to the deepest part of the Upper Alpujárride succession represented by the TGC, but at low-P/medium-T metamorphic conditions (Fig. 9.14) that Sánchez-Navas et al. (2016) interpreted in relation to a late Variscan instead of a late Alpine extensional collapse as it is usually accepted for Alpujárride pre-Mesozoic rocks in most studies.

In the western Upper Alpujárrides and in their Sebtide counterparts the downwards-increasing metamorphic grade of the pre-Mesozoic crustal successions is evidently related to the emplacement of the Ronda and Beni Bousera ultramafic bodies. The crustal succession is detached from the underlying peridotites and related mafic rocks by an extensional ductile shear zone (Balanyà and García-Dueñas 1987; Balanyà et al. 1993, 1997; Argles et al. 1999). It is totally equivalent to the Alpujárride pre-Mesozoic successions above discussed. Actually, it constitutes a quite complete but extremely thinned crustal section (ca. 6 km at present), with transitional boundaries among successive lithotypes that are, from top to bottom, the same as those present in eastern areas. Nonetheless, in addition to the low-to medium-to high-grade metapelites and quartzites, a thick belt of pelitic migmatitic gneisses and Grt-bearing granulitic gneisses, associated with leucocratic magmatic differentiates, surrounds the contact of the crustal succession with the peridotites (Loomis 1972, 1975; Torres-Roldán 1981). The Grt-bearing rocks close to the peridotites are coarse-grained and banded, show granoblastic and blastomilonitic textures, and are made of Qz-Fds-Bt-rich bands bearing Kfs, Ky, Grt and Plg, with late prismatic Sil, Crd and hercinite replacing Grt. The Grt in these rocks shows the same zoning observed in Grt from the equivalent rocks previously mentioned. Monazite dating of the Grt cores demonstrates that the high-grade rocks attached to the peridotites are also polymetamorphic, and that they record a late Variscan high T event; in addition, Grt rims record a sharp Alpine P-increase followed by T-rise reaching its peak during the Miocene (Massonne 2014; see Montel et al. 2001 for equivalent rocks in the Moroccan Sebtides).

The Grt-bearing rocks at the hangingwall of the peridotites gradually evolve upwards to pelitic migmatitic gneisses by decreasing Fd and increasing Bt, by disappearing Sil, which is replaced by Fi, and by appearing Ms and St, the latter always wrapped around by a foliation (probaly

S_{2V}) and frequently replaced by And. The associated migmatite layers are stromatic and fluidal metatexites (Barich et al. 2014). The same rocks are found in the pelites located at the footwall of the peridotites (Blanca-type units). They show medium-to coarse-grained leucosomes forming cm- to dm-thick layers that are usually parallelized to the most evident foliation or that are slightly discordant to it. The melanosomes are medium- to fine-grained, and richer in Bt, Pl and Sil, with Grt mostly replaced by Crd (Torres-Roldán 1983; Acosta-Vigil et al. 2014, 2016; Bartoli et al. 2016). They evolve gradually outwards of the contact with the peridotites to the same schistose succession above discussed.

Both the Upper Alpujárride crustal succession and, especially, the peridotites located in its footwall, constitute the hangingwall of the Intermediate Alpujárrides (*Blanca-type units*) and of the Frontal *Nieves Unit* (Martín-Algarra 1987; Mazzoli and Martín-Algarra 2011). As a consequence of this lithosphere scale hot thrust, the Triassic to Lower Miocene locally fossiliferous successions below the peridotites were transformed to high-grade marbles and calc-schists, and their associated pelitic intervals, especially those of pre-Mesozoic age, were partially melted under high-grade conditions (Mazzoli and Martín-Algarra 2011, 2014; Mazzoli et al. 2013). Consequently, the presence of migmatitic gneisses with leucocratic differentiates is quite common not only in the crustal succession above the peridotites but also in rocks belonging to the tectonic units at the footwall of the Ronda peridotites, both pre-Mesozoic and Meso-Cenozoic. This high-T metamorphism and the related peridotite emplacement and deformation within the Alpujárride Complex and also in the Sebtides are evidently Alpine (Bouybaouène et al. 1998; Michard et al. 2006; Afiri et al. 2011; Ruiz-Cruz and Sanz de Galdeano 2012a). Moreover, stratigraphic data demonstrate that the age of this Alpine event is latest Aquitanian-early Burdigalian (Martín-Algarra and Estévez 1984) then at about 22–18 Ma, and this is confirmed by different radiometric dating methods applied to different Alpujárride-Sebtide rocks either above or below the peridotites (Priem et al. 1979; Zeck et al. 1989, 1992; Monié et al. 1991, 1994; Morillon et al. 1996; Sosson et al. 1998; Sánchez-Rodríguez and Gebauer 2000; Platt and Whitehouse 1999; Whitehouse and Platt 2003; Platt et al. 2003, 2005, 2006, 2013; Esteban et al. 2004, 2011a, b, 2013; Janots et al. 2006; Negro et al. 2006; Rossetti et al. 2010; Azdimousa et al. 2014; Sánchez-Navas et al. 2014; Gueydan et al. 2015) and also to the mafic layers within the peridotites themselves (Zindler et al. 1983; pyroxenite layers in Beni Bousera provide slighghtly older ages: Blichert-Toft et al. 1999). Nevertheless, zircon and monazite dating and detailed petrological and isotopic studies in both the peridotites (e.g. Pearson and Nowell 2004; González-Jiménez et al. 2017) and in the crustal rocks related to them reveal a much more complex picture, and provide increasingly stronger support for an early emplacement of the Ronda-Beni Bousera ultramafic slices at shallow lithospheric levels since, at least, latest Variscan time (Michard et al. 1997; Montel et al. 2001; Massonne 2014; Acosta-Vigil et al. 2014, 2016).

In the Guadaiza and Albornoque tectonic windows the migmatitic gneisses below the peridotites include abundant lithoclasts and enclaves coming from diverse Alpujárride lithologies, including also marbles (Lundeen 1978; Torres-Roldán 1983; Martín-Algarra 1987; Esteban et al. 2011a, b). The SHRIMP dating of zircon from these rocks indicates partial melting events at ca. 20–22 Ma (Esteban et al. 2011a). Downwards from the peridotites these rocks grade to dark (pre-Mesozoic) schists with abundant And that include, in their geometrically lowest part, gneissic bodies similar to the Torrox Gneissic Complex above mentioned (e.g., the *Istan gneiss*). Compositionally, these rocks are strongly deformed granites or granodiorites bearing Crd, Sil and And, and have provided magmatic pre-Alpine zircon ages (ca. 290–280 Ma: Acosta-Vigil et al. 2014).

In the highest tectonic units of the Intermediate Alpujárrides of the western (Blanca-type) and central sectors (Tejeda Unit), the dark schists unanimously considered of pre-Mesozoic stratigraphic age support a thick lithologic ensemble of much controversial age. It is made of light-coloured, gray to greenish schists and quartzites very rich in Chl and epidote and locally bearing Grt, blue amphibole, Cld and Als, which gradually changes upwards to a thick carbonate succession rich in tremolite and phlogopite and locally bearing also scapolite (Torres-Roldán 1974, 1978). The light-coloured schists and quartzites are classically interpreted as Permian to earliest Triassic probably vulcanosedimentary continental to coastal metasediments, and the overlying marbles as marine Triassic deposits (Blumenthal 1927; Torres-Roldán 1974; Delgado et al. 1981; Vera 2004). They are sometimes thicker than 500 m and include, in its lower part, leucocratic differentiates and, upwards, metabasite bodies that are also present within the lower part of the overlying carbonate succession, and that are usually interpreted in relation to the Triassic rifting. Nevertheless, these rocks have also been considered to be of pre-Mesozoic age, Cambrian, or pre-Cambrian, by some authors (Blumenthal 1949; Boulin 1970) although they have provided, up to now, only Alpine geochronological ages (Monié et al. 1991, 1994; Platt and Whitehouse 1999; Platt et al. 2003; Esteban et al. 2011a).

9.6 Pre-Mesozoic Successions of the Maláguide Complex

A. Martín-Algarra, R. Rodríguez-Cañero, P. Navas-Parejo, A. Jabaloy-Sánchez, S. Mazzoli, V. Perrone

9.6.1 General Features of the Maláguide Complex

A. Martín-Algarra, A. Jabaloy-Sánchez

The Maláguide Complex (Blumenthal 1927; Fallot 1948; Durand-Delga 1968; Vera 2004) includes the highest tectonic units of the Alpine thrust stack formed by the Betic Internal Zones (Fig. 9.1). It is made, essentially, of Ordovician (and older?) to Carboniferous deep marine turbiditic terrigenous clastics, with subordinate horizons of conglomerates, pelagic carbonates (sometimes conodont-bearing) and radiolarian cherts (Fig. 9.15). Shallow marine Paleozoic rocks are only represented by clasts of Frasnian to Bashkirian limestones included in uppermost Devonian and Upper Carboniferous conglomerates (Herbig 1984; Rodríguez-Cañero and Martín-Algarra 2014). Above them, an angular unconformity below Triassic continental red beds testifies for tectonic events related to the Variscan Orogeny (Foucault and Paquet 1971; Cuevas et al. 2001; Martín-Algarra et al 2009b). However, this was neglected in the Vélez Rubio area due to the intensity of the Alpine deformation that affects the Maláguide successions (Roep 1974).

The main outcrops of the Maláguide Complex locate in the Malaga area and in the Costa del Sol but also forming a narrow belt from W to E (Fig. 9.1) along the Internal-External Zone boundary (IEZB) in the Serranía de Ronda (Blumenthal 1927, 1930, 1949), in the Cogollos Vega Zone (Blumenthal 1928; Blumenthal and Fallot 1935), in the Velez Rubio Corridor and in Sierra Espuña (Fallot 1948). Small, scattered Maláguide outcrops S of the Sierras Nevada and Filabres and around the Sierras Cabrera and Alhamilla (Bodenhausen et al. 1967; Durand-Delga 1968), testify for its presence onto the thrust stack in the eastern Internal Zones before recent erosion (Fig. 9.1).

