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Episodic exhumation of HP rocks inferred from structural data and P-T paths from the southwestern Massif Central (Variscan belt, France)

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Abstract

In the southwestern Massif Central (France), HP rocks are found in lower allochthonous units (i.e. HP units), sandwiched between upper allochthonous units (i.e. LP units) and the Parautochthonous unit, both devoid of HP rocks. HP rocks were formed at minimum conditions of 1.7 GPa/700 °C in response to the northward subduction of Gondwana beneath Armorica during the Silurian (430–390 Ma). Their exhumation results from an 80-Ma-long sequence of tectonic processes effective at various scales and active specifically at each main stage of the evolution of the Gondwana lithosphere. HP rocks were first exhumed by upward, southwestward extrusion during the Early Devonian (390–380 Ma), in relation to late stage of subduction. In the northern area, exhumation is achieved by rifting during the Upper Devonian (∼365 Ma). The most obvious event corresponds to the northwestward thrusting of LP units above HP units during early stages of continental collision (∼355 Ma). HP rocks were exhumed for a second time during orogen-parallel transpression in response to late stage of continental collision (350–345 Ma). Exhumation in internal zones is achieved by crustal-scale normal faulting and erosion during the Middle Carboniferous (335–315 Ma) synorogenic extension, while compression was still active in the southern external zones. Exhumation of the whole region is completed by regional uplift and erosional denudation during the Upper Carboniferous (310–290 Ma) postorogenic extension. Subduction and collision have played a key role for driving HP rocks to the surface, but their efficiency and the resulting strain pattern depend on the angle of convergence. Upper Devonian rifting and the subsequent Early Carboniferous crustal thickening prevents to consider exhumation of Variscan HP rocks as a single and continuous process and open new questions on the related Paleozoic lithosphere history.

Keywords: Exhumation; Tectonics mechanisms; HP rocks; P-T paths; Paleozoic belt; Massif Central

1. Introduction

Exhumation processes of high pressure (HP) rocks describe the mechanisms by which deep-seated rocks return to the surface in response to extensive vertical movements of the lithosphere (50–100 km) (e.g. Platt, 1993). Because the erosion alone is unable to produce these movements (Thompson and Ridley, 1987; Dewey, 1988), tectonics combined with erosion is assumed to be the main process responsible for the exhumation of HP rocks (Burg et al., 1997). Tectonic exhumation is able to occur in response to subduction, collision, and/or extension of the continental lithosphere.

Although single in extensional setting, exhumation mechanisms of HP rocks are suspected to be multifaceted in orogenic belts because of their long and complex history. In such mountain belts, most of the HP rocks are founded in allochthonous units, and/or in granite-migmatite-cored domes that crosscut allochthonous units. The deciphering of their exhumation history therefore requires a thorough knowledge of tectonics and metamorphism of allochthonous units at a regional scale. Exhumation processes of HP rocks can therefore be discussed, and models emphasizing the respective meaning of subduction, collision, and extension of the lithosphere in these mechanisms can be finally assessed.

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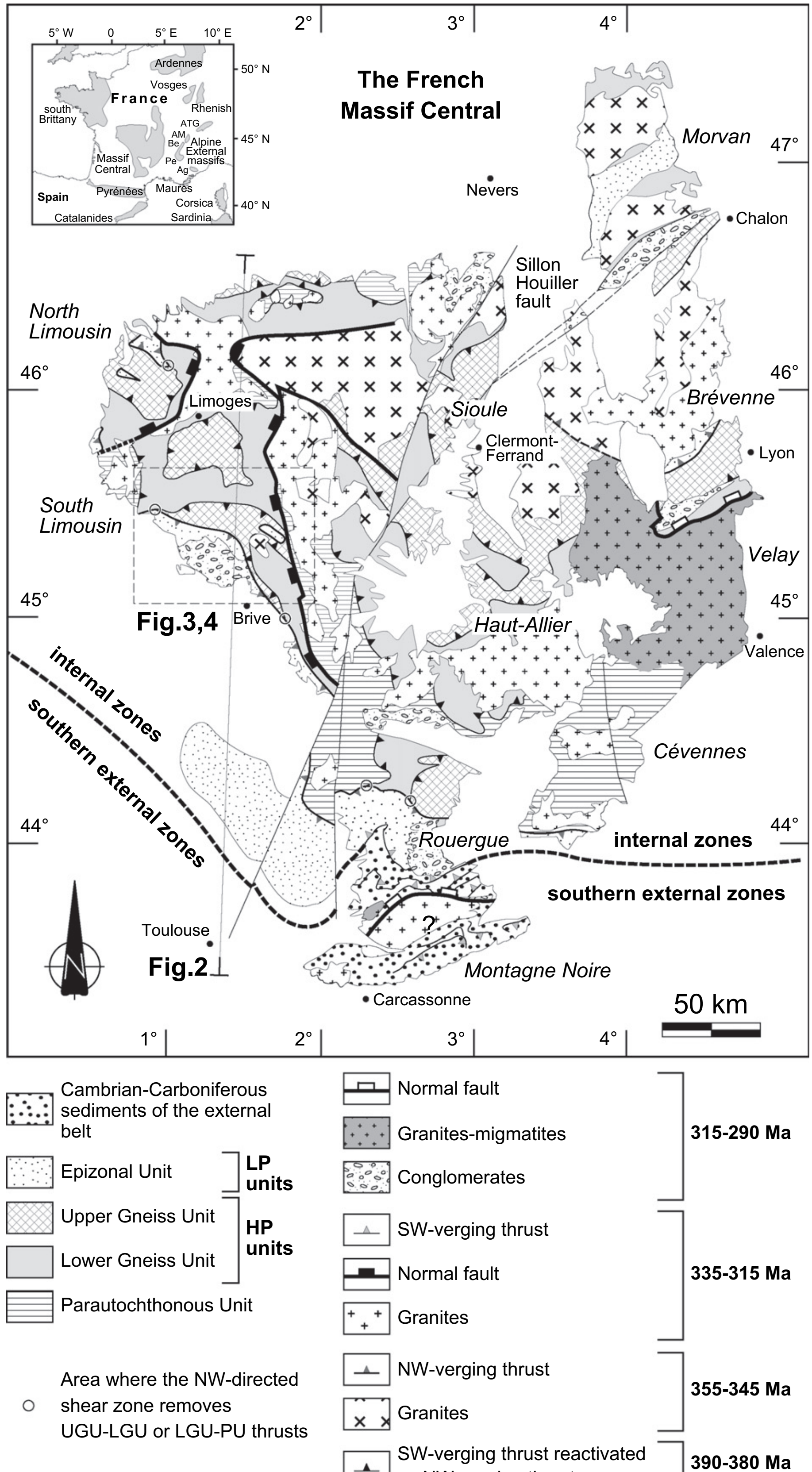

Fig. 1. Simplified map of the Massif Central region showing main compressional and extensional Variscan structures (after Ledru et al., 1994; Matte, 2001).

The aim of this study is to use the Paleozoic geology of the Massif Central region, the main Variscan outcrop, as a case study for deciphering complex exhumation processes of HP rocks in a long-lived orogenic belt (480–280 Ma). Our approach, focused on the southwestern Massif Central, reviews superimposed tectonics phases, then characterizes successive stages of metamorphisms, and finally interprets them in terms of tectonic processes in the light of available geochronological data.

2. Geological framework and previous studies

2.1. Geological framework of the Variscan belt in the Massif Central

In the main outcrop of the Variscan orogenic belt of Europe, i.e. the Massif Central region (Fig. 1), lenses of blueschist- and eclogite-type metabasites, and garnet-spinel peridotites have recorded subduction of arc/back-arc/passive margin lithospheres during Silurian (430–390 Ma) (Santallier, 1981; Dufour, 1985; Delor et al., 1986; Pin and Vielzeul, 1988; Bouchardon et al., 1989; Godard, 1990; Gardien et al., 1990; Mercier et al., 1991; Paquette et al., 1995; Matte, 1998; Lardeaux et al., 2001). These high-pressure (HP) to ultrahigh pressure rocks (Lardeaux et al., 2001) are founded in allochthonous units, made of meta-sediments (mica-schists) and meta-igneous rocks (orthogneisses) that display a pervasive Barrovian-type regional metamorphism dated Devonian to Middle Carboniferous (380–340 Ma) (Burg and Matte, 1978; Pin and Peucat, 1986; Costa, 1991–1992) with an inverted zoning (Briand, 1978). This intermediate pressure-type (IP) metamorphism is suspected to reflect southwestward thrust tectonics responsible of crustal thickening in the whole belt (Forestier, 1961–1963; Briand, 1978; Burg and Matte, 1978; Bodinier and Burg, 1980–1981; Burg et al., 1984, 1989; Ledru et al., 1994; Schulz et al., 1996; Matte, 2001; Schulz et al., 2001; Arnaud et al., 2004). LP metamorphism within La Montagne Noire and Le Velay migmatite-cored domes reflects NE-SW postorogenic extension dated Upper Carboniferous (310–290 Ma) (Duthou et al., 1984; Echtler and Malavieille, 1990; Malavieille et al., 1990; Montel et al., 1992; Lagarde et al., 1994; Faure, 1995; Gardien et al., 1997). The derived metamorphic evolution (Burg et al., 1989; Santallier et al., 1994; Gardien et al., 1997) was likely to reflect the Paleozoic convergence and assemblage of Laurentia-Baltica and Gondwana lithospheres (e.g. Ziegler, 1989; Matte, 2001) and numerous microplates between both, including Armorica (e.g. Matte, 2001) and Avalonia (e.g. Unrug et al., 1999).

This geodynamic scheme forms the base of several opposing models of exhumation of HP rocks that did not fit the whole data set (Faure et al., 2004). A model of southeastward, upward extrusion associated with continuous subduction of Gondwana (Matte, 1998) does not take into account Upper Devonian rifting in the northeastern Massif Central (i.e., the Brévenne area; Fig. 1) (e.g. Leloix et al., 1999). A model of northward subduction of Gondwana and arc-continent collision-related dextral transpression (Lardeaux et al., 2001) does not consider the occurrence of HP rocks on both sides of the Upper Devonian rift (Godard, 1990). This involves that exhumation is achieved before Upper Devonian rifting (Faure et al., 1997). Models of late-orogenic collapse (Dewey, 1988; Platt, 1993; Aerden, 1998) do not take into account that HP rocks of the Variscan belt are mainly found in the hangingwall of crustal-scale detachment faults. This supports that most of their exhumation occurs before late-orogenic extension.

Exhumation models proposed for the Variscan belt are strongly constrained by the significance of NW-verging shear zones that coincide with units boundaries. NW-verging shear zones are considered to be either thrust faults (Demay, 1948; Briand, 1978; Burg and Matte, 1978; Bodinier and Burg, 1980–1981; Brun and Burg, 1982; Floc'h, 1983; Burg et al., 1984; Bouchez and Jover, 1986; Girardeau et al., 1986; Burg et al., 1987; Friedrich et al., 1988; Burg et al., 1989; Ledru et al., 1994; Faure et al., 1997; Leloix et al., 1999; Roig and Faure, 2000) or detachment faults (Mattauer et al., 1988; Matte, 1998).

2.2. Geological units of the South Limousin region

At present-day exposure levels, the South-Limousin area, which belongs to the Variscan belt of Western Europe, displays four units as outlined below (Ledru et al., 1994) (Fig. 2).

