1
2
3
4 **Syn-collapse eclogite metamorphism and exhumation of deep crust in a migmatite dome:**
5 **the P-T-t record of the youngest Variscan eclogite (Montagne Noire, French Massif**
6 **Central)**
7
8

9 Donna L. Whitney¹, Françoise Roger², Christian Teyssier¹, Patrice F. Rey³, and J-P. Respaut²

10
11 ¹ Department of Earth Sciences, University of Minnesota, Minneapolis MN 55455 USA
12 (dwhitney@umn.edu)
13

14 ² Laboratoire Géosciences Montpellier (CNRS-UMR 5243), Université Montpellier 2, 34095
15 Montpellier Cedex 5, France
16

17 ³ School of Geosciences, University of Sydney, Sydney, NSW 2006, Australia
18
19
20
21
22

Abstract

In many orogens, high-pressure (HP) metamorphic rocks such as eclogite occur as lenses in quartzofeldspathic gneiss that equilibrated at much lower pressures. The pressure-temperature-time (P-T-t) history of eclogite relative to host gneiss provides information about mechanisms and timescales of exhumation of orogenic crust. The Montagne Noire of the southern Massif Central, France, is an eclogite-bearing gneiss (migmatite) dome located at the orogen-foreland transition of the Variscan belt. Results of our study show that it contains the youngest eclogite in the orogen, similar in age to migmatite and granite that crystallized under low-pressure conditions. P-T conditions for an exceptionally unaltered eclogite from the central Montagne Noire were estimated using a pseudosection supplemented by garnet-clinopyroxene and Zr-in-rutile thermometry. Results indicate peak $P \sim 1.4$ GPa and $T \sim 725^\circ\text{C}$ for Mg-rich garnet rim (50 mol% pyrope) + omphacite (36 mol% jadeite) + rutile + quartz. U-Pb geochronology (LA-ICP-MS) of 16 zoned zircon grains yielded ~ 360 Ma (4 cores) and ~ 315 Ma (12 rims and cores). Rare earth element abundances determined by LA-ICP-MS for dated zircon are consistent with crystallization of ~ 315 Ma zircon under garnet-stable, plagioclase-unstable conditions that we interpret to indicate high pressure; in contrast, the ~ 360 Ma zircon core corresponds to crystallization under lower pressure plagioclase-stable conditions. Based on garnet zoning and inclusion suites, rutile textures and Zr zoning, P-T results, and zircon petrochronology, we interpret the ~ 315 Ma date as the age of eclogite-facies metamorphism that only slightly preceded dome formation and crystallization at 315-300 Ma. This age relation indicates that eclogite formation at high pressure and migmatite dome emplacement at low pressure were closely spaced in time. We propose that collapse-driven material transfer from the hot orogen to the cool foreland resulted in thickening of the orogen edge, leading to eclogite facies

metamorphism of the deep crust. Soon after, the low-viscosity partially molten crust flowed from the plateau toward the foreland, incorporating and exhuming eclogite. The P-T-t history of the Montagne Noire eclogite shows that some dome material ascended from >40 km depth to shallow crustal levels, likely in a single decompression event, and that migmatite domes are therefore very efficient at exhuming the deep crust.

Keywords: crustal flow, eclogite, French Massif Central, gneiss dome, migmatite, Montagne Noire

1. Introduction

Partially molten crust flows laterally in the deep crust in response to pressure gradients, such as occur at the transition from a thick orogenic plateau to its adjacent foreland (Fig. 1). This hot crust may ascend to shallow crustal levels in response to instabilities such as upper crustal extension, forming domal structures comprised of migmatite (the crystallized partially molten crust) and crustally derived granite (e.g., Teyssier and Whitney, 2002; Rey et al., 2011) (Fig. 1). This mechanism can transport high-pressure (HP) rocks from great depth (>30-40 km) to very shallow levels (<2-10 km) (Whitney et al., 2004). Different rock types experience different degrees of equilibration during decompression, resulting in the common occurrence of HP bodies (eclogite or HP granulite) within host gneiss/migmatite that records significantly lower pressures (e.g., Stipska et al., 2008). Unresolved questions are when, where, and how eclogite metamorphism occurred, and how eclogite pressure-temperature-time (P-T-t) history relates to that of the host rock.

In the case of HP rocks exhumed as part of a collisional orogenic event, HP metamorphism may have occurred by pre-collision oceanic subduction, syn-collision continental subduction, or at depth in thickened orogenic crust. In the case of high-pressure bodies in migmatite domes, HP metamorphism may significantly pre-date migmatization and the dome-forming event (e.g., if two events occurred during different orogenies or during very different temporal stages of the same orogeny), or HP metamorphism may be more closely related in time with migmatization and doming. In the latter case, mineral compositions, zoning, textures, and age data may provide information about the orogenic events that produced both HP rocks and the host migmatite.

In this paper, we consider the exhumation of meter- to kilometer-scale eclogite inclusions in a migmatite dome: the eclogite-bearing Montagne Noire dome, French Massif Central (Figs. 2, 3). We document the petrology (including pressure-temperature conditions) of the least-altered eclogite collected as part of this study, present U-Pb geochronology and trace element abundances for zircon in eclogite, and discuss the new P-T-time (*t*) results in the context of dome dynamics and orogenic evolution. We show that the Montagne Noire eclogite is the youngest in the Variscan belt, postdating the contractional history of the orogen, requiring an explanation for crustal thickening at the orogen-foreland boundary zone during collapse of the orogenic crust, followed rapidly by ascent of eclogite in partially molten crust >40 km upward to the shallow crust.

2. Montagne Noire geologic setting and metamorphic petrology

Located at the southern edge of the French Massif Central (Figs. 2, 3a), the Montagne Noire dome occupies the transition between the Variscan orogen to the north and the foreland to the south (e.g., Faure, 1995, 2009). To the north of the Montagne Noire, the eastern Massif Central

crust (east of the Sillon Houiller, Fig. 2) can be described in terms of an upper unit derived from oceanic assemblages (with numerous eclogite pods), a middle unit dominated by metasedimentary rocks (especially micaschist, such as the schist units of the Cévennes region), and a lower unit consisting of quartzofeldspathic migmatite and peraluminous granitoids exposed in domes (Vanderhaeghe et al., 1999). This region of the Variscan orogen was accreted during the Carboniferous, with contraction starting around 340-335 Ma and lasting until ~325 Ma based on ages of synorogenic flysch deposits in the orogen foreland. According to Faure et al. (2009). The core of the southern Massif Central orogen underwent extension during 320-295 Ma (e.g., review by Lardeaux, 2014), while the outer zone represented by the Montagne Noire and the Pyrenees remained in contraction (Faure et al., 2009). Recent structural and geochronological work on the core of the Montagne Noire indicates dome emplacement, internal deformation, and melt crystallization between ~315 and 300 Ma (Roger et al., 2015).

The Montagne Noire consists of three units (Fig. 3b). The northern unit is a SW-directed thrust imbricate involving low-grade lower Paleozoic metasedimentary strata. The central unit (Axial Zone) consists of a granite-migmatite dome, mainly comprised of migmatitic orthogneiss (augen gneiss), paragneiss, and crustally derived granite within two elongate domes (e.g., Aerden, 1998; Bouchardon et al., 1979; Demange et al., 1996; Géze, 1949; Schuiling, 1960), the Espinouse and Caroux domes (or subdomes; Rey et al., 2011). Cordierite-bearing granite and gneiss occur in the cores of the domes, and cordierite-bearing schist ($\pm$ andalusite, sillimanite) occurs in the mantling units (Fig. 3b), indicating that the later stages of dome formation were associated with low-pressure/high-temperature (LP/HT) metamorphism, up to the conditions of partial melting (e.g., Thompson and Bard, 1982). Elongate sillimanite + quartz nodules occur in

high melt-fraction migmatite (Brunel and Lansigu, 1997), indicating deformation at high temperature ($> 650^{\circ}\text{C}$).

