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The Variscan Belt of Western Europe 1

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Early Metamorphisms and Deformations in the French Variscan Belt

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4.1. Introduction

This chapter deals with the metamorphic and structural aspects of the principal events responsible for the early architecture of the French Variscan belt, before the tectono-metamorphic events coeval with its late- and post-orogenic dismantling. The Variscan metamorphic units are mainly located in the “Moldanubian” domain of the belt (see Chapter 2), corresponding to the North Gondwana margin, represented in France by the Southern Armorica, the whole of the Massif Central, the Central and Southern Vosges (see Figure 4.1). Metamorphic rocks were also found in the Léon block, which is part of the Saxothuringian domain. There is also a weak metamorphism in the prehnite-pumpellyite and greenschist facies in the Franco-Belgian Ardennes, which is part of the Rhenohercynian domain. In addition, metamorphic units are represented in the eastern branch of the belt, which includes the basement of the Alps, as well as the Massif des Maures and the Corsica–Sardinia massif in which the Alpine footprint is more limited. The early metamorphism identified in the Pyrenean Variscan massif is not presented here.

For a color version of all the figures in this chapter, see www.iste.co.uk/denele/variscan1.zip.

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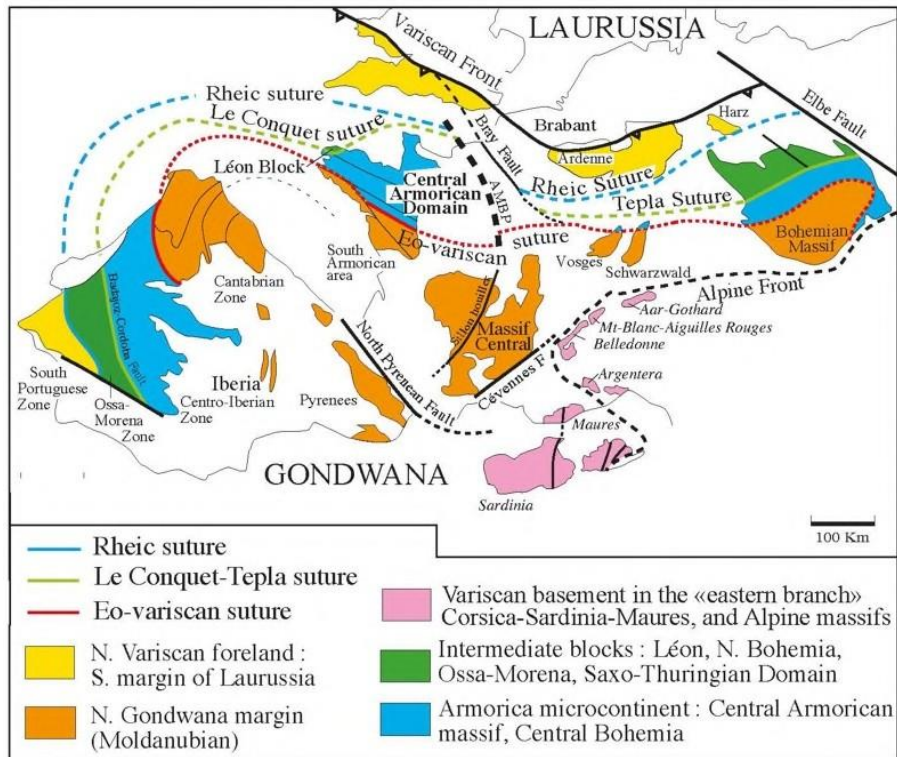


Figure 4.1. Schematic map of the various litho-tectonic domains of the western European Variscan massifs

METAMORPHISM AND DEFORMATION

Metamorphism is generally a syn-tectonic process that is accompanied by structural transformations. Metamorphic minerals are thus arranged according to surfaces, called either foliation planes or slaty cleavage depending on the metamorphic grade. The preferential crystallization of phyllosilicates with planar habitus such as micas, as well as the intra-crystalline deformation of the other minerals, are responsible for the easy splitting of micaschists or gneisses. The crystallization of acicular (amphibole) or fibrous (quartz, calcite, chlorite) minerals gives rise to a mineral lineation, which is most often parallel to the stretching lineation. Since Mattauer and Etchecopar (1976), in the study of the early stages of orogenic construction, it is considered that the low-dip syn-metamorphic foliation

is the structural response to large crustal-scale ductile shear zones responsible for the thickening of the continental crust. Mineral and stretching lineations are parallel to the transport direction of the material. Knowing its orientation, and the associated kinematic criteria, makes it possible to establish the sense of movement of the litho-structural units. Metamorphic crystallizations are highly dependent on protolith lithology and chemistry, chemical reaction kinetics, pressure (P) and temperature (T) conditions, and fluid circulation. Thus, the succession of observed paragenesis is not always representative of the entire metamorphic history of the structural units. Moreover, the accuracy of geothermometers, at best of the order of $\pm 50^\circ\text{C}$, and of geobarometers, of the order of ± 0.1 GPa, makes determining P–T paths uncertain. Finally, one of the consequences of superimposed events is the erasure of early events to a greater or lesser extent. Nevertheless, the combination of thermo-barometric, structural and geochronological studies makes it possible to establish the P–T–t–d paths followed by each of the units.

4.2. Metamorphisms and deformations in the Moldanubian domain

In this area, four early tectono-metamorphic events, D₀, D₁, D₂, D₃ contributed to the formation of the belt (see, for example, Faure *et al.* 2005, 2009; Lardeaux 2014). We do not consider pre-orogenic metamorphisms below, as the contact metamorphism developed around Cambrian or Ordovician plutons, which were later transformed into orthogneiss. Similarly, here we do not detail the oceanic metamorphisms described in the Chamrousse ophiolite (Guillot *et al.* 1992) and in Limousin (Berger *et al.* 2005). Most of the data presented in this section come from large amounts of work carried out in the Massif Central and Southern Armorica. Additional information is given for the Central and Southern Vosges, but the correlations between these three massifs are still uncertain. In order to facilitate reading, a summary of the structure of the Massif Central is given below (for more details, see Ledru *et al.* (1989) and Faure *et al.* (2009)). A historical overview of the evolution of ideas on the structure of the Massif Central is given in Faure (2020).

NAPPE ARCHITECTURE OF THE FRENCH MASSIF CENTRAL

It is now well accepted that the French Variscan Massif Central was formed by a stack of ductile and syn-metamorphic nappes all belonging to the domain known as the “Moldanubian” which, paleogeographically, corresponds to the northern margin of Gondwana (see Chapters 1 and 2). Several units are distinguished by their lithological contents and ages (Ledru *et al.* 1989; Faure *et al.* 2009; Faure *et al.* 2020). From the top to the bottom of the nappe stack, the following

litho-tectonic units are widely distributed throughout the FMC (see Figures 4.2 and 4.3).

The *Upper Gneiss Unit (UGU)* is the most original unit of the nappe stack. This is the unit in which most of the centimetric to metric alternations of acidic gneisses and amphibolites called “*leptyno-amphibolite complex*” (LAC) containing high-pressure metamorphic relics (eclogites and granulites) occurred. In general, the UGU is exposed as klippes: Central Limousin, Uzerche synform, Haut-Allier and Marjevols, Lézérou.

The *Lower Gneiss Unit (LGU)*, tectonically underlying the UGU, hardly differs from a lithological point of view. In the Tulle region, felsic–mafic alternations formed a leptyno-amphibolite complex, but unlike those of the Upper Gneiss Unit, it does not contain metamorphic HP relics. Micaschists and paragneisses of the LGU were metamorphosed in the amphibolite facies.

The *Para-Autochthonous Unit (PAU)* tectonically underlies the previous one. It is located in the central Cévennes, the upper Lot region and La Châtaigneraie. To the west of the Sillon Houiller, it formed the deepest part of the Millevaches massif. In Limousin, it appeared in the core of the tectonic windows of Sussac, Drone, Saint-Goussaud, Plateau d'Aigurande and Sioule, east of Sillon Houiller. Lithologically, the PAU is characterized by the abundance of sericitous and quartz-feldspar micaschists.

The *Fold-and-Thrust Belt Unit* forms the southern and northern slopes of the Montagne Noire, the Lodévois horst and Viganais in the Southern Cévennes. Lithologically, this complex contains mainly sedimentary rocks that can be locally epimetamorphic: limestones, sandstones and argillites whose age ranges from the Lower Cambrian (possibly Ediacaran) to the lower Carboniferous (Serpukhovian).

The *southern foreland basin* outcrops in the SE part of the Montagne Noire and extends under the Mesozoic and Cenozoic sediments of Languedoc to reappear in the Mouthoumet massif, south of Carcassonne, and then in the Pyrenees. It is also present in Sardinia and Minorca in the Balearic Islands. It is formed by turbidites (flyschs) dating from the Viséan to the Serpukhovian or even Bashkirian (Namurian-Westphalian) (see Chapter 3 of Volume 2).

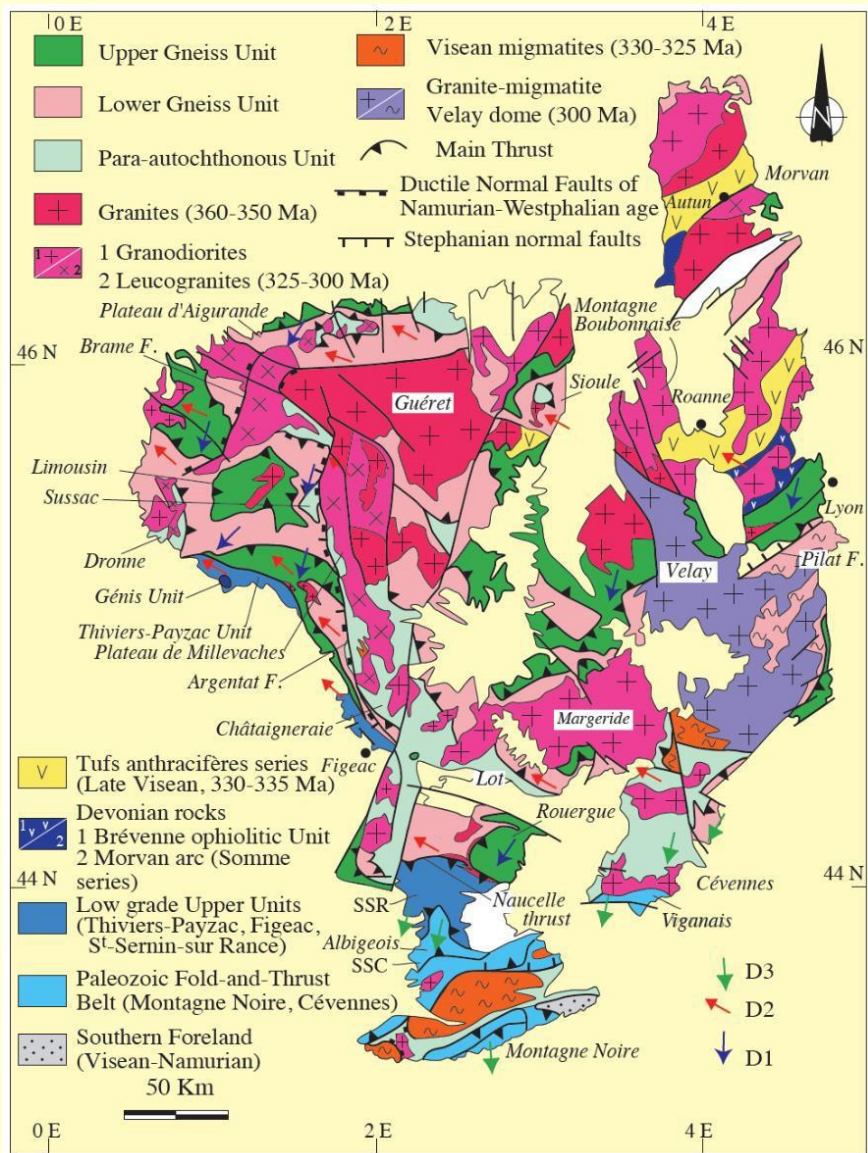


Figure 4.2. Structural map of the Massif Central showing the main lithotectonic units as well as the three compressive events (D_1 , D_2 , D_3) described in the text. SSR: Saint-Sernin-sur-Rance nappe; SSC: Saint-Salvi-de-Carcavès nappe (from Faure et al. 2009)

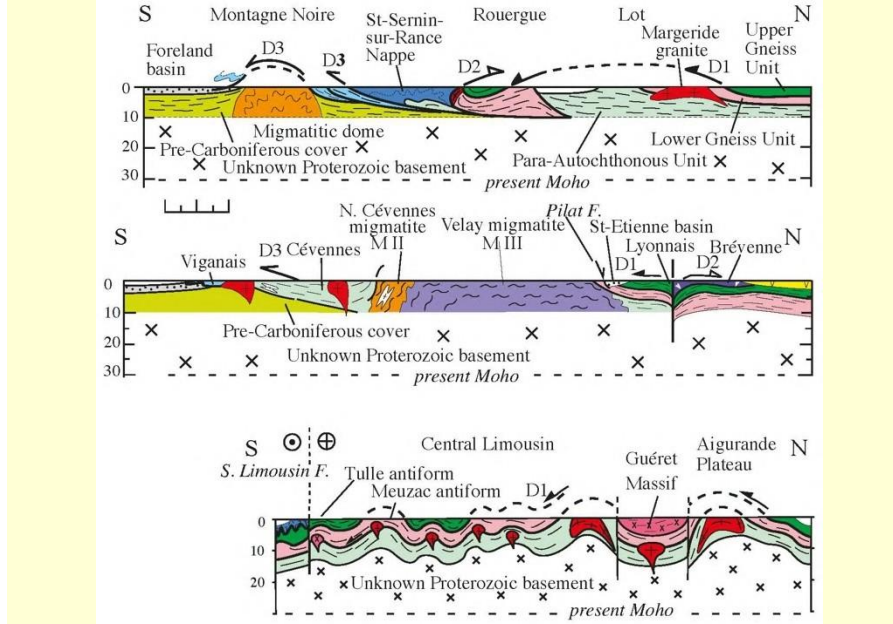


Figure 4.3. Simplified general cross-sections of the Variscan structure of the Massif Central showing superimposed deformations of the nappe stack folded into antiforms and synforms in kilometer scale and intersected by granitic intrusions. Top: N–S cross-section from the S of the Montagne Noire to the Margeride; middle: N–S cross-section of the SE of the French Massif Central from Viganais to Brévenne; bottom: N–S cross-section in Limousin (from Faure et al. 2009). Captions are the same as for the structural diagram in Figure 4.2

In addition to this succession, it is worth mentioning some peculiar units limited to some regions.

– The *Thiviers-Payzac Unit (TPU)* forms an elongate strip developed along the border of the SW Massif Central, from Limousin to Quercy. To the East of the Sillon Houiller, the St-Sermin-sur-Rance nappe, although lithologically similar to the TPU is ascribed to the PAU on the basis of structural and metamorphic features. The sedimentary and felsic volcanic formations are well developed. The TPU is marked by a regional metamorphism developed in greenschist or amphibolite facies conditions.

– The *Brévenne Unit* is located in the NE of the French Massif Central, north of the Monts du Lyonnais and in the Beaujolais. The existence of basaltic lavas (sometimes in the form of pillow lavas), diabase, gabbro, serpentized peridotites, sodic granites

(albitites), basic volcano-sedimentary rocks (grauwackes, detrital pyroxene sandstones), acid lavas (keratophyres) and siliceous sediments (argillites and radiolarites) is known for a long time (Peterlongo 1960). The tholeiitic chemistry of magmatic rocks is consistent with an ophiolitic series of Upper Devonian age. The Brévenne unit constitutes an ophiolitic nappe thrust upon chlorite gneisses (Affoux gneisses) equivalent to the UGU but already exhumed and retromorphosed before

the thrusting. The Brévenne nappe and its substratum are unconformably covered by lower Visean sandstones.