The Epizonal Unit (EU) corresponds, from top to bottom, to the Ordovician-Devonian Génis Unit formation (GF), the Cambrian-Devonian Thiviers-Payzac formation (TPF), and the St. Salvadour formation (SF). The Génis formation includes Middle Devonian limestones (Roques, 1941; Guillot and Lefevre, 1975), un-metamorphosed gabbro, dolerite, basalt and cherts, and Ordovician-Devonian sericite-bearing schist (Guillot and Doubinguer, 1971). The Thiviers-Payzac formation consists, from top to bottom, of Ordovician quartzite resting on amphibole-plagioclase-bearing schist and epimetamorphic slate (the Engastine complex), and of rhyodacitic tuffs of assumed Cambrian-Late Proterozoic age. These rocks were intruded by pretectonic calk-alkaline dolerite dykes (Guillot, 1981) of Upper Devonian age (363 ± 10 Ma, whole rock K-Ar) (Cantagrel, 1973). The St. Salvadour formation is made of metabasites distributed into levels of amphibolites, pyroxene-bearing layered amphibolites, metagabbros, and metadolerites (Santallier, 1981; Bellot, 2001) those the composition of major and trace-elements very similar to N-MORB to E-MORB suggests their origin in a back-arc setting (Ezzayani, 1991). The Epizonal Unit has experienced low-grade metamorphism as evidenced from the normal zoning of chlorite, biotite, garnet, and staurolite isograds (Guillot, 1981) (Figs. 3–5). But the relationships between mineral growth and the regional tectonic phases are not yet documented.

The Upper Gneiss Unit (UGU) consists of mafic eclogite boudins, and of mafic and feldsic granulites lenses hosted by migmatitic paragneiss (Floc'h, 1983). Eclogites give evidence for HP metamorphism. In the eastern Massif Central, these rocks have been dated Silurian (430–415 Ma, zircon U-Pb) (Matte, 1998) and preserved coesite in garnet core that indicates ultra-high-pressure (UHP) metamorphism (≥3 GPa) (Lardeaux

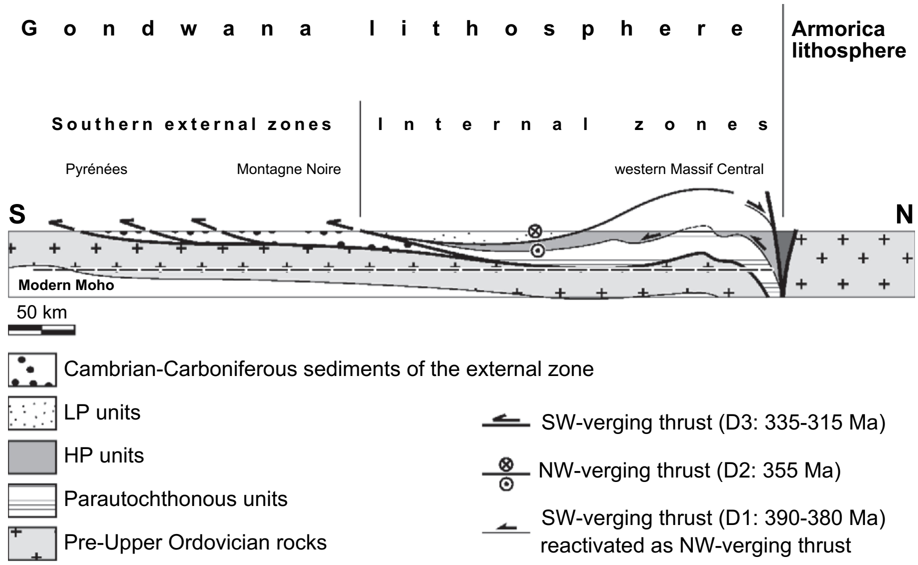

Fig. 2. Our interpretation of a pre-Mesozoic crustal-scale cross-section through the western Massif Central. No vertical exaggeration. Southern and northern parts of the section from Choukroune et al. (1990) and Bellot (2001), respectively.

et al., 2001). At the UGU base occur gabbros and peridotites interpreted as crustal and mantle parts of an oceanic lithosphere supposed to be Early Ordovician in age (Dubuisson et al., 1989). Mafic eclogites include kyanite or zoisite, and represent a co-magmatic suite involved in a common metamorphic history characterized by a P-T peak (1.5 GPa/750 °C) followed by retrogression to amphibolites facies conditions (Santallier, 1981). Migmatites include syn-D_1 sillimanite + garnet + biotite + K-feldspar crystallizing at 0.7 GPa/700 °C (Roig and Faure, 2000). Paragneiss comprise common kyanite + staurolite + garnet ± sillimanite (Autran and Guillot, 1975; Mouthier, 1976; Lamouille, 1979; Guillot, 1981; Floc'h, 1983; Ezzayani, 1991) but the relationships between mineral growth and the regional tectonic phases are undocumented.

The Lower Gneiss Unit (LGU) is mainly formed of orthogneiss, corresponding to Cambrian-Ordovician (550–450 Ma) calk-alkaline granite and granodiorite that have intruded paragneiss and mica-schist (Bernard-Griffiths, 1975; Duthou et al., 1984; Lafon, 1986; Alexandrov, 2000). In the lower LGU, lenses of eclogites and garnet-spinel peridotites are evidence for HP metamorphism (Santallier, 1981; Santallier and Floc'h, 1978; Egal et al., 1986) dated Lower Devonian in the eastern Massif Central (410–390 Ma, zircon U-Pb and whole-rock Sm-Nd) (Paquette et al., 1995). Paradoxically, this unit is considered to be free of eclogite in the South Limousin (Ledru et al., 1994) but includes syn-D_1 cordierite-sillimanite-K-feldspar-bearing migmatites of Devonian age (382 ± 5 Ma; zircon U-Pb) (Lafon, 1986). Paragneiss and mica-schist have common sillimanite + garnet + biotite (Floc'h, 1983) but the relationships between mineral growths with respect to the regional tectonic phases are again undocumented. This unit is assumed to be lesser metamorphic than the UGU, emphasizing inverted metamorphism due to thrusting of the UGU on the LGU (Ledru et al., 1994; Roig and Faure, 2000).

The Parautochthonous Unit (PU) is composed of quartzite, mica-schist and scarce Ordovician orthogneiss (467 ± 8 Ma, whole rock Rb-Sr) (Monier, 1980). This unit has experienced two superimposed metamorphic stages at IP and LP (Feix, 1988). P-T conditions of post-D_2 partial melting in the Millevaches dome are estimated at 0.40–0.60 GPa/650–700 °C (Monier, 1980).

2.3. Structures of the South Limousin region

Structures observed in meta-sedimentary (paragneiss, mica-schist) and meta-igneous rocks (metabasalt, metagabbro, metaperidotite) result from four superimposed deformation phases (Floc'h, 1983; Ledru et al., 1994; Roig and Faure, 2000; Bellot, 2001). These are related to the Upper Paleozoic tectonics, including polyphased low-angle shear tectonics (D_1 and D_2), transcurrent tectonics (late D_2) and extensional tectonics (D_3 and D_4) (Figs. 3 and 4).

The D_1 deformational phase, that affects UGU and LGU only, is responsible for the S_1-L_1 rock fabric, including F_1 sheath folds and top-to-the SW shearing. Development of migmatite (382 ± 5 Ma, zircon U-Pb) (Lafon, 1986) and retrogression of mafic (Santallier, 1981) and ultramafic rocks (Girardeau et al., 1986) are associated with this tectonic phase dated Early Devonian and interpreted as a large-scale SW-verging thrust tectonics (Girardeau et al., 1986; Roig and Faure, 2000).

The D_2 deformational phase, that affects all units, is responsible for the regional S_2-L_2 rock fabric, including F_2 isoclinal folds parallel to the NW-SE-trending L_2 lineation and associated with a main top-to-the NW shearing. The main shear zones are encountered in the uppermost UGU and the lowermost LGU, emphasizing the crustal dimension of this shearing tectonics (Fig. 3). The D_2 phase is likely to be Early Carboniferous in age, based on ^{40}Ar-^{39}Ar dating on muscovite extract from mylonites occurring at the top (352 ± 7 Ma) (Costa, 1991–1992) or the bottom of the LGU (355 ± 8 Ma) (Costa, 1991–1992). Deformed and recrystallized quartz diorite of the South Limousin has also yielded an Early Carboniferous age

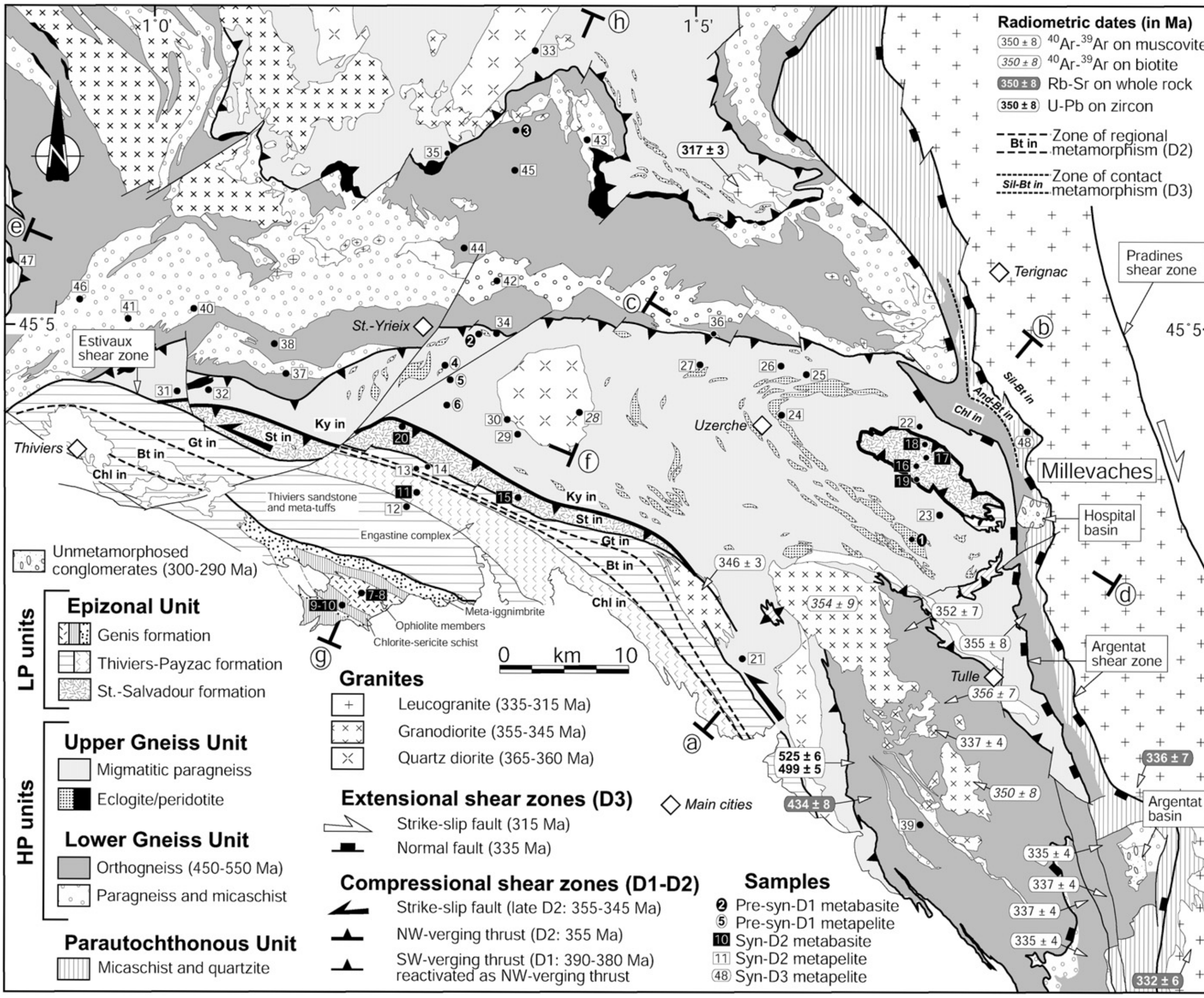

Fig. 3. Simplified geological map of the South-Limousin area (after Guillot, 1981; Santallier, 1981; Floc'h, 1983 and new observations), with location of samples used for thermobarometry and trace of cross-sections of Fig. 5. Isograds of regional metamorphism in the Epizonal Unit are taken from Guillot (1981) and Floc'h (1983) and modified from new observations. Isograds of contact metamorphism are drawn based on our observations. References for radiometric dates are given in the text.