The Axial Zone also includes minor micaschist, marble/calc-silicate, and metamorphosed mafic to ultramafic layers and pods, including a few localities with eclogite-facies assemblages in metabasaltic rocks (Demange, 1985; Faure et al., 2014). Rare occurrences of relict (texturally early) kyanite in and near the migmatite dome core (Bouchardon et al., 1979) are another indicator that higher-pressure metamorphism preceded LP/HT metamorphism.

The southern unit of the Montagne Noire consists of the “Nappes du Versant Sud” (nappes of the southern flank, Fig. 3b), in which Paleozoic overturned sequences form the inverted limbs of a pile of thin nappes emplaced at 335-325 Ma (Engel et al., 1980) or at < 320 Ma (Franke et al., 2011). Doublier et al. (2015) used the results of their petrologic and geochronologic investigation to show that the nappes record a metamorphic imprint at 310-300 Ma in proximity to the Axial Zone, whereas ages of metamorphism away from the Axial Zone are 20 million years older.

3. Eclogite petrology

In the Montagne Noire dome, metabasalt occurs primarily as amphibolite. It is also present as rare layers and lenses of eclogite (Fig. 3b, c) consisting of garnet + omphacite + quartz + rutile (Demange, 1985; Faure et al., 2014). A garnet + enstatite-bearing rock from the eastern Caroux region (Airette locality; Fig. 3b) has been interpreted as an “eclogite-facies ultramafite” originally formed as a cumulate from basaltic magma (Demange, 1985). Retrogressed mafic rocks that crop out in the eastern Caroux, some with garnet, may be former eclogite (Faure et al., 2014). Other eclogite localities are in the central region (3 sites: Le Jounié, Terme de Fourcaric/Peyrambert, Cantausse) and the southwest (Cabardès) (Fig. 3b).

In this paper, we report new petrographic and mineral composition data for the Terme de Fourcaric eclogite locality (Figs. 3b, c), as this is where we have found the least-altered eclogite. In most cases, eclogite was not observed in outcrop; samples were collected from areas with high concentrations of eclogite boulders (Fig. 3c), most several meters in diameter. Boulders of eclogite are clustered in elongate strips that correspond to the map patterns on the published geologic map (Alabouvette et al., 1993), and eclogite float is also found in a small (meter-wide) creek that flows ~250-200 m from the elongate, concentrated area of eclogite boulders to a marshy area (Fig. 3c).

The Terme de Fourcaric locality (Figs. 3b, c), near the Peyrambert farm (site of a sample dated by Faure et al., 2014; Fig. 3c), contains eclogite with garnet + omphacite + rutile + (minor) quartz (Fig. 4). The studied eclogite (sample MN13-11A), which is foliated and lineated, consists of garnet (45 modal%), omphacite (35%), rutile (4%), and fine-grained amphibole-plagioclase symplectite (16%) (Fig. 4a). Quartz is an abundant inclusion in garnet (Fig. 4b) and is more rare in the matrix. Zircon and rutile also occur as inclusions in garnet and are present in the matrix, along with texturally late apatite and patches of ilmenite that occur in rutile.

Garnet is typically subhedral, 2-5 mm in diameter, and zoned (Figs. 4c, d). Garnet core composition (corresponding to the quartz inclusion-rich region) is 43 mol% almandine, 2 mol% spessartine, 32 mol% pyrope, and 23 mol% grossular. The rim (corresponding to the rutile inclusion-bearing region) is more pyrope-rich and lower in both almandine and grossular components: 36 mol% almandine, 0 % spessartine, 50 mol% pyrope, and 14 mol% grossular (Fig. 4c, d; Table A1).

Matrix clinopyroxene is omphacite (Appendix A) and occurs as tabular grains that are ~1-3 mm long. Some have thin ($\leq 100 \mu\text{m}$) diopside-rich rims, but otherwise the jadeite content of the omphacitic part of the grains is quite homogeneous at ~35-36 mol%.

Rutile has three habits (Figs. 5a-c): inclusions in omphacite and in the near-rim region of garnets (typically rounded and in some cases in clusters in garnet), blocky matrix grains that have straight grain boundaries with omphacite, and more irregular, elongate grains (commonly with apatite, ilmenite, and zircon) along grain boundaries. Rutile commonly contains ilmenite lamellae and patches (Figs. 5d, e).

4. Thermobarometry

To estimate pressure and temperature of metamorphism, we constructed a bulk-composition specific phase diagram (pseudosection) for the Terme de Fourcaric eclogite. Because this method is dependent on accurate assumptions about the effective bulk composition, which may change during progressive metamorphism, we also applied garnet-clinopyroxene Fe-Mg exchange thermometry and Zr-in-rutile trace element thermometry. Previous studies estimated pressure at (1) $0.9 \pm 0.2 \text{ GPa}$, using the jadeite content of omphacite for eclogite from the Cabardès locality (28 mol% jadeite) (Demange, 1985), or (2) $>1.4 \text{ GPa}$ (Franke et al., 2011), based on an unpublished pseudosection described as having been calculated using an eclogite whole-rock composition of Demange (1985). Previous temperature estimates are (1) 700-800°C (grt-cpx thermometry, Le Jounié and Cabardès eclogites; Demange, 1985), and (2) $>650^\circ\text{C}$ (Franke et al., 2011, unpublished pseudosection).

4.1 Pseudosection analysis

For phase-diagram construction, we used the compositional system MnNCFMAS(H)Ti and Perple_X software (Connolly, 2005; hpver11 database); solution models were applied for garnet (Holland and Powell, 1998) and omphacite (Holland and Powell, 1996). The bulk composition of the Terme de Fourcaric eclogite (MN13-11A) was determined using a Philips PW-2400 X-Ray Fluorescence Spectrometer at Macalester College, Saint Paul, Minnesota, USA (Fig. 6). K is negligible (0.02 wt% K₂O) and was excluded from the calculation. There is no evidence (e.g., from matrix or inclusion mineral assemblages or mineral stoichiometry) that Fe³⁺ was a factor during eclogite facies metamorphism, so we assumed all Fe was ferrous. This assumption may be invalid for the retrograde part of the metamorphism (e.g., during formation of the plagioclase-amphibole-ilmenite assemblage), and possibly also the prograde history, but in this paper we focus on attempting to model the conditions of eclogite facies metamorphism.

The eclogite facies assemblage consists entirely of anhydrous minerals (and the loss-on-ignition value – assumed to represent the volatile content – was less than detection limits), so we assumed that H₂O was not present in excess, although it likely was at earlier stages of metamorphism. Exploratory T-X_{H₂O} diagrams (not shown) constructed with the XRF-determined whole-rock composition suggest that prograde assemblages likely involved a pargasitic amphibole, perhaps similar to those observed in the retrograde symplectite.