– The *Somme series* and the Devonian–Dinantian formations are known only in the western part of the central Morvan (Delfour 1989). It is a succession of volcanic rocks (andesites, dacites, rhyodacites and rhyolites), and sedimentary rocks (grauwackes, conglomerates, pyroclastites, tuffs), as well as massive sulfide deposits. The *Somme series* is neither metamorphic nor ductilely deformed; it is paleontologically dated from the lower Devonian (Frasnien-Famennien, 380–370 Ma), but the basal contact of the series with its substratum is unknown. Lower to middle Devonian (Emsian to Givetian, 407–385 Ma) limestone pebbles reworked in the Carboniferous suggest that the base of the *Somme series* may have reached the middle Devonian. Magmatic rocks have a geochemical calc-alkaline signature characteristic of a magmatic arc. The *Somme series* is unconformably covered by non-metamorphic and undeformed lower Visean detrital formations. UGU gneisses, eclogites and migmatites are known to be present in the Autun region, but the relationships between these metamorphic rocks and those of the *Somme series* are unknown. In spite of the lack of field arguments, it is often assumed that Devonian sedimentary rocks unconformably overlie this metamorphic substratum.

– The “Tufs anthracifères” series exists only in the north of the French Massif Central, where it is located in the form of two “bands” in the northern Morvan to the N of Autun and in the Roanne region. To the west, the formation of the “Tufs anthracifères” series continues in the Montagne Bourbonnaise, then south of the Sioule region and in the northern Limousin, where it forms the deepest part of the Bosmoreau-les-Mines basin. The “Tufs anthracifères” series consists of conglomerates with various pebbles, including fragments of metamorphic rocks and granites, sandstones, silts with coal intercalations. It also contains volcanic flows (rhyolites, dacites) and microgranite or dolerite veins. It is paleontologically dated from the Late Visean by plants. Rb–Sr and U–Pb radiometric dates at approximately 330 Ma confirm this early Carboniferous age. The “Tufs anthracifères” series unconformably covers some granites (e.g. the St-Gervais-d’Auvergne massif in the Sioule) and also the *Somme arc series* or the Brevenne ophiolites. This formation is therefore an essential chronological marker for the tectonic evolution of the French Massif Central because in the north of the Massif, it seals the syn-metamorphic

deformation D₂ (see Chapter 1 of Volume 2. However, it is important to note that in the south of the FMC, the D₃ deformation of the para-autochthonous units and the Fold-and-Thrust Belt is synchronous with the deposition of the “Tufs anthracifères” formation.

4.2.1. *Eo-Variscan events*

4.2.1.1. *HP and UHP metamorphism (D₀)*

Known since the end of the 19th century by the works of Lacroix and his students (Brière 1920, see Godard 2001 for more details), high-pressure (HP) rocks are mainly found in the klippe forming the UGU of the French Massif Central.

These are boudins of metabasites and more rarely ultrabasites included in plagioclasic gneisses (or grey gneisses = metagrauwackes) that are often migmatitic. Eclogites of the Axial Zone of the Montagne Noire (see section 4.2.2.3) do not belong to this event. HP rocks are also exposed in the south of the Armorican Massif in three places: (i) in the Champtoceaux complex, (ii) in the Audierne bay and (iii) in Vendée in the Essarts complex. Blue schist units of Southern Armorica – namely, Groix Island and Bois de Céné – are added to this list. HP rocks are also found in the “Gneiss Variés” unit in Middle Vosges (see Figure 4.4).

Coronitic textures of magmatic origin, preserved in some eclogites, indicate that protoliths are often gabbros, but basaltic, doleritic and volcano-sedimentary protoliths are also likely. Geochemical analyses of metabasites show various signatures: oceanic tholeiites, alkaline basalts and continental tholeiites. The bimodal magmatism characteristic of leptyno-amphibolite complexes fits well with a geodynamic context of distal continental margin (thinned continental crust or ocean–continent transition zone). Groix Island and Bois de Céné blue schist rocks are derived from gabbros and basalts included in metapelites and metagrauwackes derived from the erosion of continental detrital rocks. An accretionary prism setting in an active continental margin is a plausible interpretation (Cogné 1974; Matte 1991; Bosse *et al.* 2005; Ballèvre *et al.* 2009).

Metamorphic assemblages vary according to the lithology of the protolith. Although often retromorphosed to garnet amphibolites, metabasites contain clinopyroxene-plagioclase symplectites derived from the destabilization of omphacite (Lasnier 1968; Briand 1973; Santallier 1975; Nicollet and Leyreloup 1978; Santallier and Floc’h 1978, 1979; Lardeaux *et al.* 1989, 2001; Mercier *et al.* 1989; Ballèvre *et al.* 1994; Godard 2001; Berger *et al.* 2010; Lardeaux 2014). Thus, we distinguish between eclogites with omphacite-quartz-rutile-zoisite, eclogites with kyanite and eclogites with coesite. Corundum and sapphirine assemblages are

known in the Haut-Allier region (Lasnier *et al.* 1982). Some garnet-clinopyroxene (diopside) rocks, or pyrigarnites and clinopyroxene (diopside)-orthopyroxene-aluminous-plagioclase-garnet rocks, or pyriclasites, belong to the high-pressure granulite facies.

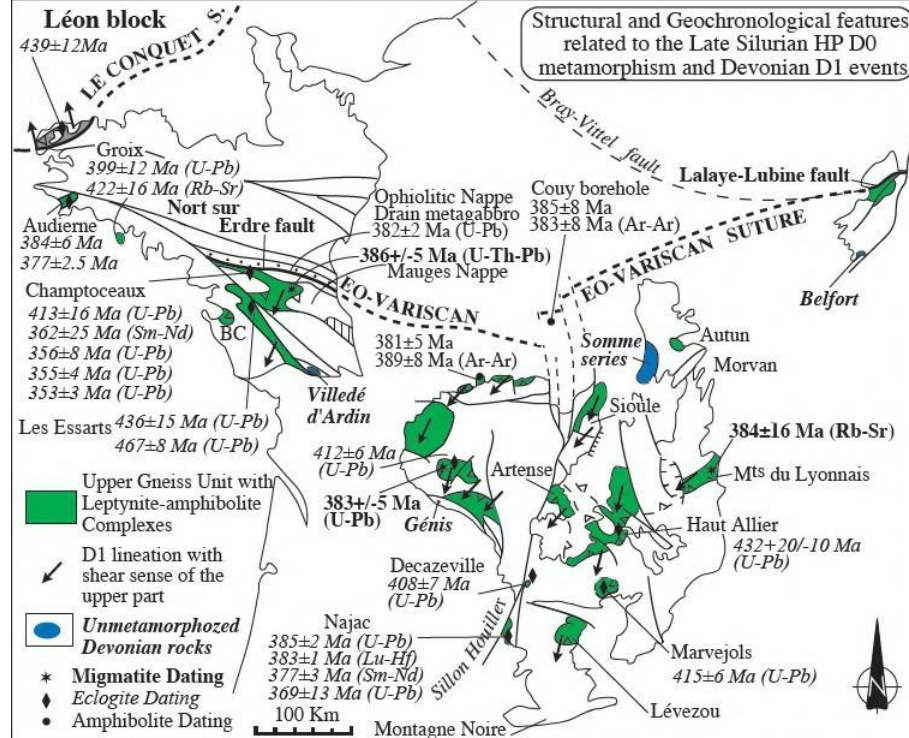


Figure 4.4. Structural map of the Armorican, Central and Vosges Massifs locating high-pressure rocks and the structural and geochronological data of D_0 and D_1 events. BC: Bois de Céné

Ultrabasites are also recognized, the largest masses of which are found in the La Bessenoit klippe in Rouergue, in central Limousin, in the Monts du Lyonnais and in the Audierne bay. These rocks are highly serpentinized lherzolites or harzburgites. In the Lyonnais mountains, the Bois des Feuilles peridotites contain a garnet-spinel paragenesis (Lardeaux *et al.* 1989).

Sodium amphibole rocks (blue schist or glaucophane eclogite facies) are rare in the French Variscides. The most famous ones are the garnet-lawsonite-epidote

glaucophanites of Groix Island, but they also exist in Vendée in the Bois de Céné klippe (Triboulet 1991) and in Rouergue in the Najac klippe (Delor *et al.* 1986).

Quartz-feldspathic rocks metamorphosed under HP conditions are not well identified. Observation of garnet coronitic textures around biotite or crystallized clinopyroxene from magmatic hornblende are criteria for recognition of an HP metamorphism developed from magmatic protoliths. The La Picherais orthogneiss, which is part of the St-Mars-du-Désert orthogneiss in the Champtoceaux complex (Lasnier *et al.* 1973; Ballèvre *et al.* 1994, 2009) and an orthogneiss in the Vibal klippe, Rouergue, are metamorphosed under high pressure conditions (Burg and Leyreloup 1989; Maurel *et al.* 2003). In Vendée, in the Les Essarts complex, orthogneiss and paragneiss forming the host of basic eclogites have preserved relics of an HP metamorphism, as well as pre-Variscan HT metamorphism (Godard 1988, 2001; see Figure 4.5). Biotite-garnet-cordierite±sillimanite paragenesis show a pre-clogitization event at 770°C and 0.5 GPa. In these same rocks, HP metamorphism conditions are estimated to be 650°C and 1.65 GPa.

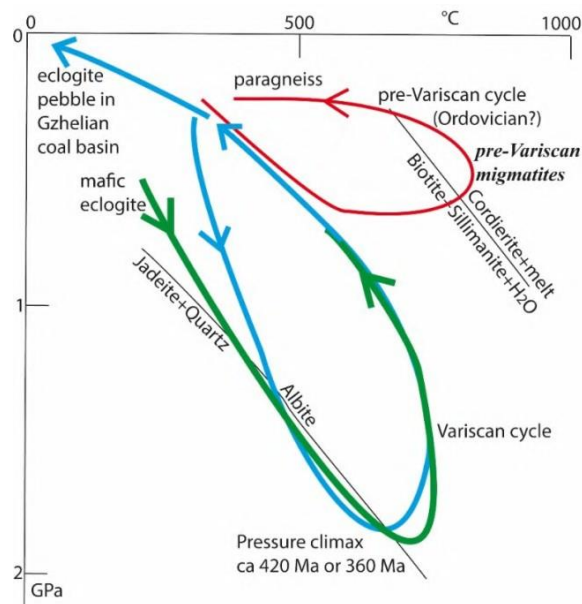


Figure 4.5. Proposed P - T paths for HP rocks of the Essarts complex in Vendée, equivalent to the UGU of the French Massif Central (from Godard 2001, 2009). Red: pre-Variscan P - T path of gneisses, supposedly related to Ordovician rifting. Sedimentary rocks were metamorphosed under HT/LP conditions producing cordierite migmatites. Blue: Variscan P - T path for sedimentary and magmatic felsic rocks. Green: P - T path for mafic rocks (gabbros)

Detailed studies have documented geothermal gradients between 7 and 15°C/km and a pressure peak of 3 GPa compatible with burial at more than 90 km for these high-pressure rocks (Ballèvre *et al.* 1994; Godard 2001; Lardeaux *et al.* 2001; Berger *et al.* 2010; Lardeaux 2014; see Figure 4.6). Nevertheless, the P - T paths are ambiguous because the synchronism between pressure (P_{\max}) and temperature (T_{\max}) peaks is not established. Eclogites of the French Massif Central and the Armorican Massif indicate P_{\max} and T_{\max} conditions between 2 and 3 GPa and between 700 and 750°C, respectively. HP granulites reached values of 1.5 GPa and 800–850°C. It can therefore be proposed that after having passed the pressure peak, some eclogites were partially heated.

In the Middle Vosges, the “Gneiss Variés” and “Granulites Acides” unit rocks, probably equivalent to the UGU, contain garnet-kyanite-K-feldspar assemblages indicating thermo-barometric conditions of 800–880°C and 1.4–1.6 GPa and 700–900°C and 1.2–1.5 GPa, respectively (Skrzypek *et al.* 2014; references included). Garnet±spinel peridotites, preserving mantle conditions (5 GPa, >1,000°C), were incorporated tectonically into crustal rocks under HP/HT conditions (1–1.2 GPa, 800–900°C; Altherr and Kalt 1996), close to those determined in the Gneiss Variés of ≈1–1.2 GPa, 800–900°C (Rey *et al.* 1989; Latouche *et al.* 1992).

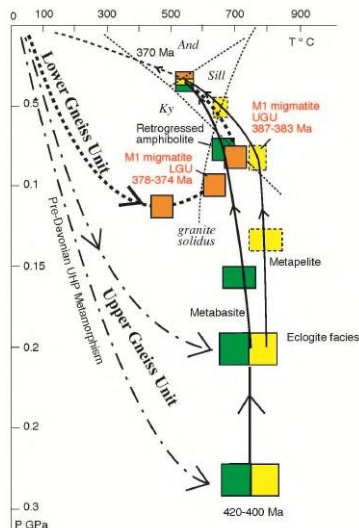


Figure 4.6. Synthetic pressure–temperature–time paths of the Upper Gneiss Unit and Lower Gneiss Unit for events D_0 and D_1 . Green: basic protolith; yellow: felsic protolith; orange: sedimentary or magmatic aluminous protolith (from Faure *et al.* 2008)

Thermo-barometric conditions calculated from the metapelites and metabasites of the Groix Island klippe range between 400–450°C and 1.4 and 1.6 GPa for the lower unit and between 450–500°C and 1.6–1.8 GPa for the upper unit (Ballèvre *et al.* 2003; Bosse *et al.* 2005). Glaucophanites of the Bois de Cené klippe were metamorphosed in the epidote-blueschist facies under thermo-barometric conditions of 500°C and 1.1 GPa (Triboulet 1991). HP micaschists are then partially retromorphosed during syn-tectonic exhumation to achieve low pressure conditions of about 0.2 GPa.

Currently, there is very little data on planar or linear microstructures developed in eclogites during the D_0 event. In the Les Essarts complex in Vendée, some garnets contain oriented omphacite inclusions defining an early foliation (Goddard 2001). In the eclogite matrix, the omphacite also exhibits a preferred orientation with a “subvertical” planar fabric and a NW–SE striking mineral lineation (Godard and Van Roermund 1995; Mauler *et al.* 2001). Groix Island metabasites have an NW–SE striking stretching lineation (Quinquis and Choukroune 1981; Cannat 1985). Ductile shearing to the SE was coeval with HP metamorphism, while that to the NW was coeval with the exhumation of the blueschist unit (Philippon *et al.* 2009). In the Bois de Cené klippe, glaucophane crystals are aligned in the NNE–SSW direction (Godard *et al.* 2009).

The age of the early HP event, D_0 , also known as Eo-Variscan, remains to be discussed (Pin and Peucat 1986; Lardeaux *et al.* 2014). In the SE French Massif Central, U–Pb dates on eclogitic zircon populations indicate ages of 415±6 Ma (Pin and Lancelot 1982) and 432±20 Ma (Ducrot *et al.* 1983). In Limousin, ICP-MS U–Pb dating of zircon monograins gives an age of 421±10 Ma (Berger *et al.* 2010). The Sm–Nd chronometer on metagabbros of the La Bessenoit klippe gives an age at 408±7 Ma (Paquette *et al.* 1995), and the zircon Pb–Pb evaporation method yields an age at 413±23 Ma (Paquette *et al.* 1995). The monazite core gives a U–Th–Pb age of 415±15 Ma. However, retromorphosed eclogites of the Najac’s klippe provided Lu–Hf and Sm–Nd isochronous ages of 383±1 Ma and 377±3 Ma,

respectively, while the age of late apatite crystallized in garnet fractures is 369 ± 13 Ma (Lotout *et al.* 2018).