(355 ± 2 Ma; zircon U-Pb) (Bernard-Griffiths et al., 1985) suspected to reflect the age of the D_2 tectonic phase, since undeformed and unmetamorphosed quartz diorite emplaced in the upper crust (0.3–0.4 GPa) (Bouvier, 1985) of the North Limousin has given a Late Devonian magmatism age (365–360 Ma; zircon U-Pb) (Pin and Paquette, 2002). The D_2 phase is responsible for the reworking of the S_1-L_1 fabric of Lower Devonian migmatites (Floc'h, 1983; Roig and Faure, 2000) and of retrogressed eclogites (Santallier, 1981; Girardeau et al., 1986) in UGU and LGU. This tectonics also results in development of migmatites in the Tulle antiform (356 ± 7 Ma; biotite ^{40}Ar-^{39}Ar) (Costa, 1991–1992). Even though the lack of convincing metamorphic data, the main steps of the D_2 tectonic phase ($\sim$355 Ma) are considered to reflect crustal thickening by thrusting and nappes stacking (Floc'h, 1983; Bouchez and Jover, 1986; Ledru et al., 1994; Roig and Faure, 2000). Late steps of the D_2 phase (350–345 Ma) resulted in a regional-scale transpressional tectonics. This involves sinistral strike-slip shearing along the Estivaux shear zone (346 ± 3 Ma; biotite ^{40}Ar-^{39}Ar) (Roig, 1997), regional-scale folding (Uzerche synform, Tulle and Meuzac antiforms), and granodiorite emplacement along these structures (Roig et al., 1996, 1998).

The D_3 deformational phase is responsible for the normal movement along the Argentat fault associated with emplacement of the Millevaches plutonic complex in a NW-SE trending extensional strain field (Ledru et al., 1994; Roig et al., 2002; Bellot, 2007). Whole-rock Rb-Sr dating of leucogranite (332–336 Ma) (Monier, 1980), muscovite ^{40}Ar-^{39}Ar dating of hydrothermalized gneiss (335–337 Ma) (Roig et al., 2002) associated with the Argentat fault, and zircon U-Pb dating of leucogranite occurring in the St.-Germain-les-Belles synform (317 ± 3 Ma) (Lafon, 1986) indicate a time span for synorogenic extension ranging from Late Visean ($\sim$335 Ma) to Late Namurian ($\sim$315 Ma).

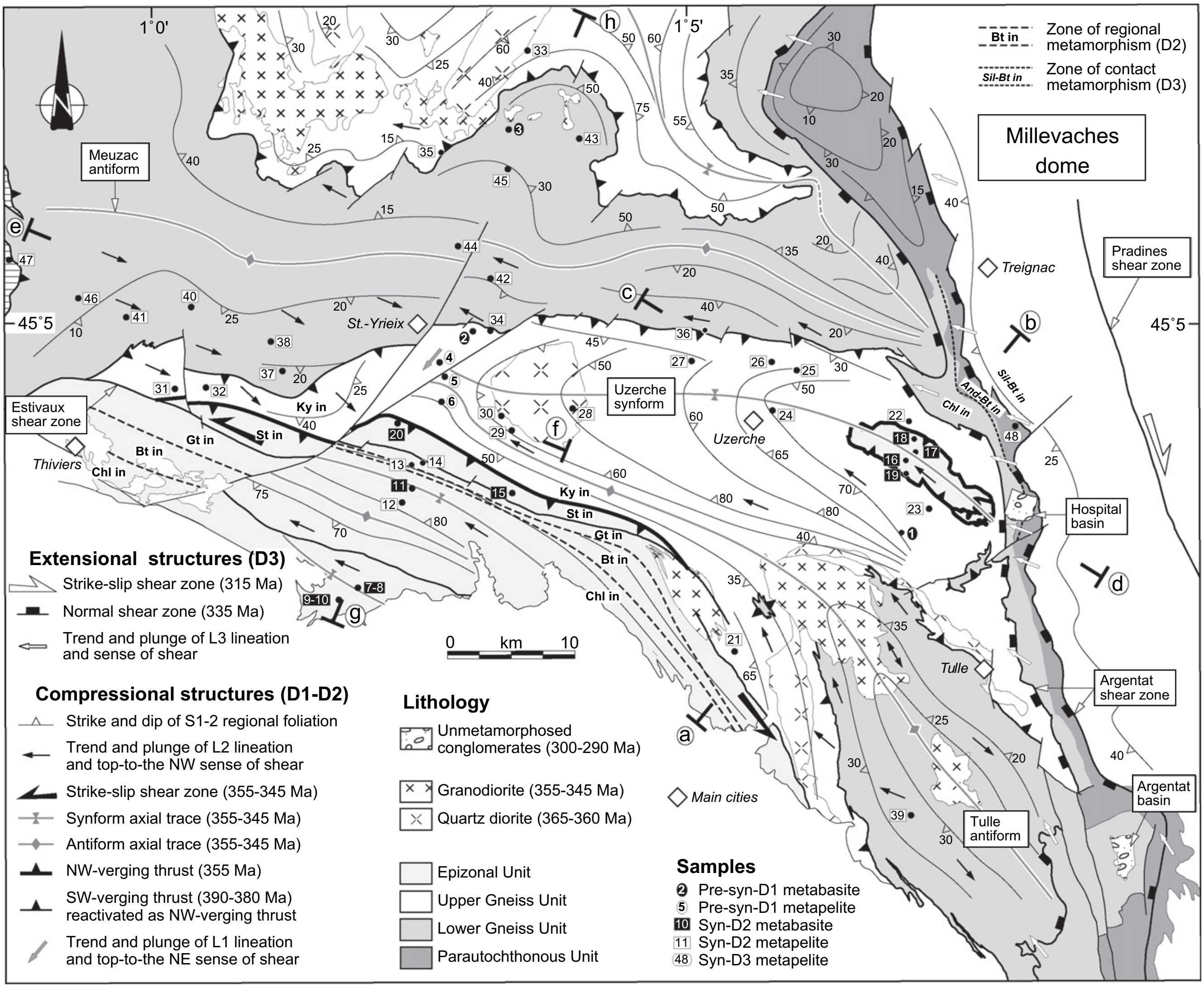

Fig. 4. Simplified structural pattern of the South-Limousin area, with the setting of samples used for thermobarometry and trace of cross-sections of Fig. 5. Leucogranites not included for clarity. The boundary between garnet and staurolite zone, staurolite and kyanite zone, andalusite-biotite and sillimanite-biotite zone, coincides with shear zones.

The D_4 deformational phase is responsible for the left-lateral movement along the Argentat fault, associated with fracturing in orthogneiss (Roig et al., 2002), folding in metapelites and metabasites (Bellot et al., 2005), deposition of syntectonic stibnite lodes along the Biards fault (Bellot et al., 2003) and deposition of coal interbedded with conglomerates in the Argentat and Hospital basins (Bellot et al., 2005). This phase is related to the Upper Carboniferous, postorogenic extension, beginning in the Late Westphalian (~310 Ma) and finishing in the Upper Stephanian (~290 Ma), in a NE-SW trending extensional strain field (Malavieille et al., 1990; Burg et al., 1994; Lagarde et al., 1994; Faure, 1995).

3. Methodology and analytical procedure

Exhumation processes of HP rocks in relation to the Paleozoic lithosphere history are deciphered by interpreting and comparing pressure-temperature-deformation-time (P-T-D-t) paths established for each unit that previously formed the different structural levels of the Variscan orogenic belt (e.g. Spear et al., 1984; Duchène et al., 1997). So, different tectonic units can be identified, variation in timing and grade of metamorphism established, and tectonics mechanisms deduced.

A P-T-D-t path per metamorphic unit was obtained by combining structural and metamorphic analyses in chosen representative samples of both metapelites and metabasites (Figs. 3–5). These samples have been investigated by transmitted and reflected optical microscopy, and by scanning electron microscopy. P-T trajectories were achieved by combining semi-quantitative and quantitative evaluations of the P-T conditions of metamorphism. Semi-quantitative evaluations were obtained using a simplified grid of the NCFMASH system for metabasites (after Goscombe and Hand, 2000 and references therein), and of the KFMASH system for metapelites (Spear, 1993; Spear et al., 1999 and references therein). Quantitative P-T estimates were obtained using conventional barometers and