The Maláguide Complex thrusts over the Alpujárride Complex, although the original tectonic contact between them has been strongly modified by extensional tectonics in most sites, and now usually corresponds to an extensional detachment (Aldaya et al. 1991; Lonergan and Platt 1995; Fernández-Fernández et al. 2007). The top of the Maláguide Complex is an unconformity below Burdigalian deep clastic deposits (Viñuela Group: Martín-Algarra 1987) that predate back-thrusting of the Campo de Gibraltar Flysch Complex during the mid-Miocene Internal-External continental collision.

The structure of the Maláguide Complex is not fully understood in most sites. Several Alpine thrust slices are certainly present in several areas, but the main Maláguide outcrops of the Montes de Malaga and Costa del Sol (Fig. 9.1) seems to constitute, essentially, a single Alpine unit, with only some Alpine thrusts of limited lateral continuity and overstep. On the contrary, in the Maláguide outcrops close to the IEZB several Alpine thrusts bound at least three independent Alpine tectonic units (Roep and Mac Gillavry 1962; Geel 1973; Martín-Algarra 1987; Martín-Algarra et al. 2009b). Nevertheless, the exact number, continuity and extent, detailed stratigraphic features, correlation and kinematics of these tectonic units are not completely solved. The lowest one was slightly touched by the Alpine metamorphism and bears transitional stratigraphic and tectonometamorphic signatures towards the underlying Alpujárride terrains (Sanz de Galdeano et al. 1995a, b, 1999). The Alpine metamorphism is absent, or of very low-grade, in the overlying Maláguide units (Ruiz-Cruz 1997; Ruiz-Cruz and Nieto 2002; Ruiz-Cruz and Rodríguez-Jiménez 2002).

Stratigraphic and structural criteria clearly demonstrate the existence of pre-Alpine deformation of the Maláguide Paleozoic successions: an unconformity below continental redbeds (Saladilla Formation: Roep 1972) and above much more intensely deformed Paleozoic deep marine sediments is locally well preserved (Fig. 9.16a) in spite of later intense Alpine deformation and thrusting (Fig. 9.16b). Moreover, remnants of Variscan thrust slices (Fig. 9.17) are yet locally preserved *within* some Maláguide Alpine thrust masses in the Ardales area (Martín-Algarra et al. 2009a, b). These Alpine and Variscan thrust slices along the IEZB show slightly different Palaeozoic stratigraphy and facies, making evident that Maláguide successions come from diverse palaeogeographic realms. Studies on the Maláguide Paleozoic succession confirm that most horizons of fossiliferous Ordovician (Hirnantian) to Devonian (Famennian) limestones come from outcrops located in the outermost Maláguide outcrops and in the highest tectonic units (both Alpine and pre-Alpine) located along the IEZB and in the northern and western sectors of the largest Maláguide outcrops of the Montes de Málaga and the Costa del Sol (Herbig 1983, 1984, 1985; Rodríguez-Cañero et al. 2010; Navas-Parejo 2012). According to lithofacies criteria these outcrops represent the proximal (shallower) sectors of the Paleozoic basin. On the contrary, the distal (deep) area of the basin corresponds to the classical Maláguide stratigraphy, dominated by turbidites (Fig. 9.15). The later stratigraphy is also found in the structurally lowest Maláguide Alpine (and pre-Alpine) units outcropping along the IEZB and in the southern and eastern outcrops of this complex in the Montes de Malaga and the Costa del Sol.

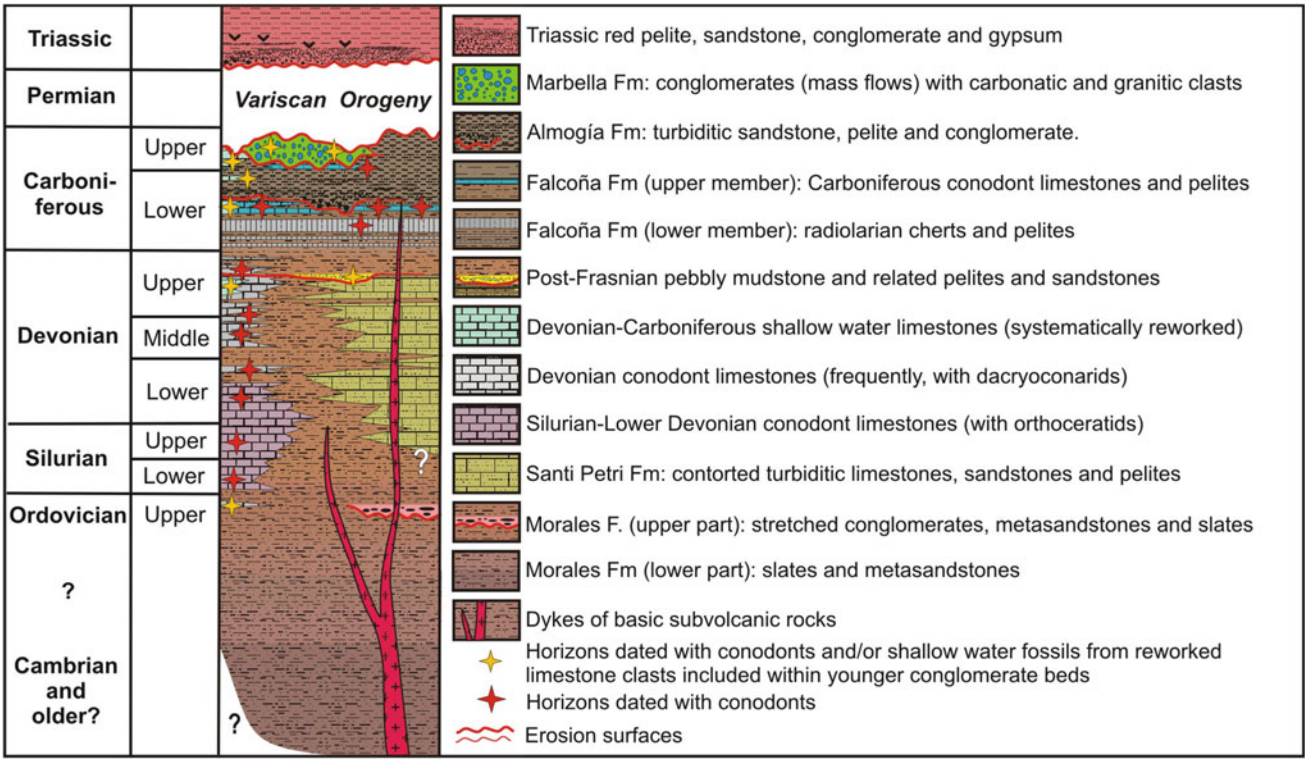

Fig. 9.15 Synthetic stratigraphy of the Maláguide Complex

9.6.2 Paleozoic (Piar Group)

A. Martín-Algarra, R. Rodríguez-Cañero, P. Navas-Parejo, V. Perrone

The Maláguide Paleozoic rocks are included in the Piar Group (Martín-Algarra 1987), previously considered as a formation (Soediono 1971; Geel 1973). This includes several thick Ordovician (and older?) to Carboniferous formations constituted by deep marine pelites and distal turbiditic litharenitic sandstones (Fig. 9.15). Subordinate fine- and coarse-grained conglomerate horizons essentially correspond to feeder channels of the turbidite systems. In addition to these predominant deep-sea terrigenous-clastics there also appear other much thinner formations made of pelagic carbonates, with different facies depending on their age, and of radiolarian cherts. The carbonates provide autochthonous fossils, mainly conodonts, from the Upper Ordovician (Hirnantian: Rodríguez-Cañero et al. 2010) up to the Upper Carboniferous (Bashkirian: Navas-Parejo et al. 2012b). They have provided also other biostratigraphically useful macro and microfossils: orthoceratid cephalopods (Silurian-lowermost Devonian: Blumenthal 1930), tintinnids (Silurian: Hermes 1966) and dacryoconarids (Devonian: Blumenthal 1930; Geel 1973; Herbig 1985; Navas-Parejo 2012). Radiolarian cherts have provided Lower Carboniferous radiolarians (O'Dogherty et al. 2000). Devonian-Carboniferous corals and other shallow marine fossils have been also found but exclusively within carbonate clasts from younger Paleozoic conglomerates (Blumenthal 1949; Boulin and Lys 1968; Geel 1973; Buchroithner et al. 1980; Herbig 1983, 1984, 1986, 1992; Herbig and Mamet 1985; Mamet and Herbig 1990; Rodríguez-Cañero and Martín-Algarra 2014).

The classical lithostratigraphy of the Piar Group was defined in the Montes de Malaga (Mon 1971; Herbig 1983, 1984) with five pre-Devonian to Upper Carboniferous formations named, from bottom to top: *Morales, Santi Petri, Falcoña, Almogía and Marbella formations* (Fig. 9.15). The Morales and Santi Petri Fms lack fossils but the latter underlies the well-dated beds of the Falcoña Fm: Tournaisian radiolarian cherts and Visean conodont limestones (Rodríguez-Cañero and Guerra-Merchán 1996; O'Dogherty et al. 2000). The Almogía and Marbella Fms contain Paleozoic fossils, but reworked, as they have been found within carbonate clasts in Carboniferous conglomerates.