A pseudosection constructed with the XRF-determined bulk composition in the MnNCFMAS(H)Ti system predicts the assemblage garnet + omphacite + rutile + (minor) quartz at P > 1.2 GPa and ~650°C (Fig. 6a) (the assemblage is apparently stable at T < 650°C but clinopyroxene is essentially diopside at those conditions). This bulk composition, however, is unlikely to represent the *effective bulk composition* during eclogite facies metamorphism, which garnet zoning and inclusion distribution indicate corresponds to the garnet rim composition.

To determine the effective bulk composition relevant to eclogite facies metamorphism, we used the garnet rim composition and unzoned omphacitic clinopyroxene composition as determined by electron microprobe (Appendix A) with a mineral mode calculated for garnet rim, clinopyroxene, rutile, and the assumption that all amphibole + plagioclase symplectite represents former clinopyroxene (as suggested by textural observations) (Fig. 6b). To account for inaccuracy in estimating the effective bulk composition using this method, we investigated the sensitivity of the pseudosection to effective bulk composition using a range of garnet : omphacite modal proportions (from 35 to 70 modal % garnet) and T-X diagrams ($X = \text{H}_2\text{O}, \text{MgO}$). Minor variation in rutile mode (few %) did not significantly affect the calculated diagram.

For each phase diagram, we calculated isopleths for garnet ($X_{\text{prp}}, X_{\text{alm}}, X_{\text{grs}}$) and clinopyroxene (X_{jld}) composition and mode as an additional means for evaluating the phase diagrams. The phase diagrams calculated for 55 ± 5 modal% garnet most closely represent the composition of the eclogite facies assemblage (cf. 45% modal garnet observed in thin section). There is, however, no single point on any phase diagram that we calculated at which *all* compositional isopleths intersect at values corresponding to those measured. In particular, the grossular content of garnet is not accurately modeled. Possible explanations for discrepancies include inaccurate assumptions about bulk composition or chemical system, modification (e.g., by diffusion at high temperature) of mineral compositions/zoning, or problems with the solution models (the solid solutions are at or near the maximum range of compositions for the solution models used; e.g., pyrope and grossular in garnet, jadeite in clinopyroxene). Nevertheless, the pseudosections provide some useful general information on P-T conditions.

The phase diagram for the assumed effective bulk composition – with consideration of the P-T conditions at which most garnet and clinopyroxene compositional isopleths intersect – predicts

that the assemblage garnet + omphacite + rutile + (minor) quartz is stable at $P > \sim 1.4$ GPa at 725-750 °C (Fig. 6b). Decreasing the amount of garnet relative to clinopyroxene in the inferred mode raises the minimum pressure stability of the eclogite facies assemblage by ~ 0.10 - 0.15 GPa for every 5% decrease in amount of garnet; across all calculations, the minimum P of the eclogite-facies mineral assemblage is ~ 1.2 - 1.5 GPa at $T \sim 700$ - 800°C . There is no evidence in mineral compositions or modes that pressure was considerably higher than 1.5 GPa. At lower pressures (< 1.3 GPa), albite is stable, and at lower temperature the clinopyroxene is diopside. Ilmenite is stable at $P < 1.1$ - 1.0 GPa (Fig. 6b). We therefore conclude that $P \sim 1.4$ GPa is a reasonable estimate of the maximum pressure attained during eclogite metamorphism.

4.2 Garnet-clinopyroxene Fe^{2+} -Mg exchange thermometry

We applied various grt-cpx calibrations to the Terme de Fourcaric eclogite for garnet rim-matrix cpx compositions and $P = 1.4$ GPa. Stoichiometric (charge balance) calculations of garnet and clinopyroxene indicate that Fe^{3+} is minor to absent, so we assumed that all Fe in these minerals is Fe^{2+} . Results for a wide range of calibrations were all in the range 700-750°C (e.g., Nakamura, 2009: 700°C; Ellis and Green, 1979: 740°C), similar to temperature estimates of Demange (1985) for the Cabardès ($700 \pm 50^\circ\text{C}$) and Le Jouié ($800 \pm 100^\circ\text{C}$) eclogites using garnet-clinopyroxene exchange thermometry.

4.3 Zr-in-rutile thermometry

The Zr content of rutile was measured in the Terme de Fourcaric eclogite because the Zr content of rutile varies with temperature (Zr increases with increasing T) and is therefore applicable as a geothermometer (e.g., Tomkins et al., 2007). Data from 17 matrix grains and 36 inclusions in

garnet were acquired with a JEOL 8900 electron microprobe in the Department of Earth Sciences at the University of Minnesota (Appendix B). For grains that exceed 10 μm in diameter, multiple analyses per grain were obtained..

Rutile is zoned with respect to Zr (Fig. 5). Matrix grains have a typical composition of 400-500 ppm Zr, and this increases over a distance of 150-200 μm to ~800-1300 ppm near ilmenite grains that have partially replaced rutile (Figs. 5d, e). This zoning is likely related to the fact that ilmenite excludes Zr, so rutile near ilmenite replacement is enriched in Zr. Further evidence for the effect of ilmenite replacement on rutile Zr content is seen in the relation of Zr and Fe content in rutile: away from ilmenite, there is little to no correlation of Zr and Fe, but in blocky matrix rutile that has been partially replaced by ilmenite, Fe and Zr both increase toward ilmenite (Figs. 5d₃, e₃).

For temperature calculations, we used the Tomkins et al. (2007) calibration and an estimated pressure of 1.4 GPa. Results for isolated rutile inclusions (i.e., not in cracks) in garnet near-rim regions that also contained quartz inclusions range from ~670 to 780°C, with an average of ~720°C. Blocky matrix rutile yields Zr-in-rutile temperatures of 700-715°C for the lowest values of Zr in each grain (i.e., not near ilmenite; Figs. 5d, e). Elongate (grain boundary) rutile records $T = 650\text{-}780^\circ\text{C}$; the average Zr content corresponds to $T \sim 740^\circ\text{C}$. These results indicate that rutile of all textural types records similar average temperatures (700-740°C) (also to the ~700-750°C results of grt-cpx thermometry at the same pressure). This is despite some variation in composition likely related to partial replacement of rutile by ilmenite and possible disequilibrium among rutile, zircon, and quartz (the assemblage required for the thermometer) for some textural types of rutile (e.g., elongate grain-boundary rutile).

Results of the pseudosection analysis (including consideration of key mineral compositions and modes), a major-element exchange thermometer (garnet-clinopyroxene Fe-Mg), and a trace-element thermometer (Zr-in-rutile) consistently indicate that the Terme de Fourcaric eclogite records P-T conditions of ~ 1.4 GPa at $\sim 725 \pm 25^\circ\text{C}$. Zr zoning in rutile records part of the decompression path as ilmenite partially replaced rutile (Fig. 5).

6. U-Pb zircon geochronology and trace element analysis

To determine the age of eclogite metamorphism in relation to metamorphism and crystallization of migmatite and granite in the Montagne Noire, we obtained U/Th-Pb dates and trace element (primarily REE) concentrations of zircon crystals in the Terme de Fourcaric eclogite (MN13-11A) (Figs. 7, 8; Appendix C, D, E).

6.1 Geochronology methods and results

U-Pb analyses of zircon crystals separated from the Terme de Fourcaric eclogite were analyzed using Laser Ablation–Inductively Coupled Plasma Mass Spectrometry (LA-ICPMS) at the Laboratoire Magmas et Volcans, Clermont-Ferrand (France). Analytical methods and detailed results are in Appendix C. Analyzed zircon crystals are transparent, colorless, and contain few inclusions. Grain sizes are 50-100 μm , and crystals are rounded with some facets (Fig. 7a; Fig. A1). BSE images show that many grains have brighter cores and darker rims; others have patchy zoning (Fig. 7a, Appendix D). Results from 31 analyses (core, rim) of 16 crystals were obtained (Fig. 7b; Appendix C, D).