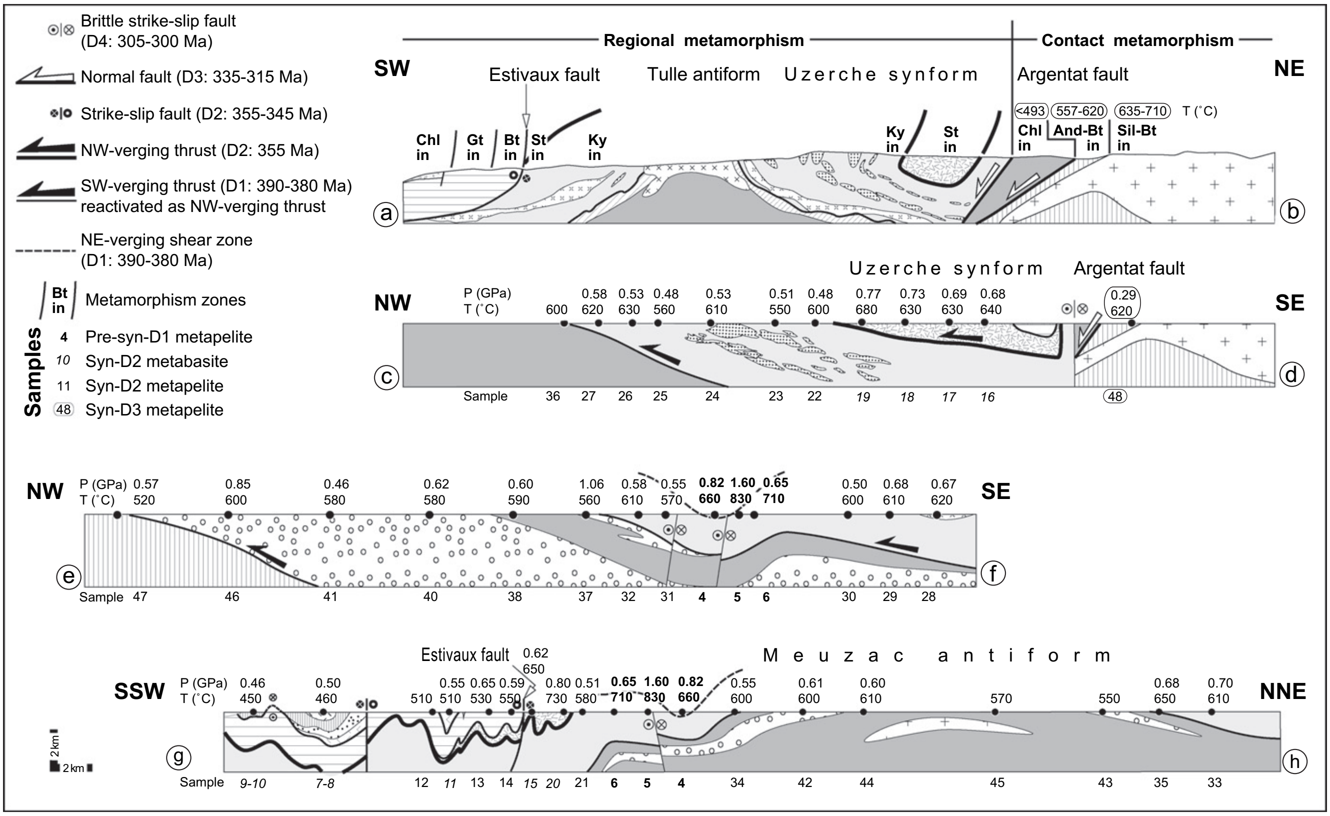

Fig. 5. Simplified cross-sections through the South-Limousin area, with location of isograds of regional and contact metamorphisms, of samples used for thermo-barometry and of maximum P-T estimates obtained for rim-rim equilibrium.

thermometers that use several experimental and/or theoretical calibrations of multi-phase equilibrium (Table 1). Thermobarometry was applied thoroughly on rim-rim compositions, and when feasible on syntectonic zoned garnet supposed to be in equilibrium with inclusion minerals or with core of matrix biotite and plagioclase, providing prograde and/or retrograde metamorphism associated with shear deformation (Spear et al., 1984). Indeed, observed zoning trends in garnet and plagioclase, together with biotite composition, were interpreted in terms of reactions that operated at various stages during the metamorphic evolution (e.g. Escuder Viruete et al., 2000). Different P-T values obtained in rocks that belongs to the same structural level and associated with the same tectonic phase were likely to reflect local equilibrium established at various stages of the metamorphic history (e.g. Spear, 1993) and were therefore used to further constrain P-T path for a given unit.

Table 1

Barometer/thermometer		References
Barometer	Jadeite content in clinopyroxene	Holland (1980)
Barometer	Aluminium content in hornblende	Holland and Blundy (1994)
Barometer	Aluminium content in hornblende	Hammarstrom and Zen (1986)
Barometer	Aluminium content in hornblende	Hollister et al. (1987)
Barometer	Aluminium content in hornblende	Johnson and Rutherford (1989)
Barometer	Aluminium content in hornblende	Schmidt (1992)
Barometer	Garnet-rutile-ilmenite-plagioclase-silica	Bohlen and Liotta (1986)
Barometer	Garnet-plagioclase-hornblende-quartz	Kohn and Spear (1990)
Barometer	Garnet-aluminosilicate-silica-plagioclase	Ghent et al. (1979)
Barometer	Garnet-aluminosilicate-silica-plagioclase	Newton and Haselton (1981)
Barometer	Garnet-mica-aluminosilicate-plagioclase	Ghent and Stout (1981)
Thermometer	Garnet-clinopyroxene	Berman et al. (1995)
Thermometer	Garnet-hornblende	Graham and Powell (1984)
Thermometer	Plagioclase-hornblende	Holland and Blundy (1994)
Thermometer	Garnet-biotite	Kleemann and Reinhardt (1994)
Thermometer	Garnet-ilmenite	Pownceby et al. (1987)
Thermometer	Biotite-tourmaline	Blamart et al. (1992)
Thermometer	Chlorite-quartz	Vidal et al. (2001)

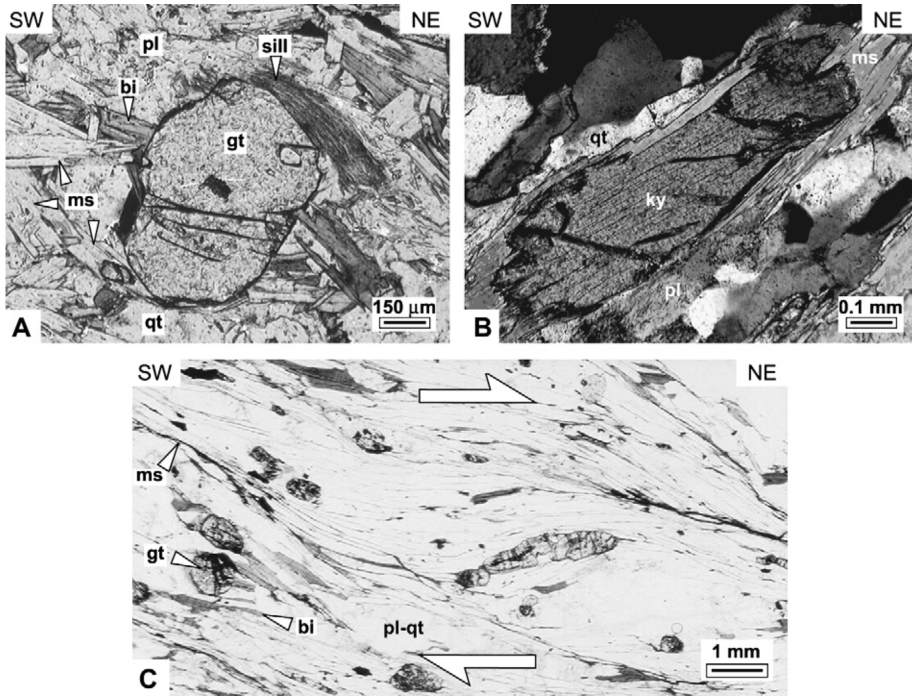

Fig. 6. Photomicrographs (transmitted light) showing key textures and mineral assemblages observed in metapelites associated with pre-syn-D_1 metamorphism (A–C), syn-D_2 metamorphism (D–G: Limousin nappes) and syn-D_3 metamorphism (H–J: Millevaches dome). A. Sillimanite + garnet + biotite + muscovite + plagioclase + quartz in a migmatitic paragneiss. B. Muscovite + quartz development from kyanite and plagioclase, suggesting the retrograde reaction kyanite + plagioclase + H_2O = muscovite + quartz. C. Garnet + biotite + muscovite + plagioclase + quartz growth during top-to-the NE shearing. D. Garnet δ-type porphyroblast with asymmetric strain shadows of biotite + plagioclase + quartz. E. Quartz + K-feldspar + sillimanite forming sigmoid aggregate or asymmetric strain shadows of garnet with kyanite and biotite inclusions. F. Isoclinal folds within the S_{1-2} foliation made of biotite + sillimanite + quartz. G. Development of sillimanite, garnet and biotite, suggesting the prograde reaction staurolite = sillimanite + garnet + biotite. H. Andalusite δ-type porphyroblast with asymmetric strain shadows of biotite and helicitic inclusions of garnet, biotite, plagioclase, tourmaline, ilmenite and chlorite (andalusite-biotite zone). I. Coexisting K-feldspar, cordierite (now penninitised), and biotite, and biotite shear bands (sillimanite-biotite zone). J. Syntectonic growth of tourmaline and biotite.

4. The Epizonal Unit

The Epizonal Unit is devoid of eclogites, migmatites, and pre- to syn-D_1 assemblages. Syn-D_2 mineral or assemblages define metamorphic zones parallel to the S_{1-2} regional foliation and oblique to the Estivaux shear zone, indicating a syn-D_2 metamorphism with normal zoning developed prior to Estivaux sinistral shearing: chlorite, biotite, garnet, staurolite, and kyanite in metapelites (Figs. 3–5); pumpeleite + chlorite + quartz + albite, chlorite + epidote + albite + actinote, and hornblende + andesine ± garnet ± biotite in metabasites. Prograde metamorphism responsible for normal metamorphic zoning is evidenced from chemical zoning in syn-D_2 garnet, amphibole and plagioclase, their mineral inclusions, and thermobarometry. In the garnet zone, garnet + biotite associated with relict chlorite (Guillot, 1981) is likely to have occurred through the continuous reaction: chlorite + muscovite + quartz + anorthite = garnet + biotite + albite + H_2O, indicating >450 °C. In the staurolite zone, growth of staurolite with inclusion of biotite, partial dissolution of garnet, and continued dissolution of chlorite (Floc'h, 1983) probably occurred by the reaction: garnet + chlorite = staurolite + biotite, indicating 0.50–0.90 GPa/570–670 °C. Syn-D_2 metamorphism peak is estimated at 0.50 GPa/460 °C, 0.65 GPa/530 °C, and 0.80 GPa/670 °C for GF, TPF, and SF respectively (Fig. 7) consistent with the results of Duguet (2003) on the EU of the Rouergue area.

5. Upper Gneiss Unit and Lower Gneiss Unit

5.1. Pre- to syn-D_1 metamorphism in metabasites

Eclogites from the upper part of the UGU display an equigranular texture involving four assemblages. An assemblage A1 of coarse garnets with omphacite + rutile + quartz inclusions indicates HP metamorphism increasing from blueschist (Fig. 7, box 1/1) to eclogite facies conditions (1.70 GPa/700 °C) (Fig. 7, box 1/2). Assemblages A2 and A3, that correspond to matrix clinopyroxene + plagioclase symplectites developed between garnet and omphacite, and to matrix clinopyroxene + quartz coronas developed around omphacite, indicate HP (Fig. 7, box 1/3) and IP (Fig. 7A, box 1/4) granulites facies metamorphism associated with near-isothermal decompression. An assemblage A4 of matrix amphibole (katophorite) + plagioclase associations developed between clinopyroxene and plagioclase indicates HT amphibolites facies metamorphism associated with decompression and cooling (Fig. 7, boxes 1/5 and 1/6). The P-T path derived

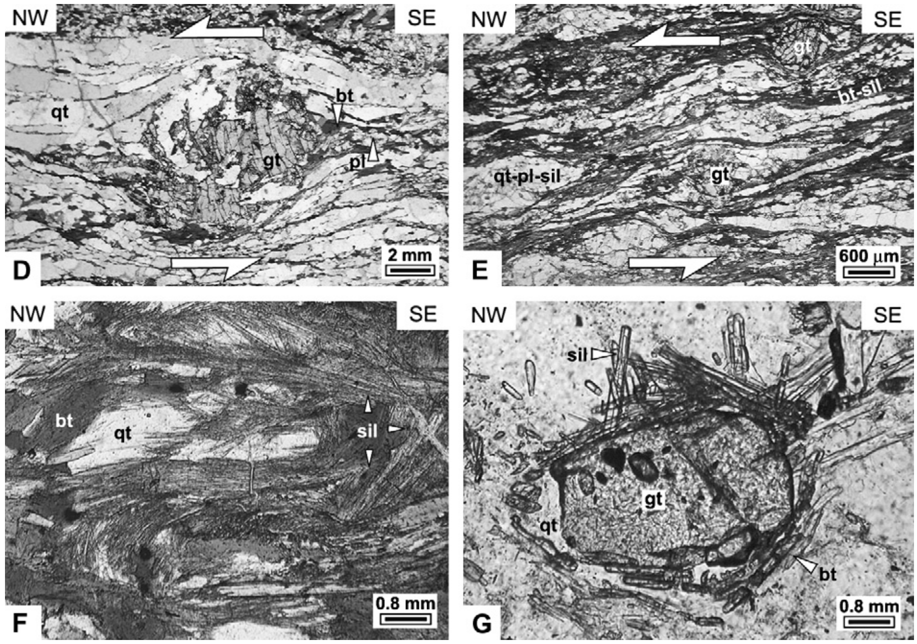

Fig. 6 (*continued*).

from thermobarometry shows prograde then retrograde metamorphism similarly to other eclogites of the Massif Central (Mercier et al., 1991).