The existence of fossiliferous Ordovician, Silurian and Devonian rocks in addition to Carboniferous formations is only demonstrated from thin horizons of condensed conodont-bearing limestones with diverse pelagic facies that are occasionally found associated with pelites and fine-grained sandstones (Kockel 1959; Kockel and Stoppel 1962; Rodríguez-Cañero 1993a). The lithostratigraphic correlation of these limestones with the classical Maláguide

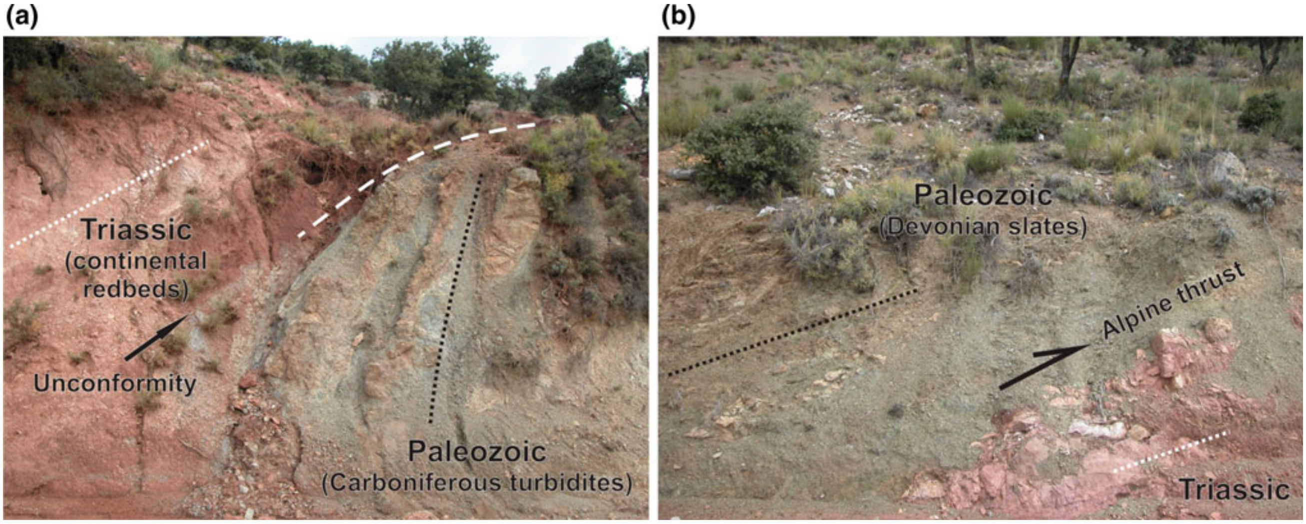

Fig. 9.16 **a** Unconformity below Triassic continental redbeds (Saladilla Fm) and steeply dipping Carboniferous turbiditic conglomerates and sandstones alternating with pelites (Almogía Fm). **b** Alpine thrust of Devonian deep-marine pelites onto Triassic continental redbeds. Both images come from La Solana outcrop in the Cogollos Vega Zone (see Fig. 9.1 and, for details, Fig. 9.2b in Navas Parejo et al. 2015)

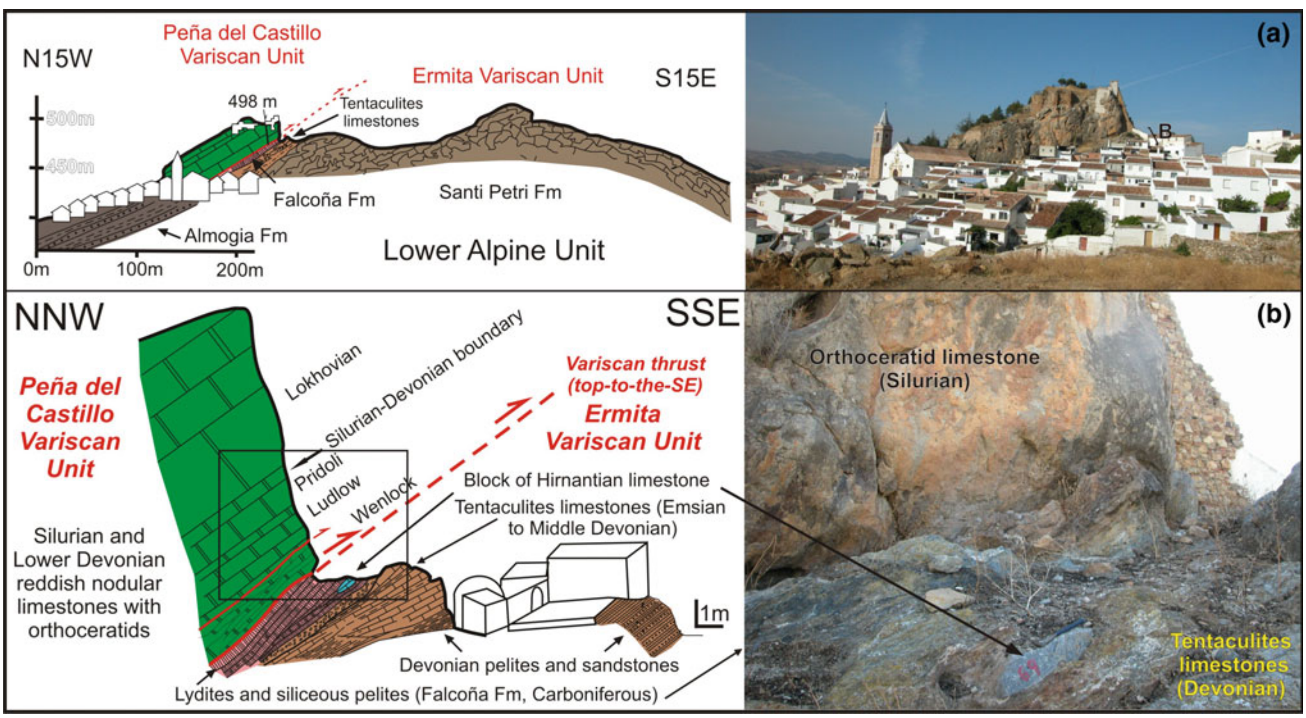

Fig. 9.17 **a** Geological section through the Ardales Castle Rock, showing two Variscan thrust units within the lower Maláguide Alpine Unit in the area. **b** Close-up of the thrust contact between the lower (Peña del Castillo) and upper (Ermita) Variscan units. This Variscan thrust emplaces Silurian limestones (dated by conodonts) onto bedded cherts correlated to the Carboniferous Falcoña Fm. The cherts stratigraphically underlie Visean conodont limestones (not present in this section but dated in a site located a few hundred of metres to the West: Navas-Parejo et al. 2012a). The Falcoña Fm cherts stratigraphically overlie a brecciated and laterally discontinuous lens of Tentaculites limestone, dated by conodonts as Emsian to Middle Devonian, which overlies foliated pelites and sandstones from the uppermost beds of the Santi Petri Fm. At that site, the beds just below the cherts, which are brecciated and rich in Fe-crusts, include one block of limestone that has provided the youngest Ordovician (Hirnantian) conodont association in Iberia (labelled with the red number 69, longest arrow, see text for details) Modified from Martín-Algarra et al. (2009a, b), and from Rodríguez-Cañero et al. (2010)

lithostratigraphy below the well-dated Falcoña Formation is far to be completely solved. Actually, these pre-Carboniferous limestones should be included in independent formations, which remain yet formally undefined. In addition, these pre-Carboniferous conodont-bearing limestones only appear in the highest tectonic units of the Maláguide Complex outcropping along the IEZB, in the outermost Maláguide outcrops in the Montes de Malaga (Almogía area) and in the western Costa del Sol (Marbella).

The stratigraphy of the Piar Group reveals two main evolutionary stages. The first stage comprises pre-Hirnantian to upper Visean-lower Serpukhovian) pre-orogenic sediments (with respect to the Variscan Orogeny). Their mainly deep marine facies, dominated by siliciclastic and calciclastic turbidites (Morales and Santi Petri Fms) relates most Maláguide outcrops to distal areas of a continental margin during its post-rift and syn-drift evolution. The sedimentation in the proximal areas of this margin was dominated by fine-grained clastics including the above-mentioned conodont-bearing condensed limestones. The basin became deepest, tectonically stable and morphologically homogeneous during the deposition of the Tournaisian-Visean cherts and conodont limestones (Falcoña Fm). The siliciclastic turbidites and conglomerates (Culm facies: Almogía and Marbella Fms) were deposited during the second stage, in relation to the Variscan synorogenic evolution of a convergent margin before its final orogenic deformation in Late Carboniferous time.

Dykes of mafic subvolcanic rocks (dolerites) that are usually interpreted to be arc tholeites of Tertiary age (Torres-Roldán et al. 1986; Turner et al. 1999; Esteban et al. 2013) crosscut the Benamocarra Unit and the lowest Maláguide formations (Morales to Falcoña), either in the southern and eastern Maláguide outcrops in the province of Malaga (Fig. 9.15) or in the lowest Maláguide units along the IEZB in the Velez Rubio Corridor (Fernández-Fernández et al. 2007). Nevertheless, the dikes have never been found along the IEZB in the provinces of Malaga and Granada nor in the upper Maláguide units of the Velez Rubio Corridor, in the western Maláguide outcrops of the Costa del Sol or in the northwestern areas of the Montes de Malaga (Almogía area). The dikes never crosscut the Almogía and Marbella Fms, or conodont-bearing limestones older than those of Visean age of the Falcoña Fm in the central Montes de Malaga. Mafic volcanics and dikes of subvolcanic rocks (but different to those above-mentioned) are locally found within Triassic continental red beds, however.

9.6.3 Pre-Orogenic Evolution

A. Martín-Algarra, R. Rodríguez-Cañero, P. Navas-Parejo

The ***Morales Formation*** (Mon 1971) includes the lowest deposits of the Piar Group. It is usually considered pre-Devonian and, maybe, pre-Silurian in age and constitutes a thick (>1 km in the eastern Montes de Malaga) monotonous pelitic-psammitic succession (Fig. 9.18a). It shows intense deformation and low-grade metamorphic overprint, both pre-Alpine and Alpine, which certainly increase downwards from anchimetamorphic to low-grade to medium-grade metamorphic rocks (Ruiz-Cruz 1997; Ruiz-Cruz and Nieto 2002).