Two inherited core analyses are discordant (concordance = 69-74%) with similar $^{238}\text{U}/^{206}\text{Pb}$ ages of ~ 450 - 460 Ma and higher Th/U ratios (0.12 and 1.34) compared to other analyses of these

zircons (Fig. 7b; Appendix C). These Ordovician dates are identical to the zircon U-Pb ages obtained for augen gneiss of the Axial Zone (Cocherie et al., 2005; Faure et al., 2010; Roger et al., 2004, 2015). Discordance is probably the result of common Pb contamination; all other spot analyses are concordant to sub-concordant (Fig. 7b).

There are two distinct age populations, as seen on the concordia diagram (Fig. 7b): one population is concordant, with a concordia age of 315.3 ± 1.6 Ma ($\pm 2\sigma$). This population represents the majority of analyses (25) and consists of both cores and rims. A smaller population is represented by the core analyses of 4 zircons. These are concordant to sub-concordant, with a weighted average $^{206}\text{Pb}/^{238}\text{U}$ age of 359.5 ± 4.7 Ma. Zircons with ~360 Ma core ages have ~315 Ma rims (Fig. 7a; Appendix D).

Both core and rim populations are characterized by low Th (1-22 ppm) and moderate U (89-327 ppm) concentrations, and very low Th/U ratios (0.01-0.10) (Appendix C). Such low ratios are typical of non-magmatic zircon, such as those crystallized in metamorphic or hydrothermal settings (Rubatto, 2002).

6.2 Zircon trace element analysis: methods and results

Using LA-ICP-MS, REE analyses were acquired on the dated zircon crystals to infer the geochemical context of zircon crystallization (Appendix E). Because the core regions are small (Fig. 7a; Appendix D), in some cases the spot was not completely within the core when attempting to analyze the ~360 Ma zircon core regions.

With only two exceptions (24-Zr38c and 35-Zr12c) out of 30 spot analyses, the chondrite normalized REE patterns show flat HREE slopes ($\text{Lu}_\text{N}/\text{Dy}_\text{N} = 0.63\text{-}1.55$) and no significant Eu anomalies ($\text{Eu}^* = -0.18$ to 0.12) (Fig. 8). In our dataset the two exceptions are both in the core

regions of zoned zircons; these cores are more enriched in HREE ($\text{Lu}_\text{N}/\text{Dy}_\text{N} = 7.20$ and 4.75) (Fig. 8). One of these cores has been dated and yielded an age of ~360 Ma (Fig. 7a; Appendix C).

7. Discussion

Results of this study provide new information about the P-T-t path of eclogite in the central region of the Montagne Noire. These data can be compared with previous results from the host rocks and other aspects of the Montagne Noire tectonic history to understand the relationship between eclogite facies metamorphism and the formation and exhumation of the migmatite dome. In addition, results are the basis for a general discussion of the processes of eclogite metamorphism and exhumation in the context of crustal flow and migmatite dome emplacement.

7.1 The age of eclogite metamorphism

Eclogite – including coesite-bearing (ultrahigh-pressure, UHP) eclogite – is spatially associated with other migmatite domes of the Variscan belt, although there are structural and age differences between these eclogites and those described here for the Montagne Noire dome. UHP eclogites of the eastern French Massif Central are of oceanic affinity and occur with mafic and felsic granulite in a gneissic unit that structurally overlies the migmatitic Velay dome (e.g., Lardeaux et al., 2001). The age of U/HP metamorphism of Massif Central eclogite-facies rocks from at least four regions has been determined as 420-400 Ma (Limousin: Berger et al., 2010; La Borie –Haut Allier: Ducrot et al., 1983; Monts Lyonnais: Lardeaux et al., 2001; La Bessenoits massif – Lot: Paquette et al., 1995).

The age of eclogite metamorphism in the Montagne Noire was also previously thought to be ~430 Ma (U-Pb zircon lower intercept age) with a basaltic protolith age of ~1.9 Ga to 600 Ma

(Gebauer and Grünenfelder, 1982). Recent work by Faure et al. (2014), however, determined a 3-point (grt-cpx-whole rock) Sm-Nd isochron age of 358 ± 9 Ma for an eclogite boulder collected near Peyrambert (Fig. 3c). The age of Montagne Noire eclogite metamorphism is therefore most likely Variscan. The Sm-Nd date of Faure et al. (2014) is similar to a ~ 360 Ma population in zircon core U-Pb results determined in our study.

Faure et al. (2014) also reported U-Pb SHRIMP and SIMS ages for zircon and rutile for the Peyrambert eclogite: 311 ± 2 Ma (U-Pb zircon), and 308 ± 4 Ma (U-Pb rutile), similar to our U-Pb results for zircon rims and some cores. Because their U-Pb ages were younger than U-Pb ages of migmatite and granite in the Montagne Noire, Faure et al. (2014) interpreted the ~ 358 Ma Sm-Nd age to represent the timing of high-pressure metamorphism. The younger U-Pb zircon and rutile ages were interpreted to indicate a “hydrothermal” event associated with significant retrogression of the anhydrous eclogite-facies minerals to amphibole-plagioclase symplectite $\pm$ biotite. The eclogite sample analyzed by Faure et al. (2014), however, is significantly retrogressed; e.g., see their Fig. 6a, which shows a garnet crystal in which the rim has been entirely replaced by amphibole-plagioclase symplectite. This raises the question of whether garnet rims (which we interpret as the HP part of the garnet in the Terme de Fourcaric eclogite) were adequately represented in their Sm-Nd analysis, and therefore what P-T conditions correspond to the ~ 358 Ma and ~ 315 Ma events.

The Terme de Fourcaric eclogite that contains the zircon that we dated shows only minor retrogression (Fig. 4a) and is essentially anhydrous in the whole-rock analysis (Fig. 6a). It is therefore difficult to ascribe widespread zircon crystallization at ~ 315 - 310 Ma to a hydrothermal event for the Terme de Fourcaric eclogite. Some Zr may have been released by rutile during incipient replacement by ilmenite during decompression from HP conditions to $P < \sim 1$ GPa.

However, it is difficult to account for widespread zircon crystallization by this process owing to the challenge of transporting and precipitating Zr in a largely anhydrous rock. Most rutile, whether as inclusions in garnet and omphacite or as blocky matrix grains with straight grain boundaries with garnet and omphacite, appears to have been part of the eclogite facies assemblage, whereas textures (Fig. 5) and the calculated phase diagram (Fig. 6) indicate that ilmenite was part of the retrograde assemblage. If most or all rutile was part of the eclogite facies assemblage, the 308 ± 4 Ma U-Pb rutile age of Faure et al. (2014) may indicate the age of eclogite metamorphism or a cooling age that is likely only slightly younger than peak eclogite metamorphism owing to rapid exhumation and cooling of the dome rocks and included eclogite.

Perhaps the most compelling evidence for the timing of eclogite metamorphism is the REE concentrations of ~315 Ma zircon. REE signatures such as those documented in these zircons (flat HREE, no Eu anomaly; Fig. 8) are typically interpreted to indicate the presence of garnet and the absence of plagioclase during zircon crystallization; i.e., eclogite-facies conditions (Rubatto, 2002). The conclusion that eclogite metamorphism occurred at ~315 Ma – making the Terme de Fourcaric locality the youngest Variscan eclogite – is consistent with garnet zoning trends and inclusion distribution (rutile in pyrope-rich garnet rim zones), the interpretation that rutile was part of the eclogite facies assemblage, and the lack of significant retrogression in the dated Fourcaric eclogite.