Mafic granulites from the lower part of the UGU are made of pre- and syntectonic assemblages. A pretectonic assemblage B1 of unzoned spinel, plagioclase, clinopyroxene, quartz, amphibole (pargasite), and garnet with inclusions of quartz and ilmenite, indicates IP granulites metamorphism (1.0 GPa/805 °C) (Fig. 7, box 2/1). A syntectonic (D_1) assemblage B2 of zoned amphibole (pargasite, hastingsite or tschermakite core to hornblende rim), zoned plagioclase, epidote, ilmenite, and quartz indicates retrograde metamorphism from HT (Fig. 7, box 2/2) to LT amphibolites facies conditions (0.40 GPa/525 °C) during southwestward shearing (Fig. 7, box 2/3). The P-T path derived from thermobarometry shows retrograde metamorphism similarly to other granulites of the Massif Central (Dufour, 1985).

Eclogites from the lower part of the LGU exhibit an equigranular texture involving four assemblages. An assemblage C1 of rounded garnet, omphacite, interstitial phlogopite, zoisite, and rutile indicates HP metamorphism increasing from blueschist (Fig. 7, box 3/1) to eclogite facies conditions (1.56 GPa/700 °C) (Fig. 6B), as well as barroisite also indicates eclogite facies metamorphism (Fig. 7, box 3/2). An assemblage C2 of plagioclase + spinel + clinopyroxene corona between garnet and clinopyroxene indicates IP granulites facies metamorphism (1.22 GPa/850 °C) (Fig. 7, box 3/3). An assemblage C3 of plagioclase + clinoamphibole (barroisite core to hornblende rim) symplectites developed around garnet indicates retrograde metamorphism to HT amphibolites facies conditions (Fig. 7, boxes 3/4 and 3/5). The P-T path derived from thermobarometry shows prograde then retrograde metamorphism similarly to other eclogites of the Massif Central (Mercier et al., 1991).

5.2. Pre- to syn-D_1 metamorphism in metapelites

Slightly foliated gneiss (sample 5) from the top of the UGU occurs as relic in syn-D_1 migmatites. It contains matrix kyanite + biotite + muscovite + plagioclase and Ca-rich garnet with retrograde zoning pattern (e.g. Spear and Daniel, 2001). Rutile + biotite and ilmenite + quartz inclusions in weakly and strongly zoned garnets indicate respectively $P > 1.2$ GPa and $P < 1.2$ GPa for 750 °C. The P-T path derived from these reactions combined to thermobarometry shows retrograde metamorphism in the kyanite field from HP granulites (1.60 GPa/830 °C) to amphibolites facies conditions (Fig. 7).

Migmatites from the upper part of the UGU (sample 6) develop from slightly foliated gneiss like sample 5. Kyanite is relictual. An early syn-D_1 assemblage of garnet + sillimanite + biotite + K-feldspar is preserved in boudins and indicate 0.6–1.0 GPa/700–800 °C, in agreement with P-T values given by Roig and Faure (2000) on a similar assemblage. In the common syn-D_1 assemblage of garnet + sillimanite + biotite + muscovite + plagioclase + ilmenite + quartz (Fig. 6A), garnet has typical retrograde zoning pattern, K-feldspar is pseudomorphosed by muscovite + quartz, and destabilised kyanite + plagioclase is associated with muscovite + quartz (Fig. 6B). These textural relationships probably reflect both reaction $Al_2Si_2O_5$ (kyanite) + K-feldspar + H_2O = muscovite + quartz, and $Al_2Si_2O_5$ (kyanite) + plagioclase + H_2O = muscovite + quartz. In the foliation plane, abundant biotite + sillimanite are associated with relict muscovite and garnet, and may produce by the reaction muscovite + garnet = biotite + sillimanite, that indicates <0.50–0.60 GPa/<600–700 °C. The P-T path derived from these reactions combined to thermobarometry displays retrograde metamorphism from 0.65 GPa/710 °C to 0.42 GPa/570 °C, associated with syn-D_1 migmatites development (Fig. 7).

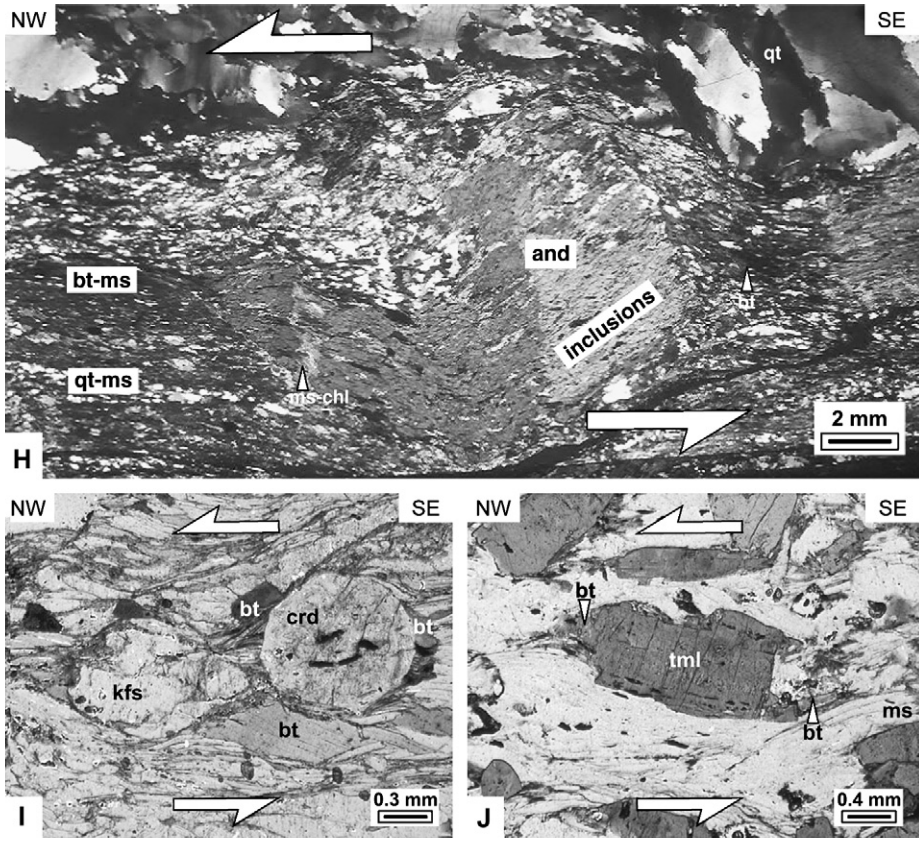

Fig. 6 (*continued*).

The shear zone located at the uppermost UGU has experienced a northwestward movement associated with the isoclinal refolding of a previous S_1-L_1 fabric (Bellot, 2004). However, several outcrops along the shear zone show mylonitic paragneiss with a flat-lying foliation and a N040 trending lineation associated with a northeastward movement (sample 4). Kyanite + biotite + plagioclase + muscovite + apatite + quartz + garnet with retrograde zoning pattern growth during shear deformation (Fig. 6C). The P-T path derived from these reactions combined to thermobarometry shows retrograde metamorphism in the kyanite field from 0.82 GPa/655 °C to 0.48 GPa/465 °C (Fig. 7).

5.3. Syn-D_2 metamorphism

Both UGU and LGU are in the kyanite zone. In UGU, common kyanite + biotite, without garnet, may reflect the reaction staurolite + muscovite + chlorite = biotite + Al_2SiO_5 (kyanite) + H_2O, indicating >0.60 GPa/>620 °C. In LGU, incomplete dissolution of staurolite in addition to garnet growth with inclusion of biotite and kyanite (Floc'h, 1983; Bellot, 2001) can occur by the reaction: staurolite = garnet + biotite + Al_2SiO_5 (kyanite). In both UGU and LGU, syntectonic growth of quartz + biotite + plagioclase ± kyanite + garnet (Fig. 6D) with prograde zoning pattern indicates prograde metamorphism. A common syn-D_2 P-T peak is estimated at 0.70 GPa/610 °C for UGU (Fig. 6D). A common syn-D_2 P peak is estimated at 1.06 GPa/560 °C for the LGU, although syn-D_2 T peak is estimated at 0.90 GPa/630 °C for the upper part of the LGU, and at 0.72 GPa/680 °C for its lower part (Fig. 7).

The common syn-D_2 retrograde metamorphism is evidenced from syntectonic growth of biotite + sillimanite + muscovite + plagioclase + ilmenite + quartz + garnet with retrograde zoning pattern, deformation of kyanite, and syntectonic development of partial melting increasing downward the allochthonous units (Fig. 6E). Muscovite in presence of quartz rimming relict kyanite and plagioclase suggests the reaction: Al_2SiO_5 (kyanite) + plagioclase + H_2O = muscovite + quartz. Sillimanite + biotite + quartz (Fig. 6F) associated with relict garnet + muscovite suggest the reaction: garnet + muscovite = biotite + Al_2SiO_5 (sillimanite) + quartz, indicating <0.55 GPa/600–700 °C. Highly resorbed staurolite rimmed by sillimanite + biotite + garnet (Fig. 6G) can reflect the reaction: staurolite = garnet + biotite + Al_2SiO_5 (sillimanite), indicating 0.40–0.55 GPa/590–650 °C. Lack of K-feldspar in migmatites suggests $T < 700$ °C at 0.50 GPa. P-T paths derived from these reactions combined to thermobarometry shows retrogression to 0.40 GPa/550 °C for UGU, to 0.52 GPa/540 °C for the upper part of the LGU, and to 0.46 GPa/580 °C for its lower part (Fig. 7).

