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The Palaeozoic evolution of the Maures massif (France) and its potential correlation with others areas of the Variscan belt: a review

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alaeozoic history for the Maures-Tanneron massif is proposed ology, tectonics, metamorphism and radiometric ages, and by making a comparison with its surrounding Variscan outcrops, mainly the Massif Central, the Pyrénées, Bohemian, the Alpine crystalline massifs, Sardinia, Corsica, and Italian massifs (Toscany, Catabria, and Silicia).

The Maures-Tanneron massif is formed of three zones. The western Maures includes Cap Sicié, Fenouillet, Maurette, and Loli units that correspond to Ordovician-Early Carboniferous low-metamorphosed metasediments devoid of HP rocks. These units belong to the southern external zone of the Variscan belt including Pyrénées, Mouthoumet, southern Montagne Noire, Catalanides, south Sardinia, and Tuscany. They were formed at the north-Gondwana margin. The central Maures includes Collobrières, Bormes, Cap Nègre Unit, and Cavalaire units that correspond to portions of a Precambrian to Early Ordovician supra-subduction zone lithosphere involved in continental subduction of probable Ordovician-Silurian age. These units belong to an intermediate zone of the Variscan belt including South Brittany, the Massif Central, central Sardinia, and Belledonne Alpine External massif. The eastern Maures-Tanneron, that includes Cavalières and Petites Maures units, corresponds to portions of an Ordovician back-arc lithosphere involved in the Silurian continental subduction. These units belong to an internal zone of the Variscan belt including the Massif Central, northeast Bohemia (?),Alpine External massifs, northeast Sardinia, Corsica, Tuscany, Sicilia, and Calabria.

These zones have possibly suffered early specific tectonics, but a common Carboniferous polyphased tectonics is inferred from superimposed metamorphic fabrics, composite foliations and various lineations. During the Tournaisian (350-340 Ma), WNW-verging thrusting (S1-L1) is responsible for stacking of units and a first regional IP/HP metamorphism. This event is well preserved in the western Maures. During the Middle-Upper Visean (340-330 Ma), SE-verging back-thrusts (S2-L2) reworked the nappes pile at the climax of the regional metamorphism. This event is well preserved in the central Maures. SE-verging thrusting evolved into, or combined with, a main orogen-parallel transcurrent tectonics (S3-L3) that produces upright folding and left-lateral shearing of the nappes pile (S0-1-2 foliation). During the Namurian (325-315 Ma), synorogenic extension thins the nappes pile by combining orogen-parallel, top-to-the NNW flat-lying shearing and dextral shearing (S4-L4), leading to exhumation of the central Maures. During the Upper Westphalian-Lower Stephanian (310-300 Ma), postorogenic extension produces a thermal overprint in the eastern Maures associated with orogen-parallel sinistral transcurrent tectonics, development of coal basins, granite emplacement, and rapid exhumation of eastern Maures. The orogenic evolution is achieved with deposition of Permian volcano-sedimentary deposits. All along the orogenic process, the Grimaud fault, that separates central and eastern Maures, plays a major role in the partitioning of deformation and seems to be the root of WNW-verging nappes.

The Paleozoic history of the Maures massif is interpreted as the formation of continental/oceanic rifts at the north-Gondwana margin during the Cambrian-Ordovician, their subduction at mantle depths during the Ordovician-Silurian, and their exhumation coeval with their thrusting during the Carboniferous collision between Gondwana and Baltica-Laurentia. Similarities of lithology and differences in tectonics between Maures and its surrounding Variscan areas may reflect the irregular shape of the north-Gondwana margin involved all along the orogenic process.

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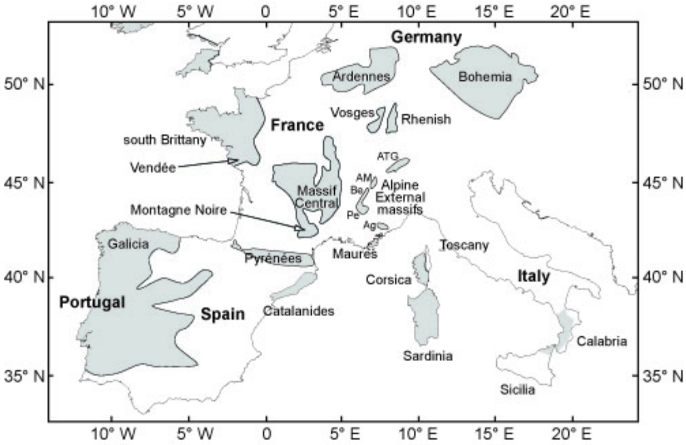
Introduction

The Variscan belt of Western Europe was a Palaeozoic (480-290 Ma) mountain belt formed in response to the closure of two oceans by subduction and collision of two majors continents, Laurentia-Baltica to the North and Gondwana to the South, and numerous microplates including Armorica and Avalonia (e.g., Ziegler, 1989; Franke, 2000; Matte, 2001).

A very simplified history of the southern side of this belt, on the example of the Massif Central, can be summarizing as follow. During the Cambrian-Ordovician, rifting at the north-Gondwana margin formed oceans (e.g., Ménot et al., 1988) that separate microblocks. During the Late Silurian-Devonian, oceanic/continental subduction leads to HP/UHP metamorphism (e.g., Matte, 1988; Lardeaux et al., 2001). During the Middle Devonian, limited extensional environments developed in a back-arc position while convergence continued (Faure et al., 1997; Pin and Paquette, 1997). During the Upper Devonian-Lower Carboniferous, continent-continent collision leads to crustal thickening and inverted IP metamorphism (Briand, 1978; Burg et al., 1984, 1989) by combined thrust and wrench tectonics (e.g., Burg and Matte, 1978; Brun and Burg, 1982; Ledru et al., 1994; Matte, 2001). During the Middle Carboniferous, orogen-parallel extension thinned internal zones while compression evidenced by southward thrusts was active in southern external zones (Burg et al., 1994; Faure, 1995; Roig et al., 2002). From Upper Carboniferous to Permian, postorogenic, also orogen-perpendicular, extension is widespread through out the belt (Burg et al., 1994; Faure, 1995; Faure et al., 2002; Roig and Faure, 2002). Postorogenic extension resulted in development of a lower layered crust (Rey, 1992), intense magmatism, partial melting, and crustal-scale normal faulting, leading to crustal re-equilibration (Echtler and Malavieille, 1990; Malavieille, 1993; Brun and Van Den Driessche, 1994; Lagarde et al., 1994; Ledru et al., 2001).

This geological history has not been yet recognized in the southern part of the Variscan belt which outcrops as isolated and small massifs, partly rotated and displaced during the Alpine orogeny (Figure 1). Despite post-Palaeozoic events, these massifs, that include Bohemia, Alpine crystalline massifs, Maures, Sardinia, Corsica, Tuscany, Catabria and Sicilia, seems to have experienced a common history due to their probable position at the north-Gondwana margin. These massifs may form a segment extending for ~1000 km (Matte, 2001) those the probable N-S trend may contrast with the general E-W trend of the Variscan belt. These geometrical features suggest contrasting tectonics history due to an eastern virgation of the belt similarly to the Ibero-Armorican virgation. However, correlations of lithology and tectonics between them are not well established. In this paper, we reassess the geology of the Maures massif and put forward potential correlations of lithology and tectonics with others Variscan areas, providing a new regard on the geodynamics of the southern Variscan belt.