If zircon grew during the retrogression in which plagioclase crystallized, it should have a steep HREE pattern and a negative Eu anomaly. Such a pattern is observed only for the ~360 Ma zircon core. The nature of this ~360 Ma event is not known, and it is puzzling why the ~360 Ma event recorded in eclogite is not recorded in Montagne Noire migmatite or granite. Faure et al. (2014) interpreted ~360 Ma as the age of high-pressure metamorphism and proposed that

intracontinental subduction occurred at this time. In the context of our interpretation of ~315 Ma eclogite metamorphism, however, two possibilities are that 360 Ma is (1) the protolith age (crystallization of mafic magma); or (2) the age of a pre-eclogite facies (likely amphibolite facies) metamorphism. The second scenario is more consistent with petrologic and geochemical evidence; i.e., zircon has low Th/U, and garnet zoning (lower pyrope in core relative to rim) and inclusion suites (primarily quartz) are more consistent with metamorphism than mafic magmatism.

If eclogite facies conditions were attained at ~315 Ma, high-pressure metamorphism was part of the orogenic event that created the Montagne Noire dome and was closely followed by the high-temperature, dome formation-exhumation event that led to crystallization of migmatite and granite at relatively shallow crustal levels between ~315 and 300 Ma (e.g., Franke et al., 2011; Roger et al., 2015). In the following section, we explore tectonic scenarios that could explain eclogite facies metamorphism and exhumation in the context of migmatite-dome formation.

7.2 Eclogite in migmatite: metamorphism and exhumation of HP rocks in a migmatite dome

Many different models have been proposed for the formation of the Montagne Noire dome, including diapirism (e.g., Charles et al., 2009) during extension/collapse (Soula et al., 2001), shortening during compression or transpression ($\pm$ erosion) (e.g., Demange, 1999; Matte et al., 1998), gravitational collapse (Aerden, 1998), and Cordilleran-style core complex development by extension (e.g., Echtler and Malavieille, 1990; van den Driessche and Brun, 1992). These same models have also been debated for other migmatite domes worldwide (Whitney et al., 2004).

Although a variety of mechanisms can account for incorporation of high-pressure rocks in gneiss/migmatite and eventual exhumation of these rocks together, ascent of migmatite in an extensional or transtensional setting results in the most rapid and direct path from deep to shallow crustal levels. According to the Rey et al. (2011) interpretation, which was based on integrated 2D numerical modeling and field-based studies of migmatite domes, deep crustal rocks exhumed in a migmatite dome first experience isothermal decompression in a steep high-strain zone that forms below a localized region of extension in the upper crust. This phase of exhumation may bring formerly deep, still-hot rocks to very shallow levels (< 10 km), where they are transferred to the footwall of an extensional detachment system or incorporated in one of the subdomes that flank the high-strain zone. In this model for exhumation of HP rocks in a migmatite dome, the degree of preservation of HP mineral assemblages and compositions may vary as a function of the path of the rock through the ‘dome system’.

7.3 Eclogite metamorphism and migmatite dome emplacement at a plateau-foreland transition

We present a model for the formation of the Montagne Noire dome, taking into account three significant parameters: (1) the Montagne Noire dome developed in the Variscan orogen foreland (e.g., Franke et al., 2011) or at the transition between the orogen and the foreland, (2) eclogite facies metamorphism occurred at ~ 315 Ma as recorded in mafic fragments dispersed in the dome (this study); and (3) exhumation of eclogite was nearly synchronous with dome formation and migmatite emplacement that occurred at 315-300 Ma (Roger et al., 2015). We cannot be certain that eclogite facies metamorphism occurred in the deeper parts of what is now the Montagne Noire dome – that is, at the plateau-foreland transition zone (Figs. 1, 2). Nevertheless, the close temporal association of eclogite metamorphism and host migmatite crystallization is consistent

with a “local” origin for the eclogite, as opposed to the eclogites being “far traveled” from a different region (likely north) and transported to the southernmost Massif Central by flowing deep crust.

The location of the Montagne Noire dome in the foreland of the Massif Central has been presented as a paradox (e.g., Franke et al., 2011) because the presence of migmatite and associated granite bodies that crystallized between 315 and 300 Ma begs the question of the heat source and tectonic setting that allowed such high-grade metamorphism and deformation to develop. In many ways (protolith age, P-T-t history, and structure), the Montagne Noire has significant affinity with the Variscan Pyrenean belt to the south (Fig. 2), where the age of metamorphism and magmatism is similarly clustered at 315-300 Ma. The Pyrenees belt is characterized by dominantly wrench tectonism (transpression and transtension), like the Montagne Noire, where both contraction and extension (transpression and transtension) coexisted spatially and/or developed sequentially in a wide wrench zone. Therefore, the Montagne Noire may have developed its own dynamics within a wrench corridor (Variscan Pyrenees style) where contraction and extension components led to crustal thickening and the formation of eclogite, followed by exhumation and the emplacement of migmatite/granite domes. As a result, our interpretation of the Montagne Noire dome considers both its wrench setting and its location at the edge of the collapsing Variscan orogen in Late Carboniferous time. We envision a tectonic scenario similar to the present-day Altyn Tagh and Kunlun wrench systems at the northern edge of the Tibetan plateau.

Orogenic collapse takes various forms, including collapse during convergence or divergence, or while boundaries remain stationary (Rey et al., 2001). Collapse can be achieved by upper crustal extension, lower crustal flow, or both. As long as the upper crust does not extend and rift,

the low viscosity crust flows laterally, but if upper crustal extension occurs, the low viscosity crust immediately flows upward and fills the gap that opens in the upper crust. In fact, the presence of a low viscosity lower crust enhances localization of extension in the upper crust and the development of migmatite-cored metamorphic core complexes (e.g., Rey et al., 2009; Whitney et al., 2013).

In the case of fixed boundary collapse, the gravitational potential energy stored in the plateau region can trigger motion of material toward the foreland (low land) (Figs. 1, 9). This may happen as the boundaries of the orogen transition from converging to non-converging mode. During fixed boundary collapse, the foreland is in compression while the plateau is in extension, and significant crustal thickening can occur at the edge of the plateau (Fig. 1). This phenomenon has been evaluated in 2D numerical models (e.g., Clark and Royden, 2000; Rey et al., 2010): modeling results show that the viscosity distribution in the plateau crust as well as the strength of the foreland crust control the evolution of the system. In most cases the plateau edge moves toward the foreland, putting the foreland-plateau boundary in compression and thickening mode, while the low viscosity lower crust flows at depth from the plateau to the foreland (Fig. 1).

In this part of the Variscan orogen, orogenic collapse may have developed as the southern edge of the orogenic plateau evolved to strike-slip or wrenching mode. This switch in tectonism effectively forms a tectonic ‘wall’ at the edge of the plateau, resulting in two fundamental processes: (1) the gravitational potential energy stored in the plateau imparts a compressive force against the strike-slip system, resulting in transpression; and (2) the low-viscosity lower crust of the orogenic plateau to the north flows southward and has the potential of feeding the wrench/transpressional system (Fig. 9). In our proposed model, eclogite metamorphism is generated by burial of mafic rocks in a transpressional system at the edge of the plateau, owing

to the added push from the collapsing orogen to the north, and this is closely followed by eclogite exhumation owing to the flow of partially molten crust from north to south; eclogite and migmatite are exhumed together in the Montagne Noire dome (Fig. 9).