In addition to the dikes of basic subvolcanics mentioned above, the pelites of the lower part of the Morales Fm typically include mm- to cm-sized graphite-rich black spots with increasing size downwards, that correspond to sericiticed andalusite, chloritoid and garnet (Orozco and Gálvez 1979; Gálvez and Orozco 1980). These Alpujárride-like spotted schists are identical to those included in the Benamocarra Unit that outcrops below the Maláguide Complex in the eastern Montes de Malaga (Sánchez-Navas et al. 2016). This suggests gradual lithological transition from Paleozoic Alpujárride (Benamocarra) to Maláguide rocks in this region where, actually, it is practically impossible to precisely locate in the field a tectonic contact between typical Maláguide and Alpujárride rocks.

In spite of widespread pre-Alpine foliations (see below), the psammites of the Morales Formation locally preserve sedimentary structures (scours marks, cross lamination, graded beds) and locally intercalate decimetre to metre-thick beds of black turbiditic limestones (Fig. 9.18c) made of crinoid bioclasts, testifying for the marine origin of this formation. The stratigraphic upper part of the succession includes metre- to decametre thick horizons of *stretched conglomerates* (Fig. 9.18b) constituted mainly by clasts of quartzites, but pelites, sandstones, black cherts, recrystallized limestones, schists and gneisses are also present.

The ***Santi Petri Formation*** (Mon 1971) lies above the Morales Fm in rapid but gradual lithological transition. It is up to several hundred of metres thick but it laterally changes notably due to tectonic causes. Its most typical lithofacies is known as ***calizas alabeadas***, that is, warped/contorted limestones (Fig. 9.18d, e). These are intensely folded, finely terrigenous turbiditic limestones that vertically and laterally change to calcareous greywackes and pelites showing very similar features to those of the upper part of the Morales Fm. Typically, these rocks are black or dark-grey coloured when fresh, brown-greenish when weathered, thin to medium-bedded (Fig. 9.18f) and crosscut by numerous white veins filled with calcite and/or quartz. They contain frequent incomplete Bouma sequences, rippled surfaces (Fig. 9.18g) and sole marks (Fig. 9.18h, i) that evidence their distal turbiditic origin. They have not provided yet any fossil and remain undated, but are usually considered to be of late Devonian age, as they are always located stratigraphically below the Falcoña Fm, of well-dated Carboniferous age (Tournaisian-Visean).

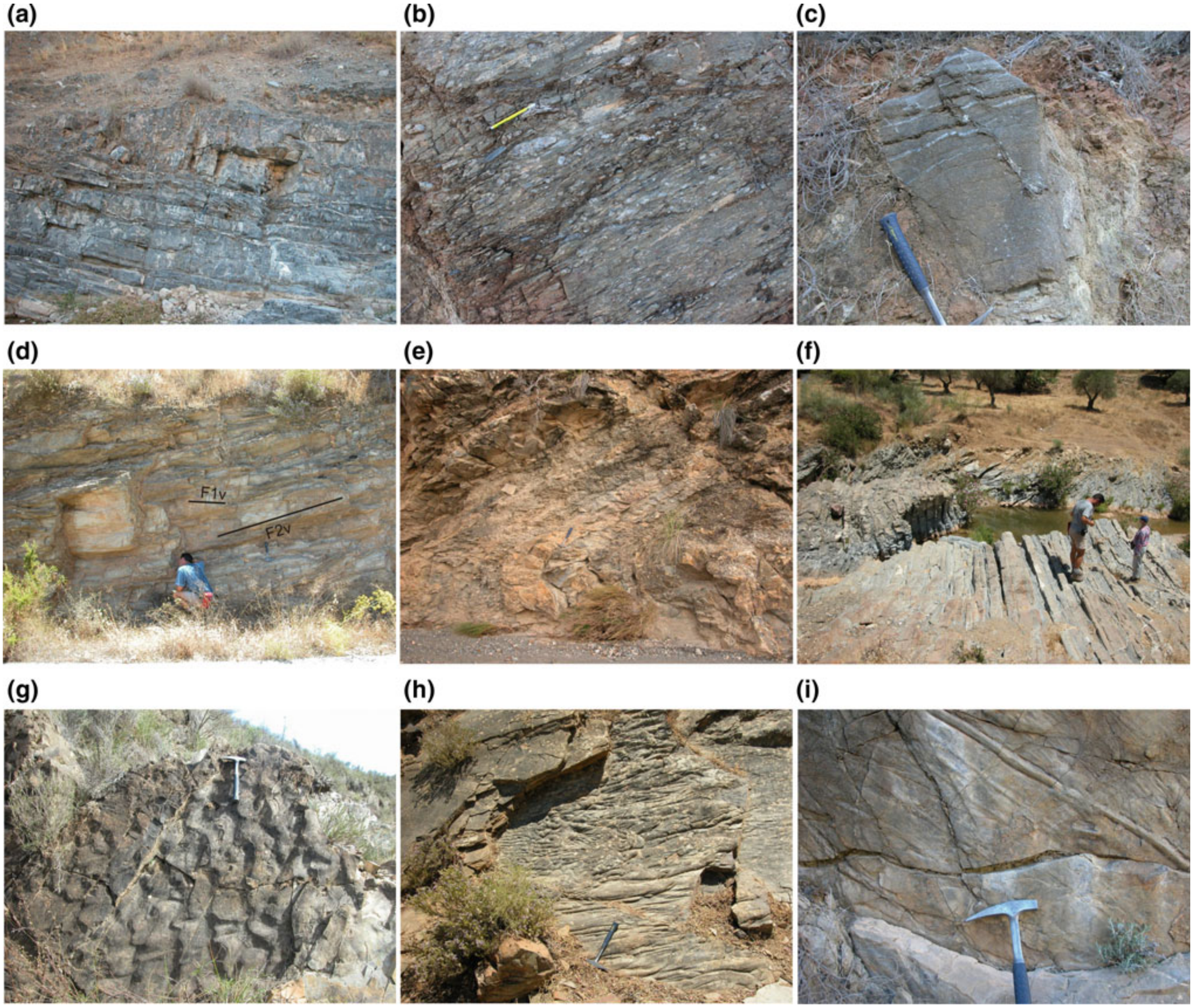

Fig. 9.18 Field views of the Morales (**a**–**c**) and Santi Petri (**d**–**i**) Formations. (**a**–**d**) and (**f**) come from the Montes de Málaga; (**e**, **g**–**i**) from the Vélez Rubio Corridor. **a** Well-bedded sandstones. **b** Stretched conglomerate. **c** Turbiditic limestone rich in crinoid debris. **d** Typical aspect of the "calizas alabeadas" (contorted limestones) facies, which is, mainly, a result of superimposed folding of variscan age (F1v, older, and F2v, younger). **e** Recumbent fold with well-developed axial plane foliation in alternating pelites and limestones. **f** Moderately folded distal turbiditic succession of alternating limestones, sandstones and pelites. **g** Rippled bed surface atop of a turbidite bed. **h** Sole marks, mainly flute and bounce casts. **i** Grooved sole of a turbidite bed

The Falcoña Fm has been also well characterized in all the outermost Maláguide outcrops in the western Montes de Malaga (Almogía area: Kockel 1959, Kockel and Stoppel 1962; Rodríguez-Cañero 1993a, b, 1995), the westernmost Costa del Sol (Marbella area, Arroyo de la Cruz: Herbig 1985) and, especially, in those mainly belonging to the highest Alpine Maláguide units outcropping along the IEZB, from west to east: Algatocín and Ardales-Chorro areas (Kockel 1959, 1964; Kockel and Stoppel 1962; Rodríguez-Cañero et al. 1997; Martín-Algarra et al 2009a; Cogollos Vega Zone Navas-Parejo et al. 2011, 2015a, b; Velez Rubio Corridor (Navas-Parejo 2012). In the higher Alpine Maláguide units of the latter areas, the Morales and Santi Petri Fms do not exist stratigraphically below the Falcoña Fm: they are replaced by a succession dominated by pelites, with subordinate thin- to medium-bedded sandstones that occasionally include laterally discontinuous ***conodont-bearing limestone lenses*** of Silurian and Devonian age. The original stratigraphic thickness of these conodont-bearing limestones has been strongly reduced by intense tectonic stretching in most sites but, locally (e.g. the Ardales Castle Rock: Fig. 9.17), they are up to several tens of metres thick. Their existence demonstrates that the Maláguide palaeogeography changed laterally from deeper, basinal siliciclastic or mixed siliciclastic/carbonatic turbiditic environments (represented by the Morales Fm and, especially, by the Santi Petri Fm) to less deep successions including condensed pelagic limestones and alternating platy limestones and pelites. Regional outcrop distribution suggests that deepest successions were located towards the SE (in present-day coordinates) and that they evolved to less deep areas towards the NW within the same continental margin.

Ordovician conodonts have been only found at Ardales (Rodríguez-Cañero et al. 2010; Sarmiento et al. 2011), in *one* sample coming from a boulder of recrystallized bioclastic limestone (Fig. 9.17b), firstly observed and differentiated, but not dated, by Blumenthal (1930). It is included within siliceous–ferruginous–calcareous claystones associated with a laterally discontinuous calcareous breccia horizon lying below well bedded cherts that correlate to the Falcoña Fm and that lie onto a decametre-sized Devonian (Emsian) *Tentaculites* limestone lens (Rodríguez-Cañero 1993a), which probably resulted from submarine sliding and resedimentation (Martín-Algarra et al. 2009a). The original stratigraphic location within the Maláguide succession of the Ordovician beds from which this block was eroded is unknown. Interestingly, the block contains a rich Hirnantian conodont fauna showing remarkable differences with coeval Iberian conodont associations, but strong palaeobiogeographic affinities with Austroalpine-South-Alpine coeval faunas found in the Uggwa and Wolayer Fms (Rodríguez-Cañero et al. 2010; see also Sarmiento et al. 2011, Schönlaub 1988). This Hirnantian fauna indicates deposition (of the limestone bed where the block come from) in cold-water marine environments at high latitudes like coeval deposits in the Alps, but not in glaciomarine settings like those present in the Variscan areas of the Iberian massif at that time (Rodríguez-Cañero et al. 2010; Sarmiento et al. 2011).