Figure 1. Main outcrops of the Variscan belt of Western Europe

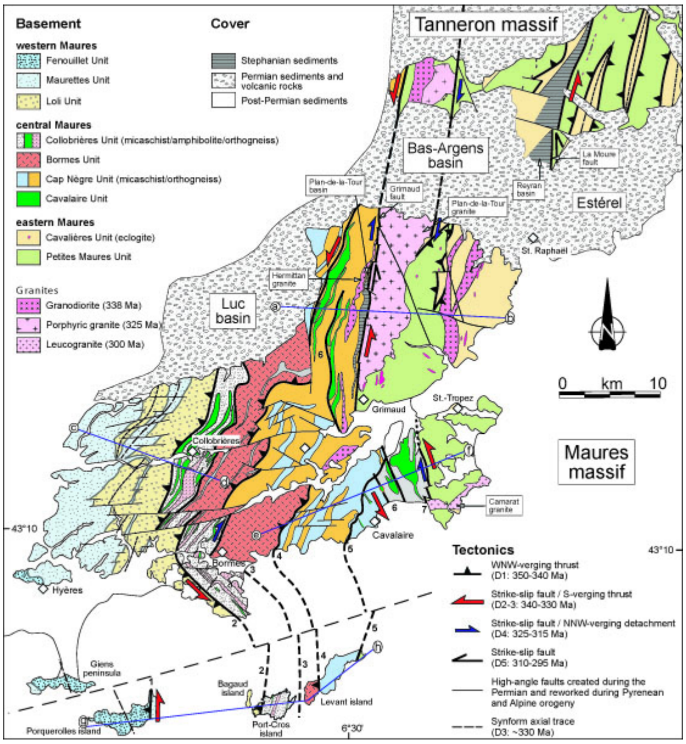


Abbreviations for Alpine External massifs: ATG = Aar-Tavetsch-Gotthard massifs, AM = Aiguilles Rouges / Mont Blanc, Be = Belledonne massif, Pe = Pelvoux massif, Ag = Argentera massif.

Lithology: subdivision of units

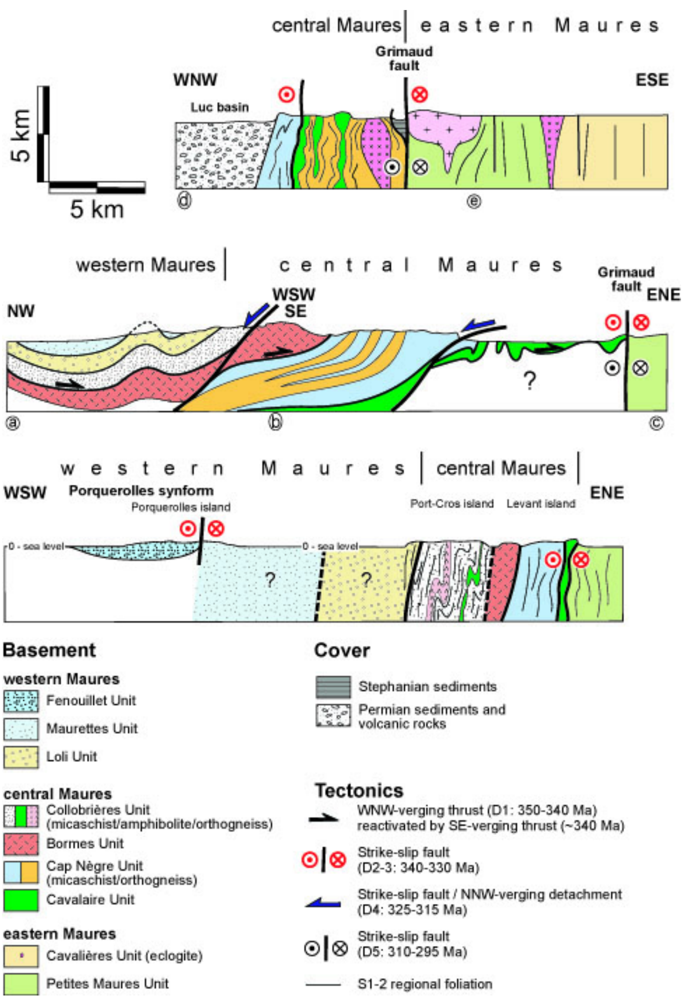
The Maures massif consists of Pre-Permian low- to high-grade metamorphic rocks of uncertain age intruded by granites and overlie by coal basins. A new subdivision is proposed here based on (1) lithology, (2) radiochronometric or stratigraphic age, and (3) large-scale geometric relationships between units and ductile shear zones. We have paid particular attention to the presence of high-pressure (HP) mineral assemblages and mafic-ultramafic rocks which may be markers of major thrusts. For a long time, three mafic-ultramafic units of the Maures massif were considered to be equivalent and therefore used as a marker-horizon (Bordet, 1957, 1969; Bordet and Gueirard, 1967; Caruba and Turco, 1976; Bard and Caruba, 1981; Seyler and Crévola, 1982; Seyler, 1983; Seyler, 1986; Crévola and Pupin, 1994; Morillon, 1997; Briand et al., 2002). The resulting interpretation has emphasized orogen-parallel, regional-scale isoclinal folds, following the hypothesis of Demay (1926a, 1926b). Unfortunately, this interpretation is in opposition to (1) geochemical analyses of the mafic rocks that emphasize their differences in nature and origin (Ricci and Sabatini, 1978; Avice, 1995), (2) the asymmetrical structure of the Maures-Tanneron massif inferred by geological mapping (Gueirard, 1960; Buscail, 2000) and structural analyses (Bronner, 1986; Vauchez, 1987; Morillon, 1997). Several units are distinguished here and are presented from west to east (Figures 2 and 3). They mainly correspond to the zones defined by Gueirard (1957) and detailed by Orsini (1968), Bordet (1969), Seyler (1975), Bordet (1976), Le Marrec (1976), Maquil (1976), Crévola (1977), Conti (1978), Seyler and Boucarut (1979), Olives Banos (1979), Seyler and Crévola (1982), Golberg (1983), Caruba (1983), Seyler (1986), Vauchez (1987) and Buscail (2000).

Figure 2. Simplified geological map of the Maures massif



Simplified geological map of the Maures massif (after Gueirard, 1960; Orsini, 1969; Crévola, 1977; Seyler, 1986; Bronner, 1998; Buscail, 2000) with trace of cross-sections of Figure 3 .

Figure 3. Simplified cross-sections



Simplified cross-sections through northern, central, and southern parts of the Maures massif. See Figure 2 for location.

The Cap Sicié Unit

The Cap Sicié Unit corresponds to very low-grade metamorphic rocks outcropping along the coast southwest of the Maures massif. This peculiar unit has northeastward thrusted the Permian-Trias during Pyrenean-Provencal events (Mattauer and Proust, 1963). By its anchizonal metamorphism of probable Paleozoic age, this unit may view as the uppermost unit of the Maures massif. From bottom to top, the Cap Sicié Unit is made of quartzite (intruded by dolerite sills) and conglomerates, then volcano-sedimentary deposits made of green schist and black schist that include black phtanite and radiolarian, then quartzite and black schist (Gouvernet, 1963).

The Fenouillet Unit

The Fenouillet Unit outcrops in the core of the Porquerolles synform (Bronner et al., 1971; Bellot, 2004) and in the Presqu'île de Giens (Gueirard, 1960; Bordet, 1976). It consists of dominant schist (metapelites) and quartzite (microconglomerate and sandstone) with frequent normal or reverse graded-beddings (Olives Banos, 1979). Metamorphism is anchizonal and no HP rock was found.

The Maurettes Unit

The Maurette Unit, 10 to 12 km wide, consists to the East, of dominant limonite-bearing schist (Fe-rich metapelites) with minor quartzite, calc-schist, crinoids-bearing limestone, arkoses (Tempier, 1978), and levels of metadolerite (Gueirard, 1960). The westernmost Maurette Unit consists of massive quartzite associated with graptolites, graphite-bearing schist, crinoids-bearing limestone and sandstone, indicating a Middle Silurian (Upper Llandoverian to Lower Tarannonian) age for its unit (Schoeller, 1938; Gueirard et al., 1970). Stable biotite and chloritoid indicate epizonal metamorphism (Buscail, 2000) and no HP rock was found (Olives Banos, 1979).