More specifically, exhumation of the migmatite dome occurred in a dextral wrench corridor as a pull-apart system formed in the upper crust (Echtler and Malavieille, 1990; Roger et al., 2015). This interpretation reconciles apparently disparate observations and results, and allows crustal thickening (HP metamorphism) and exhumation of partially molten crust to be part of a single tectonic event that is fundamentally related to dynamics at the edge of a collapsing orogen.

Given the general nature of these processes, it is expected that this type of plateau-edge tectonism is common, although difficult to recognize in the geologic record. In the Montagne Noire, the mobile partially molten crust incorporated and exhumed fragments of HP rocks, revealing this otherwise poorly documented tectonic mechanism. The presence of eclogite the same age as or only a few million years older than the migmatite in which they are nested affirms the minimum depth as well as the mobility of the channel of the orogenic partially molten layer, and therefore its deep-crust (>40 km) sampling capabilities during the collapse of orogens.

8. Conclusions

Inclusions of eclogite in migmatite in the cores of gneiss domes provide information about dome dynamics because they retain a memory of P-T conditions that are otherwise not recorded in the host quartzofeldspathic rocks. In the Montagne Noire migmatite dome, eclogite P-T conditions are highly relevant to understanding crustal flow and dome emplacement because eclogite metamorphism was an immediate precursor to dome formation. The Terme de Fourcaric eclogite is dated at ~315 Ma, making it the youngest eclogite identified in the Variscan orogen. Eclogite

peak P-T conditions of ~1.4 GPa, 725°C may have been related to crustal thickening at the orogen-foreland transition driven by flow of the collapsing crust toward the foreland during a switch from convergent to strike-slip tectonism. Development of a dextral pull-apart in the Montagne Noire upper crust allowed upward flow of partially molten crust and entrainment of eclogite that was exhumed together with the migmatite dome. We propose that mass transfer from orogen to foreland during orogenic collapse thickened the cool foreland, leading to eclogite facies metamorphism, while low-viscosity partially molten crust flowing from the plateau to the foreland incorporated and exhumed eclogite. If so, these results show that some dome material comes from ~40-50 km depth, although most of it equilibrates at shallow crustal levels. This study reconciles apparently paradoxical elements of Montagne Noire geology, including contraction (producing eclogite), wrenching, and emplacement of high-temperature metamorphic rocks (migmatite), all within ~10 million years in a foreland setting.

Acknowledgments

This research was supported by NSF grant EAR-1050020 to Teyssier and Whitney. We thank Jessie Shields (Macalester College) for acquiring the whole-rock (XRF) analysis, Anette von der Handt for developing the protocols for the Zr-in-rutile analysis, Côme Lefebvre for mineral separation (zircon), and Katherine Fornash for initial Perple_X experiments. We also thank Valérie Bosse and Jean-Louis Paquette (Laboratoire “Magmas et Volcans”, Clermont-Ferrand) for their participation in the geochronological-geochemical analyses. We thank two anonymous reviewers for constructive comments that helped improve the manuscript.

References

- Aerden, D.G.A.M., 1998. Tectonic evolution of the Montagne Noire and a possible orogenic model for syncollisional exhumation of deep rocks, Variscan belt, France. *Tectonics*, 17, 62–79.
- Alabouvette B., Demange M., and Echtler H., 1993. Carte géol. France (1/50 000), feuille Saint-Pons (1013). Orléans : BRGM, 123 p.
- Berger, J., Féménias, O., Ohnenstetter, D., Bruguier, O., Plissart, G., Mercier, J-C. C., and Demaiffe, D., 2010. New occurrence of UHP eclogites in Limousin (French Massif Central): Age, tectonic setting and fluid-rock interactions. *Lithos*, 118, 365-382.
- Bouchardon, J.L., Déchomets, R., and Demange, M., 1979. A propos du disthène en roche dans les micaschistes et les gneiss du synclinal de Rosis et du flanc sud, zone axiale de la Montagne Noire (Massif Central français). *C.R. Acad. Sci., D*, 288, 1067-1071.
- Brunel, M., and Lansigu, C., 1997. Déformation et cinématique de mise en place du dôme de la zone axiale de la Montagne Noire: signification des nodules à quartz-sillimanite (Massif central français). *C.R. Acad. Sci. IIA*, 325(7), 517-523.
- Charles, N., Faure, M., and Chen, Y., 2009. The Montagne Noire migmatitic dome emplacement (French Massif Central): insights from petrofabric and AMS studies. *J. Struct. Geol.*, 31, 1423-1440.
- Clark, M.K., and Royden, L.H., 2000. Topographic ooze: Building the eastern margin of Tibet by lower crustal flow. *Geology*, 28, 703-706.
- Cocherie, A., Baudin, T., Autran, A., Guerrot, C., Fanning M., and Laumonier, B., 2005. U-Pb zircon (ID-TIMS and SHRIMP) evidence for the early Ordovician intrusion of metagranites in the late Proterozoic Canaveilles Group of the Pyrenees and the Montagne Noire (France).

548 Bull. Soc. géol. France, 176, 269-282.

549 Connolly, J.A.D., 2005. Computation of phase equilibria by linear programming: a tool for
 550 geodynamic modeling and an application to subduction zone decarbonation. *Earth Planet. Sci.*
 551 *Lett.*, 236, 524–541.

552 Demange, M., 1985. The eclogite-facies rocks of the Montagne Noire, France. *Chem. Geol.*, 50,
 553 173-188.

554 Demange, M., 1999. Evolution tectonique de la Montagne Noire: un modèle en transpression.
 555 *C.R. Acad. Sci.*, 329, 823-829.

556 Demange, M., Guérangé-Lozes, J., and Guérangé, B. 1996. Carte géologique de Lacaune et sa
 557 notice. Carte géologique de la France au 1:50 000 n8987. Bureau de Recherches Géologiques
 558 et Minières, Orléans, France.

559 Doublier, M.P., Potel, S., Wemmer, K., 2015. The tectono-metamorphic evolution of the very
 560 low-grade hanging wall constrains two-stage gneiss dome formation in the Montagne Noire
 561 example (Southern France). *J. Metamorph. Geol.*, 33, 71-89. doi: 10.1111/jmg.12111.

562 Ducrot, J., Lancelot, J.R., and Marchand, J., 1983. Datation U-Pb sur zircons de l'eclogite de la
 563 Borie (Haut-Allier, France) et consequences sur l'évolution ante-hercynienne de l'Europe
 564 occidentale. *Earth Planet. Sci. Lett.*, 62, 385-394.

565 Echtler, H., and Malavieille, J., 1990. Extensional tectonics, basement uplift, and Stephan-
 566 Permian collapse basin in a late Variscan metamorphic core complex (Montagne Noire,
 567 southern Massif Central). *Tectonophysics*, 177, 125-138.

568 Ellis, D.J., and Green, D.H., 1979. An experimental study of the effect of Ca upon garnet-
 569 clinopyroxene Fe-Mg exchange equilibria. *Contrib. Mineral. Petrol.*, 71, 13-22.