In the Armorican Massif, recent data on HP rocks of the Champtoceaux and Vendée complex do not give any Silurian–Devonian Eo-Variscan age (Paquette *et al.* 2017). A Champtoceaux eclogite provides a zircon U–Pb age at 413 ± 16 Ma (Paquette *et al.* 1985), currently considered “meaningless” (Paquette *et al.* 2017). The La Picherais HP orthogneiss gives a U–Pb age of 423 ± 10 Ma (Vidal *et al.* 1980). The age at 467 ± 8 Ma of a Vendée eclogite is interpreted as a contamination effect of a basic protolith by Proterozoic crustal zircons. The U–Pb at 355 ± 4 Ma and

353 ± 3 Ma ages of the Champtoceaux eclogites are close to the Sm–Nd and U–Pb ages, respectively, at 362 ± 25 Ma and 356 ± 8 Ma (Bosse *et al.* 2000).

U–Pb, Rb–Sr and $^{40}\text{Ar}/^{39}\text{Ar}$ dating were carried out on the Groix Island blueschists. Early measurements on detrital zircon indicate ages of 422 ± 16 Ma and 399 ± 12 Ma (Peucat 1986). On the contrary, the Rb–Sr and $^{40}\text{Ar}/^{39}\text{Ar}$ dates on phengites provide more recent ages around 360–350 Ma acquired during the retrograde path of the HP units during their exhumation (Bosse *et al.* 2005). These results are easily interpretable when considering the closing temperatures of chronometers (see box below).

In the Middle Vosges, eclogites of the Gneiss Variés Unit and HP Granulites Acides contain zircons dated between 345 and 335 Ma. These ages largely post-date the pressure peak (Skrzypek *et al.* (2014) and references included).

SIGNIFICANCE OF RADIOMETRIC AGES

Since the early 1980s, radiometric dating in the Variscan belt has been essential for tectonic and geodynamic interpretations. Some methods (Rb/Sr, Sm/Nd) are based on isotopic measurements on several supposedly co-genetic rocks or on the whole rock and separated minerals. The alignment of analytical points defines an isochron, which may correspond to the crystallization of magma or to a subsequent event (metamorphism, hydrothermalism, etc.). These techniques are open to criticism to the extent that it is not certain that the samples and minerals analyzed are contemporary. Other methods have been developed by improving analytical conditions of very small quantities of material. For example, the laser ablation technique makes it possible to sample and then to assay isotopes contained in a few cubic microns (μm^3) of a mineral. This is called “spot age determination”. It is sometimes confused with “in-situ” dating, which is also spot-age, but carried out on rock samples and not on separate minerals. The advantage of in situ dating is that it respects the textural relationships between minerals in a rock, or sometimes allows for dating inclusions in other minerals. Isotopic dating requires the measurement of daughter products (resulting from the decay of a parent element), but the amounts of daughter and parent isotopes of elements contained in a mineral may have varied throughout its geological history. Temperature, fluid flow and grain size are the most important factors to consider when discussing the significance of radiometric data. It is important to take into account the “closing temperature” of the isotopic system. For example, for the biotite, the *closing temperature* of the $^{40}\text{Ar}/^{39}\text{Ar}$ chronometer is 350–400°C, which means that the obtained age corresponds to the moment when the biotite has passed under the 350–400°C isotherm, when the isotopic system has become closed and the

chronometer has been triggered. Thus, geochronometers are also geothermometers. Sometimes, the term of “thermo-chronology” is preferred instead of geochronology. The effects of fluid circulation or changes in grain size (mylonization, recrystallization) are more difficult to assess, and require careful textural study of a rock and its constituent minerals. Dated minerals might have had a long and complex geological history. For example, a zircon grain present in a granite may be a refractory grain (a residual lump) inherited from the magma source protolith. This is the reason why it is common to find grains in magmatic rocks with inherited cores surrounded by concentric aureoles crystallized during the cooling of magma. In addition, the magmatic rock can be transformed in the solid state into orthogneiss, or be percolated by fluids or be melted to give rise to migmatites, or even anatectic granites. During these stages, the zircon grain acquires new overgrowths. It is therefore very important, for interpreting radiometric ages, to know the internal structure of grains, imaged by various techniques (scanning electron microscopy, cathodoluminescence). In the years 1990–2000, a “chemical” dating method of monazite was developed, notably in France by Cocherie at BRGM (Cocherie *et al.* 1998; Cocherie and Albarède 2001). This method is based on the hypothesis, verified a posteriori, that this mineral incorporates less than 5% of common Pb during its crystallization. An age can thus be calculated by simply measuring, with an electronic microprobe, the total U, Th, and Pb concentrations resulting from the disintegration of U and Th elements. This method has the advantages of being fast, inexpensive, and involving very small volumes of material (almost 300 times less than for laser ablation). It is therefore possible to clearly demonstrate the inheritance or poly-metamorphism phenomena. On the contrary, the absence of isotopic measurements gives inaccurate, often overestimated, ages. Finally, as any physical measurement is flawed, in discussing the significance of radiometric ages, it is also necessary to consider uncertainties that must be given at $\pm 2\sigma$.

The age of the Eo-Variscan HP D_0 event is therefore poorly constrained, since the most modern geochronological methods indicate rather Upper Devonian–lower Carboniferous ages. However, the interpretation of certain data remains difficult because the behavior of isotopic chronometers in minerals with respect to deformation and hydrothermal fluids remains highly underestimated. It is difficult to determine whether the ages obtained reflect the pressure peak or already indicate a retrograde time interval of the P–T path. In particular, the Upper Silurian to Lower Devonian ages are questionable because they were carried out on grain populations. Mixing effects between grains that have preserved the age of protolith (generally Lower Ordovician) and that of HP metamorphism at 390–380 Ma may be responsible for these intermediate ages. On the contrary, the most recent dates, obtained with the ICP-MS method with laser ablation on monograins, are also

inaccurate and associated with a significant interpretative bias. Thus, the ~380 Ma ages of Najac eclogites (Lotout *et al.* 2018) could date their exhumation because the Lu–Hf and Sm–Nd isochrone calculations take into account the crystallized hornblende during the retrogression of HP rocks. At present, a flaccid consensus seems to emerge to estimate the age of HP metamorphism around 390–380 Ma (Givetian). Although the Lower Devonian to the Later Silurian ages (420–410 Ma) are probably to be eliminated, ages around 360–350 Ma are difficult to reconcile with other stratigraphic (especially those of the NE of the French Massif Central) and structural (early retrogression of migmatitic gneisses of the Brevenne nappe tectonic substratum) constrains. Of course, there could be several episodes of HP metamorphism. At the scale of this small French segment of the Variscan belt, this half-hearted interpretation appears unconvincing. It is also interesting to note that these Upper Devonian (385–375 Ma) and late Lower Devonian–Carboniferous (~360–350 Ma) ages are similar to those obtained for the D₁ and D₂ tectono-metamorphic events, respectively.

4.2.1.2. MP/MT event (D₁)

Structural, metamorphic and anatectic phenomena are grouped under the term D₁ event. Structurally, D₁ is the first event identified in the field by the main foliation, S₁, often with a low dip and a mineral and stretching lineation, L₁, oriented NE–SW, associated with top-to-the-SW kinematics (see Figure 4.4). In the French Massif Central and Southern Armorican Massif, L₁, recognized in both the UGU and the LGU, is interpreted as an indication of an early emplacement of the UGU as a nappe upon the LGU of Monts du Lyonnais, Plateau d’Aigurande, Limousin, Rouergue, Artense, Sioule and Haut-Allier (Nicollet 1978; Rolin and Quenardel 1980; Floc’h 1983; Burg *et al.* 1984; Lardeaux and Dufour 1987; Faure *et al.* 1993, 2005). In the Champtoceaux complex, shearing in the NW direction is associated with the UGU exhumation (Lagarde 1978; Faure *et al.* 2005, 2008; Philippon *et al.* 2009), but due to radiometric ages at approximately 360 Ma, it could also be attributed to D₂ (see section 4.2.2.1).

Metabasites of the UGU leptyno-amphibolite complex underwent an isothermal decompression (see Figure 4.6). HP eclogites and granulites are retromorphosed to garnet amphibolites and common amphibolites (Santallier 1976; Mercier *et al.* 1989, 1991; Santallier *et al.* 1994; Berger *et al.* 2010; Lardeaux 2014). ⁴⁰Ar/³⁹Ar dates on amphiboles indicate ages between 379 and 385 Ma for this retrogression (Costa and Maluski 1988; Boutin and Montigny 1993). This latter data comes from relatively low temperature thermo-chronometers (450–350°C); thus, the anatexis age could be older.

Furthermore, the hydrated pelitic (paragneiss and micaschists) and quartz-feldspar (orthogneiss) rocks were partially melted to give rise to biotite-sillimanite-muscovite±garnet±cordierite migmatites, called “M₁ migmatites” (Faure *et al.* 2009; see Figure 4.6). Orthogneiss parent rocks, still recognizable in some migmatites, testify to the polyphase history of the LGU and the UGU. These orthogneiss are the Cambrian or Ordovician granites that emplaced during the rapture of Panotia (see Chapter 2), which were deformed and turned into orthogneiss during the early D₀ and D₁ events, then partially melted at the end of D₁. With very rare exceptions, such as the La Picherais orthogneiss of the Champtoceaux complex, or the Pomeyrols metagranodiorite in the Vibal klippe in Rouergue, the orthogneisses did not keep track of the D₀ event. This absence may be due either to a complete obliteration of the HP parageneses or to the fact that these rocks have never actually been deeply subducted under the conditions of the eclogite facies. Although this latter interpretation is acceptable for the LGU, it is difficult to accept for the UGU because orthogneiss and metabasites belong to the same contemporary bimodal magmatic set of pre-orogenic rifting. In the early stages of M₁ anatexis, the presence of garnet and K-feldspar, the destabilization of kyanite to sillimanite and rutile to ilmenite, and the absence of primary muscovite indicate pressure and temperature conditions of 0.9–1 GPa and 700–750°C, respectively. Melting then continued under lower pressure and temperature conditions, respectively at 0.4–0.6 GPa and 650–700°C as UGU and LGU are progressively exhumed (Floc’h 1983; Mercier *et al.* 1991; Roig and Faure 2000; Bellot and Roig 2007; Lardeaux 2014). U–Pb zircon and Rb–Sr total rock and mineral radiometric data indicate ages from late Middle Devonian (Givetian) to early Upper Devonian (Frasnian) at about 375 Ma (Duthou 1977; Dufour 1982; Lafon 1986; Duthou *et al.* 1994; Feybesse *et al.* 1996; Cocherie *et al.* 2005; Faure *et al.* 2008). The Morvan M₁ migmatite contains garnet amphibolites, retrogressed eclogites and spinel peridotites (Godard 1990). These rocks crop out near Autun, close to the Somme sedimentary and volcanic series (Delfour 1989; Faure *et al.* 1997), the lower part of which (near the villages of Diou and Gilly) is paleontologically dated to the Upper Devonian (Frasnian, approximately 380 Ma), but the true basis of this series is not known. Moreover, Emsian- to Givetian-aged limestone pebbles (407–385 Ma) were reworked in the lower Carboniferous formations (Delfour 1989). These observations suggest that the LGU and the UGU were placed in the supracrustal conditions as early as the end of the Devonian. Numerical modeling indicates that the anatexis accompanying the exhumation of a continental crust occurs approximately 20–30 Ma after the pressure peak (e.g. England and Thompson 1984; Thompson and Connolly 1995). This time interval coincides with the Eo-Variscan cycle described above.

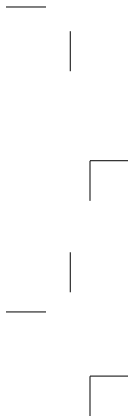
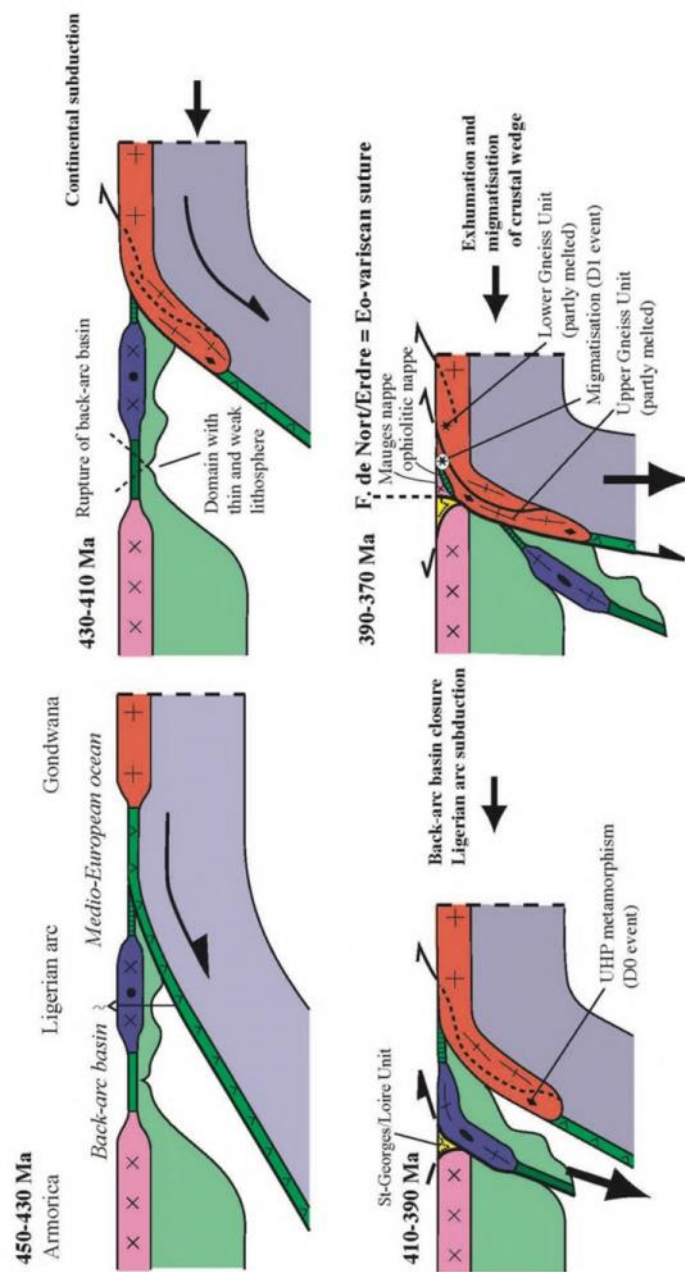


Figure 4.7. Geodynamic model of the Eo-Variscan cycle showing the location of D_0 and D_1 events in the lower plate (according to Faure et al. 2008)



In the Middle Vosges, the early metamorphic events are less constrained than in the Armorican Massif and the French Massif Central (Skrzypek *et al.* 2014). The biotite-sillimanite Monotone Gneiss series, comparable to the LGU, is overthrust by the Gneiss Variés series and Granulite Acides series, which contain peridotite, eclogite, amphibolite, marble and felsic granulite lenses (Wickert and Eisbacher 1988; Fluck *et al.* 1989; Rey *et al.* 1989). After the HP metamorphism (see the previous section), the two series were tectonically superimposed. Metamorphic conditions are quite close, respectively, from 0.7 to 0.9 GPa and 750–800°C and 0.5 to 0.65 GPa and 600–660°C for the upper and lower series. The rare structural data refers to a NE–SW stretching lineation (Wickert and Eisbacher 1988) that can be compared to the L₁ one of the French Massif Central. The M₁ migmatization has not yet been recognized in the Middle Vosges.

A geodynamic interpretation of the Eo-Variscan events is proposed in Figure 4.7. The closure of the Middle European Ocean (see Chapter 2) is accommodated by an oceanic subduction under a magmatic arc located on the Armorican margin. In this interpretation, the continental subduction of the northern margin of Gondwana was responsible for the high (or even ultra-high) pressure metamorphism in the Upper Gneiss Unit. The exhumation of the crustal root of the Eo-Variscan belt is associated with a first anatexis (M₁ migmatization) and the syn-metamorphic deformations of the D₁ event (for more details, see Faure *et al.* 2008).