The Loli Unit

The Loli Unit consists of a monotonous flysch-like formation made of quartzite and schist, fining and thinning eastward (Gueirard, 1960). However, its base, occurring in the Bagaud island and in the westernmost Port-Cros island, shows an uncommon formation called Bagaud-Malalongue (Bronner, 1986). In this formation, short, low-dipping normal limb of P2 folds preserved sedimentary features as cross-beddings, drop stones, mud balls, slumps and graded-beddings, indicating that the whole unit is overturned due to P2 folding. Tempestite and turbidite occurrences suggest a distal platform palaeoenvironment. Biotite-rich quartzites include very abundant carbonated nodules with marine acritarchs and rare phosphorous nodules with radiolarians (Bronner and Bellot, 2000). Biotite and garnet indicate mesozonal metamorphism (Buscail, 2000) and no HP rock was found.

The Collobrières Unit

The Collobrières Unit consists of mica-schist and quartzite that includes a variety of Fe-rich rocks. At their bottom, several meters of an iron ore (the so-called Collobriérite) consisting of abundant magnetite, almandine garnet, fayalite, and Fe-amphibole, is regarded as a metamorphosed sedimentary iron ore (de Groulard, 1982). Upward, mica-schist becomes Fe-rich, hyper-aluminous, and concentrated abundant graphite, white micas, garnet and staurolite. They include levels of quartzites, marbles, gneiss with silicate calcite, amphibolites (alkaline basalt), and blue quartz-bearing orthogneiss (alkaline trachyte), and are regarded as volcano-sedimentary products emplaced during early stages of continental rifting (Seyler, 1986). Zircon U-Pb dating on alkaline meta-rhyolite and alkaline trachyte yielded 561-495 Ma and 498 12 Ma, respectively (Lancelot et al., unpub), indicating a Cambrian to Late Cambrian age for rifting assumed to be formed during plume activity (Briand et al., 2002). Garnet, staurolite, and biotite indicate mesozonal metamorphism (Buscail, 2000) and no HP rock was found.

The Bormes Unit

The Bormes Unit consists of an orthogneiss that includes mica-schist and relics of a aluminous porphyritic granite (Barral granite) metamorphosed under LP granulites facies conditions (6.7 kbar/>850°C; Gueirard, 1976). Zircon Pb-Pb (~605 Ma; Chessex et al., 1967), whole rock Rb-Sr (~560 Ma; Maluski, 1971), and biotite 40Ar-39Ar (575 ± 8 Ma; Maluski and Gueirard, 1978) dating converge into a Precambrian (550-600 Ma) age for the emplacement of the Barral granite, interpreted as an undeformed equivalent of the Bormes orthogneiss (Gueirard, 1982). Monazite U-Pb on the orthogneiss yielded 345 ± 3 Ma, interpreted as the age of regional IP metamorphism (Moussavou, 1998). Biotite 40Ar-39Ar dating on the western boundary of sheared orthogneiss yielded 323-328 Ma, interpreted as the age of top-to-the NW shear deformation and retrograde LP metamorphism (Gaubert, 1994). Metapelites included in the Bormes orthogneiss preserve pre-D1 white-schist assemblages (12-16 kbar/480-550°C) that retrograded to 4-6 kbar/600-650°C during shear deformation of the Bormes orthogneiss (Leyreloup et al., 1996).

The Cap Nègre Unit

The Cap Nègre Unit consists of paragneiss, aluminous kyanite-garnet-staurolite micaschist, orthogneiss, and quartzite (Gueirard, 1960).

The Cavalaire Unit

The Cavalaire Unit consists of a tectonic melange composed of a wide range of rock types, including acid, mafic, and ultramafic igneous rocks hosted by migmatitic paragneiss and migmatitic orthogneiss. Paragneiss (meta-siltstone) includes micaschist levels and quartzite with calcic silicates. Orthogneiss are metamorphosed and deformed aluminous cordierite-garnet bearing granite, gabbros to diorite and syenite with monzonitic composition, cordierite-bearing aplite, and pegmatite (Seyler, 1975). They are evidence for continental magmatism due to crustal extension (Seyler, 1986).

A layered formation of amphibolite (meta-tholeiites with oceanic affinities; Seyler, 1986), pink orthogneiss (metamorphosed alkaline lavas; Seyler, 1986), and amphibole-biotite orthogneiss is interpreted as evidence for bimodal magmatism related to an extensional setting (Seyler, 1986) due to plume activity (Briand et al., 2002). The layered formation includes small lenses of meta-igneous rocks as parts of a supra-subduction zone lithosphere (Bellot et al., 2000b). A first group of abundant spinel peridotites (Gueirard, 1956; Bard and Caruba, 1981; Lasnier, 1977; Laverne et al., 1997), garnet-spinel peridotites (Bellot, 1998; Bouloton et al., 1998), coronitic gabbros (Lasnier, 1970), garnet amphibolites (meta-andesites), felsic amphibolites (meta-dolerites), and fine-grained amphibolites (metamorphosed transitional to tholeiitic lavas with arc affinities; Ricci and Sabatini, 1978; Seyler and Boucarut, 1979; Bard and Caruba, 1981) was interpreted as portions of lithosphere generated in a supra-subduction zone during Early Palaeozoic time. A second group of gabbros (Seyler, 1982; Caruba, 1983; Avice, 1995; Bouloton et al., 1998) and fine-grained amphibolites (meta-tholeiites with oceanic affinities; Seyler, 1986; Briand et al., 2002) was interpreted as portions of an oceanic lithosphere generated at the Cambrian-Ordovician boundary. A consensual hypothesis is to interpret all these rocks as parts of a back-arc lithosphere. This hypothesis is supported by the presence of minor limestones (Seyler, 1986). Only garnet-spinel peridotites (Bellot 1998; Bouloton et al., 1998) are evidenced for HP metamorphism (P>2.8 GPa/T>850°C). Coronitic gabbros (0.6-0.7 GPa/750-850°C; Caruba, 1983) and garnet amphibolites (0.5 GPa/550°C; Bellot et al., 2003), previously interpreted as HP rocks (Bard and Caruba 1981, 1982), most likely reflect LP granulites and amphibolites facies metamorphisms, respectively (Seyler, 1982).

Zircon U-Pb dating on alkaline orthogneiss defines a Cambrian age for bimodal magmatism (498 17 Ma and 507 ± 5 Ma; Lancelot et al., unpub; 548 +15/-7 Ma; Innocent et al., 2003). Whole rock Rb-Sr dating on amphibolite gives 348 ± 7 Ma, interpreted as the age of regional metamorphism (Innocent et al., 2003), while 40Ar-39Ar dating on amphibolites (330 ± 2 Ma and 328 ± 3 Ma), on garnet micaschist (322,9 ± 1,7 Ma), on biotite micaschist (321,1 ± 1,3 Ma and 319,5 ± 0,3 Ma), on migmatite (317,2 ± 1,0 Ma), and sheared migmatite (319,7 ± 1,3 Ma and 320,5 ± 1,4 Ma) suggest cooling of the central Maures in relation with its exhumation during the Namurian (Morillon et al., 2000).

This unit is interpreted here as relics of a Cambrian (550-500 Ma) back-arc lithosphere involved in the Silurian continental subduction and the Carboniferous continental collision.