Silurian and lowermost Devonian conodonts have been found in red, brown and grey nodular limestones (Fig. 9.19a) bearing orthoceratids and scyphocrinites, with biomicritic microfacies including bioclasts of trilobites, cephalopods, crinoids, ostracods and tintinnids (Hermes 1966). The thickest outcrop is that of the Ardales Castle Rock (Fig. 9.17a) where Rodríguez-Cañero (1993a; see also García-López et al. 1996; Martín-Algarra et al. 2009a; Rodríguez-Cañero et al. 1997, 2010) identified diverse conodont biozones, from the Llandovery to the Pragian. Equivalent facies and faunas have been found also in the Serranía de Ronda, the Montes de Málaga (Kockel 1959, 1964; Kockel and Stoppel 1962) and the Vélez Rubio Corridor (Van den Boogaard 1965). These Silurian to Lower Devonian beds were deposited in distal ramp to pelagic high bottom environments, with condensed sedimentation of moderate depth, at middle latitudes, within a divergent continental margin related to the Paleotethys opening (Herbig 1992).

Lower and Middle Devonian conodont bearing limestones also contain, commonly, abundant dacryoconarids (*Tentaculites* limestone facies). These facies are frequently formed by platy limestones interlayered with pelites (Fig. 9.19b), sometimes reddish-pinkish or varicoloured although they locally constitute a few metres thick calcareous horizons of alternating thin to medium-bedded grey limestones and grey to pinkish nodular limestones (Fig. 9.19c) including thin pelitic partings and platy limestones sometimes crowded with dacryoconarids. Geel (1973) differentiated the reddish shales as a “variegated phyllite member” within her Piar Formation that she frequently found in relation to her *Tentaculites* limestone member, but similar pelites also appear associated with conodont-bearing Carboniferous limestones belonging to the Falcoña Fm (Navas-Parejo et al. 2015b).

The Upper Devonian beds show equivalent features, but they are much poorer in dacryoconarids during the Frasnian. A stratigraphic gap (Fig. 9.19d) is systematically detected at the base of Famennian beds (Fig. 9.19f), usually the richest conodont-bearing horizons within the Maláguide Complex (Rodríguez-Cañero 1993b, 1995). This gap records the Frasnian-Famennian crisis (Kellwasser event) in the Maláguide Complex (Rodríguez-Cañero 1993b) and is associated with Fe-rich crusts and nodules (Fig. 9.19d) and with redeposited horizons (pebbly mudstones and coarse-grained carbonate turbidites). The latter provide abundant carbonate clasts with shallow marine to pelagic microfacies (including oolites, corals, stromatoporoids, ostracods, algae, crinoids, etc.), and with Frasnian conodont biofacies indicating shallow marine to moderately deep to very deep depositional settings (Fig. 9.20). These clasts were eroded from late Devonian carbonate platforms formed under warm subtropical climate, and from laterally related and deeper pelagic beds, during a period of tectonic instability of the Maláguide basin, as it is also frequently observed worldwide during the Frasnian-Famennian crisis (Rodríguez-Cañero and Martín-Algarra 2014, and references therein). The Devonian shallow carbonate platform was, however, totally dismantled.

In the northwestern Montes de Malaga, a few metres thick -sometimes reddish or pinkish, sometimes dark-coloured- pelitic interval above the Santi Petri Fm includes some horizons of calcareous greywackes (“*serie Rana*” of Kockel and Stoppel 1962). This interval gradually changes upwards to the ***Falcoña Formation*** (Herbig 1983), which includes a thin (usually a few metres, up to 20 m at most) but conspicuous Lower Carboniferous chert-limestone interval. The cherts are much more commonly found than the related limestones, and are well bedded in thin to very thin beds (Fig. 9.19f). Although their original color was black or dark gray, occassionally reddish, and yet preserve abundant but usually flattened and recrystallised radiolarians

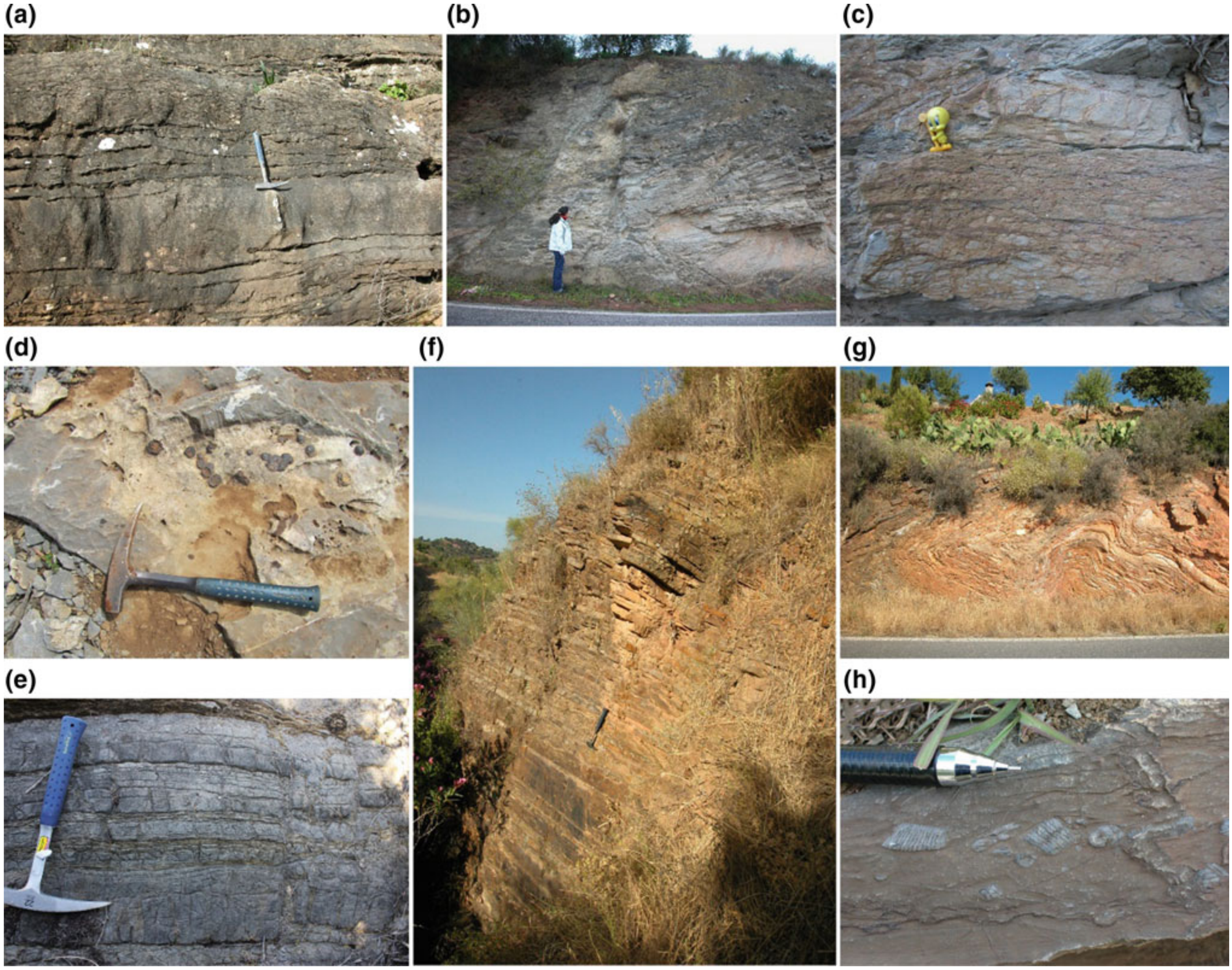

Fig. 9.19 Silurian-Devonian conodont limestones and related facies (**a**–**e**) and Falcoña Formation (**f**–**h**). **a** Silurian nodular limestones (Arroyo de las Viñas section, Ardales). **b** Lower Devonian platy and nodular limestones alternating with thin shale partings (Almogía). **c** Close-up of a nodular limestone of the same outcrop as B (Tweety for scale is 4 cm high). **d** Fe-rich discontinuity surface onto uppermost Frasnian beds encrusted by Fe-rich nodules and patinas covered by Famennian limestones, which delineates the Kellwasser event at Almogía. **e** Alternating calcarenitic and micritic slightly nodular Famennian beds (Arroyo de la Cruz, Marbella). **f** Gradual very well bedded transition from uppermost Devonian pelites to Tournaisian dark radiolarian cherts (lydites) of the lower member of of the Falcoña Fm in its type area (Cortijo de la Falcoña, Montes de Málaga). **g** Strongly folded lydites of the lower member of the Falcoña Fm (Montes de Málaga, old road Málaga-Casabermeja). **h** Crinoid stems in Visean limestones of the upper member of the Falcoña Fm (Montes de Málaga)

(O'Dogherty et al. 2000), the radiolarian cherts (lydites) are commonly transformed to fine-grained reddish-orangish quartzites, usually intensely folded (Fig. 9.19g). The lydites constitute the lower member of the Falcoña Formation and are commonly found in all areas, in spite of being laterally discontinuous due to intense folding, shearing and boudinage (Orozco and Gálvez 1979; Gálvez and Orozco 1979, 1980). Above the cherts, the upper member of the Falcoña Fm. is made of micritic limestones sometimes including dark grey chert (and quartz) nodules and stretched crinoid stems (Fig. 9.19h). Thinly bedded limestones interlayered with calcareous slates gradually change upwards to pelites overlying the Falcoña Fm. In the type section, the cherts provided Tournaisian radiolarians and the limestones Visean conodonts (Rodríguez-Cañero 1993a; Rodríguez-Cañero and Guerra-Merchán 1996; O'Dogherty et al. 2000). Consequently, this formation is well dated as Lower Carboniferous. This is confirmed in many other sites (Navas-Parejo et al. 2008, 2012a, 2015a, b).