570 Engel, W., Feist, R., and Franke, W., 1980. Le Carbonifère anté-stéphanien de la Montagne

571 Noire: rapports entre mise en place des nappes et sédimentations, Bulletin du BRGM 2, 341-
572 389.

573 Faure, M., 1995. Late orogenic carboniferous extensions in the Variscan French Massif Central.
574 Tectonics, 14, 132–153.

575 Faure, M., Cocherie, A., Bé Mézème, E., Charles, N., and Rossi, P., 2010. Middle Carboniferous
576 crustal melting in the Variscan belt: new insights from U-Th-Pb monazite and U-Pb zircon
577 ages of the Montagne Noire Axial Zone (southern French Massif Central). Gondwana Res.,
578 18, 653-673.

579 Faure, M., Cocherie, A., Gaché, J., Esnault, C., Guerrot, C., Rossi, P., Wei, L., and Qiuli, L.,
580 2014. Middle Carboniferous intracontinental subduction in the Outer Zone of the Variscan
581 Belt (Montagne Noire Axial Zone, French Massif Central): multimethod geochronological
582 approach of polyphase metamorphism. In: Schulmann, K., Martinez Catalan, J.R., Lardeaux,
583 J.M., Janousek, V., and Oggiano, G., (eds.), The Variscan Orogeny: Extent, Timescale and
584 the Formation of the European Crust. Geol. Soc. Lond. Spec. Pub., 405, doi 10.1144/SP405.2.

585 Faure, M., Lardeaux, J.-M., and Ledru, P., 2009. A review of the pre-Permian geology of the
586 Variscan French Massif Central. C.R. Geosci., 341, 202-213.

587 Franke, W., Doublier, M.P., Klama, K., Potel, S., and Wemmer, K., 2011. Hot metamorphic core
588 complex in a cold foreland. Int. J. Earth Sci., 100, 753-785.

589 Gebauer, D., and Grunenfelder, M., 1982. Geological development of the Hercynian belt of
590 Europe based on age and origin of high grade and high pressure mafic and ultramafic rocks.
591 First Int. Conf. on Geochronology, Cosmochronology, Isotope Geology, Nikko, pp. 111-112.

592 Géze, B., 1949. Etude géologique de la Montagne Noire et les Cévennes méridionales. Mém. Soc.
593 géol. Fr., Nouv. Sér., 29(62), 215 pp.

594 Holland, T.J.B., and Powell, R., 1996. Thermodynamics of order-disorder in minerals. 2.
 595 Symmetric formalism applied to solid solutions. *Am. Mineral.*, 81, 1425-1437.
 596 Holland, T.J.B., and Powell, R., 1998. An internally consistent data set for phases of petrologic
 597 interest. *J. Metamorph. Geol.*, 16, 309-43.
 598 Lardeaux, J.M., 2014. Deciphering orogeny: a metamorphic perspective. Examples from the
 599 European Alpine and Variscan belts. Part II. Variscan metamorphism in the French Massif
 600 Central – A review. *Bull. Soc. géol. France*, 185, 281-310.
 601 Lardeaux, J.M., Ledru, P., Daniel, I., and Duchene, S., 2001. The Variscan French Massif
 602 Central – a new addition the ultra-high pressure metamorphic ‘club’: exhumation processes
 603 and geodynamic consequences. *Tectonophysics*, 332, 143-167.
 604 Matte, P., Lancelot J., and Mattauer, M., 1998. La Zone axiale hercynienne de la Montagne
 605 Noire n'est pas un "metamorphic core complex" extensif mais un anticlinal post-nappe à
 606 coeur anatectique. *Geodinam. Acta*, 11, 13-22.
 607 Nakamura, D., 2009. A new formulation of garnet-clinopyroxene geothermometer based on
 608 accumulation and statistical analysis of a large experimental dataset. *J. Metamorph. Geol.*, 27,
 609 495-508.
 610 Paquette, J.L., Monchoux, P., and Couturier, M., 1995. Geochemical and isotopic study of a
 611 norite-eclogite transition in the European Variscide Belt. Implications for U-Pb zircon
 612 systematics in metabasic rocks. *Geochim. Cosmochim. Acta*, 59, 1611-1622.
 613 Rabin M., Trap P., Carry N., Fréville K., Cenki-Tok B., Lobjoie C., Goncalves P., and Marquer
 614 D., 2015. Strain partitioning along the anatectic front in the Variscan Montagne Noire massif
 615 (Southern French Massif Central). *Tectonics*, doi: 10.1002/2014TC003790.
 616 Rey, P., Vanderhaeghe, O., and Teyssier, C., 2001. Gravitational collapse of the continental

617 crust: definition, regimes and modes. *Tectonophysics*, 342, 435-449.

618 Rey, P., Teyssier, C., and Whitney, D.L., 2009. Crustal melting and core complex dynamics.
619 *Geology*, 37, 391-394.

620 Rey, P.F., Teyssier, C., and Whitney, D.L., 2010. The limit of channel flow in orogenic plateaux.
621 *Lithosphere*, 2, 328-332. doi: 10.1130/L114.1

622 Rey P.F., Teyssier C., Kruckenberg S.C., and Whitney, D.L., 2011. Viscous collision in channel
623 explains double domes in metamorphic core complexes. *Geology*, 39, 387-390.

624 Roger, F., Respaut, J.P., Brunel, M., Matte, P., and Paquette, J.L., 2004. Première datation U–Pb
625 des orthogneiss ocellés de la zone axiale de la Montagne Noire (Sud du Massif Central):
626 nouveaux témoins du magmatisme ordovicien dans la chaîne varisque. *C.R. Geosci.*, 336,
627 19–28.

628 Roger, F., Teyssier, C., Respaut, J-P., Rey, P.F., Jolivet, M., Whitney, D.L., Paquette, J-L.,
629 Brunel, M., and Matte, P., 2015. Emplacement of anatectic granite during Variscan extension
630 and exhumation of the Montagne Noire double dome, French Massif Central, France.
631 *Tectonophysics*, 640-641, 53-69.

632 Rubatto, D., 2002. Zircon trace element geochemistry: distribution coefficients and the link
633 between U–Pb ages and metamorphism. *Chem. Geol.*, 184, 123–138.

634 Schuiling, R.D., 1960. Le dôme gneissique de l’Agout (Tarn et Hérault). *Mem. Soc. géol. France*,
635 39, 1-59.

636 Soula, J.-C., Debat, P., Brusset, S., Bessière, G., Christophoul, F., and Déramond, J., 2001.
637 Thrust-related, diapiric, and extensional doming in a frontal orogenic wedge: example of the
638 Montagne Noire, Southern French Hercynian Belt. *J. Structural Geol.*, 23, 1677-1699.