The Cavalières Unit

The Cavalières Unit consists of migmatitic orthogneiss, migmatitic paragneiss and minor mica-schist that include lenses of eclogites (Le Marrec, 1976; Maquil, 1976; Crévola, 1977; Vauchez, 1987). Kyanite-sapphirine-bearing eclogites are inferred to be calk-alkaline meta-gabbro (Avice, 1995) emplaced in a back-arc setting (Buscail et al., 1999) during the Upper Ordovician (452 ± 8 Ma, zircon U-Pb; Lancelot et al., 1998), and buried at mantle depths (15-25 kbar/800-950°C; Bard and Caruba, 1982; Caruba, 1983) during the Lower Silurian (431 ± 4 Ma, zircon U-Pb; Lancelot et al., 1998). The surrounding biotite- and amphibole-rich orthogneiss are interpreted as calk-alkaline diorite and granodiorite emplaced during the Precambrian (zircon U-Pb, 612-630 Ma; Lancelot et al., 1998) and intruded by mafic magma (forthcoming eclogites). Even though HP metamorphism has not been discovered in orthogneiss, these rocks have probably been buried with metagabbros during the Silurian. They have experienced partial melting during the Upper Visean (zircon U-Pb, 334 ± 3 Ma; Lancelot et al., 1998) in relation to their exhumation (Le Marrec, 1976; Vauchez and Buffalo, 1988).

The Petites Maures Unit

The Petites Maures Unit consists, from bottom to top, of metagabbros, felsic granulites, various migmatitic paragneiss, monotonous, layered or nodules-rich, that includes abundant calcic silicate-bearing gneiss, minor migmatitic orthogneiss, lenses of marbles, eclogites and serpentinites

(Le Marrec, 1976; Maquil, 1976; Crévola, 1977; Vauchez, 1987).

Coal basins

Two coal basins of Late Palaeozoic age are located along major orogen-parallel faults. The Plan-de-la-Tour basin, 16-km-long and 1-km-wide, occurs in the central Maures along the Grimaud fault (Wallerant, 1889; Demay, 1927a and b; Gueirard, 1960; Bordet, 1967). The basin is filled by 400 m of conglomerates and arkoses (Wallerant, 1889; Bordet, 1967; Basso, 1985; Bégassat, 1985) those the base is dated Lower Stephanian (Masurel, 1964; Basso, 1985). Pebbles of mylonitic rocks from the Grimaud fault are frequent, while those of the Plan-de-la-Tour granite are lacking (Demay, 1927). Microgranite dykes emplaced within the basin are dated Upper Stephanian (290 ± 10 Ma, Rb-Sr on minerals: Roubault et al., 1970b; 295,4 ± 2,4 Ma, biotite 40Ar-39Ar; Morillon, 1997). These dates, in addition to ASM analyses (Edel, 1999), suggest that magmatism and conglomerates deposition were coeval during the Stephanian.

The Reyran basin, 12-km-long and 1-km-wide, occurs in the eastern Maures along the La Moure fault (Basso, 1985 and references therein). The basin is filled by 1000 m of sandstone, coal, pelites, and conglomerates dated Upper Westphalian to Lower Stephanian (Basso, 1985). Pyroclastic rhyolite took place during sedimentation.

Permian basins

End of the Variscan orogeny in the Maures massif is evidenced from deposition of Lower to Upper Permian sediments that unconformably overlain both basement and Upper Carboniferous coal basins in the western and northern Maures massif (Toutin-Morin et al., 1988). Basin development is associated with acid then mafic volcanisms in the Estérel area (Bordet, 1951; synthesis in Crévola and Pupin, 1994), and hydrothermal activity that generates some of the F-Ba-Pb-Zn deposits (Solety, 1964; Vervialle, 1975; Mari, 1979). 40Ar-39Ar dating on plagioclase of a mafic dyke and on adularia of a barite-fluorite lode have yielded 278 ± 0.4 Ma and 264 ± 0.7 Ma, respectively (Zheng et al., 1991-1992). In the northern massif, Thuriangian basin allows final exhumation of the Plan-de-la-Tour granite (Delfaud et al., 1989). Permian and post-Permian events have produced polyphased movements along E-W faults (Zheng, 1990; Toutin-Morin et al., 1992) between which some basement blocks are rotated, especially the Hyères-Bormes-Cavalaire block (Bronner, 1996).

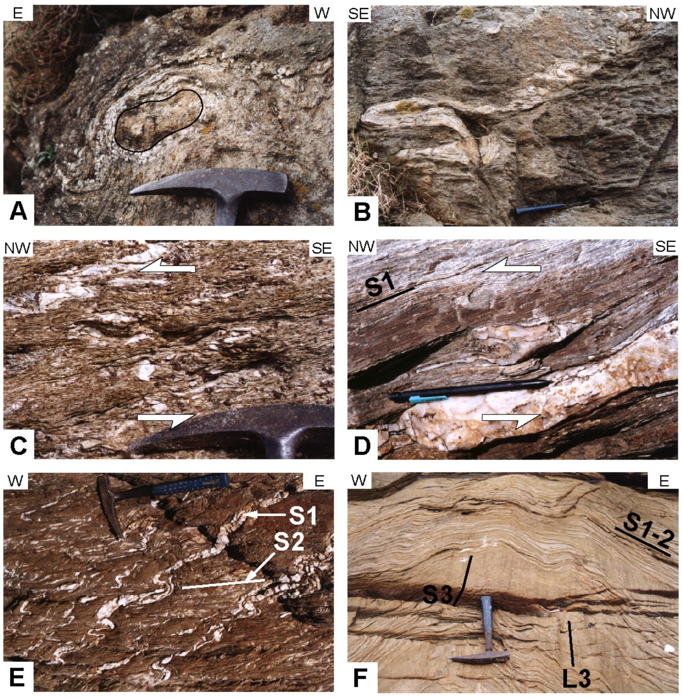
Structure, metamorphism, and plutonism

Lithological map supports a westward-dipping monoclinal structure for western and central Maures (Figure 3). However, this relatively simple structure reflects a complex tectonics history that involves several tectonics phases (Arthaud and Matte, 1966; Maluski, 1968; Bronner et al., 1971; Chabrier and Mascle, 1975; Seyler, 1975; Le Marrec, 1976; Maquil et al., 1976; Conti, 1978; Olives-Banos, 1979; Bard and Caruba, 1981; Seyler and Crévola, 1982; Caruba, 1983; Goldberg, 1983; Vauchez and Bufalo, 1985; Seyler, 1986; Vauchez and Bufalo, 1988; Morillon, 1997). Petro-structural investigations have inferred the relationships between three of them and a polyphased regional metamorphism (Maluski, 1968; Conti, 1978; Seyler, 1975; Caruba, 1983; Goldberg, 1983), and demonstrate that the D2 tectonic phase is coeval with regional IP metamorphism (Buscail, 2000 and references therein). Many authors agree that the Variscan Maures massif has experienced late-orogenic extension marked by brittle-ductile normal faults (Gaubert, 1994; Ciancalleoni, 1995; Buscail, 2000; Morillon et al., 2000; Bellot et al., 2002), but debate continues about the mode of extension, and the relative importance of one or several pre-extension tectonics that structured this Variscan outcrop (Dumoulin-Thiault et al., 1996). Petro-structural investigations inferred the close relationships between three metamorphism and ductile tectonics (Seyler, 1975; Conti, 1978; Golberg, 1983; Séno, 1986; Gaubert, 1994; Ciancaleoni, 1995; Buscail, 2000). Based on previous and new data, we attempt to propose a sequence of Paleozoic tectonic, metamorphism, and plutonic events of the Maures massif. Western, central, and eastern Maures have possibly suffered early specific tectonics, but a common Carboniferous polyphased tectonics is inferred from superimposed metamorphic fabrics, composite foliations and various lineations.