4.2.2. Lower Carboniferous Variscan tectono-metamorphic events

4.2.2.1. MP/MT event (D₂)

This is the major phase of the Variscan orogeny, recognized for a century as the “Bretonian phase” (Stille 1922 cited in Rolet *et al.* 1982; Faure *et al.* 2017). In the French Massif Central, the D₂ event is found in Limousin (Floc’h *et al.* 1977; Floc’h 1983; Girardeau *et al.* 1986; Friedrich *et al.* 1988), Rouergue, Haut Lot, Northern Cévennes, Monts du Lyonnais, Haut-Allier, Sioule, Chavanon and Plateau d’Aigurande (Kornprobst and Poulain 1972; Briand 1978; Briand and Gay 1978; Pin 1980; Burg *et al.* 1984, 1989; Quenardel and Rolin 1984; Delor *et al.* 1987; Faure *et al.* 1990, 1993, 2005; Duguet *et al.* 2004, 2007). D₂ is also recognized in the South of the Armorican Massif: Vendée (Burg 1981; Cannat and Bouchez 1986; Goujou 1992; Colchen and Rolin 2001); Champtoceaux complex and South Armorican coastal domain (Audren 1987; Turrillot *et al.* 2011). In all of these regions, D₂ concerns the LGU, the TPU and their equivalents in Vendée and Rouergue, and to a lesser extent, the UGU and the northern windows of the Para-autochthonous Unit of the French Massif Central: Dronne, Châteauneuf-la-Forêt, Plateau d’Aigurande, Sioule (see Figure 4.8). Moreover, it is worth noting that the D₂ event

never existed in the para-autochthonous Unit of the SE of the FMC (Cévennes, Southern Albigeois).

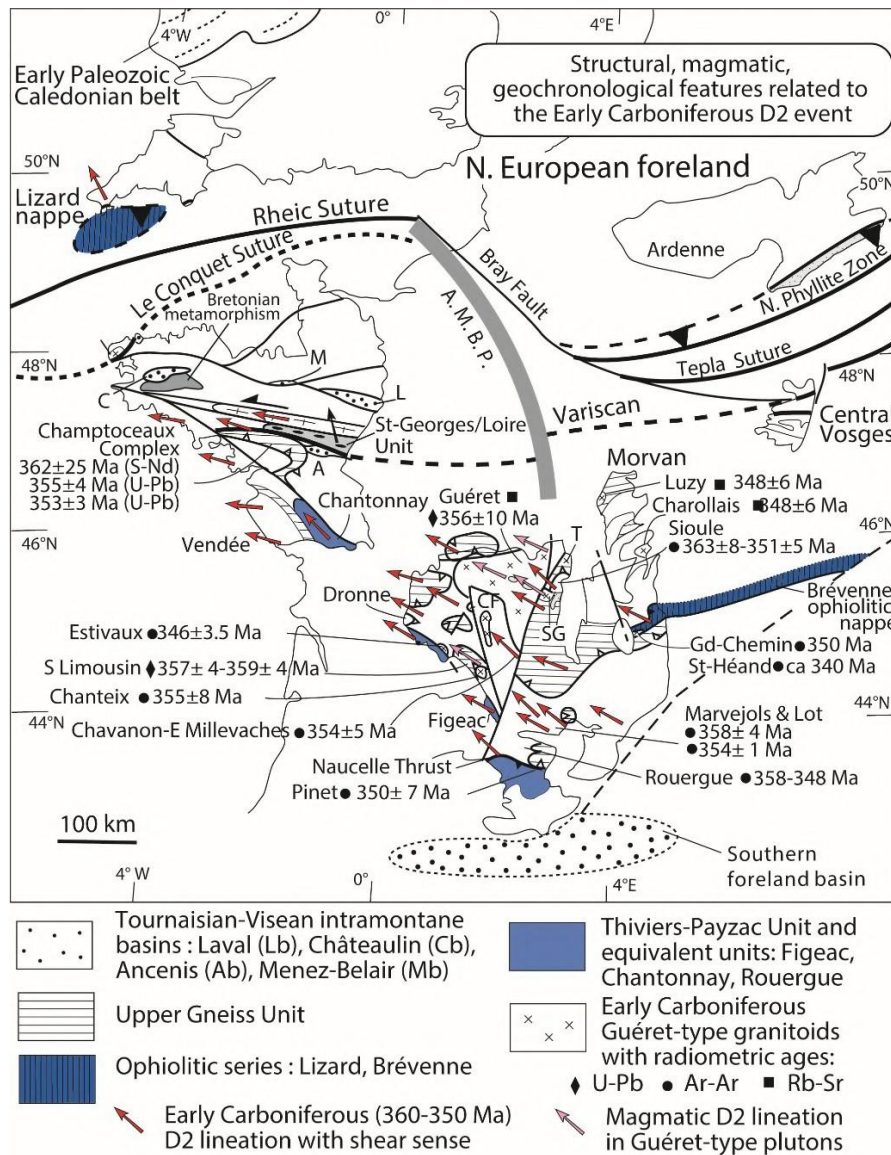


Figure 4.8. Structural map of the Armorican Massif, French Massif Central and the Vosges Massif locating the structural, magmatic and geochronological data of the D₂ event

Petrographically, the D₂ metamorphism is well recorded in aluminous protoliths; biotite-garnet-staurotide±kyanite micaschists are the most representative rocks of this metamorphism. Prograde Barrovian assemblage indicates thermo-barometric conditions of 600–750°C and 0.6–0.9 GPa (Triboulet and Audren 1988; Goujou 1992; Bellot and Roig 2007; Duguet *et al.* 2007; Do Couto *et al.* 2016; see Figure 4.9).

A systematically oriented NW–SE mineral and stretching lineation, associated with top-to-the-NW kinematics of the upper part is the characteristic structural element of this event (see Figure 4.8). Note that this orientation is incompatible with the tectonic model that favors a progressive spread of deformation from the north to the south, from the highly metamorphic mesozonal and catazonal domains of the UGU to regions with very low anchizonal metamorphism, such as the northern slope of the Montagne Noire and Viganais (Matte 1986).

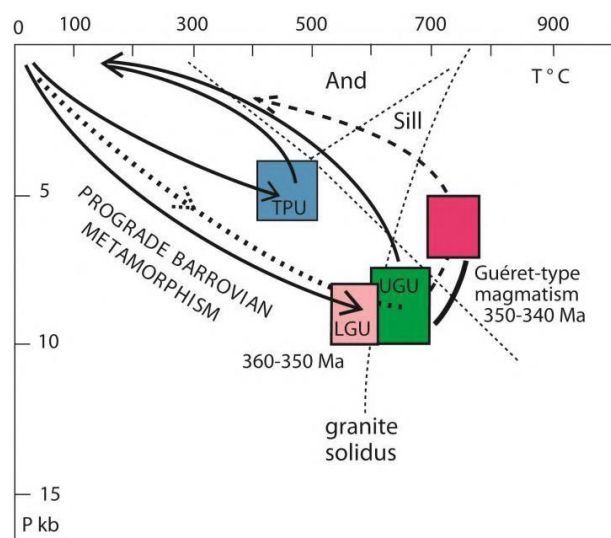


Figure 4.9. Synthetic pressure–temperature–time paths of the Upper Gneiss Unit (UGU), Lower Gneiss Unit (LGU) and the Thiviers-Payzac Unit (TPU) for the D_2 event

Although metamorphosed under low-pressure and low-temperature conditions (greenschist facies and lower amphibolite facies), the Brévenne ophiolites are ductilely deformed during their NWward thrusting upon the UGU, according to the SE–NW stretching lineation and related kinematics (Leloix *et al.* 1999; Faure *et al.* 2005). This syn-metamorphic deformation is ascribed to the D_2 event because it is contemporary with that observed in the other units of the French Massif Central. $^{40}\text{Ar}/^{39}\text{Ar}$ radiometric dating on biotite, muscovite and amphibole in the Upper Lot series give

ages between 358 ± 4 Ma and 340 ± 4 Ma (Costa 1989). Comparable ages, around 350–346 Ma, are measured on the biotites of the Pinet orthogneiss (Maluski and Monié 1988) and in the Monts du Lyonnais (Costa *et al.* 1993). Contemporary synfolial monazites developed in the Barrovian metamorphism provide identical U–Th–Pb chemical ages, within error bars, between 365 ± 5 Ma and 357 ± 4 Ma in southern Limousin (Melleton *et al.* 2009). In the Sioule series, monazite U–Th–Pb ages range from 363 ± 5 Ma to 351 ± 5 Ma (Do Couto *et al.* 2016). A $^{40}\text{Ar}/^{39}\text{Ar}$ age at 354 ± 5 Ma on biotite and muscovite was also obtained in the Chavanon series (Gébelin *et al.* 2007).

Furthermore, the biotite±cordierite Guéret-type syn-kinematic granites exhibit a preferred magmatic fabric, with an NW–SE lineation compatible with that ascribed to the D_2 event (Jover and Bouchez 1986). These plutons, dated at 355 ± 5 Ma (Zircon, U–Pb, ICP-MS) (Berthier *et al.* 1979; Cartannaz *et al.* 2007), were emplaced at the end of the D_2 event (Roig *et al.* 1996; Faure *et al.* 2005). These Tournaisian radiometric ages comply with the stratigraphic ones since the Guéret

granite and east of the Sillon Houiller, its equivalent, as the St-Gervais-d'Auvergne pluton, and its metamorphic country rocks of the Sioule series are unconformably covered by terrigenous and calcareous formations of the lower Viséan (345–340 Ma) or the upper Viséan (330 Ma) “Tufs anthracifères” volcano-sedimentary series. The same conclusion also stands in the Brévenne area, where rocks deformed and metamorphosed during D₂ are unconformably covered by the lower Viséan conglomerates and sandstones (Julien 1896; Leloix *et al.* 1999).

The D₂ event corresponds to the collision between Laurussia in the north and Gondwana in the south. The latter continent regained the Armorica continental strip after the Eo-Variscan orogeny. But in detail, the structural significance of the D₂ event is still poorly understood. An enigmatic “corner effect” has been proposed at the scale of the Iberian–Armorican orocline (Brun and Burg 1982; Burg *et al.* 1987). Since the E–W to NW–SE lineation is associated with a prograde metamorphism, it is difficult to interpret it as due to top-to-the-NW detachment faults that are, in fact, not recognized. On the England–France transect of the Variscan belt, the end of the Devonian corresponds to the period of closure of the Rheic Ocean and the beginning of the Variscan collision. The obliquity of the lineation with respect to the assumed trace of the Rheic suture under the Channel (see Figure 4.8) suggests an oblique component in the convergence. Syn-metamorphic ductile shearing is conspicuously developed in the North Gondwana margin, while in Armorica, it is located in the St-Georges-sur-Loire and Lanvaux units. Along sinistral strike–slip faults, such as the one controlling the pull-apart opening of the Ancenis basin, and possibly the Laval and Châteaulin basins, the brittle deformation dominates. The difference in rheological behavior between the Armorican and Gondwanan domains might be explained by the presence of a more rigid lithosphere in the Armorican domain

because it was already structured by the Neoproterozoic Cadomian orogeny. The Eo-Variscan suture of the Nort-sur-Erdre fault is also reworked as a strike–slip fault. At the lithospheric scale, everything happens as if the ductile shearing towards the west–northwest of the upper part could be interpreted as a lateral expulsion, or an escape of the Gondwana lithosphere towards the SE with respect to the Armorica micro-continent (see Figure 4.10 and Faure *et al.* 2017).

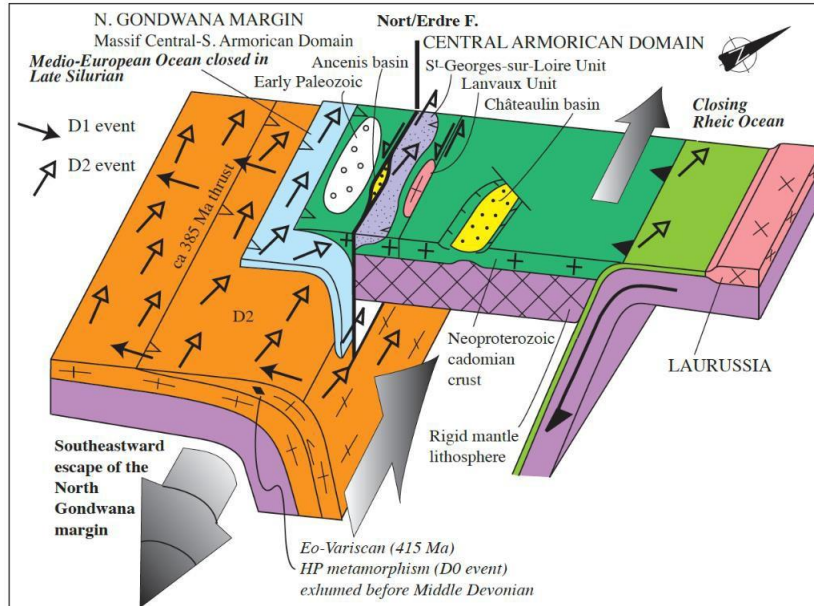


Figure 4.10. Interpretative geodynamic block diagram of the D_2 event in the Moldanubian and Armorican Domains. The syn-metamorphic ductile shearing is highly developed in the North Gondwana margin (according to Faure *et al.* 2017)

The main D_2 metamorphism is characterized by a long recognized inversion of metamorphic isograds (e.g. Jung and Roques 1936; Richard 1938; Roques 1941; Delorme *et al.* 1949; see review in Lardeaux 2014). This inverse metamorphism has been studied in the Upper Lot series (see Figure 4.11; Briand 1978; Briand and Gay 1978; Pin 1980; Burg *et al.* 1984, 1989) in the Sioule series (Grolier 1971; Grolier *et al.* 1980; see Figure 4.12), and in the Plateau d'Aigurande (see Figure 4.13, Rolin and Quenardel 1980; Quenardel and Rolin 1984). In all of these series, isograds appear inverted, with metapelites containing chlorite, biotite, garnet, staurotide and kyanite (\pm sillimanite) developing successively from bottom to top. Biotite-sillimanite \pm kyanite gneisses and migmatites still occupy the upper part of the pile. This inversion of

the isograds was initially interpreted as a consequence of post-metamorphic multi-kilometer scale recumbent folds. More recently, by analogy with the Himalayas, the inverse metamorphism has been considered to result from the thrusting of a hot upper unit on a colder lower one (Mattauer and Etchecopar 1976; Briand 1978; Burg and Matte 1978; Rolin and Quénardel 1980). It is this “ironing” effect that would be responsible, in the LGU, for the temperature decrease from the top to the bottom from the UGU basal thrust. However, it is important to note that the inversion of the biotite, garnet and staurotide isograds concerns only the LGU. In fact, in sericitic chloritosechists and micaschists, belonging to the para-autochthonous Unit, there is no geometric continuity of the inverse metamorphism. Moreover, the Barrowian metamorphism D_2 is dated to the Tournaisian (360 Ma), while in the SE part of the French Massif Central, the UGU overthrust upon the LGU is dated to the lower Viséan. In Marvejols, syn-kinematic granites in the thrust zone yield an age of 345 Ma (Pin and Lancelot 1982). The Barrowian metamorphism is associated with a top-to-the-NW shearing, whereas the UGU overthrusts upon the LGU is directed southward, as documented by the stretching lineation as observed in the Marvejols mylonites, for example (Faure *et al.* 1979; Pin 1979). Even if restricted to the LGU, the phenomenon of inverted metamorphism remains poorly understood. On a crustal

scale, D₂ could be interpreted as an eastward displacement of the para-autochthonous unit and its unknown substratum relative to the LGU and the UGU already structured during the Eo-Variscan orogeny. Therefore, it would be a type of underthrusting.