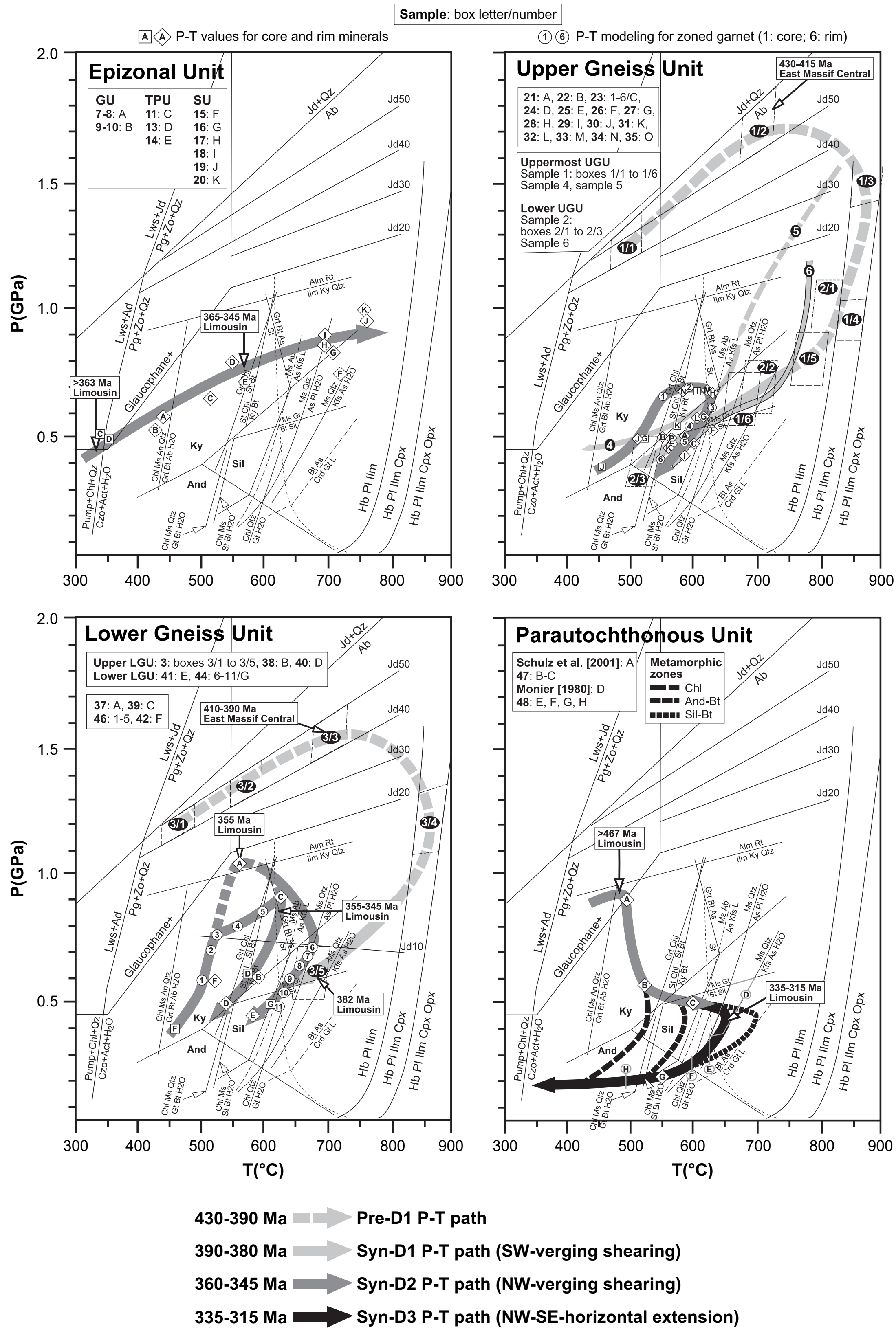

Fig. 7. P-T paths obtained on Epizonal Unit, Upper Gneiss Unit, Lower Gneiss Unit, and Parautochthonous Unit of the South Limousin area. Reactions taken from Goscombe and Hand (2000) and Lardeaux et al. (2001 and references therein) for metabasites, from Spear et al. (1999) for metapelites. Errors not included in the figure for clarity. Geothermobarometry errors are assumed to be of the order of ±1 kbar and ±50 °C.

6. The Parautochthonous Unit

6.1. Syn-D_2 metamorphism

The Parautochthonous Unit is devoid of eclogites, migmatites, and pre- to syn-D_1 assemblages. Metapelites sampled from the Parautochthonous Unit roof display syntectonic growth of inclusions kyanite + biotite + albite + quartz in garnet and matrix staurolite + garnet + biotite + sillimanite + muscovite + plagioclase + ilmenite + quartz. Stable staurolite and sillimanite indicates >0.28 GPa/500–615 °C. Abundant biotite + sillimanite developing from garnet + muscovite suggests the reaction biotite + sillimanite = garnet + muscovite, restraining P-T conditions to 0.28–0.55 GPa/500–615 °C. Schulz et al. (2001) have obtained P-T peak at ~0.90 GPa/490 °C for early syn-D_2 assemblages of the same unit in the eastern Massif Central. The P-T path derived from these reactions combined to thermobarometry indicates decompression with slight heating from 0.90 GPa/490 °C to 0.57 GPa/520 °C, then heating with slight decompression to 0.49 GPa/605 °C during syn-D_2 shearing tectonics (Fig. 7).

6.2. Syn-D_3 metamorphism

In the footwall of the Argentat fault, metamorphic zones of syn-D_3 minerals are organized parallel to the fault trace. This relationship indicates syn-D_3 metamorphism increasing inwards the Millevaches dome (Figs. 3–5), from chlorite in orthogneiss, to andalusite-biotite in mica-schist (Fig. 6H), and sillimanite-biotite in the granite-migmatite dome (Fig. 6I). Syn-D_3 mineral assemblages involve a suite of prograde (complete) and retrograde (incomplete) reactions in the KFMASH system. In the chlorite zone, chlorite + quartz replacing garnet can be explained by the reaction: garnet + H_2O = chlorite + quartz, indicating <655 °C at 0.30 GPa. In more pelitic rocks, the common syntectonic chlorite + muscovite + quartz assemblage may be explained by the reaction: garnet + biotite + H_2O = chlorite + muscovite + quartz, indicating <570 °C at 0.30 GPa. Chlorite + white mica replacing staurolite + biotite suggest the reaction staurolite + biotite + H_2O = chlorite + white mica, indicating <0.50 GPa/<520–550 °C. The absence of K-feldspar in this zone and prevalent, late muscovite + quartz in samples can be attributed to the growth of retrograde muscovite by the reaction: Al_2SiO_5 (sillimanite) + K-feldspar + H_2O = muscovite + quartz, indicating <680 °C at 0.30 GPa. This reaction uses water vapor exsolved from the crystallizing melt to produce muscovite from K-feldspar. In the andalusite-biotite zone, early sillimanite + biotite associated with relict garnet + muscovite suggest the reaction: garnet + muscovite = biotite + Al_2SiO_5 (sillimanite) + quartz, indicating <0.55 GPa/600–700 °C. In the sillimanite-biotite zone, sillimanite and K-feldspar in migmatite may reflect melting of muscovite from the reaction: muscovite + albite (plagioclase) = K-feldspar + Al_2SiO_5 (sillimanite) + melt, indicating >635 °C at 0.40 GPa. Cordierite and K-feldspar in migmatites (Fig. 6I) may form by the reaction: garnet + sillimanite + quartz = cordierite + K-feldspar + melt, indicating <0.50 GPa/650–750 °C. Stable biotite and the lacking cordierite + garnet assemblage suggest that line of the univariant biotite melting reaction (biotite + sillimanite = cordierite + garnet + melt) has not been passed, thus limiting the conditions at <710 °C at 0.40 GPa.

Thermobarometry was performed on a metapelite sampled close to the Millevaches dome in order to test the genetic link between Argentat normal shearing and emplacement of the Millevaches granite. The sample displays a range of assemblages that grew during top-to-the NW shearing related to the normal movement along the Argentat fault. Garnet with ilmenite inclusions (Fig. 7, box E), tourmaline with asymmetric strain shadows of biotite (Fig. 7, box F; Fig. 6J), and a δ-type porphyroblast of andalusite with helicitic inclusions of garnet, biotite, muscovite, tourmaline, ilmenite, and quartz (Fig. 7, box G) stretched with growth of muscovite + chlorite and chlorite + quartz in strain shadows (Fig. 7, box H; Fig. 6H) indicates progressive simple shear deformation and retrograde metamorphism during Argentat normal faulting. The P-T path derived from observed reactions combined to thermobarometry indicates retrograde metamorphism with decompression and heating from 0.49 GPa/605 °C to 0.45 GPa/650 °C (Fig. 7, boxes C–D), then decompression and cooling to ~0.20 GPa/350 °C (Fig. 7, boxes E–H).

7. Discussion

7.1. Evidence for two stages of nappes stacking

7.1.1. Structural mapping

At a regional scale, EU is made in contact indifferently with UGU, LUG or PU along a NW-directed shear zone. In the North Limousin, the South Limousin, and the Rouergue area (Fig. 1), this shear zone removes UGU-LGU thrust and LGU-PU thrust that correspond to SW-directed shear zones (390–380 Ma) reactivated as NW-directed shear zones (355 Ma). These relationships imply that EU was emplaced later than the emplacement of UGU and LGU on the PU.

7.1.2. Partitioning of metamorphism within the Variscan crust

Contrasting metamorphic history between EU, UGU and LGU, and PU, involves a partitioning of metamorphism within the Variscan crust. The Epizonal Unit, devoid of HP rocks, formed a lowly-metamorphosed, upper allochthonous unit that will referred to as LP units. UGU and LGU, which include HP rocks, both formed highly-metamorphosed, lower allochthonous units that will referred to as HP units. The Parautochthonous Unit, devoid of HP rocks, formed a lowly-metamorphosed, parautochthonous domain. The partitioning of metamorphism within the Variscan crust is repeatedly observed in many other locations in the Massif Central region (Fig. 1). It implies that LP units were emplaced later than the emplacement of HP units on the PU.

7.1.3. Middle Devonian magmatism within HP units

In the vicinity of the Tulle antiform (Fig. 2), quartz diorites: (i) cross-cut the UGU-LGU boundary underlined by a SW-verging thrust (390–380 Ma); (ii) were folded by the Tulle antiform formed at 355–345 Ma (Roig et al., 1998); (iii) were cross-cut by the Estivaux shear zone (355–345 Ma; Roig et al., 1996); and (iv) does not occur in the EU. These geometrical relationships imply that quartz diorites were emplaced later than the emplacement of UGU on the LGU and before emplacement of EU on UGU-UIG-PU (i.e. during the 380–355 Ma interval). These relationships are in favour of two stages of nappes stacking in the region.

Geochronological investigations on undeformed and unmetamorphosed quartz diorites in the North Limousin, confirm a Late Devonian age for this magmatism (365–360 Ma; zircon U-Pb) (Pin and Paquette, 2002) inferred to take place in the upper crust (0.3–0.4 GPa) (Bouvier, 1985). In the South Limousin, quartz diorites have recrystallized during under amphibolites-facies metamorphism (0.7 ± 0.1 GPa) (this study). The Early Carboniferous age obtained on recrystallized quartz diorites (355 ± 2 Ma, zircon U-Pb; Bernard-Griffiths et al., 1985) is therefore interpreted to reflect burial of these rocks during the D_2 deformation stage. Moreover, quartz diorites have been undetectable from gravity (Bellot, 2001), suggesting unrooting of the pluton due to thrusting after their emplacement. The history of quartz diorites therefore argue for crustal thickening later than 365–360 Ma, due to NW-verging thrusting of EU on previously stacked UGU-LGU-PU at 355 Ma.