D1 deformational phase

The D1 deformational phase is responsible for the S1-L1 rock fabric, sheath F1 folds (Figure 4 A), isoclinal F1 folds (Figure 4 B), and and top-to-the WNW shearing (Figures 4C and 4D). It affects western and central Maures. The main shear zone occurs within the Collobrières Unit, the uppermost Bormes orthogneiss dated 345 ± 3 Ma, the lowermost Loli Unit interpreted as a Tournaisian syntectonic deposit, and the Cavalaire Unit those sheared amphibolites are dated 348 ± 7 Ma. Because D1 shearing ceased before intrusion of the Hermittan granite (338 ± 6 Ma), it can therefore be assigned to the Tournaisian (~350-340 Ma). D1 shearing was associated with IP/HP-LT prograde metamorphism (Pmax) evidenced from garnet inclusions (Buscail, 2000). D1 structures and mineral assemblages are well-preserved in the western Maures where S0-1 composite foliation and L0-1 intersection lineation are commonly observed in flysch-like meta-sediments (Figure photo). In the northern Cavalaire Unit, lenses of metagabbros, orthogneiss, and metaperidotites forming a tectonic melange indicate a WNW-ESE stretching axis and preserved top-to-the WNW shearing in amphibolites facies conditions. In the northeastern Cavalaire Unit, a WNW-directed, eastward dipping shearing is described at the boundary between Cavalaire and Cap Nègre Units and interpreted as a WNW-directed thrust (Morillon, 1997). WNW-directed shearing is likely to be WNW-verging thrusting responsible for stacking of units leading to crustal thickening during the Early Carboniferous.

Figure 4. Superimposed fabric and kinematic criteria



Superimposed fabric and kinematic criteria for NW-verging thrust tectonics in the western-central Maures. A. N110°E-trending sheath F1 fold (Cavalaire Unit). B. Meta-aplite foliated and WNW-trending folded during D1 tectonics (Cavalaire Unit). C. Development of F1 foliation and shear bands associated with syntectonic growth of garnet, staurolite, biotite and kyanite, indicating a top-to-the WNW sense of shear (Cap Nègre Unit). D. Sigmoid-shaped quartz developed from isoclinal folds, indicating northwestward shearing (Fenouillet Unit). E. Low-angle, superimposed S1 and S2 foliations in the Porquerolles island (Fenouillet Unit). F. Composite S1-2 foliation reworked by an upright N-S-trending F3 fold (Maurette Unit) that the axis is parallel to the stretching L3 lineation.

D2 deformational phase

The D2 deformational phase is responsible for the regional S2-L2 rock fabric (Figure 4 E), F2 isoclinal and sheath folds and top-to-the SE shearing. It affected the central Maures and is particularly intense at top and bottom of the Cap Nègre Unit (Bellot and Bronner, 2000; Buscail, 2000). Observed asymmetric microstructures are: sigma-type garnet (Buscail, 2000) and staurolite with asymmetric strain shadows of biotite (Gaubert, 1994), preferred orientation of quartz sub-grains in ribbons (Séno, 1986), red biotite strain-shadows around plagioclase porphyroblasts (Bellot and Bronner, 2000), -type plagioclase with inclusions of muscovite-biotite-garnet-sillimanite (Bellot et al., 2002a). Following these observations, D2 shearing is assumed to be associated with IP-HT metamorphism (Seyler, 1975; Conti, 1978; Golberg, 1983; Gaubert, 1994; Ciancaleoni, 1995; Buscail, 2000) that decreases from 0.8-1.0 GPa/500-600°C to 4-5 kb/600-700°C in the Central Maures (Bellot et al., 2002a). D2 metamorphism increases eastward, i.e. downward the nappes pile, from anchizonal to amphibolites facies conditions.

D3 deformational phase

The D3 deformational phase is responsible for orogen-parallel sinistral shearing of the S1-2-L2 rock fabric. It affected the whole Maures massif and increases eastward. It was associated with IP-HT retrograde metamorphism (Tmax), and is likely to have accommodated exhumation of the nappes pile (Bellot et al., 2000a). In the eastern Maures, orogen-parallel sinistral shearing associated with sheath folds deformed the nappes pile (Vauchez and Buffalo, 1988) during its partial melting (Le Marrec, 1976) at c.a. 334 ± 5 Ma (Morillon, 1997). In the central Maures, a similar strain pattern has been found in the Cavalaire Unit in which the asymmetric shape of peridotites lenses and the en echelon pattern of amphibolites lenses both indicate sinistral shearing at a regional-scale (Bellot et al., 2000a;

2002b). In the western Maures, a high-angle sinistral shearing is superimposed on the boundary between Fenouillet and Maurettes Units (Bellot, 2004). These shear zones are likely to reflect crustal-scale sinistral shearing centered on the Grimaud fault that produces asymmetric folding on both sides (Vauchez and Buffalo, 1988). Sinistral shearing combined with regional-scale, orogen-parallel upright folding (Figure 4 F), as the Porquerolles synform (Bellot, 2004) and synforms and antiforms of the eastern Maures (Vauchez and Buffalo, 1988) and the Tanneron (Crévola, 1977).

The D3 orogen-parallel sinistral tectonics was coupled with intrusion of syntectonic granites in the middle crust (3-6 kbar; Amenzou, 1988) close to the Grimaud fault during its sinistral movement (Vauchez and Bufalo, 1988; Bellot et al., 2002b). They are the Hermittan granite in the western Maures (338 ± 6 Ma, zircon U-Pb; Moussavou et al., 1998) and the Réverdi quartz diorite in the eastern Maures (Gueirard, 1964; 333 ± 4 Ma, zircon U-Pb; Moussavou et al., 1998). They are likely sheet-dykes complexes rather than sills emplaced in the core of an orogen-parallel upright antiform of the nappes pile (Morillon, 1997; Buscail, 2000). Shearing of amphibolites along the Grimaud fault (330 ± 2 Ma and 328 ± 3 Ma) likely reflect lastest stages of the D3 tectonics. By their relative age and their kinematics, SE-directed shearing and sinistral shearing are likely to be either two stages of a progressive transpressional tectonics, or combined into a single transpressional tectonics of Upper Visean age (340-330 Ma).

D4 deformational phase

The D4 deformational phase is responsible for top-to-the NNW flat-lying shearing, NNW-post metamorphic folding, brittle-ductile normal faulting and that crosscut the nappes pile in the central Maures, combined to dextral shearing close to and along the dextral Grimaud fault. This deformation is well dated Namurian (325-315 Ma) by biotite 40Ar-39Ar dating on western boundary of the sheared Bormes orthogneiss (~325 Ma; Gaubert, 1994), and by amphibole, biotite, muscovite 40Ar-39Ar dating on mylonites of the Cap Nègre Unit (319-321 Ma; Morillon, 1997). The Plan-de-la-Tour porphyric monzogranite (Serment and Triat, 1967), dated Namurian (324 ± 5 Ma, zircon U-Pb; Moussavou, 1998; 325 ± 10 Ma, whole rock Rb-Sr; Roubault et al., 1970a; 325 ± 10 Ma, whole rock Rb-Sr; Maluski, 1972), was emplaced in the eastern Maures in a relay of two branchs of the Grimaud fault during late stages of its dextral movement (Onezime et al., 1999). According to our field investigations, the root zone seems to be close to the center pluton. Shearing of amphibolites along the dextral Grimaud fault (312 ± 3 Ma and 314 ± 5 Ma) possibly reflects latest stages of the D4 tectonics.