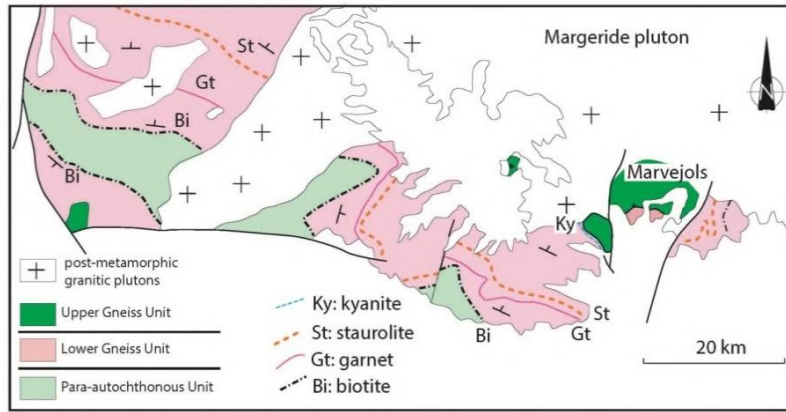


Figure 4.11. Structural map of the inverted metamorphic series in the Upper Lot Valley (redrawn according to Burg et al. 1984; and Briand et al. 2009). Since the biotite isograd is very close to the tectonic contact between the para-autochthonous and Lower Gneiss Units, the two contours were not separated

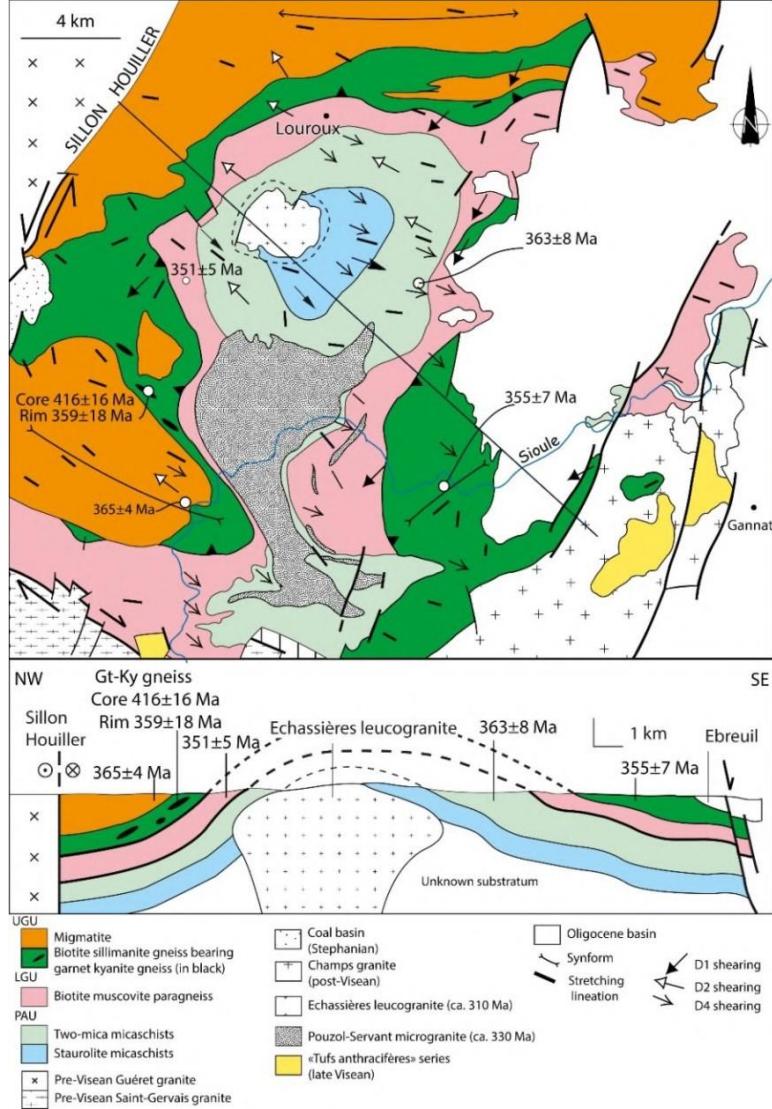


Figure 4.12. Structural map of the Sioule Series showing superimposed deformations and geochronological data (according to Faure et al. 1993; Do Couto et al. 2016). It should be noted that D_2 is contemporary with the Barrowian metamorphism of lower Carboniferous age, whereas D_4 is synchronous with the metamorphic retrogression in an extensional setting, and with the emplacement of the Echassières pluton (for the relationships between deformation and emplacement of the post-nappe plutons, see Faure 1995)

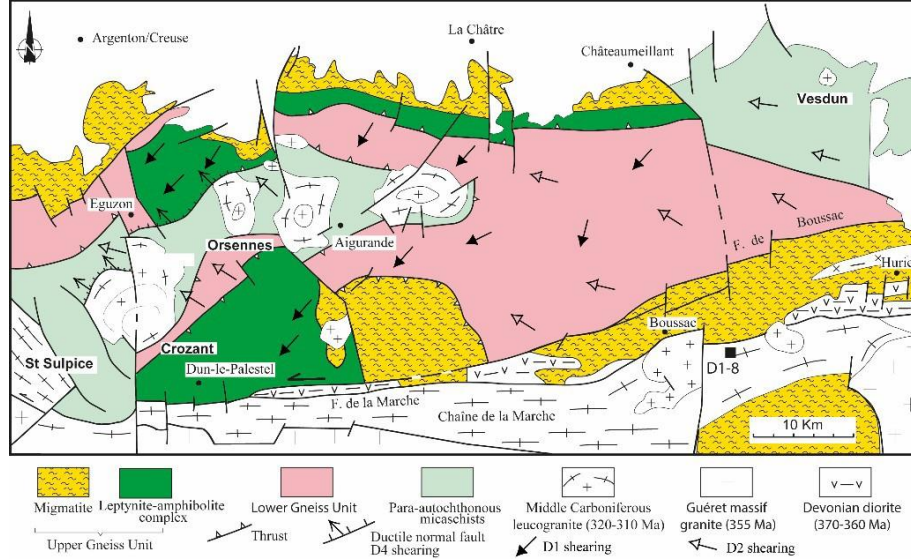


Figure 4.13. Structural map of the Plateau d'Aigurande (redrawn from Rolin and Quenardel 1980; Quenardel and Rolin 1984; Faure *et al.* 1990)

4.2.2.2. LP/LT event (D_3) 345–320 Ma

The D_3 event is the last tectono-metamorphic event that accompanied a thickening of the Gondwana margin in the Moldanubian domain, after the Variscan collision. It is located in the south of the French Massif Central (Cévennes and Montagne Noire), in the coastal Vendée, and probably in the Mouthoumet Massif and also in the Pyrenees (Soliva *et al.* 1989).

In the north of the French Massif Central, north of the Margeride pluton, rocks already structured and metamorphosed by early events, D_0 , D_1 , D_2 , are often retromorphosed in the greenschist facies during D_3 . In the North Limousin (Guéret massif), Sioule, Morvan and Vosges, the metamorphic series have already been exhumed since the lower Visean (see the previous section). The upper Visean “Tufs anthracifères” formation (see Figure 4.14) is a key marker for analyzing the “early” tectono-metamorphic evolution of the Variscan belt. This formation, well exposed in the Vosges and the NE part of the French Massif Central, is recognized in the Limousin only in the Bosmoreau-les-Mines coal basin (Bruguier *et al.* 1998). In the South of the Armorican Massif, turbiditic sandstones and pelites alternations seen in

Brétagne-sur-Mer in the coastal Vendée, ascribed to the Upper Visean, are severely folded and overturned towards the south. However, this structural context is closer to that of the southern domain of the Montagne Noire (see section 4.2.2.3) than to the Tufs anthracifères.

The compressive D_3 event is well developed in Cévennes, Châtaigneraie, Western Rouergue and South Millevaches. In Cévennes, the S_3 foliation, with a general slight north dip, bears a N–S to N50E striking stretching lineation L_3 , easily visible in the quartz-feldspar levels or the metaconglomerates. These planar-linear structures are at the origin of the “Himalayan model” (Mattauer and Etchécopar 1976). Kinematic criteria mainly indicate a shearing of the upper part towards the south (or towards the SW), even though opposite directions are also observed (Arnaud and Burg 1993; Faure *et al.* 2001; Montmartin *et al.* 2021). These

syn-metamorphic ductile deformations were contemporaneous with the stacking of the litho-tectonic units and the repetition of the quartz-feldspar and micaschist series (Brouder 1968; Mattauer and Etchécopar 1976; Arthaud and Matte 1977; Faure *et al.* 2001, 2009; Brouder *et al.* 2009). Extensional shearing coeval with the contact metamorphism develops around the late granitic plutons. It superimposes on the structures formed during the D₃ event.

Micaschists in the Southern Para-autochthonous Unit of the French Massif Central experienced a LP/MT metamorphism in the greenschist and lower amphibolite facies (Rakib 1996). From the SE to the NE, in the same pelitic sequence, the following zoneography is observed: (i) chlorite-chloritoid-albite; (ii) chlorite-biotite-muscovite-K-feldspar; (iii) garnet-chlorite-oligoclase; and (iv) garnet-biotite. The estimated thermo-barometric conditions were 450–580°C and 0.3–0.6 GPa, respectively (Rakib 1996). Despite the absence of chemical variations in index mineral, the existence of post-folial biotite and garnet suggests that the Cévennes Para-autochthonous Unit underwent a poly-metamorphism. The D₃ metamorphism, which is synchronous with crustal thickening, was partially erased by a post-nappe thermal event. Isotherms for this second event are oblique to the regional foliation and nappe tectonic contacts (see Figure 4.15, Faure *et al.* 2001). The origin of this late thermal phenomenon remains poorly understood. East of the Villefort fault in Cézarenque, where the post-nappe thermal event appears to be absent, the ⁴⁰Ar/³⁹Ar dates on biotite and muscovite indicate ages between 340 and 325 Ma, considered to be those of the D₃ metamorphism (Caron 1994; Montmartin 2021).

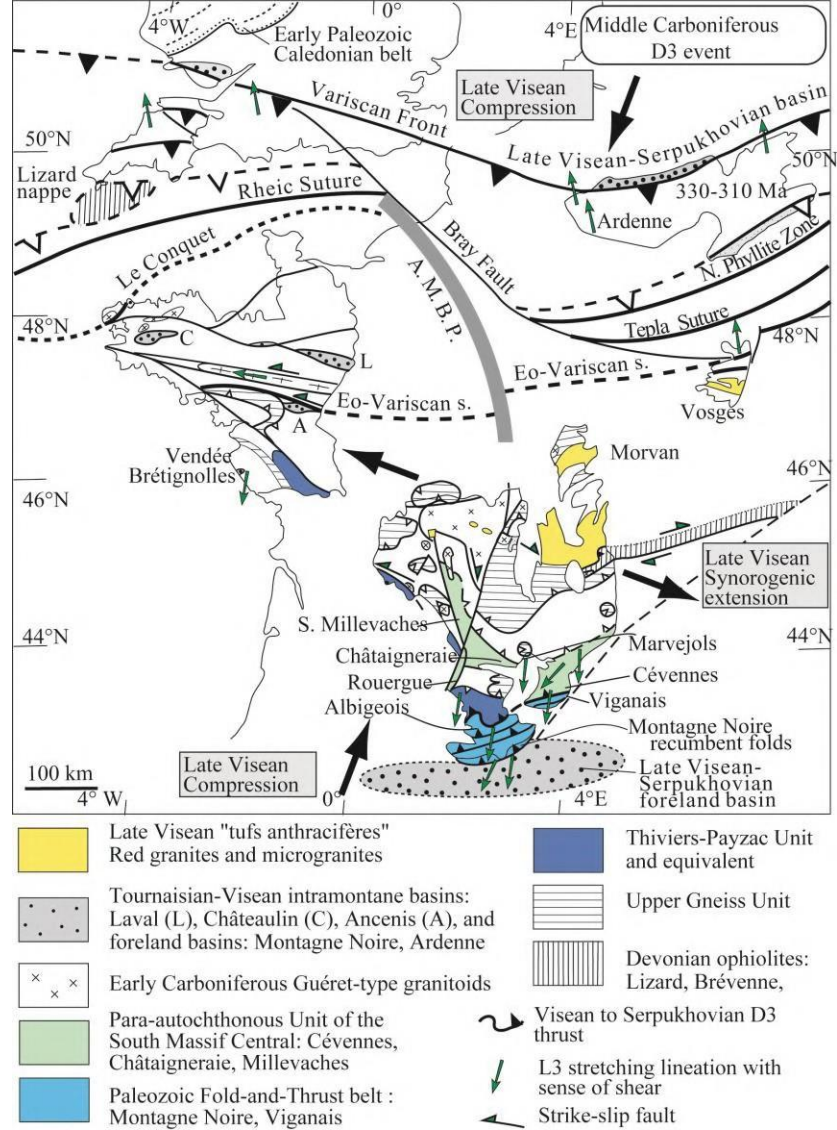


Figure 4.14. Structural map of the Armorican Massif, Massif Central and Vosges Massif locating the structural data of the D₃ event. In the south part of the French Massif Central, D₃ corresponds to a crustal thickening that spreads in space and time from the Visean (Marvejols) to the Serpukhovien (Montagne Noire). In the north part of the French Massif Central, the D₃ event is characterized by the deposition of "Tufs anthracifères" series and extensional or transtensional deformation

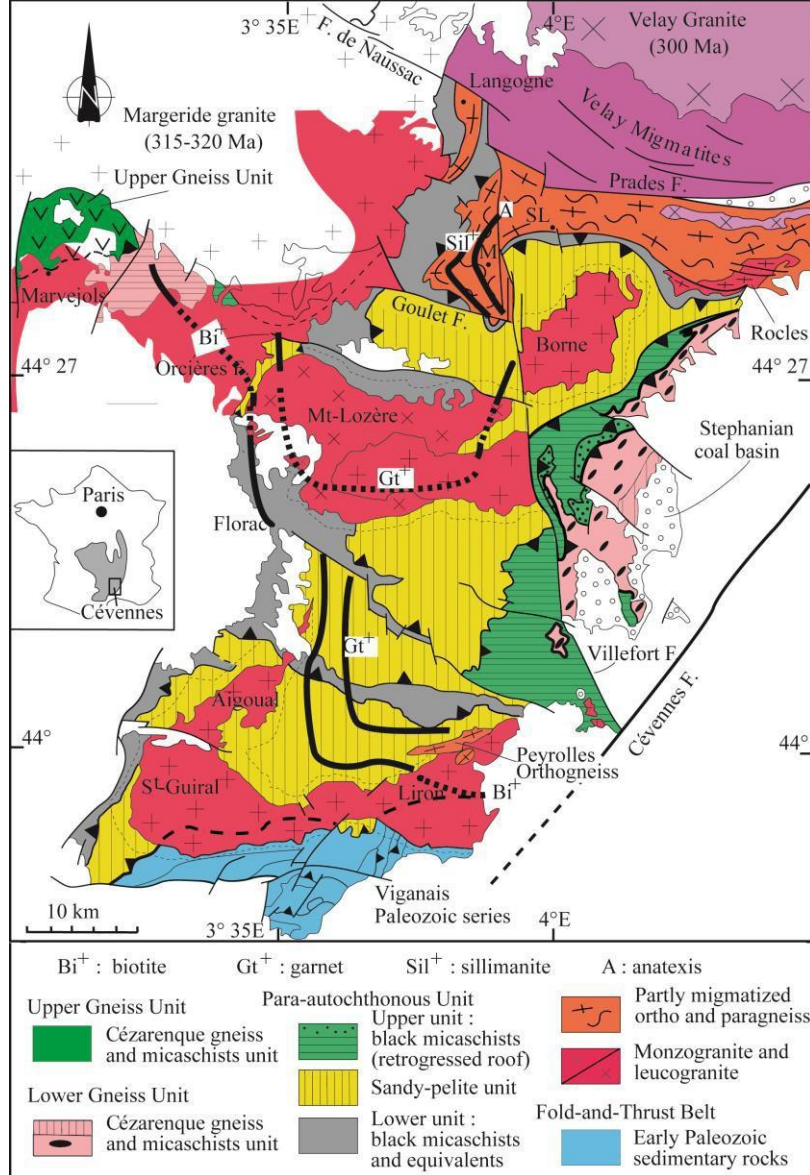


Figure 4.15. Structural and metamorphic map of the para-autochthonous unit of the Cévennes (according to Rakib 1996 and Faure *et al.* 2001)

In Albigeois and Viganais, a ductile top-to-the-south deformation was contemporary with a greenschist facies metamorphism (Guérangé-Lozes and Burg 1990). Raman spectrometry measurements of the graphite crystallinity indicate temperatures of approximately 300°C (Montmartin *et al.* 2021). The fold-and-thrust belt underlying the Cévennes Para-autochthonous Unit is virtually free of metamorphism with only prehnite-pumpellyite facies. In the Paleozoic series of the southern slope of the Montagne Noire, temperatures measured by the illite crystallinity, conodonts alteration color, fluid inclusions and the Raman spectrometry of carbonaceous matter crystallinity are around 250°C to 300°C (Engel *et al.* 1981; Guiraud *et al.* 1981; Wiederer *et al.* 2002; Montmartin *et al.* 2021). However, these anchizonal temperatures measured in the Paleozoic carbonaceous pelites of the southern slope of the Montagne Noire are related to the emplacement

of the granitic and migmatitic dome of the Axial Zone that superimposed on the D₃ event.