7.1.4. Middle Devonian magmatism within LP units

Metabasites from LP units reflect a common calk-alkaline magmatism (Cabanis et al., 1983; Thieblemont and Cabanis, 1994) emplaced during the Upper Devonian, as stated by whole-rock K-Ar dating on dolerite of the South Limousin area (363 ± 10 Ma) (Cantagrel, 1973), and zircon U-Pb dating on a metagabbro of the Rouergue area (367 ± 10 Ma) (Pin and Piboule, 1988). These data imply that LP units were tectonically emplaced after ∼365 Ma, later than the emplacement of UGU and LGU on the PU (Fig. 8).

All these data suggest two stages of nappes stacking for explaining the present-day crustal structure of the southwestern Massif Central. Two stages of nappes stacking involved that exhumation processes in the Variscan orogenic belt are not continuous.

7.2. Geological differences within the Massif Central

Because the presented model involved the whole Massif Central region, apparent differences in crustal structures, lithology, and metamorphism between western and eastern parts of the region have to be examined from the Paleozoic orogeny point of view.

The Massif Central region is divided into western and eastern parts by the NNE-trending Sillon Houiller fault that has produced a sinistral movement of 70-km-long offset during the Upper Carboniferous (Feybesse, 1981). In a pre-Upper Carboniferous reconstitution, outcropping Paleozoic structures (mainly thrust and strike-slip faults) of both sides of the Massif Central exactly coincide. This is particularly noticeable for Epizonal Unit in the southern Massif Central (Fig. 1). The eastern Massif Central has experienced similar superimposed tectonics events than the western one (Feybesse and Teygey, 1987; Feybesse et al., 1988; Leloix et al., 1999). Refraction-seismic investigations inferred similar crustal structures on both sides of the Sillon Houiller fault (Zeyen et al., 1997).

The main difference between western and eastern sides of the Massif Central concerns lithology of the Lower Gneiss Unit that includes more orthogneisses in the western side (Ledru et al., 1994). Because these orthogneisses are interpreted as parts of granite plutons lately metamorphosed and sheared during polyphased thrusts tectonics (Duthou et al., 1984), these differences in lithology haven't geodynamic significance.

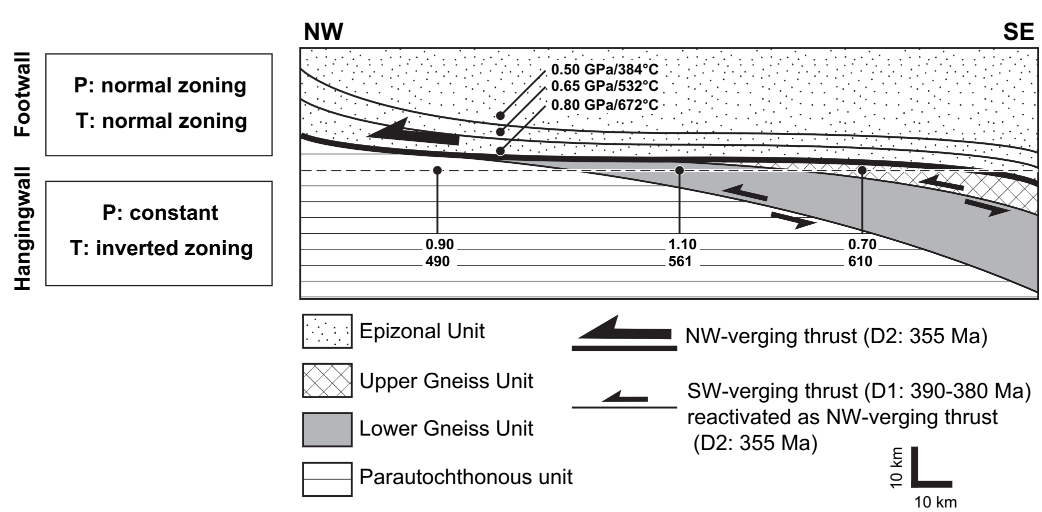

Fig. 8. Our tectonic model at crustal scale attempting to reconstitute crustal-scale structures developed during the D_2 tectonic phase experienced by Paleozoic metamorphic rocks of the Variscan orogeny in Western Europe.

As already pointed out by Gardien et al. (1997), Matte (1998), and Schulz et al. (2001), similar P-T paths were obtained on UGU, LGU, and PU of different sites of the eastern Massif Central. They are evidencing for large-scale tectonics processes occurring at the regional scale (Schulz et al., 2001). Differences in P-T paths mostly reflect variety in exhumation rates and/or the position of sample within the exhumed unit (Mercier et al., 1991).

7.3. Deciphering exhumation processes

7.3.1. Pre- to syn-D_1 exhumation

Pre-D_1 metamorphism of eclogites, with its prograde evolution at LT, its HP-HT peak (1.70 GPa/700 °C) and its retrograde evolution at HT, involves all rock types forming both UGU and LGU, corresponding to pieces of arc/back-arc/passive margin lithospheres (Girardeau et al., 1986; Matte, 1998; Pin and Vielzeul, 1988). Pre-D_1 metamorphism therefore reflects oceanic and/or continental subduction of these domains, and exhumation of HP rocks in the lower crust, during the Silurian (430–390 Ma).

Syn-D_1 metamorphism corresponds to tectonic exhumation of HP rocks to the middle-upper crust and migmatite development (382 ± 5 Ma) in relation to southwestward shearing on HP units bottom (390–380 Ma) (Faure et al., 1997), and northeastward shearing on HP units top, during the Early Devonian. These features are best explained with a model of upward, southwestward extrusion of HP units (e.g. Platt, 1993; Tenczer and Stüwe, 2003), during which they thrusted the Parautochthonous unit (Fig. 9A). Upward extrusion refer to models of subduction-related exhumation, based on physical analogue modeling that predicts breaking and upward extrusion of the continental slab driven by the buoyancy forces and erosion during continental subduction, and scarping of the sediments in front of the slab, (Chemenda et al., 1995, 1996). These models give an explanation for the coeval development of NE-directed detachment and SW-verging thrust faults during plate frontal convergence.

A part of the sediments resulting from erosion of the extruded units may correspond to Early Devonian sandstones that overlie Ordovician turbidites in the Montagne Noire, i.e. a part of the southern external zone (Quémart et al., 1993; Feist et al., 1994). In the northeastern Massif Central, exhumation is achieved before the Middle Devonian (~370 Ma) deposition of limestone on HP rocks (Delfour, 1989). The abundance of these Devonian limestones overlying HP rocks (Delfour, 1989) emphasize that exhumation is mainly accommodated by tectonic processes associated with limited erosion.

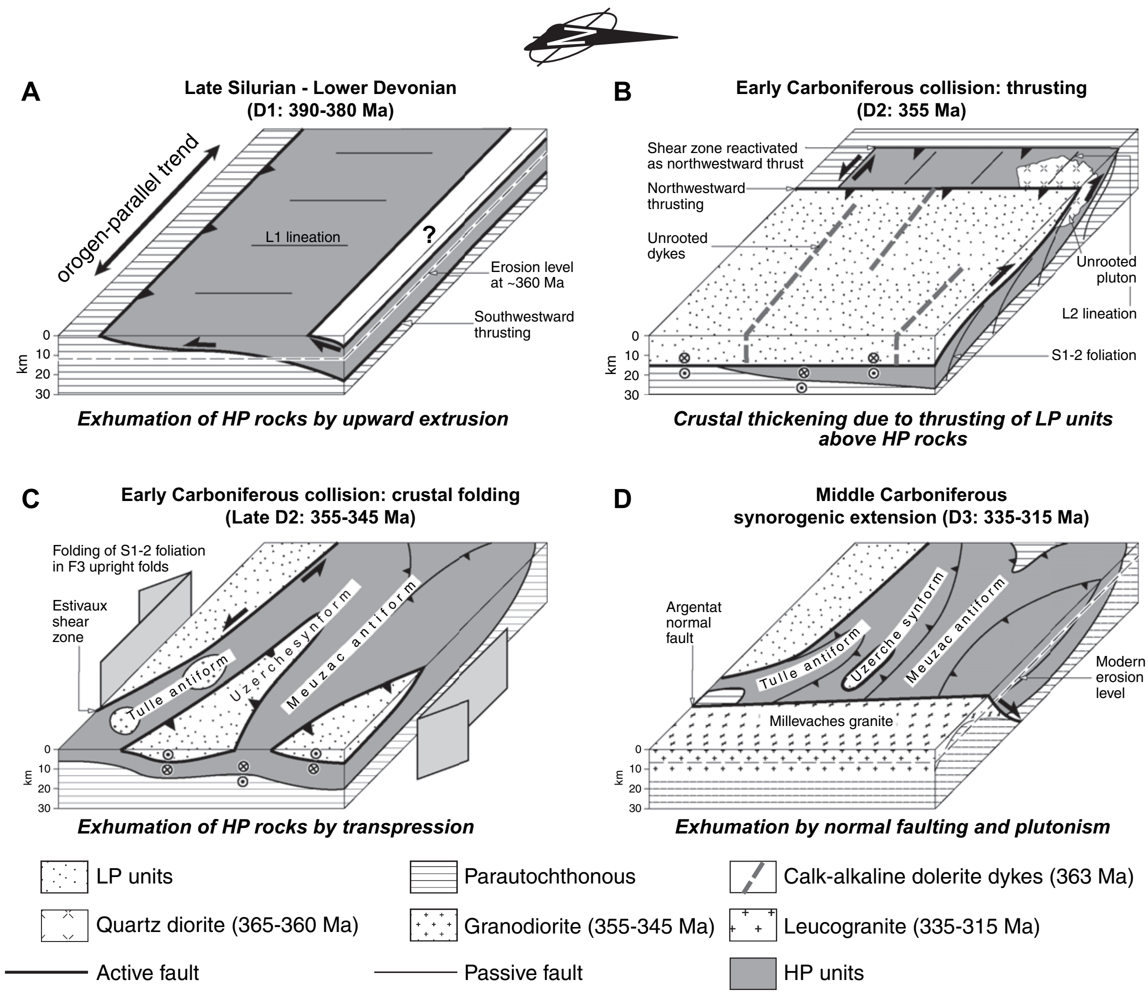

Fig. 9. Our tectonic model at crustal scale attempting to explain exhumation of Paleozoic HP rocks in the context of the Variscan orogeny in Western Europe.

7.3.2. Post-D_1, pre-D_2 rifting

Post-D_1, pre-D_2 events corresponds of calk-alkaline magmatism through the Massif Central. In the northeastern area (i.e. the Brévenne area; Fig. 1), magmatism is particularly intense and associated with rift-type sedimentation (Sider et al., 1986; Wickert, 1988). Extensive geochemical investigations on mafic rocks supports tholeiitic/calk-alkaline magmatism originated from rifting developed within a supra-subduction zone lithosphere (Briand et al., 1994; Pin and Paquette, 1997, 2002). Zircon U-Pb dating on keratophyre (365 ± 10 Ma) (Milési and Lescuyer, 1993), metarhyolite (366 ± 5 Ma) and trondhjemite (358 ± 1 Ma) (Pin and Paquette, 1997) support a common Upper Devonian age for supra-subduction zone rifting. In the Massif Central, the Brevennes Unit is made in contact on UGU or LGU along a NW-directed fault (Feybesse et al., 1988; Leloix et al., 1999). Equivalents of the Brévenne rift (Leloix et al., 1999 and references therein) are found in the eastern Variscides, in the Vosges and Black Forest massifs (Schneider et al., 1989), in the Taillefer and Rioupéroux-Livet formations of the Alpine external massifs (Ménot, 1988), and in the Argentalla formations of northern Corsica (Beaudelot et al., 1976). These correlations support the large extend of the Brévenne rift of which development ceased when syn-D_2 tectonics began. Exhumation of HP rocks may achieve by rifting during the Upper Devonian (~365 Ma).