The D4 deformational phase was associated with LP (0.2-0.3 GPa) metamorphism those the T increases eastward in both metapelites (Buscail and Leyreloup, 1999) and metabasites (Bellot et al., 2003). Involved mineral assemblages in metapelites are chlorite-quartz-muscovite in the western Bormes Unit (Gaubert, 1994), andalusite-biotite in the Cap Nègre Unit (Buscail and Leyreloup, 1999), and sillimanite-quartz-K-feldspar in the Cavalaire Unit (Buscail, 2000). Retrogression of previous garnet, staurolite and kyanite occurs too (Bellot et al., 2002a). Sills of syntectonic leucogranite and pull-apart pegmatite in the northern and southern Cavalaire Unit took place along normal/dextral shear zones. Centimeter-thick pseudotachylite layers and/or a cataclastic strain along high-angle normal faults are superimposed on ductile structures (Morillon, 1997; Bellot et al., 2002a). They indicate that extensional deformation has continued to brittle-ductile conditions. These data support rapid cooling and exhumation of the Central Maures during the Middle Carboniferous (Morillon et al., 2000).

The component of wrench tectonics increases eastward and culminates along the dextral Grimaud fault that produces chlorite-muscovite-bearing ultramylonites (Morillon, 1997; Onezime et al., 1999). These relationships suggest that the Grimaud fault plays a role of transfer fault of deformation permitting to the eastern Maures to be less affected by extension. By all its features, the D4 deformational phase typifies synorogenic extension experienced by internal zones of the Variscan belt during the Middle Carboniferous.

D5 deformational phase

The D5 deformational phase is responsible for fracturing along orogen-parallel Grimaud and La Mourre faults, deposition of coal-interbedded conglomerates and arkoses in the Plan-de-la-Tour basin (Lower Stephanian) and Reyran basin (Upper Westphalian-Lower Stephanian) along these faults. It was also coupled, in the eastern Maures, with posttectonic emplacement of granites (~300 Ma) in the upper crust close to the Grimaud fault: cordierite-bearing Moulin Blanc granite (301,6 ± 0,8 Ma on biotite and muscovite), Camarat granite (300,2 ± 0,2 Ma on biotite and

299,4 ± 0,5 Ma on muscovite), and dykes of pegmatite (303,6 ± 1,5 Ma on muscovite).

Amphibole, biotite, and muscovite 40Ar-39Ar dating on a wide range of igneous and metamorphic rocks of the eastern Maures yielded a reduce and very concordant time span (305-300 Ma; Morillon et al., 2000). 40Ar-39Ar ages obtained on metamorphic rocks and syntectonic granites are likely to be cooling ages: amphibolitized eclogite (303 ± 3 Ma), migmatitic gneiss (304 ± 1,5 Ma, 306 ± 2 Ma, 301 ± 0,6 Ma on biotite), Reverdi quartz diorite (300,5 ± 0,6 Ma on amphibole, biotite, muscovite), St. Pons les Mûres granodiorite (302,5 ± 2 Ma on biotite and muscovite), and apatite U-Pb age on migmatitic gneiss (301 ± 2 Ma; Lancelot et al., 1998). Similarly, ages obtained on sheared samples of the Plan-de-la-Tour granite located along the Grimaud fault (301,6 ± 1,6 Ma on biotite; 301,7 ± 1 Ma and 295 ± 0,5 Ma on muscovite; Morillon et al., 2000) are similar to those of undeformed samples of the same granite more eastward (304,4 ± 2,7 Ma on muscovite; Morillon et al., 2000) and are therefore interpreted as cooling ages, rather than "deformation" age. As a summary, 40Ar-39Ar dating reflects a major thermal overprint and/or rapid cooling (~305-300 Ma) of the eastern Maures associated with magmatism (~300 Ma) during the Lower Stephanian (Morillon et al., 2000). The D5 deformational phase can therefore be assigned to the Upper Westphalian-Lower Stephanian (310-300 Ma). This phase typifies postorogenic extension experienced by internal zones of the Variscan belt during the Upper Carboniferous, as well-described in the French Massif Central (Malavieille, 1993; Burg et al., 1994; Faure, 1995; Ledru et al., 2001; Roig et al., 2002; Bellot et al., 2005).

Geometry of the Plan-de-la-Tour basin and its precise tectonic setting in relation with the Grimaud fault is a matter of debate: syncline (Demay, 1927a and b; Bordet, 1967; Edel, 1998), pull-apart developed during a late stage of Grimaud dextral faulting (Onezime et al., 1999), deposition in a basement flexure during Grimaud sinistral faulting and refolding during Grimaud dextral faulting (Toutin-Morin and Bonijoly, 1992), graben lately refolded during Grimaud dextral faulting (Morillon et al., 2000), hemi-graben formed during the Lower Stephanian and refolded during the Upper Stephanian (Basso, 1985), continuous deposition and deformation in a basement flexure in relation to Grimaud sinistral faulting (Castanet, 1998). Our field investigations point out a monoclinal structure formed during folding of basement and fracturing along the Grimaud fault, and support the model of Castanet (1998) in relation to a sinistral movement of the brittle Grimaud fault. The Reyran basin, which also displays a monoclinal structure (Basso, 1985), was interpreted here to be a hemi-graben developed in relation to a sinistral movement of the brittle La Moure fault.

Large-scale correlations

The most likely pre-Permian position of the Sardinia-Corsica block is based on paleomagnetic data (Gattacceca, 2001), in addition to structural and gravity data (Gueguen, 1995). It involves an anticlockwise rotation of ~70° coeval for both Sardinia and Corsica. The resulting Variscan structures trend NNE-SSW and imply continuity of Sardinia-Corsica into the Maures-Tanneron-Estérel massifs. Gravimetric data also suggest continuation of these massifs to the Alpine External crystalline massifs (Masson et al., 2000). For this hypothesis to be confirmed and extended to others isolated massifs, correlations of lithology, structures, and metamorphism between the Maures massif and their surrounding Variscan areas are required. New detailed correlations are proposed here based on the presented lithological and structural data. They also take into account previous correlations (Chabrier and Mascle, 1974, 1975; Westphal et al., 1976; Arthaud and Matte, 1977; Tempier, 1978; Ricci and Sabatini, 1978; Tongiorgi, 1978; Bagnoli et al., 1979; Chabrier and Mascle, 1979; Orsini, 1979; Bourrouihl et al., 1980; Rau and Tongiorni, 1981; Seyler, 1986; Rey et al., 1997; Matte, 2001; Elter et al., 2004). Our correlations attempt to relate the Maures massif to a Sicilia-Calabria-Tuscany-Sardinia-Corsica-Alpine External massifs-Bohemia branch on one hand, and to a Massif Central-Pyrénées-South Brittany-Spain branch on the other hand.

Lithology

The Cap Sicié Unit displays a range of similarities with the Upper Ordovician of the axial zone of Pyrénées (pers. observation), the Ordovician-Lowermost Silurian formations of southern Sardinia (Tempier, oral communication), and the Upper Ordovician-Lower Devonian Réaumur Group of Vendée (Wyns et al., 1989).

Middle Silurian graptolites of black schist located at the top of the Maurette Unit are similar to graptolites of the southern Gondwana margin, including Portugal, Morocco, western Pyrénées, Sardinia (Gueirard et al., 1970 and reference therein). Fenouillet and Maurette Units are similar to the Barousse units of Pyrénées (Waterlot, 1969; Capdepont, 1982), the Silurian of the southern Montagne Noire (e.g., Arthaud, 1970; Vidal, 1952; Engel et al., 1981; Feist et al., 1994), the Ordovician-Silurian of Catalanides (Bourrouihl et al., 1980;), the Goceano s.l. Group of North central Sardinia (Conti et al., 2001), the Upper Ordovician Galleria Group of Corsica (Naud, 1979), the Bocchegiano formation and the Bitu Group of Tuscany (Gianelli et al., 1978; Bagnoli et al., 1979; Seyler, 1986; Elter and Pandeli, 1990; Pandeli et al., 1994). More precisely, Fenouillet and Maurette Units are very similar to the Sarrabus and Gerrei Units (San Vito formation) of Central Sardinia, noted 62-63-64 on the 1/250 000 geological map (Carmignani et al., 2001).