639 Stipska, P., Schulmann, K., and Powell, R., 2008. Contrasting metamorphic histories of lenses of

640 high-pressure rocks and host migmatites with a flat orogenic fabric (Bohemian Massif, Czech
 641 Republic): a result of tectonic mixing within horizontal crustal flow? *J. Metamorph. Geol.*,
 642 26, 623-646. doi: 10.1111/j.1525-1314.2008.00781.x
 643 Sun, S.S. and McDonough, W.F., 1989. Chemical and isotopic systematics of oceanic basalts;
 644 implications for mantle composition and processes. In: *Magmatism in the ocean basins*.
 645 Saunders, A.D. and Norry, M.J. (Eds.), *Geol. Soc. London*, 42, 313-345.
 646 Teyssier, C., and Whitney, D.L., 2002. Gneiss domes and orogeny. *Geology*, 30, 1139-1142.
 647 Thompson, P.H., and Bard, J.-P., 1982. Isograds and mineral assemblages in the eastern axial
 648 zone, Montagne Noire (France): implications for temperature gradients and P-T history. *Can.*
 649 *J. Earth Sci.*, 19, 129–141.
 650 Tomkins, H.S., Powell, R., and Ellis, D.J., 2007. The pressure-dependence of the zirconium-in-
 651 rutile thermometer. *J. Metamorph. Geol.*, 25, 703-713.
 652 van den Driessche, J., and Brun, J.P., 1992. Tectonic evolution of the Montagne Noire (French
 653 Massif Central): a model of extensional gneiss dome. *Geodinam. Acta*, 5, 85-99.
 654 Vanderhaeghe, O., Burg, J.-P., and Teyssier, C., 1999. Exhumation of migmatites in two
 655 collapsed orogens: Canadian Cordillera and French Variscides. In: *Exhumation Processes:*
 656 *Normal Faulting, Ductile Flow and Erosion*. (Eds) Ring, U., Brandon, M., Lister, G.S., and
 657 Willett, S.D., *Geol. Soc. Lond. Spec. Pub.*, 154, 181-204.
 658 Whitney, D.L., Teyssier, C., and Vanderhaeghe, O., 2004. Gneiss domes and crustal flow, *in*
 659 Whitney, D.L., Teyssier, C., and Siddoway, C.S., eds., *Gneiss domes in orogeny*, *Geol. Soc.*
 660 *Am. Spec. Paper* 380, 15–33.
 661 Whitney, D.L., Teyssier, C., Rey, P.F., and Buck, W.R., 2013. Continental and oceanic core
 662 complexes. *Geol. Soc. Am. Bull.*, 125, 273-298. doi: 10.1130/B30754.1

Figure captions

Fig. 1. Conceptual model of the flow of partially molten crust laterally (in a deep crustal channel) and vertically (to form migmatite domes) at the transition between the thick crust of an orogenic plateau (including a ductile layer in the deep crust) and the thinner, cooler foreland during extension (figure based on 2D numerical models and modified from Rey et al., 2010).

Fig. 2. Schematic block diagram (after Faure, 1995) and cross-section (after Vanderhaeghe et al., 1999) of the main late Carboniferous units in the French Massif Central. Inset sections represent two stages of orogenic development: crustal accretion and plateau development followed by orogenic collapse, emplacement of migmatite domes, and deposition of extensional sedimentary basins.

Fig. 3. (a) Map of some of the Variscan massifs in western Europe: Armorican (ARM), Iberian (IB), the Massif Central (MC), and the Pyrenees (PYR). The rectangle at the southern end of the Massif Central shows the location of the Montagne Noire; (b) Simplified geologic map of the Montagne Noire (foliation trajectories after Rabin et al., 2015, and Roger et al., 2015, for the eastern end of the dome). The locations of eclogite are shown, and abbreviated as follows: A = Airette, CAB = Cabardès, CT = Cantausse, LJ = Le Jounié, TF = Terme de Fourcaric. Anatectic granites are labeled (V = Vialais); (c) Map of Terme de Fourcaric/Peyrambert area, showing the location of the analyzed eclogite, MN13-11A (after Alabouvette et al., 1993; Demange et al., 1996).

Fig. 4. Terme de Fourcaric eclogite MN13-11A; (a) thin section scan (4 cm long) showing garnet, omphacite, and minor retrogression (gray = amphibole-plagioclase symplectite); (b) Photomicrograph (plane light) of garnet with quartz inclusions in the core and rutile inclusions near the garnet rim; (c) Element maps showing zoning in Ca, Mg, and Mn; (d) BSE image of garnet with accompanying compositions along the line shown on the image. Mineral abbreviations: Grt = garnet; Omp = omphacite; Qz = quartz; Rt = rutile. Component abbreviations: alm = almandine; grs = grossular; prp = pyrope; sps = spessartine.

Fig. 5. (a, b, c) BSE images of rutile in the Terme de Fourcaric eclogite MN13-11A: (a) rutile inclusions in garnet; (b) blocky matrix rutile (with ilmenite), see also Fig. 5d₁; and (c) elongate rutile along a grain boundary between garnets. (d, e) Zr in rutile. (d₁, e₁) BSE images of two blocky (matrix) rutile crystals showing partial replacement by ilmenite; locations of analytical spots are marked, with corresponding Zr content (ppm). Analyses indicated by the red circles are plotted in Figs. 3d₂, e₂ (on the distance axis, 0 is the location of the point furthest from ilmenite (near the rutile rim) and the highest value is closest to ilmenite). Ilmenite contains no Zr. Additional rutile analyses are shown on the BSE image (e₁) as black circles with Zr (ppm) indicated. In the BSE images, ilmenite exsolution (bright lines) can be seen. (d₃, e₃) ZrO₂ vs. FeO weight %, showing that the highest Zr content (near ilmenite) corresponds to the highest Fe content. Zr and Fe decrease away from ilmenite. The region of the graph showing scatter of Fe relative to Zr corresponds to the lowest Zr values, at the greatest distance from ilmenite.

Fig. 6. (a) Pseudosection determined using Perple_X software for (a) the XRF-determined bulk composition, in weight %: SiO₂ = 45.79; TiO₂ = 1.42; Al₂O₃ = 16.35; Fe₂O₃ = 11.78; MnO =

0.18; MgO = 11.49; CaO = 10.48; Na₂O = 2.16; K₂O = 0.02; P₂O₅ = 0.18; loss on ignition = < detection limit; and (b) pseudosection determined for an effective bulk composition that does not include garnet core regions (garnet mode = 55%). The ellipse in (b) indicates the region most consistent with garnet rim composition (Mg, Fe), matrix omphacite composition (Na), and inferred peak mineral mode ($\pm$ 5%, see text), as well as temperatures determined by grt-cpx and Zr-in-rutile thermometers. Abbreviations: Ab = albite; Cpx = clinopyroxene; En = enstatite; Grt = garnet; Ilm = ilmenite; Omp = omphacite; Rt = rutile.

Fig. 7. Zircon geochronology (a) BSE images of zircon with ages in Ma, showing the location of geochronology analysis spots (red circles) and trace element (REE) analysis spots (white circles); (b) Concordia plot of zircon core and rim results (data in Appendix C; images of analytical spots in Appendix D).

Fig. 8. Chondrite-normalized REE plot for dated zircons in MN13-11A eclogite (data in Appendix E; images of analytical spots in Appendix D). Normalization values from Sun and McDonough (1988).

Fig. 9. Proposed stages of evolution of the Montagne Noire region from burial (eclogite formation at > 40 km in the thickened crust) to exhumation of the migmatite dome. (a) Schematic map view of proposed wrench corridor in which burial and eclogite metamorphism take place (left panel; e: eclogite) as well as exhumation and unroofing of dome rocks (right panel); (b) Lithosphere-scale cross sections showing burial (left) and exhumation (right) stages of Montagne Noire. The proposed interpretation involves a zone of wrenching at the edge of the

733 orogen in which contraction is enhanced by collapse-induced convergence in a fixed-boundary
734 collapse mode (Rey et al., 2001).

735

736

737

738

739

740

741

742

743

744

745

746

747

748

749

750

751

752

753

754

755

756

Figure

Figure 1

Figure

Figure 2

Figure

Figure 3

Figure

Figure 4

Figure

Figure 5

Figure

Figure 6

Figure

Figure 7

Figure

Figure 8

Figure

Figure 9