The Falcoña Fm is present in all Maláguide outcrops and tectonic units. Actually, it is a key horizon for stratigraphic correlation among Maláguide Paleozoic successions bearing the Morales and Santi Petri Fms and those without them but bearing Ordovician-Silurian-Devonian limestones with

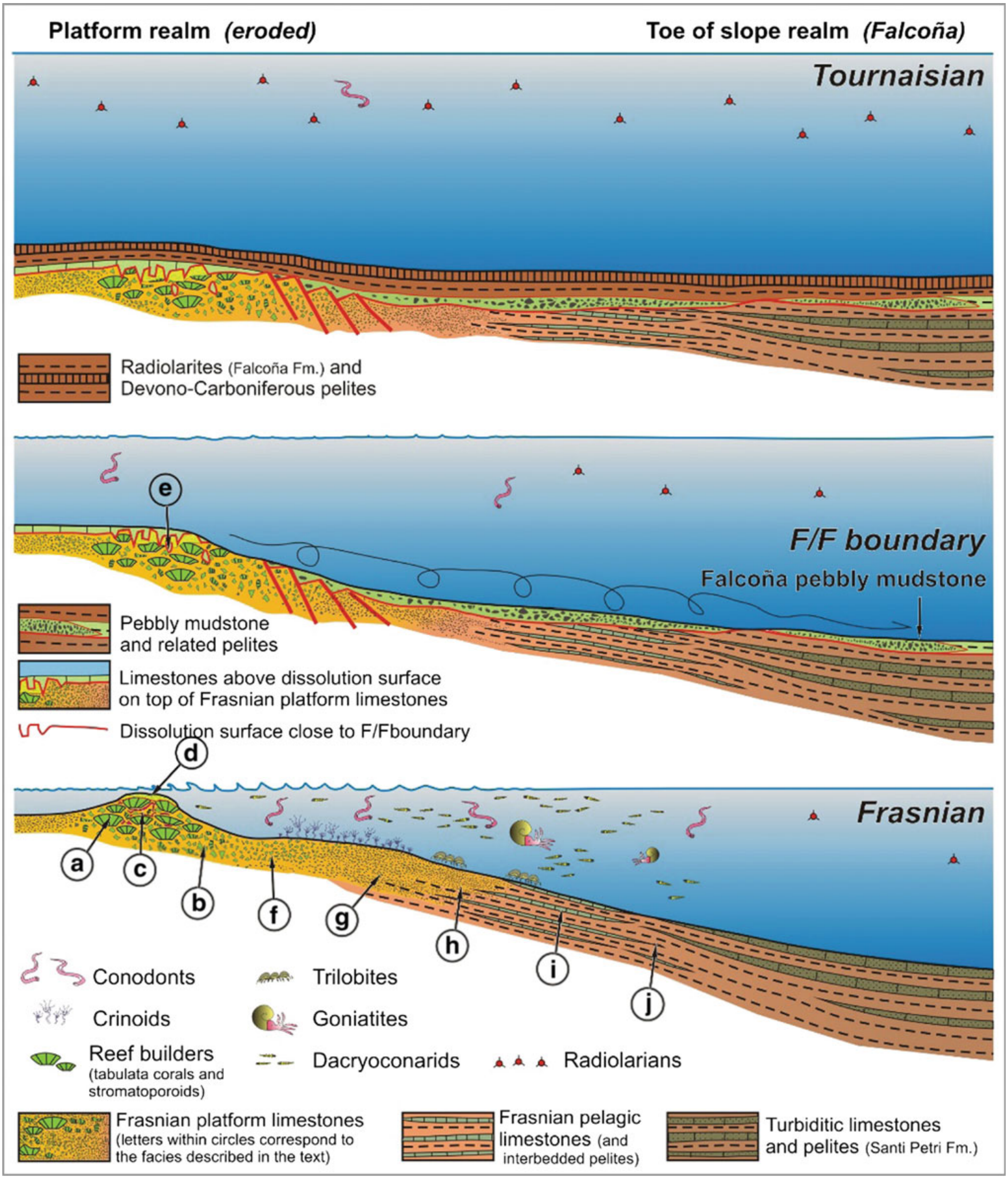

Fig. 9.20 Model on the sedimentary evolution of the Maláguide basin in the frame of a Paleotethyan margin during the Late Devonian and Early Carboniferous, with a shallow marine proximal area distally connected with a turbiditic deep basin where the Santi Petri Fm and related facies were being deposited. The shallow water area collapsed during a tectonic event related to the Frasnian/Famennian crisis, and was dismantled to produce the pebbly mudstone horizon at the base of the Falcoña Fm. During the successive Tournaisian deepening the bottom irregularities attenuated and a widespread deposition of radiolarites accounted everywhere in the Maláguide basin (modified from Rodríguez-Cañero and Martín-Algarra 2014)

conodonts (Fig. 9.15). In the central-eastern Montes de Malaga, in the Costa del Sol and in the tectonically lower Maláguide units along the IEZB the Falcoña Fm is well dated but the stratigraphic age of the underlying Santi Petri and Morales Fms is controversial or at least uncertain. Most field data point to a late Devonian age for the Santi Petri Fm and probably also for the upper part of the Morales Formation (Rodríguez-Cañero and Martín-Algarra 2014). Nevertheless, intense tectonic folding does not preclude other possibilities. Actually, the existence of reverse limbs affecting the contact between the Santi Petri and Falcoña Fms can be locally demonstrated and, in many areas, the lithological similarities among the Almogía (see below) and the Morales Fms make difficult to distinguish among them if the Falcoña Fm is absent and the Santi Petri Fm stratigraphically reversed.

After the tectonic disturbances recorded by the Kellwasser event, the deposition of the Falcoña Fm represents a period of tectonic quiescence dominated by thermal subsidence in the Maláguide basin (Fig. 9.20). The previously formed marine bottom irregularities were levelled and a marked deepening of the whole basin led to radiolarian-rich sedimentation below the CCD and to deposition of the siliceous member (lydites) of the Falcoña Fm during the Tournaisian (O'Dogherty et al. 2000). After siliceous sedimentation a lowering CCD was responsible for deposition of pelagic limestones sometimes including chert nodules and isolated crinoid stems with Visean conodonts and deep (*Gnathodus*) biofacies (Rodríguez-Cañero and Guerra-Merchán 1996). A similar vertical facies evolution is commonly observed in many other European Variscan regions.

When the successions are more complete and tectonically less disturbed, the upper member of the Falcoña Fm changes, rapidly but gradually, to a pelitic succession with identic features to those of the finer-grained facies of the Almogía Fm. In addition, a carbonate horizon intercalated between pelites similar to those of the Almogía Fm has recently provided the youngest conodont association found in the Maláguide Complex, of early Bashkirian age (Navas-Parejo et al. 2012b). Then, the Maláguide sedimentation shows deep-water basinal facies from the Tournaisian up to the earliest Bashkirian. In addition, the existence of Visean to Early Bashkirian shallow marine carbonate platform environments in the Maláguide realm (now dismantled) or close to it is also revealed by carbonate clasts with shallow marine facies and fossils found, ocassionally, in the conglomerates of the overlying Almogía Fm and, especially and systematically, in those of the Marbella Fm (Buchroithner et al. 1980; Herbig 1984, 1986; Herbig and Mamet 1985; Mamet and Herbig 1990).

9.6.4 Syn-orogenic Evolution

A. Martín-Algarra, R. Rodríguez-Cañero, P Navas-Parejo, S. Mazzoli, V. Perrone

The Variscan evolution in the Maláguide Complex started with deposition of the *Almogía Fm* and culminated after deposition of the *Marbella Fm*, which are usually attributed to the synorogenic Culm facies. Their deposition evidence the progressive erosional dismantling of *previously deformed* Paleozoic and/or older successions, as well as that of Devonian-Carboniferous carbonate platforms probably formed on them and/or on laterally related successions now outcropping in the Rif (Chalouan 1986; Chalouan and Michard 1990, 2004). These sediments were redeposited within turbiditic synorogenic basins from the latest Early Carboniferous to the Late Carboniferous. After that, the main (late) Variscan deformation accounted in the Maláguide Complex, as revealed by a marked unconformity below the Triassic beds of the Saladilla Fm.

The starting deposition of the Almogía Fm was associated with the post-Visean onset of the (frequently coarse-grained) siliciclastic sedimentation everywhere within the Maláguide basin (Figs. 9.15 and 9.21a–e). Moreover, although stratigraphic and tectonic relationships between the Falcoña and Almogía Fms are frequently masked by poor outcrop exposures or by faulting, a marked contrast in the tectonic style and/or in the intensity of tectonic deformation can be observed between them: the Falcoña Fm beds are usually much more intensely folded than, and are laterally discontinuous below, the Almogía Fm. This suggests that an unconformity exists between them and that the Maláguide basin was probably affected by important tectonic events since the Serpukhovian or, more probably after the early Bashkirian, although this is not precisely dated. Actually, the Almogía Fm only provides redeposited Devonian and Early Carboniferous conodont faunas that have been extracted from carbonate clasts included in the conglomerate horizons that frequently (but not always) characterize its stratigraphic lowest part. Recent zircon dating indicates predominant Cambro-Ordovician and older magmatic ages for most clasts of crystalline rocks within these conglomerates although a few zircon grains with Permian ages have also been obtained (Esteban et al. 2017), but these younger ages need to be confirmed and/or re-evaluated.