The lower part of the Loli Unit, that includes carbonated nodules associated with phosphorous nodules, is very similar to the Tournaisian of the Castelnau-Durban, Alet, and Cierp formations of Pyrénées (Perret, 1988; Chernoff1 and Orris, 2002), the Mont Peyroux and Serre formations of the southern Montagne Noire (Gèze, 1949; Vidal, 1952; Maurel, 1956; Andrieu and Matte, 1966; Feist and Schoenlamb, 1973; Vachard, 1974; Vignard, 1976; Engel et al., 1981; Feist et al., 1994), the Lowermost Carboniferous of the Mouthoumet massif (Bourrouihl et al., 1980), and the lowermost Carboniferous of Catalanides (Fontbole and Julivert, 1954). It could also correspond to the culm-type flysch of the Riu Gruppa and Castello Madusa Unit of central Sardinia (Conti et al., 2001), noted 61 on the 1/250 000 geological map (Carmignani et al., 2001), and to some parts of the Bretignolles formation of southern Vendée that also includes phosphorous nodules (Combe et al., 1985). These similarities suggest a correlation of the Loli Unit with other synorogenic flysch-like deposits that took place in a southern foreland basin. Difference in nappes verging, WNW in the Maures massif and South in the Pyrénées and the Montagne Noire, may reflect the irregular shape for the northern Gondwana margin involved in the Early Carboniferous collision.

The Collobrières volcano-sedimentary Unit displays a range of similarities with the Lower Ordovician (Arenig: 485-470 Ma) of South Brittany (Chauvel, 1968; de Groulard, 1982), the Middle Ordovician volcano-sedimentary formation of the southern Montagne Noire (Robardet et al., 1994), the volcano-sedimentary Ordovician of the Pyrénées (André, 1985; Nicol, 1997 and references therein), the Arenig of the Barradian terrane of Bohemia (Matte et al., 1990), and the Barbagia Unit of Central Sardinia (Seyler, 1986; Conti et al., 2001), noted 60-59-58 on the 1/250 000 geological map (Carmignani et al., 2001).

The Bormes Unit seems to be similar to the Precambrian basement (550-580 Ma) of the Moravian terrane of southeastern Bohemia (Matte et al., 1990 and references therein), and to Cadomian Canignou-type granite of the Pyrénées (Laumonier et al., 2004).

The Cap Nègre Unit, of mica-schist and orthogneiss of probable Precambrian to Cambrian age, displays a range of similarities with the Allemont series of the Belledonne Alpine External massif (Ménot, 1988), the Medium grade metamorphic complex of Central Sardinia (Conti et al., 2001), noted 56 on the 1/250 000 geological map (Carmignani et al., 2001), and the Micaschist Group of Tuscany (Elter and Pandeli, 1990).

The Cavalaire Unit displays a range of similarities with serpentinites and gabbros of the Medium grade metamorphic complex of Central Sardinia (Ricci and Sabatini, 1978; Seyler, 1986), the Chamrousse-Séchilienne unit of the Aar-Tavetsch-Gotthard (von Raumer et al., 1999) and Belledonne Alpine External massifs (Ménot, 1988).

This Cavalières Unit displays a wide range of similarities with the Lower Gneiss Unit of the Massif Central (Burg and Matte, 1978; Ledru et al., 1994; Matte, 2001), with the Münchberg-Tepla terranes of Bohemian (Matte et al., 1990), the High grade metamorphic complex (the Gallura Group; Naud, 1979) of northeastern Sardinia (Conti et al., 2001), and the Belgodere gneiss of central Corsica (Palagi et al., 1985). In particular, zircon U-Pb dating on fresh eclogites and retrogressed eclogites have yielded 460 5 Ma and 352 3 Ma, respectively interpreted as the age of emplacement for the mafic protolith and the age for amphibolites facies metamorphism. These two

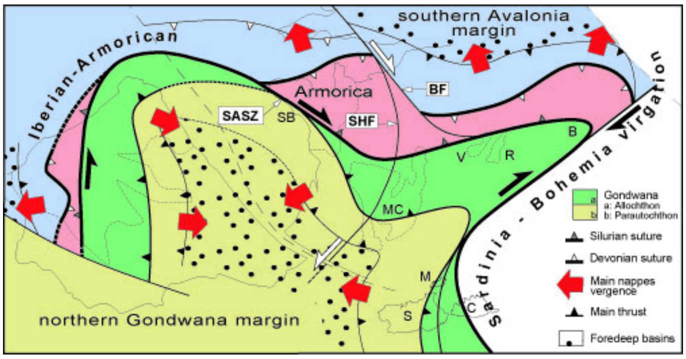
The Petites Maures Unit displays a wide range of similarities with the Upper Gneiss Unit of the Massif Central (Burg and Matte, 1978; Ledru et al., 1994; Matte, 2001), the Sila area of Calabria (Atzori et al., 1984; Liotta et al., 2004), and central Corsica (Bourrouihl et al., 1980). Cavalières and Petites Maures Units are very similar to units of the Alpine External massifs (von Raumer et al., 2002) of Argentera/Mercantour (Bogdanoff, 1986), and Aiguilles Rouges/Mont Blanc (Dobmeier, 1996).

Upper Permian basins of Provence display a wide range of similarity in facies and paleoenvironments with those from Argentera (Vinchon and Toutin-Morin, 1987), although similar Permian volcanism is described in northwestern Corsica (Vellutini, 1977).

Overall, the Maures-Tanneron massif is likely to exposed three main contrasting zones as part of three zones of the Variscan belt (Figure 5).

- The western Maures includes Cap Sicié, Fenouillet, Maurette, and Loli units. It corresponds to Ordovician-Early Carboniferous metasediments, low-metamorphosed and devoid of HP rocks. These units belong to an external zone of the Variscan belt including Pyrénées, Mouthoumet, southern Montagne Noire, Catalanides, south Sardinia, and Tuscany. They were formed at the north-Gondwana passive margin.
- The central zone includes Collobrières, Bormes, Cap Nègre Unit, and Cavalaire units that correspond to Precambrian to Early Ordovician magmatic rocks having experienced HP metamorphism and polyphased regional metamorphism. These units belong to an intermediate zone of the Variscan belt including South Brittany, the Massif Central, central Sardinia, and Belledonne / Aar Alpine External massifs. They likely correspond to pieces of a Cambrian back-arc lithosphere involved in the Ordovician-Silurian continental subduction.
- The eastern Maures-Tanneron, that includes Cavalières and Petites Maures units, is made of Cambrian-Ordovician metagranites and Ordovician metagabbros having experienced Silurian HP metamorphism and Upper Visean partial melting. These units belong to an internal zone of the Variscan belt including the Massif Central, northeast Bohemia (?), Argentera/Mercantour and Aiguilles Rouges/Mont Blanc Alpine External massifs, northeast Sardinia, Corsica, Tuscany, Sicilia, and Calabria. They likely correspond to pieces of an Ordovician back-arc oceanic basin developed from a Precambrian lithosphere, and involved in the Ordovician-Silurian continental subduction.

Figure 5. An attempt of reconstitution



An attempt of reconstitution for the southern Variscan belt based on lithological and structural correlations between Maures massif and other Variscan areas (modified after Matte, 2001). Abbrevations of Variscan outcrops: B = Bohemia, V = Vosges, R = Rhenish, S = Sardinia, C = Corsica, M = Maures, MC = Massif Central, SB = south Brittany. Abbrevations of Variscan faults: SHF = Sillon Houiller fault, BF = Bray fault, SASZ = South Armorica shear zone.