7.3.3. Syn-D_2 thrusting

Syn-D_2 prograde-retrograde metamorphism at IP (0.8–1.0 GPa) in EU, UGU, and LGU, associated with NW-verging shearing, indicate progressive simple shear at crustal scale during the Early Carboniferous. Syn-D_2 prograde metamorphism in the kyanite field is associated with contrasting zoning on both sides of the main NW-verging shear zone: (1) normal zoning in T and P, in the hangingwall (i.e. LP units); and (2) inverted zoning in T and invariable P peak, in the footwall (i.e. HP units + PU units) (Fig. 8). These relationships indicate that HP units correspond to an autochthon relative to LP units during NW-verging shearing. Syn-D_2 prograde metamorphism is therefore best explained by crustal thickening due to northwestward thrusting of LP units on HP units, during the Early Carboniferous (~355 Ma) (Figs. 8 and 9B).

The structural position and the thickness of the Epizonal Unit (~20–25 km) estimated from the regional map (Fig. 1), both support this hypothesis. These results confirm the Himalayan-type nappes stacking model (e.g. Le Fort, 1975; Pécher, 1989; Hubbard, 1996; Goscombe and Hand, 2000) commonly assumed for the Variscan Massif Central (Briand, 1978; Burg and Matte, 1978; Burg et al., 1984, 1989; Ledru et al., 1994). However, our data point out the northwestward vergence of thrust tectonics responsible for an Early Carboniferous crustal thickening in internal zones of the Variscan belt. This tectonic event reactivated previous low-angle shear zones as northwestward thrusts. The setting of both LP units and NW-verging D_2 tectonics south to the Armorica lithosphere suggests that NW-verging thrusting combined with dextral wrenching along the northern branch of the South Armorica shear zone (Brun and Burg, 1982). We propose that the S-dipping, crustal-scale shear system imaged by a seismic line though the Armorican massif results from this northwestward thrust tectonics (Bitri et al., 2003).

7.3.4. Syn-D_2 collisional exhumation

Syn-D_2 retrograde metamorphism in the sillimanite field reflects tectonic exhumation of HP units (~18 km) during northwestward shearing of HP units, sinistral wrenching along the Estivaux shear zone (345 Ma) (Roig et al., 1996), southeastward shearing in LP units (Roig, 1997), and early stages of regional-scale upright folding (350–345 Ma) (Roig et al., 1998). Accordingly, the Estivaux shear zone is suspected to play a key role in syn-collisional exhumation of HP units. Exhumation by orogen-parallel transpression is predicting by models of oblique collision-related exhumation which are based on the statement that the angle of convergence influences elevation rate, the amount of shortening, and the lateral displacement (Thompson et al., 1997).

The theoretical displacement ($D = 17$ km) along the shear zone is estimated using the shear strain and the thickness of shear zone ($L = 6$ km) from the equation: $D = \int_L \gamma \, dx$, the shear strain ($\gamma = 2.8578$) in mylonitic rocks devoid of steady state recrystallization being accessed using the angle ($\alpha = 7°$, estimated using field analyses and the Fig. 4) between S and C planes from the equation: $\gamma = 2\text{cotg}\ (\alpha)$ (Ramsay and Graham, 1970). For an angle of convergence of 35°, numerical models of Thompson et al. (1997) predict transpression-related exhumation of a 17 km-thick crustal piece in 5 Ma, the resulting elevation rate being 3.3 mm yr^{-1}, and a lateral displacement of 17 km for the exhumed crust. Owing the good fit between numerical models, field observation, and metamorphism, syn-D_2 exhumation is best explained by a model of transpression, controlled by Estivaux shearing, in relation to oblique collision with a low angle of convergence, during the Late Tournaisian-Early Visean interval (350–345 Ma) (Fig. 9C).

7.3.5. Syn-D_3 extensional exhumation

Syn-D_3 metamorphism in the footwall of the Argentat fault corresponds to decompression and heating, followed by decompression and cooling at LP. This metamorphic evolution associated with extensive partial melting and granite emplacement is related to tectonic exhumation in relation to Argentat normal faulting then thermal re-equilibration, in relation to synorogenic extension of Middle Carboniferous age (Fig. 9D). The P-T path implies a minimum value of ~12 ± 3 km for the finite vertical displacement along the Argentat fault, suggesting the crustal scale of extensional tectonics. The Millevaches dome is therefore likely to be a piece of middle crust having experienced granite emplacement and partial melting during its exhumation to the upper crust. Exhumation to the surface of the eastern most hangingwall is completed before the Upper Visean (332 ± 4 Ma) deposition of conglomerates in the northern Argentat fault (Brugier et al., 1998). Extension-related exhumation of HP rocks occurred in the internal zones while crustal thickening caused by southwestward thrust tectonics was still active in the southern external zones.

According to Bellot (2007), exhumation in the Millevaches dome is controlled by normal shearing along the Argentat fault, due to crustal extension, and diapirism, due to addition of syntectonic melt within the dome. Emplacement of a large quantity of syntectonic melt (leucogranite) within the dome led to decrease of the footwall density relative to the hanging-wall density, favouring more efficient exhumation than that produced by normal shearing slip only. Exhumation itself favours partial melting of the fertile rocks and ongoing production of syntectonic melt.

Models of (late-orogenic) extension-related exhumation predict whole lithospheric thinning (e.g. McKenzie and Bickle, 1988), or crustal collapse (e.g. Dewey, 1988) possibly driven by slab break-off, eduction of the subducted continental crust (e.g. Andersen et al., 1991; Michard et al., 1994), lithospheric root detachment (e.g. Platt, 1993), or convective removal of subcontinental lithosphere (e.g. Platt et al., 2003 and references therein). Synorogenic extension of the Variscan belt in the Massif Central seems to be originated from lithospheric root detachment beneath the Brévennes area (Faure et al., 2002).

Exhumation to the surface of both footwall and hanging-wall is achieved by erosional denudation of a crustal thickness of 6 ± 3 km before or during the Middle-Upper Stephanian deposition of conglomerates in Argentat and Hospital basins (Bellot et al., 2005). These events reflect postorogenic extension of Upper Carboniferous age in response to slab break-off (Lardeaux et al., 2001).

Overall, exhumation of Variscan HP rocks in the Massif Central is sequentially originated from: (1) Early Devonian (390–380 Ma) southwestward, upward extrusion during continental subduction; (2) Early Carboniferous (350–345 Ma) orogen-parallel transpression during continental collision; and (3) Middle Carboniferous (335–315 Ma) normal faulting and erosion during orogen-parallel synorogenic extension. Upper Devonian rifting (365–360 Ma) and the subsequent Early Carboniferous (355 Ma) northwestward thrusting of LP units above HP units both prevent to consider exhumation of Variscan HP rocks as continuous. For to be interpret at lithosphere scale, these key results should be confronted to previous models of the Variscan belt.

7.4. Consequences for the Variscan lithosphere history

Two end-member models are proposed for the Variscan belt based on the number (one or two) of orogenic cycles (i.e. subduction-collision-extension stage) for the Silurian-Carboniferous interval. They imply major differences in: (1) the vergence and age of thrust tectonics responsible for crustal thickening; (2) the metamorphism history; (3) the significance of NW-verging shears; (4) the significance of LP units; and (5) polarity of subduction. The one-cycle model (Brun and Burg, 1982; Behr et al., 1984; Burg et al., 1987; Rey, 1992; Ledru et al., 1994; Boutin et al., 1995; Dias and Ribeiro, 1995; Franke, 2000; Matte, 2001) involves: (1) a long and progressive collision during which crustal thickening results from SW-verging thrusts migrated southward from Devonian to Middle Carboniferous; (2) a single P-T history for all units; (3) NW-verging shear zones as detachment faults (Mattauer et al., 1988); (4) LP units originated from meta-sediments of a flexural basin (Rey, 1992); and (5) the northeastward subduction of Gondwana responsible for closure of the Massif Central ocean. The two-cycles model (Cogné, 1977; Sider et al., 1986; Eisbascher et al., 1989; Ziegler, 1989; Faure et al., 1997; Roig and Faure, 2000; Pin and Paquette, 2002) involves: (1) a short Carboniferous collision during which crustal thickening results from NW-verging thrust tectonics; (2) two P-T cycles in internal zones; (3) NW-verging shear zones as thrust faults; (4) LP units originated from a back-arc basin associated with a thinned lithosphere resulting from either the collapse of a previous subduction-collision belt or back-arc spreading (Faure et al., 1997); and (5) the southward subduction of Armorica responsible for closure of the Rheic ocean.

These opposing models explain specifically some of the data, but none consider the whole data set. On one hand, the one-cycle model that involves a northward subduction best explains: (1) S-verging thrusting early in the internal zones (390–380 Ma) and lately in the southern external zones (365–310 Ma) (Souquet et al., 2003; Arnaud et al., 2004); and (2) the synchronous end (315 Ma) of S-verging thrusting in the southern external zones and of synorogenic extension in the internal zones, suggesting detachment of the northward subducted slab (Lardeaux et al., 2001). On the other hand, the two-cycle model takes into account: (1) back-arc spreading within the northeastern Massif Central; and (2) the two exhumation stages of HP rocks of the southwestern Limousin, separated by Early Carboniferous NW-verging thrusting (this study). A consensual model at a large scale is required for explaining all these features, and particularly the origin of the Devonian rift (those width and geometry are unknown), and the roots of Early Carboniferous NW-verging thrusts, in a context of N-S-trending lithosphere convergence history. The resulting model will therefore best explain how the mechanical evolution of the lithosphere has controlled exhumation of Paleozoic HP rocks.

8. Conclusion

Constrained by microstructural data and broadly extended by large-scale correlations, the tectonic interpretation of metamorphic assemblages is a powerful method for deciphering tectonic processes responsible for exhumation of HP rocks and for reconstituting the corresponding lithosphere history. In internal zones of an orogenic belt like the Massif Central, exhumation of HP rocks is likely to be episodic and results from an 80-Ma-long sequence of tectonic processes effective at various scales and active specifically at each main stage of the evolution of the involved lithosphere. Upward extrusion exhumes at lithosphere scale during continental subduction, while transpression exhumes at crustal scale during continental collision, depending on the angle of convergence, and normal faulting exhumes at upper-middle crustal scale during synorogenic extension, depending on the extension rate. Although postorogenic extension considerably reworked the lithosphere architecture by slab break-off, it slightly contributes to exhumation by regional uplift and erosional denudation, and only concerns shallows crustal levels. Most of the exhumation of HP rocks achieved before postorogenic collapse concludes of the key role played by subduction and collision of the continental lithosphere for the return of deep-seated rocks to the surface in internal parts of mountain belts like the Variscan orogenic belt of Europe. The efficiency of exhumation mechanisms depends on the angle of convergence, and so on the location of HP rocks along the irregular continental margin.

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