4.2.2.3. *MP to HP/LT (D₂ or D₃?) metamorphism in Montagne Noire*

Migmatites of the Montagne Noire Axial Zone dome and its micaschist envelope developed during an HT and LP metamorphism in the upper Carboniferous. These rocks present relics of one or more metamorphic events of higher pressure. Migmatites contain “restitute blocks” of orthogneiss, paragneiss, marbles, amphibolites and eclogites (Demange and Gattoni 1976; Demange and Léger 1980; Demange 1985; Demange and Jamet 1985; Alabouvette *et al.* 2003). Metapelites with biotite-garnet-staurotide and rare kyanite form the dome envelope (Bouchardon *et al.* 1979; Fréville *et al.* 2016).

The field context of eclogites is difficult to observe. In the Héric gorges, the protoliths are basalt and dolerite dykes, or gabbros intrusive in the ortho- and paragneisses. U–Pb dating of zircon cores of garnet amphibolites (retrogressed eclogite) gave Ordovician ages (Cocherie, unpublished data). In the Caroux, metabasites (eclogitic gabbro) are blocks included in migmatites globally dated around 320–315 Ma. Zircons of those eclogites yielded a U–Pb age around 311–312 Ma (Faure *et al.* 2014a; Whitney *et al.* 2015), which is incompatible with the anatexis age of the Axial Zone estimated at 325–315 Ma. Poorly constrained Sm–Nd dating of the same eclogite indicates an age of 357±9 Ma. In addition, monazites included in biotite in kinzigite gave chemical ages of 357±13 Ma and 352±10 Ma. Thus, in the current state of knowledge, the age of early metamorphism in the Montagne Noire would be closer to the D₂ event than D₃, but as in other points of the Variscan belt, these geochronological constraints remain to be refined.

This high to medium pressure and medium temperature metamorphism is associated with isoclinal folds ranging from centimetric- to metric-scale. In the

Axial Zone, the existence of penninic-style kilometric-scale recumbent folds has been proposed (Demange 1975; Alabouvette *et al.* 2003). If their existence is confirmed, these early syn-metamorphic structures could correspond to a deep stage, slightly older than that which formed the Serpukhovian to Bashkirian recumbent folds which developed in the southern slope of the Montagne Noire, but belonging to the same D₃ event (for recent information, see Faure *et al.* 2010a, 2014a; Faure 2019).

From the Upper Visean, around 330 Ma, the poly-metamorphic Moldanubian domain of the Variscan belt underwent several episodes of crustal melting characterized by migmatitic dome formation and several generations of granitoids accompanied by ductile deformations, presented in Chapter 2 of Volume 2.

ACHIEVEMENTS AND PROBLEMS OF THE MOL DANUBIAN DOMAIN

The brief overview of early deformations and metamorphisms recognized in the Moldanubian domain of the French Massif Central, the southern part of the Armorican Massif and Central Vosges set out in section 4.2 clearly shows the noticeable progress made since the 1980s. This includes the following, in particular.

- 1) The existence of nappes responsible for the thickening of the domain.

2) The presence of several tectono-metamorphic events.

3) The absence of a Neoproterozoic basement formed during the Cadomian orogeny, unlike in the central and northern Armorican domain. In the South Armorican region, which belongs to the Moldanubian domain, the Mauges nappe overthrusts the Eo-Variscan ophiolitic suture and the metamorphic units of the Champtoceaux nappe complex. Mutatis mutandis, this architecture is similar to the Austro-Alpine Dent Blanche nappe that overthrusts the “Schistes Lustrés” with ophiolites of the Western and Eastern Alps. Unlike the Alps, but in a similar way to the Himalayas, in the Moldanubian domain of the French Massif Central there is no Cadomian basement, either in the substratum or in the nappes. However, this does not preclude the existence of Neoproterozoic sedimentary or magmatic rocks, the deformation of which took place during the Variscan orogeny.

4) The abundance of orthogneiss derived from alkaline or calc-alkaline porphyritic granites. These granitoids, emplaced partly in the Lower Cambrian and mainly in the Ordovician, represent a context of pre-orogenic rifting (see discussion in Chapter 3).

5) The leptyno-amphibolite complexes with bimodal magmatism of Lower Ordovician age are characteristic of a very thin passive continental margin of the

continent–ocean transition type (see discussion in Chapter 3, and also Faure (2020, 2021)).

6) The evolution of the French Moldanubian domain can be explained by a polyorogenic evolution with an early cycle “Eo-Variscan” and a Variscan cycle *stricto sensu*.

7) The *Variscans* evolution is diachronous in the French Massif Central. It has a north–south polarity since, on the one hand, the Visean–Bashkirian D₃ event does not exist in the north of the French Massif Central where the D₂ event is sealed by the sedimentary and volcanic deposits of the Visean, in particular the “Tufs anthracifères” formation. On the other hand, in the southern part of the French Massif Central, the D₃ metamorphic and tectonic structures become younger and younger towards the south. In Marvejols, pegmatite boudins in the UGU/LGU contact are dated from the Lower Visean (345±10 Ma, Pin and Lancelot 1982), while in the Montagne Noire, syn-sedimentary folds and thrust formed in a Serpukhovian–Bashkirian (320–315 Ma) olistostrome.

However, several points remain unresolved.

1) The precise age of the high-pressure metamorphism, around 390 Ma or later, remains to be established.

2) The age of the tectonic contacts between the different units (UGU, LGU, PAU) is unknown, except in the Marvejols area (at approximately 345 Ma). It is likely that even for a given contact, its age varies from one region to another.

3) The tectonic and geodynamic significance of the D₂ event and the NW–SE lineation are not clearly established.

4.3. Metamorphisms and deformations in the Saxothuringian domain

In France, the Saxothuringian domain is limited to the Léon block in the NW of the Armorican Massif (see Figure 4.1). It is now well accepted that this block corresponds to a stack of ductile and syn-metamorphic nappes thrust from the south to the north then folded into a foliation antiform (Rolet *et al.* 1994; Faure *et al.* 2005, 2010b; see Figure 4.16). From the top to the bottom of the structure, the following units are recognized: (i) the Upper Unit formed by Paleozoic sedimentary rocks; (ii) an ophiolitic nappe; (iii) a Lower Unit composed of paragneisses and micaschists containing eclogite boudins retrogressed into amphibolites; and (iv) an autochthonous or para-autochthonous lowermost unit formed by orthogneisses and

their paraderived metamorphic country rocks. The rocks of the latter unit are highly migmatized.

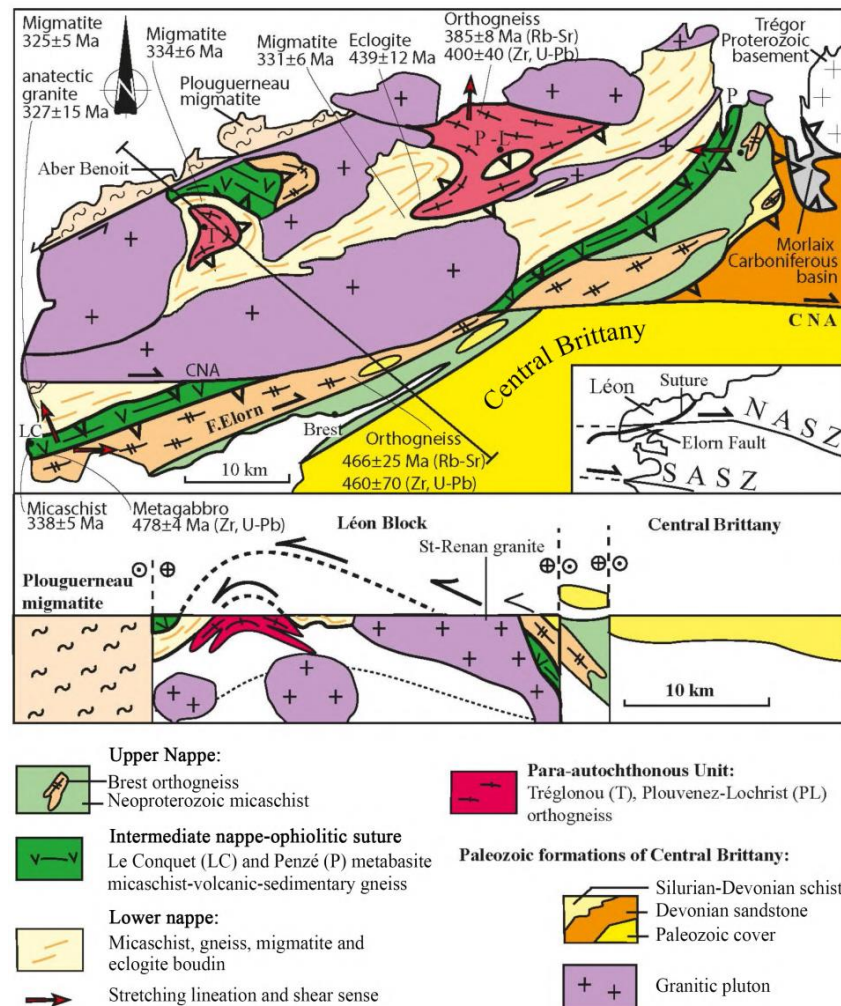


Figure 4.16. Structural map and cross-section of the Léon Block (from Faure *et al.* 2010b)

The thermo-barometric conditions of the eclogitic metamorphism are estimated to be 650°C–700°C and 1.4–1.5 GPa, respectively (Cabanis and Godard 1987). Zircons in these rocks yield a U–Pb age of 439±12 Ma (Paquette *et al.* 1987).

Metapelites of the lower unit and the ophiolitic nappe exhibit biotite-garnet-staurolite paragenesis characteristic of an MP/MT metamorphism, but thermo-barometric data are not available. This Barrowian metamorphism is contemporaneous with a ductile deformation with a top-to-the-north shearing, associated with the nappe thrusting. A monazite U–Th–Pb chemical dating of a micaschist from the ophiolitic nappe at Le Conquet provides an age of 338±5 Ma for this event (Faure *et al.* 2010b). Post-nappe anatexis superimposed on the nappe stack is dated, also on monazites present in migmatite leucosomes, between 334±6 and 325±5 Ma.

By analogy with the German Saxothuringian domain, the Léon block is interpreted as a micro-continent subducted to the south and then exhumed during the Variscan tectonics. Contrary to a previous hypothesis (Balé and Brun 1986), the stack of syn-metamorphic nappes of the Léon block, thrust from south to north, cannot be considered a fragment of the Moldanubian domain of Southern Armorica translated to the north by the Elorn fault.

4.4. Metamorphisms and deformations in the eastern Variscan branch

This branch of the Variscan belt contains several metamorphic massifs (see Figure 4.1). Since the correlations with the different units of the main branch are not yet well established, this area is presented separately. Two parts can be distinguished: (i) the Maures–Tanneron massif and the Corsica–Sardinia Massif in which the Alpine superimposition is low or absent; and (ii) the Variscan basement of the Alps, exposed in the external and internal crystalline massifs and the Briançonnais zone in which the Alpine deformation and metamorphism superimposed to varying degrees on the Paleozoic events. Syn-metamorphic deformations of the Sesia zone and the Dent Blanche klippe are not considered here.

4.4.1. Massifs free of alpine superimposition: Maures–Tanneron and Corsica–Sardinia Massifs

4.4.1.1. Maures–Tanneron Massif

Like the other Variscan massifs, this group consists of polymetamorphic rocks, migmatites and granitoids. Its tectono-metamorphic evolution can be subdivided into two major stages: (i) an early stage responsible for crustal thickening; and (ii) a late stage contemporaneous with a major crustal melting, associated with the

emplacement of migmatitic domes and granitic plutons. This chapter is limited to presenting the different metamorphic and structural events of the early stage.

4.4.1.1.1. HP/MT event (450 < – < 360 Ma)

Basic HP rocks, derived from various protoliths: basic lavas, metagabbros, metagranites and peridotites, are found in the central Maures (La Croix-Valmer, Cavalaire or Garde-Freinet units, see Figure 4.17), and have been known for a long time (Seyler and Boucarut 1978; Bard and Caruba 1981, 1982; Crévola and Pupin 1994; Bellot *et al.* 2003, 2010; Bellot 2005; Rolland *et al.* 2009; Schneider *et al.* 2014). These formations are quite comparable to the leptyno-amphibolite complexes of the Moldanubian domain. In comparison with the HP rocks of northern Sardinia, thermo-barometric conditions of 600–700°C and 1.5–2 GPa are proposed (Schneider *et al.* 2014; references included). Small masses of serpentized peridotites are also known. The garnet-spinel peridotite of La Croix Valmer is an ultrabasic magmatic cumulate crystallized at low pressure and metamorphosed under HP conditions of approximately 1.6 and 1.8 GPa and 850°C (Bouloton *et al.* 1998) or even under ultra-high pressure of 2–2.8 GPa (Bellot *et al.* 2010).

In the eastern Maures, eclogites with kyanite-sapphirine-hypersthene, frequently retromorphosed into HP granulites and amphibolites, form blocks (probably restites) within migmatites (Bard and Caruba 1982; Bellot 2005). Zircons extracted from eclogites of the Eastern Maures yield U–Pb ages at 452±8 Ma and 431±4 Ma, the interpretation of which in terms of protolith age, metamorphism or mixture between the two phenomena is delicate (Moussavou 1998; Schneider *et al.* 2014).

Furthermore, orthogneisses metamorphosed in the HP granulite facies are recognized in the Central Maures (Barral granite in the orthogneiss of Bormes; Maluski and Gueirard 1978) and in the Tanneron massif (Bois de Bagnols granite; Crévola and Pupin 1994; Demoux *et al.* 2010). In the latter, monazites of a granulitic orthogneiss range in age from 441±1 Ma to 410 Ma (Demoux *et al.* 2008). A U–Th–Pb chemical age on monazite at 410±10 Ma has also been obtained for a migmatitic orthogneiss of Tanneron (Oliot *et al.* 2015), which may correspond to the age of HP metamorphism. A migmatitic paragneiss from the eastern Maures provides a U–Th–Pb age on monazite at 382±11 Ma, interpreted as that of a Devonian metamorphism (Oliot *et al.* 2015).