Tectonics and metamorphism

Early Carboniferous WNW-verging thrusts (D1) associated with IP metamorphism, that characterizes western and central Maures, have been found in the central and northern Massif Central (Bouchez and Jover, 1986; Faure et al., 1997; Roig and Faure, 2000; Bellot, 2001), western Bohemia (e.g., Matte et al., 1990), Alpine External massifs (Fernandez et al., 2002), and Sardinia (344 ± 7 Ma, whole rock Rb-Sr; Ferrara et al., 1978; Elter et al., 2000; Conti et al., 2001).

Lower Visean (~345 Ma) SE-verging thrusts (D2) associated with HT-IP metamorphism, that also characterizes the central Maures, have been described in the southern Massif Central (Gébelin, 2004), eastern Bohemia (Matte et al., 1990), and Alpine External massifs (Guillot and Ménot, 1999).

Upper Visean sinistral/top-to-the SSW shearing (D3), that followed or combined to SE-verging thrusts in the Maures massif, is also documented in northeastern Sardinia (Elter et al., 1999) where modern top-to-the SE shearing coeval with partial melting corresponds to top-to-the SSW in a pre-Permian position of Sardinia.

Namurian synorogenic extension (D4), that characterizes the central Maures, trends NW-SE in the Massif Central (Burg et al., 1994; Faure, 1995; Roig et al., 2002), Bohemia (e.g., Matte et al., 1990), Maures (Morillon et al., 2000; Bellot et al., 2000), and Sardinia (Elter et al., 2000;

Carosi and Palmeri, 2002), but trends SW-NE in Alpine External massifs (Guillot and Ménot, 1999) and in the Vosges massif (Rey, 1992).

The late Carboniferous (~300 ± 10 Ma) emplacement of I-type granite associated with heat advection then cooling, strike-slip brittle faulting, and deposition of coal basins (D5) characterize the eastern Maures. Equivalent of this event may be, on one hand, domes of the axial zone of the eastern Massif Central (e.g., Faure, 1995), Montagne Noire (e.g., Soula et al., 2001), Pyrénées (Vissers et al., 1995), and Catalonia. On the other hand, the eastern Maures granite seems to belong to an orogen-parallel K-granite belt extending from the Alpine External massifs (Aiguilles Rouges-Mont Blanc, Aar-Tavetsch-Gotthard massifs) (von Raumer et al., 1999), to the Corsica-Sardinia batholith (Orsini, 1979; Rossi and Cocherie, 1991; Gattacceca et al., 2004), and Calabria (Schenk, 1990 Gräßner et al., 2000). This granite belt, associated with extensive LP-HT metamorphism and partial melting, therefore reflects the NNE-SSW trending postorogenic extension of the Variscan lithosphere (Carmignani et al., 1994; Gattacceca et al., 2004).

Discussion

Problems on correlations of structures

The main structure of the Maures massif, i.e. the Grimaud fault (Demay, 1927a and 1927b; Vauchez and Bufalo, 1985; Morillon et al., 2000; Bellot et al., 2002b), prolongs likely to the South to the Posada-Asinara fault zone in northern Sardinia (Vauchez and Bufalo, 1988; Elter et al., 1990; Cappelli et al., 1992; Carmignani et al., 1994; Onézime et al., 1999; Carosi and Palmeri, 2002). In both cases, the shear zones (1) have experienced a ductile dextral slip at middle to low temperatures during which syntectonic granite emplace, and (2) separates two contrasted metamorphic and lithological blocks. The western block displays regional metamorphism increasing eastward from anchizonal to high-temperature amphibolites facies conditions (Arthaud and Matte, 1977; Vai and Cocozza, 1986; Buscail, 2000; Elter et al., 2000; Carosi and Pameri, 2002), although the eastern block is made of eclogite, granulite rocks and migmatites. However, the Grimaud fault has experienced a polyphased tectonics history (Vauchez and Bufalo, 1988; Onezime et al., 1999) including an Upper Visean sinistral movement, a Namurian dextral movement, and a Stephanian sinistral movement, although the Posada-Asinara fault zone seems to have experienced only dextral movement during the 330-300 Ma interval (Elter et al., 1990; Carosi et al., 2002). The Grimaud fault prolongs to the North to orogen-parallel strike-slip shear zones in the Alpine External massifs (von Raumer et al., 1989 and references therein), and the Central Bohemian shear zone in Bohemian (Rajlich, 1987) interpreted as a NW-verging thrust reactivated as a dextral fault during the Westphalian (Matte et al., 1990).

Top-to-the N to NE ductile shearing are found along the Bohemia-Sardinia branch and are interpreted as Middle Carboniferous thrust in the Belledonne massif (Fernandez et al., 2002), or Upper Carboniferous detachment faults in central Bohemia (Matte et al., 1990) and central Sardinia (Conti et al., 2001; Carosi and Palmeri, 2002). They are also found in the eastern Maures (Vauchez and Bufalo, 1988) but interpreted as syntectonic of the sinistral movement of the Grimaud fault, and thus probably associated with Middle Carboniferous S-verging thrusting in central Maures (Bellot et al., 2002b). Because of their contrasting interpretations and their unknown relationships with regional metamorphism, the significance of top-to-the N to NE ductile shearing is uncertain.

Plate-scale geodynamic model

The southern Bohemia-Sardinia Variscan branch probably formed a 1000 km-long and N-S to NE-SW-trending branch should be considered as a fundamental structure that should be included in the reconstitution of plate motions and the geodynamic history of the Variscan belt. This Variscan branch has experienced Precambrian back-arc spreading then Late Cambrian rifting. These small oceanic and continental domains were subducted, probably eastward, during the Late Ordovician-Silurian. Continental collision was initially frontal and absorbed by westward thrusting leading to crustal thickening. It rapidely became oblique, leading to southeastward back thrusting combining or evolving into sinistral orogen-parallel shearing that marks the waning stages of collision (Bellot et al., 2002b). Synorogenic extension leads to crustal thinning, but it is largely transferred along by orogen-parallel dextral movements. Postorogenic extension leads to crustal reequilibration but it is also transferred along by orogen-parallel sinistral movements. These fast changes likely reflect the progressive involvement of the north irregular margin of Gondwana in the collisional process.

Conclusion

Although some parts of its tectonics history remain unresolved, it is inferred that the Maures massif has a particular place in the Variscan belt of Western Europe. It corresponds to a small zone having experienced a similar history (with some exceptions) of external, internal, and suture zones associated with the north-Gondwana margin, but where wrench tectonics led to large-scale shearing that have considerably reduced in width the involved units.

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A. Tables

Table A.1. Stratigraphic and radiometric ages available on the Maures-Tanneron massif

Event dated	Tectonic	Dates	Method	Mineral	Other	References
Sedimentation	pre-D1	Middle Silurian (~430 Ma)				
Magmatism	pre-D1	~605 Ma	Pb-Pb	Zr		Chessex et al. (1967)
Granulite metamorphism	pre-D1	~560 Ma	Rb-Sr	WR		Maluski (1971)
Granulite metamorphism	pre-D1	575 ± 8 Ma	40 Ar/39 Ar	Bt		Maluski and Gueirard (1978)
Top-to-the SE shearing	syn-D2	344 ± 15 Ma	U-Pb	Zr	Discordia	Moussavou et al. (1998)
Top-to-the SE shearing	syn-D2	339 ± 16 Ma	U-Pb	Zr	Discordia	Moussavou et al. (1998)
Top-to-the SE shearing	syn-D2	345 ± 3 Ma	U-Pb	Mz	Concordia	Moussavou et al. (1998)
Top-to-the NW shearing	syn-D4	323,4 ± 3,3 Ma	40 Ar/39 Ar	Bt		Gaubert (1994)
Top-to-the NW shearing	syn-D4	328,1 ± 3,3 Ma	40 Ar/39 Ar	Bt		Gaubert (1994)