The ***Almogía Fm*** is a several hundred of metres thick succession of siliciclastic turbidites dominated by fine-grained pelites and greywackes with undeterminable plant remains and Bouma sequences, which commonly include also subordinate conglomeratic horizons (Blumenthal 1930, 1949; Soediono 1971; Mon 1971; Geel 1973; Herbig

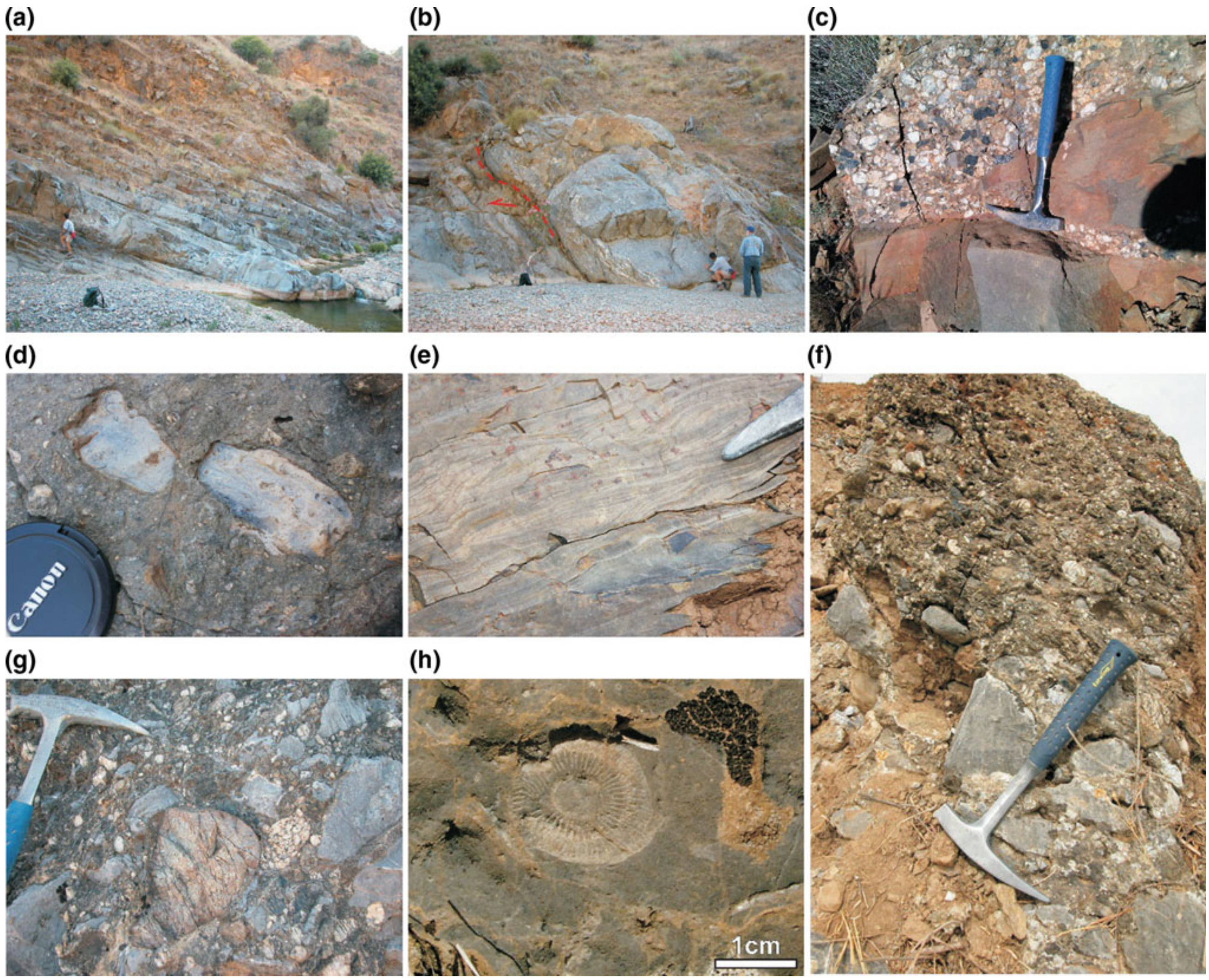

Fig. 9.21 Field views of the Almogía Formation, mostly Retamares member (**a**–**d**), in the Montes de Malaga (**a**, **b**, **e**), Costa del Sol (**d**) and Vélez Rubio Corridor (**c**), and of the Marbella Formation in the Montes de Málaga (**f**, **g**), and Vélez Rubio (**h**). **a** Thinning and fining-upwards turbiditic succession (mainly greywackes with subordinate conglomerates. **b** Channeled conglomerate bed that buries a synsedimentary normal fault. **c** Channeled conglomerate beds including black clasts of lydite eroded from the Falcoña Fm) and white clasts of quatzites and of granites. **d** Clasts of Tentaculites limestone included in a microconglomeratic matrix. **e** Fine-grained rippled sandstone in a stratigraphically reversed turbidite bed. **f** Typical outcrop of coarse-grained and normally graded Marbella Fm conglomerates including clasts of Visean to lowermost Bashkirian shallow marine limestones. **g** Close-up of a Marbella Fm conglomerate, where rounded clasts of granites and angular clasts of limestones are clearly visible. **h** Detail of a coral included in a limestone clast

1983, 1984; Martín-Algarra 1987; Herbig and Statteger 1989; Mayoral et al. 2018). The succession is organized as a thinning- and fining-upwards turbiditic megasequence, whose lower part is dominated by greywackes sometimes forming normally graded beds up to several metres thick stacked in metre to decametre thick thinning and fining upwards sequences (Fig. 9.21a) with commonly channeled bases (Fig. 9.21b) filled with coarse- to very coarse-grained conglomerates (Fig. 9.21c) and/or finer conglomeratic to coarse-grained amalgamated sanstone beds with commonly scoured bases. Its upper part is dominated by fine-grained olive-green pelites with subordinate interlayered thin greywacke beds. Laterally discontinuous coarser-grained conglomeratic beds (from granule up to cobble, rarely boulder size) with clasts usually subrounded to well rounded, are also frequently present in the greywackes of the lower part of the formation, which is known as Retamares Member (Kockel and Stoppel 1962). The later beds can be up to several metres thick and include clasts eroded from the underlying succession, among them clasts of pelagic limestones with Devonian and Carboniferous conodonts (Fig. 9.21d) and, especially, of black chert (Fig. 9.21d) and of greywackes eroded from the

Falcoña Fm and from older formations. In addition, clasts of crystalline rocks, mostly granitoids and gneisses of unknown provenance are frequently present. The conglomerate and thickest sandstone beds, with erosional (channeled and scoured) bases and that are normally graded, indicate deposition in channelled areas of a deep clastic system, whereas the finer-grained and thinner bedded greywackes and the related pelites indicate outermost areas of deep sea fans and/or basin plain environments. Heavy mineral spectra and clastic provenance analysis indicate a tectonically active source that was in very rapic tectonic rise, certainly forming a nascent orogenic belt related to the development of a rapidly evolving convergent margin (Herbig and Statteger 1989). Nevertheless, basin unstability atenuated gradually with time as indicated by the thinning and fining upwards megasequential evolution. Recent palaeoichnological data from the eastern Montes de Málaga indicate that the sequence became shallower upwards (Mayoral et al. 2018).

The ***Marbella Fm*** is dominated by coarse to very coarse-grained olistostromic conglomerates that systematically include granitic, gneissic and especially carbonatic clasts of cobble to boulder size, frequently with reverse grading although normal grading is also present (Fig. 9.21f). Finer grained facies are constituted by siliciclastic turbidites mixed with some carbonate components. The Marbella Fm conglomerates shows deep erosional bases on the Almogía Fm, whose upper part is systematically much finer-grained, and were certainly deposited by debris flows (Herbig 1984; Martín-Algarra 1987). The carbonate clasts are commonly abundant and sometimes huge (Fig. 9.21f, g), and contain shallow marine fossils like corals (Fig. 9.21h), bivalves, benthic foraminifera, algae etc., that allow precise biostratigraphic dating and reconstructing diverse Visean to Early Bashkirian shallow marine environments that are now totally dismantled (Boulin and Lys 1968; Boulin 1970; Geel 1973; Buchroithner et al. 1980; Herbig 1984, 1986; Herbig and Mamet 1985; Mamet and Herbig 1990). The common presence -and locally abundance- of clasts of gneisses, granites (Fig. 9.21g), quartzites and metapelites in the Marbella Fm is particularly remarkable, as well as that subordinate of rare metabasites and mafic plutonic rocks that are also occasionally present. This reveals a deep and renewed erosion of a pre-Variscan (Early Paleozoic and/or older) crystalline basement that was being affected by rapid tectonic rise and orogenic deformation after a period of tectonic quiescence during sedimentation of the upper part of the Almogía Fm. Again, this is a typical feature of synorogenic deposits, as also revealed by the spectrum of heavy mineral data that, as in the case of the Almogía Fm, show contrasting clastic provenance signatures with respect to those of the underlying preorogenic deposits (Herbig and Statteger 1989; Henningsen and Herbig 1990).

9.6.5 Variscan Tectonics and Metamorphism in the Maláguide Complex

A. Martín-Algarra, S. Mazzoli, V. Perrone

The importance and intensity of the Variscan versus Alpine tectonics in the Maláguide Complex have been debated (Foucault and Paquet 1971; Roep 1974; Orozco and Gálvez 1979; Gálvez and Orozco 1979, 1980; Cuevas et al. 2001). Nonetheless, in spite of the intense Alpine deformation affecting basement and cover together, the Maláguide Paleozoic shows strong tectonic deformation and mild metamorphic recrystallization that are absent in the cover. The marked depositional contrasts between deep marine Paleozoic successions and overlying Triassic continental red beds also indicates that the Maláguide Paleozoic was affected by intense Variscan deformations, whose kinematics, timing and relationships with the tectonometamorphic evolution at deeper structural levels (Alpujárride Complex) are still poorly understood.