There are no structural data to characterize the contemporary tectonic evolution of the HP metamorphism. According to most authors, it is associated with a Silurian or Lower Devonian subduction towards the east. After an oceanic subduction, the central and western Maures continental block was buried under the eastern Maures

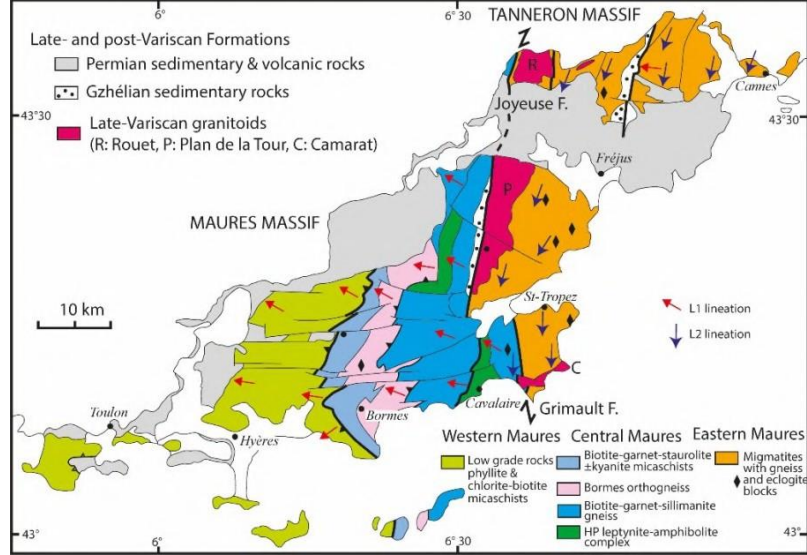


Figure 4.17. Structural map of the Maures–Tanneron Massif (redrawn according to Crévola and Pupin 1994; Bellot 2005; Corsini and Roland 2009; Rolland *et al.* 2009; Schneider *et al.* 2014; Oliot *et al.* 2015)

4.4.1.1.2. Barrowian MP/MT Event (360 < – < 340 Ma)

This is the major tectono-metamorphic phase of the Maures–Tanneron Massif (D_1 - M_1 by Schneider *et al.* 2014). From the Toulon–Hyères region to the Grimaud–Joyeuse fault, the succession of chlorite, biotite, garnet, staurolite, kyanite and sillimanite isograds is recognized from west to east (Seyler and Boucarut 1978; Crévola and Pupin 1994; Buscaïl 2000; Bellot 2005). The current geometry of foliation and west dipping isograds (see Figures 4.17 and 4.18) results from “return” folds (phase D_2 of Figure 4.18). The Rouet antiform complies with a migmatitic and granitic dome.

The Hermitan granite, dated to 338 ± 6 Ma (Moussavou 1998), occupies the center of the Rouet dome. But this geometry results from two deformations. The antiformal structure is posterior or contemporary with back thrusts and post-folial folds with east vergence, attributed to the upper Carboniferous (about 330 and 310 Ma).

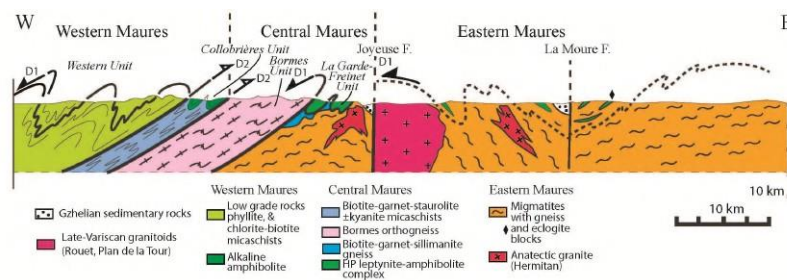


Figure 4.18. Synthetic cross-section of the Maures–Tanneron segment (redrawn according to Oliot *et al.* 2015). East-directed back thrusts and back folds distort the early structures (D_1) and precede the granite-gneiss dome

Amphibolites of the Central Maures give thermo-barometric constraints of 700–750°C and 0.7–0.8 GPa, respectively (Bellot *et al.* 2010). The foliation synchronous

of the Barrowian metamorphism contains an N–S or NNW–SSE mineral and stretching lineation, associated with top-to-the-N or -NW shearing (Vauchez and Buffalo 1988; Corsini and Rolland 2009; Rolland *et al.* 2009). This major event is dated to around 360 Ma by syn-metamorphic monazites in paragneisses and micaschists from central and eastern Maures (Oliot *et al.* 2015). The Bormes orthogneiss gives U–Pb ages between 345 and 339 Ma (Moussavou 1998) interpreted as those of the syn-tectonic metamorphism. Furthermore, the Rb–Sr dating of an amphibolite gives an age of 348 ± 7 Ma, also interpreted as the metamorphic age (Innocent *et al.* 2003).

Corsica–Sardinia Massif

Paleogeographic reconstructions prior to the opening of the Gulf of Lions show analogies between the Maures–Tanneron Massif and northern Sardinia (see Figure 4.19). In particular, the Central Maures are lithologically and structurally similar to the metamorphic internal domain of the nappe zone in the NW part of Sardinia. The Grimaud–Joyeuse fault is correlated to the Posada–Asinara shear zone, and the NE Sardinia is a migmatitic dome, with blocks of eclogitic relics, comparable to the Eastern Maures (Carmignani *et al.* 1994; Elter *et al.* 1999; Rossi *et al.* 2009; Faure *et al.* 2014b). Thus, the Variscan segment of Corsica–Sardinia represents the eastern extension of the Maures–Tanneron segment.

In Corsica, Variscan complexes are dispersed into 1–20 km-long septa within the Corsica–Sardinia batholith (see Figure 4.20; Durand Delga and Rossi 1991; Rossi and Cocherie 1991; Rossi *et al.* 1995). These mega-enclaves can be divided into four groups: (i) the Argentella and Agriates domain characterized by a Cadomian

neoproterozoic basement, unconformably covered by a Paleozoic sedimentary series going from Ordovician to Carboniferous; (ii) a domain located south of Argentella (West of Ajaccio) formed of small migmatitic septa; (iii) the Zicavo, Solenzara–Fautea and Porto Vecchio septa in which the metamorphic rocks are well developed; and (iv) the Belgodere domain in the NE of the Variscan Corsica also characterized by partially migmatized metamorphic rocks. Only the two latter ensembles are presented here. The Argentella domain does not contain Variscan metamorphic rocks. It consists of Neoproterozoic metamorphic rocks unconformably covered by Ordovician and Silurian sedimentary rocks (sandstones, conglomerates) and Upper Devonian and lower Carboniferous limestones and sandstones. The Ajaccio migmatitic ensemble, dating back to the Visean (~345 Ma, Li *et al.* 2014), relates to the late Variscan events (see Chapter 2 of Volume 2).

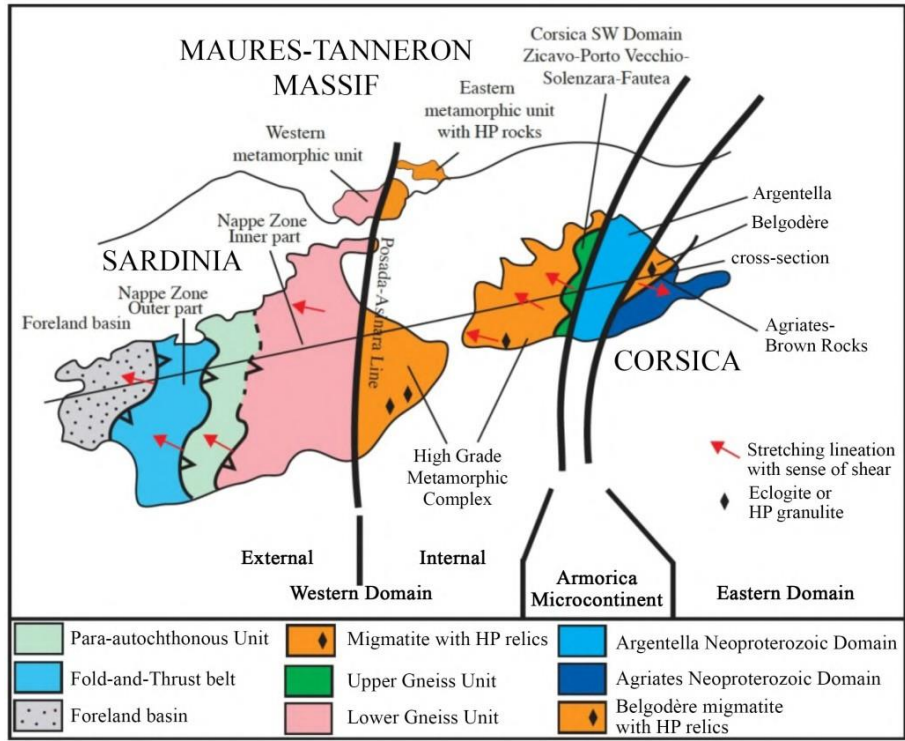


Figure 4.19. Structural correlations between the Maures–Tanneron and Sardinia–Corsica Massifs, the Meso-Cenozoic formations of Corsica (see Figure 4.20) and Sardinia are not represented (according to Faure et al. 2014b). The position of Corsica and Sardinia at the end of the Variscan orogeny was restored by considering a counterclockwise Cenozoic rotation of 55° of the whole massif

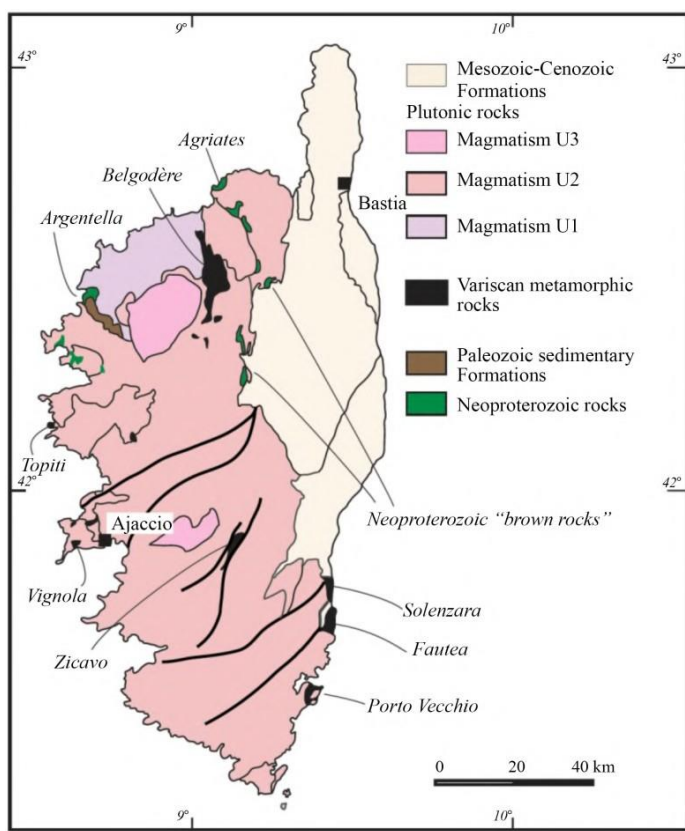


Figure 4.20. Structural map of Variscan Corsica showing the main Variscan and pre-Variscan Septa in the Corsica Batholith (according to Rossi *et al.* 1995; Faure *et al.* 2014b)

The *Zicavo septum* contains various lithologies: augen orthogneiss, leptyno-amphibolite alternations with some serpentinite blocks, biotite-garnet-staurolite±kyanite micaschists, metaconglomerates, biotite-muscovite-garnet-staurolite micaschists and black quartz sericite-chlorite±graphite micaschists. Paragenetic differences lead to distinguishing three units separated by tectonic contacts: (i) a lower para- and orthogneissic ensemble; (ii) an intermediate leptyno-amphibolite complex; and (iii) a higher metapelitic ensemble (Faure *et al.* 2014b). Syn-metamorphic structures indicate that the *Zicavo septum* was deformed by at least two tectono-metamorphic phases. The first one, characterized by NE–SW striking stretching lineations and top-to-the-SW kinematics, is responsible for the tectonic stacking (see Figures 4.19 and 4.21 with lineation strike restored

considering the Cenozoic rotation of the Corsica–Sardinia block). The second deformation, observed only in the upper ensemble, is characterized by a NW–SE-oriented mineral and stretching lineation and top-to-the-SE kinematics. There are no precise thermo-barometric data, but the biotite-garnet-staurolite±kyanite paragenesis present in the metapelites of the leptyno-amphibolite unit indicates metamorphic conditions in the amphibolite facies (P ~0.8 GPa, T ~600°C; Figure 4.19; Thévoux-Chabuel *et al.* 1995). The second deformation, contemporary with conditions of the greenschist facies, developed in a retrograde context which may correspond to the exhumation of metamorphic series. The monazite of a biotite-garnet micaschist with the NW–SE lineation gives a chemical U–Th–Pb age of 336±5 Ma, interpreted as the age of the second deformation phase (Faure *et al.* 2014b).

The *Porto Vecchio septum* consists of a lower orthogneissic complex, tectonically overlain by a leptyno-amphibolite complex and biotite-muscovite-garnet-sillimanite paragneisses containing kyanite relics and cordierite-spinel reaction corona (Libourel 1985; Giacomini *et al.* 2008). This septum presents great structural analogies with that of *Zicavo*. A first ductile deformation with top-to-the-SW shearing under MP/MT conditions is followed by a top-to-the-E or -SE ductile shearing. But unlike the *Zicavo septum*, the *Porto Vecchio* one shows a HT metamorphism coeval with an early migmatization, contemporary with the second event.

The *Solenzara-Fautea septum* consists essentially of migmatites dated to around 345±3 Ma (Li *et al.* 2014). The migmatitic foliation defines a dome structure intersected by granitic intrusions. These anatexites contain blocks of pyrigarnite, often retromorphosed into amphibolites, kyanite-garnet HP granulite, also retromorphosed into amphibolite, gneiss and marble (Giacomini *et al.* 2008), but no early structure is recognized. The eclogites contain zircon with a core dated between 484 Ma and 440 Ma, interpreted as the age of the protolith. In contrast, zircon rims give ages of 418±6 Ma and 407±6 Ma, interpreted as close to the age of the HP metamorphism (Rossi *et al.* 2009). In addition, the zircons of the HP granulites give an age of 361±3 Ma, also considered as that of the HP metamorphism (Giacomini *et al.* 2008). However, this age may also correspond to that of the MP–MT metamorphism developed during the retrogression of the HP rocks.

The *Belgodère septum* is the largest of the Variscan Corsican domain. It also consists of metatexites dated between 345 ± 1 Ma and 338 ± 1 Ma (Paquette *et al.* 2003; Li *et al.* 2014). In this septum, the foliation exhibits an antiformal structure overturned to the east. It contains various blocks of orthogneiss, paragneiss, biotite-muscovite-garnet-staurolite micaschists, metaconglomerates and amphibolites (some associated with acidic gneisses to constitute a leptyno-amphibolite complex). Garnet

amphibolites show plagioclase-amphibole-epidote symplectite textures around garnet, indicating these rocks have undergone an early eclogitic stage (Palagi *et al.* 1985; Lardeaux *et al.* 1994). But there are no detailed thermo-barometric, structural or geochronological data on these pre-migmatitic metamorphic rocks.

In conclusion, the Corsica–Sardinia segment of the Variscan belt presents numerous analogies with the architecture of the Moldanubian–Armorican segment described in sections 4.2 and 4.3 (see Figure 4.21). A stack of tectonic units, displaced to the south (in the present geometry) with S to N metamorphic increase is recognized. The Argentella domain, characterized by a Cadomian Neoproterozoic basement, is similar to the Armorica block, and the Belgodère area might be correlated with the Saxothuringian domain (see detail in Faure and Ferrière 2022).

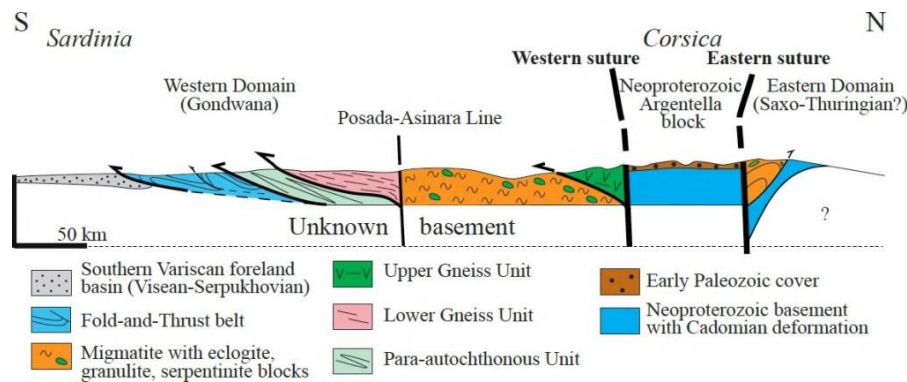


Figure 4.21. Schematic cross-section of the Corsica–Sardinia Massif. For clarity, the various generations of granites have not been represented (according to Conti 2001; Rossi *et al.* 2009; Faure *et al.* 2014b)

4.4.2. Alpine Variscan substratum

4.4.2.1. External crystalline massifs

These massifs are mainly formed by migmatites and numerous granitic plutons of late Carboniferous age. However, metamorphic MP/MT or LP/LT series are also identified. Furthermore, the migmatites contain blocks of orthogneiss, eclogite or HP granulite, often retromorphosed to amphibolites.

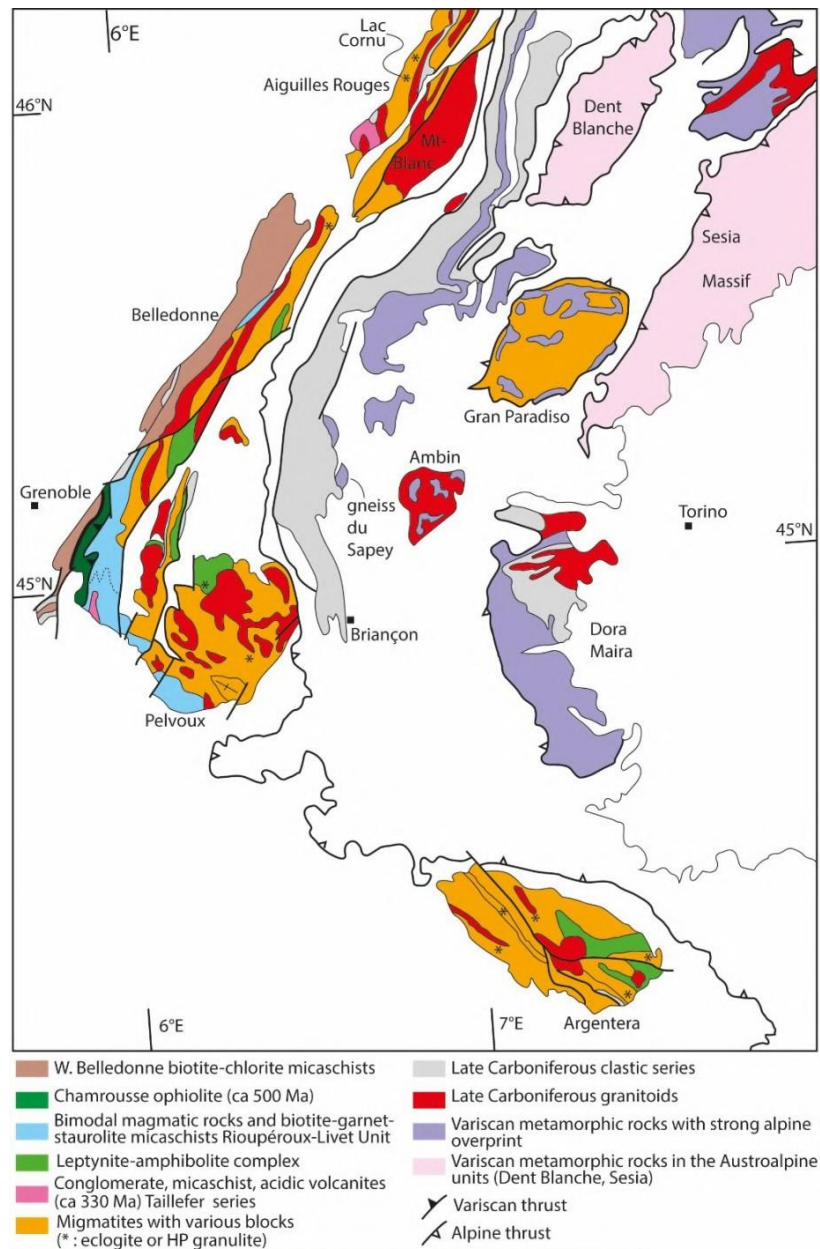


Figure 4.22. Structural map of the Variscan basement in the Alps (redrawn according to Fernandez *et al.* 2002; Guillot *et al.* 2009; Ballèvre *et al.* 2018; Fréville *et al.* 2018)

The Belledonne–Pelvoux massifs have the greatest lithological diversity. They have been the subject of numerous studies (Carme 1965; Carme and Pin 1987; Ménot 1988; Ménot *et al.* 1988; Guillot and Ménot 1999, 2009; Fernandez *et al.* 2002; Fréville *et al.* 2018, 2022; see Figure 4.22). The Belledonne massif is subdivided into three parts: (i) the western domain (or “*série satinée*”) formed of

sandstone-pelite alternations of Ordovician age, deformed by upright folds with NE–SW striking axes, and metamorphosed in greenschist facies conditions; (ii) the central domain formed by several metamorphic units detailed in the next paragraph; and (iii) the eastern domain that also includes the Grandes Rousses and the Pelvoux massifs, essentially consisting of gneisses and migmatites. The boundaries between these three domains are ductile or brittle strike–slip or thrust faults, active from the late Variscan to the Alpine times.

The central domain of Belledonne consists of at least two tectono-metamorphic units tectonically stacked, from top to bottom: (i) the Chamrousse ophiolitic nappe with mafic protoliths dated around 500–496 Ma (Pin and Carme 1987; Ménot *et al.* 1988). These ophiolites are weakly metamorphic, but have preserved a pre-orogenic oceanic metamorphism (Guillot *et al.* 1992); (ii) the Rioupéroux-Livet unit is a bimodal volcano-sedimentary and plutonic ensemble with a partially migmatitic base. In addition, the Taillefer volcano-sedimentary series, paleontologically dated to the Upper Visean, and free of MP/MT metamorphism unconformably covers the two underlying units (Carme 1965).

Petrological and structural studies of the Rioupéroux-Livet unit have documented its polymetamorphic and polyphase character. A biotite-garnet-kyanite MP/MT metamorphism characterized by thermo-barometric conditions of $427\pm 16^\circ\text{C}$ and 0.68 ± 0.1 GPa is contemporaneous with the tectonic emplacement of Chamrousse ophiolites upon the Rioupéroux-Livet unit (Fréville *et al.* 2018; see Figure 4.23). An amphibole yields an $^{40}\text{Ar}/^{39}\text{Ar}$ age of 376 ± 7 Ma (Guillot *et al.* 2009). A second biotite-garnet-staurolite plurifacial metamorphism (D_1 – D_2) with a top-to-the-NE ductile shearing represents the main event. The pressure peak during D_1 is not settled; D_2 developed under $600\pm 14^\circ\text{C}$ and 0.58 ± 0.66 GPa conditions (Fréville *et al.* 2018). Monazite crystallized during this Barrowian metamorphism gives an ICP-MS age of 337 ± 7 Ma. The deformation continued in an isothermal decompression context up to pressures of 0.7 ± 0.2 GPa (see Figure 4.23). Aluminous rocks of the lower part of the Rioupéroux-Livet unit underwent the onset of anatexis, giving rise to sillimanite-cordierite leucosomes. In northern Belledonne, migmatites, dated on zircon single grain at 326 ± 3 Ma, contain retromorphosed eclogite blocks. Initially dated to 395 ± 2 Ma, these rocks have recently been re-examined. Zircons yields ICP-MS ages between 332 and 326 Ma, while rutile gives an age of 340 ± 11 Ma (Paquette *et al.* 1989; Jacob *et al.* 2021 and work in progress).

The Pelvoux massif is essentially composed of migmatites and granites, probably structured as a metamorphic dome. However, orthogneiss blocks derived from Ordovician granites and garnet amphibolites, probably derived from eclogites, are recognized in migmatites (Grandjean *et al.* 1996; Fréville *et al.* 2022; references included).

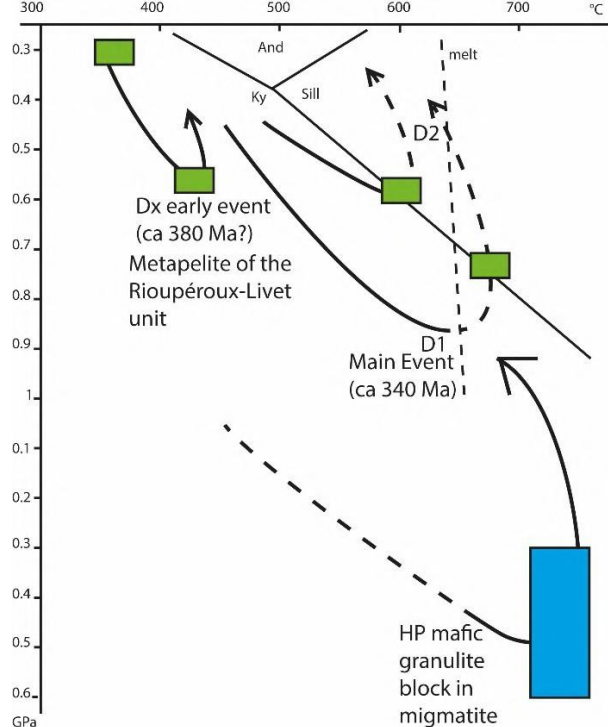


Figure 4.23. *P–T* paths established for the Rioupeyroux-Livet in the Belledonne Unit (Fréville *et al.* 2018) and HP granulites in migmatites of Belledonne NE (Jacob *et al.* 2021)

The Mont Blanc–Aiguilles Rouges Massif is similar to that of Belledonne. Volcano-sedimentary formations of Viséan age, metamorphosed in greenschist facies, are comparable to the Taillefer series, but because of Alpine tectonics, the foliations are turned vertical and lithological boundaries are deformed, so that it is not possible to assess the unconformable or tectonic nature of the contact. However, these series overlie a set of micaschists, gneisses and migmatites intruded by late Carboniferous granites, yielding zircon ages at 303 ± 2 Ma and 307 ± 2 Ma for the Mont Blanc and Vallorcine granites, respectively. The migmatites and associated

anatectic granites are dated at 320 ± 1 Ma and 332 ± 2 Ma for the Pormenaz and 331 ± 2 Ma for the Montées Péliissier granites, respectively (Bussy *et al.* 2000). Migmatites contain blocks of orthogneiss, marble, ultrabasites, garnet amphibolites and eclogite. Eclogite formation conditions in Lac Cornu are estimated at 780°C and 1.1 GPa (Liégeois and Duchesne 1981). A U–Pb zircon age of these eclogites at 453 ± 3 Ma is interpreted as that of basic protolith magmatism (Paquette *et al.* 1989). Biotite-garnet-staurolite-kyanite (or sillimanite) metapelites crystallized under 650°C and 0.4–0.6 GPa conditions (von Raumer and Bussy 2004). Monazites crystallized during this Barrovian metamorphism yield ICP-MS U–Pb age of 327 ± 2 Ma (Bussy *et al.* 2000). In addition, other monazite grains are dated by the U–Th–Pb chemical method to 440 Ma (Schulz and von Raumer 2011). The structural analysis of the Aiguilles Rouges massif shows the existence of east-directed syn-metamorphic ductile thrusting, accompanied by a dextral shearing component (Dobmeier 1998; von Raumer and Bussy 2004; Simonetti *et al.* 2020).

The outer crystalline massif of Argentera is formed by two sets of gneisses and migmatites (Latouche and Bogdanoff 1987; Compagnoni *et al.* 2010) intruded by granitoids and separated by a dextral ductile shear zone active between 340 and 320 Ma (Simonetti *et al.* 2018), reworked by colder deformations during the Alpine orogeny between 34 and 22 Ma (Corsini *et al.* 2004; Sanchez *et al.* 2011). Migmatites contain various retromorphosed blocks of orthogneiss, marble, amphibolite and eclogite. The thermo-barometric conditions for eclogitization are estimated to be 735°C and 1.4 GPa (Ferrando *et al.* 2008). A zircon U–Pb age from eclogites at 471±40 Ma is interpreted as that of the protolith, while the Devonian–Carboniferous ages at 424±4 Ma, 351±1 Ma and 327±1 Ma are interpreted in terms of polymetamorphism (Paquette *et al.* 1989). The HP granulites contain ICP-MS U–Pb zircon ages at 341±4 Ma and 336±4 Ma (Rubatto *et al.* 2010). The Variscan syn-metamorphic ductile deformations, prior to migmatization, are not elucidated.

4.4.2.2. *Internal alpine domains: internal crystalline massifs and Briançonnais*

In these massifs, the very strong Alpine overprint largely erased the Variscan events. Nevertheless, various relics of early pre-Alpine metamorphism are recognized (for a recent and detailed review, see Ballèvre *et al.* 2018). Only a few highlights are summarized below.

In a metapelite from the Gran Paradiso Massif, the core of Variscan garnet in equilibrium with a biotite-staurolite-muscovite assemblage gives thermo-barometric conditions of 620°C and 0.6 GPa (Le Bayon *et al.* 2006). Variscan biotite-garnet-sillimanite micaschists crystallized at 650–700°C and 0.4–0.7 GPa are recognized in the north of the Dora Maira Massif (Bouffette *et al.* 1993). Comparable rocks are

found in the Ambin massif where micas yield $^{40}\text{Ar}/^{39}\text{Ar}$ ages between 350 and 340 Ma (Monié 1990).

In the Briançonnais zone, the micaschists country rocks of the Sapey orthogneiss (see Figure 4.22) contain relics of garnet, staurolite and kyanite that are considered of Variscan age (Detraz and Loubat 1984; Bertrand *et al.* 1998). Thus, it seems that the Variscan substratum of the Briançonnais zone and the Internal Crystalline Massifs have undergone at least one MP/MT metamorphic episode in the lower Carboniferous (around 350–340 Ma), prior to the late Carboniferous crustal melting and pluton emplacement. Due to the importance of the Alpine deformations and metamorphisms, there are no available detailed structural data on the Variscan syn-metamorphic events.

4.5. Conclusion

As in all mountain ranges, the Variscan orogeny was built progressively by the stacking of ductile and syn-metamorphic crustal nappes leading to crustal thickening. These processes can be well investigated in the Moldanubian domain. Although it is convenient to distinguish several tectono-metamorphic events, all the phenomena are part of a continuous and progressive evolution, particularly characterized by variations of the thermo-barometric conditions. The earliest episodes, known as Eo-Variscan ones are still poorly known, both in terms of the precise age of their formation, Upper Silurian–Lower Devonian or lower Carboniferous, as well as the mechanical and thermal modalities of their evolution.

An important feature of the French Variscan belt is the existence of a polycyclic evolution (Pin 1990; Faure *et al.* 1997; Lardeaux 2014; Lardeaux *et al.* 2014; Faure 2020, 2021). The current architecture of the Variscan belt formation can only be understood by considering the existence of an Eo-Variscan cycle responsible for the collision of the North Gondwana margin with the Armorica micro-continent in the Devonian, followed by a Carboniferous Variscan cycle *stricto sensu*, responsible for the collision of this continental mass with the northern Laurussian continent and the Saxothuringian micro-continent (or Mid-German Crystalline rise). During the Lower and middle Carboniferous period, the Variscan thermo-mechanical events, *stricto sensu*, mainly involved continental crust. It is only since the Upper Viséan, around 330 Ma, that the very high thermal gradient was at the origin of numerous episodes of crustal melting, giving rise to anatexis and syn-kinematic pluton emplacement leading to the “un-thickening” of the belt. In addition, many uncertainties remain on the precise thermo-barometric conditions of early events.

As far as the Léon block is concerned, the stacking of syn-metamorphic nappes directed towards the north and the plurifacial metamorphic evolution are well recognized. The attribution of this block to the Saxothuringian domain is also well accepted. However, the small exposure surface scarcely makes it possible to propose an overall geodynamic interpretation for this domain.

Finally, because of the high intensity of Alpine deformations and metamorphisms, as well as the various displacement senses that have occurred since the Permian, the place of the eastern branch of the French Variscan belt in the European Variscan puzzle remains speculative. A detailed discussion is beyond the scope of this chapter. Some answers are proposed in the article by Faure and Ferrière (2022).

4.6. References

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