In relation to the Variscan tectonometamorphic evolution, the composition of chlorite in the rocks on both sides of the unconformity reveals an abrupt increase in the temperature of the metamorphism in the rocks successions belonging to the Paleozoic Piar Group with respect to the Triassic Saladilla Formation (Ruiz-Cruz 1997). In addition, the Alpine metamorphic overprint is demonstrated by the gradual evolution of the crystalchemical parametres of other mineral phases in both successions, in particular by the IC index and by the evolution of the kaolinite polymorphs (Ruiz-Cruz 1997). Concerning the Variscan tectonometamorphic evolution, IC values = 0.32–0.25 corresponding to the deep anchizone characterize the Morales Fm, whose lowest beds, bearing vermiculite, biotite, paragonite or chloritoid, are of of evident metamorphic origin. In areas close to Malaga, in rocks bearing ghosts of andalusite and garnet at the very base of typical Maláguide successions the IC values <0.25 clearly indicate epizonal metamorphism (Ruiz-Cruz and Rodríguez-Jiménez 2002). Mineral associations in the same rocks indicate $T \approx 500$ ºC and $P \approx 5$ kbar for the pre-Alpine metamorphism (Ruiz Cruz and Nieto 2002), compatible with those deduced for the underlying Benamocarra rocks (Sánchez-Navas et al. 2016).

The Variscan tectonics in the Maláguide Complex is evidenced by major scale structures (folds, thrusts and faults) that are not yet well featured neither cartographically nor regionally, and by several generations of meso- and microstructures like isoclinal and crenulation folds and other associated ductile deformation structures (axial plane foliations, stretching and crenulation lineations) and by fragile structures that are absent in the Saladilla Formation (Martín-Algarra et al. 2009b). In addition, as stated above, the

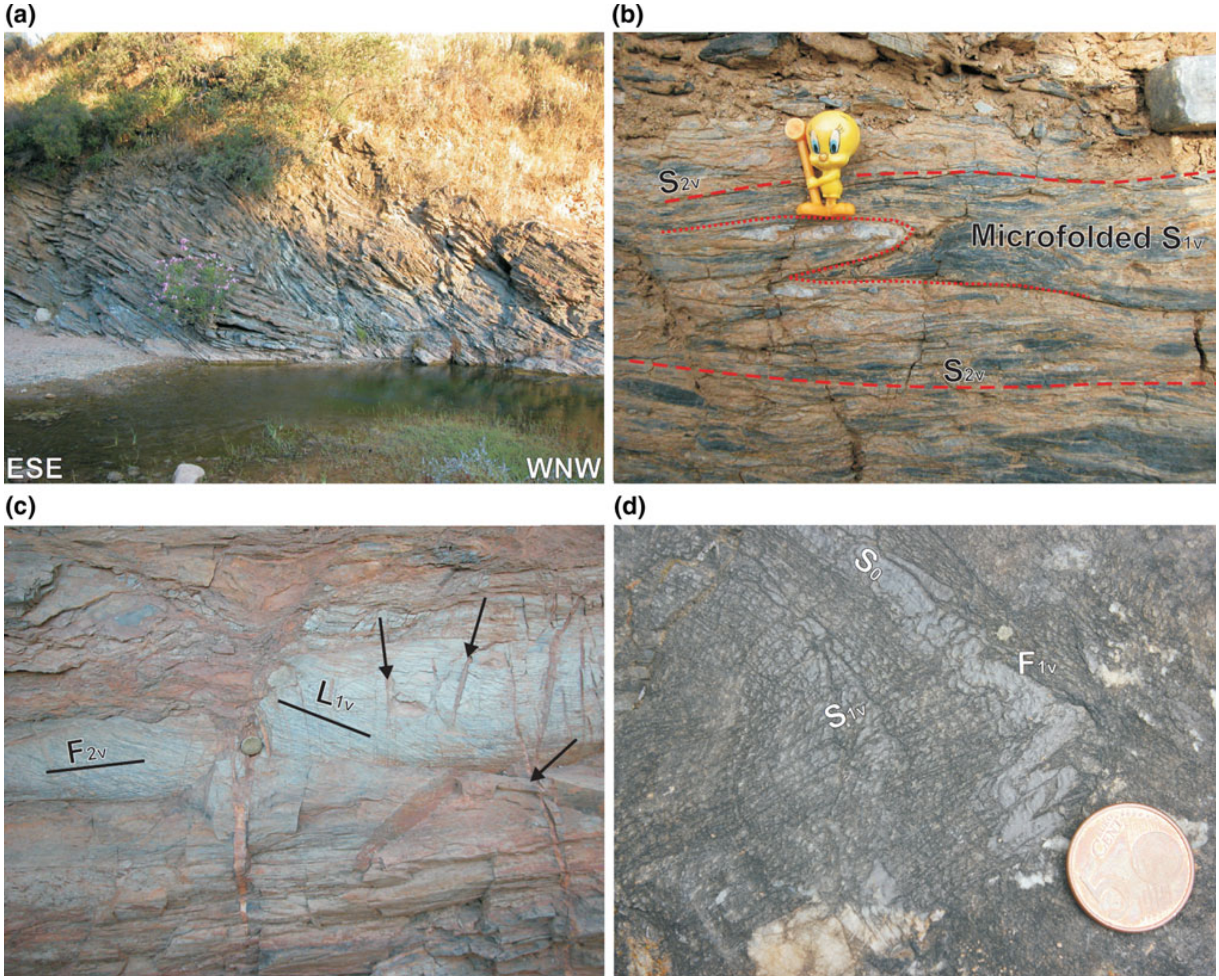

Fig. 9.22 Variscan structures of the Maláguide Complex. **a** Markedly assymmetric (E-vergent) fold by heterogeneous simple shear oblique to S_0 (top-to-the-SE reverse motion) affecting thin-bedded mixed carbonatic siliciclastic turbidites of the Santi Petri Fm. **b** Two variscan foliations within the Morales Fm metapelites: the slaty cleavage S_{2v} is roughly axial planar with respect to hosting folds of the S_{1v}. **c** Lineation (L_{1v}) affected by the second phase of folding F_{2v}; the arrows point to extensional quatz-filled joints related to F_{2v} folds. **d** Microfolds (F_{1v}) folding the bedding (S_0) and related foliation (S_{1v}) in Visean limestones dated with conodonts at that site (Falcoña Fm)

Maláguide Paleozoic succession shows two parts with different style and intensity in deformation between Visean and older pelagites and turbidites and post-Visean clastics. Pre-Culm deposits (pre-Almogía Fm) form deep marine successions deposited in proximal to distal areas of a divergent Paleotethyan margin that were affected by the first important Variscan deformation. The synorogenic siliciclastic sediments with Culm facies (Almogía and Marbella Fms) record marked synsedimentary instability that affected the Maláguide margin since mid-Carboniferous time, and were affected by the main latest Variscan (Hercynian) deformation.

In the Morales Fm and in the Santi Petri Fm at least two Variscan foliations are present. The Variscan deformation is characterized by a markedly asymmetric, E-vergent folding (Fig. 9.22a) related to heterogeneous simple shear oblique to S_0, with top-to-the-SE reverse motion (Martín-Algarra et al. 2009b). A slaty cleavage develops in the pelitic beds, which is roughly axial planar with respect to hosting F_{1V} folds, as well as a lineation (L_{1V}). These structures (Fig. 9.22a, b) are affected by a second phase of folding (F_{2V}), which refolded previous (F_{1V}, S_{1V}, L_{1V}) structures (Fig. 9.22c) and produced a new cleavage (S_{2V}), which locally becomes the main foliation, and a mostly W plunging lineation (L_{2V}). This occurred within the framework of the same progressive deformation, as suggested by a general coaxiality of fold structures. The same type of folds, also E-verging, are well documented in the Lower Carboniferous radiolarites (Falcoña Fm) and in the related Visean limestones (Fig. 9.22d). Nevertheless, the style and intensity of the deformation affecting the Culm-like deposits is quite different from what it is usually observed in the pre-Culm sediments. These deposits are usually much less deformed than the pre-Culm deposits and their more pelitic lithotypes document only one clear foliation, corresponding to the S_{2V} of the underlying beds. In addition, they are affected by numerous faults that document a roughly E-W extension (Fig. 9.21b).

Finally, two stratigraphically different types of pre-Carboniferous successions have been recently recognized in the Maláguide Complex, mainly in the outcrops along the IEZB (Martín-Algarra et al. 2009a; Navas-Parejo 2012; Rodríguez-Cañero and Martín-Algarra 2014; Navas-Parejo et al. 2015b), and they usually make part of independent Alpine tectonic units that overprint a previous Variscan thrust nappe structure. In the Ardales area, the pre-Alpine structure has been demonstrated by cartographic, stratigraphic and structural criteria as formed by two Variscan thrust units (Fig. 9.17), where the proximal facies constitute the hanging wall unit and the distal facies the footwall units (Martín-Algarra et al. 2009a). Structural analysis indicates that the contact between these Variscan units is an E-directed thrust and that the internal structure of the lower Variscan unit is consistent with bulk noncoaxial strain dominated by top-to-the-E shearing (Martín-Algarra et al. 2009b).

Additional evidence on the Variscan evolution of the Maláguide Complex and neighbour areas is indirectly obtained from granitic and gneissic clasts that are commonly present in much younger deposits, in particular in those of late Oligocene to Aquitanian age included in the Ciudad Granada Group. These clasts are identical to several types of late Variscan granites and related moderately alpinised metamorphic lithotypes that are commonly present within the Calabria-Peloritani pre-Mesozoic basements and that are currently dated as latest Carboniferous to earliest Permian (Martín-Algarra et al. 2000, and references therein).

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