Cambro–Ordovician ferrosilicic magmatism along the northern Gondwana margin: constraints from the Cézarenque–Joyeuse gneiss complex (French Massif Central)

Le magmatisme ferrosiliceux cambro-ordovicien de la marge Nord du Gondwana : nouvelles contraintes issues de l'étude du complexe gneissique de la Cézarenque–Joyeuse (Massif Central français)

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Abstract

It is well-acknowledged that the northern margin of the Gondwana supercontinent was affected by a major magmatic event at late Cambrian (Furongian) to early Ordovician (Tremadocian–Floian) times. However, an accurate assessment of its extent, origin, and significance is partly hampered by the incomplete characterization of the numerous gneiss massifs exposed in the inner part of the Variscan belt, as some of them possibly represent dismembered and deformed Furongian–Lower Ordovician igneous bodies. In this study, we document the case of the "Cézarenque–Joyeuse" gneiss complex in the Cévennes parautochthon domain of the French Massif Central. The gneisses form decametre- to kilometre-thick concordant massifs interlayered within a pluri-kilometric sequence of mica- and quartz schists. They encompass two main petrological types: augen gneisses and albite gneisses, both typified by their blue and engulfed quartz grains with the augen facies differing by the presence of centimetre-sized pseudomorphs after K-feldspar and the local preservation of igneous textures. Whole-rock geochemistry highlights that many gneisses have magmatic *ferrosilicic* (acidic with anomalously high FeO_t and low CaO) compositions while others are akin to greywackes. Collectively, it is inferred that the bulk of the Cézarenque–Joyeuse gneisses represent former rhyodacite lava flows or ignimbrites and associated epiclastic tuffs. Volumetrically subordinate, finer grained, and strongly silicic leucogneisses are interpreted as microgranite dykes originally intrusive within the volcanic edifices. LA–ICP–MS U–Pb dating of magmatic zircon grains extracted from an augen gneiss and a leucogneiss brackets the crystallization age of the silicic magmas between 486.1±5.5 Ma and 483.0±5.5 Ma which unambiguously ties the Cézarenque–Joyeuse gneisses to the Furongian–Lower Ordovician volcanic belt of SW Europe. Inherited zircon date distributions, Ti-in-zircon and zircon saturation thermometry demonstrate that they formed by melting at 750–820 °C of Ediacaran sediments. Zircon Eu/Eu* and Ce/Ce* systematics indicate that the melts were strongly reduced (fO_2 probably close to the values expected for the iron–wüstite buffer), possibly because they interacted during ascent with Lower Cambrian black shales. This would have enhanced Fe solubility in the melt phase and may explain the peculiar *ferrosilicic* signature displayed by many Furongian–Lower Ordovician igneous rocks in the northern Gondwana realm. We infer that crustal melting resulted from a combination of mantle-derived magma underplating in an intracontinental rift setting and anomalously elevated radiogenic heat production within the Ediacaran sedimentary sequences.

Résumé

Un important évènement magmatique a affecté la marge Nord du Gondwana de la fin du Cambrien (Furongien) au début de l'Ordovicien (Trémadocien–Floien). Afin de préciser son extension géographique et de mieux contraindre son origine et sa signification géodynamique, il est essentiel de caractériser finement les nombreux massifs de gneiss affleurant dans les zones internes de la chaine Varisque, certains d'entre eux représentant probablement d'anciens corps magmatiques cambro-ordoviciens. Cet article présente de nouvelles données sur les gneiss dits de la « Cézarenque » ou de « Joyeuse » qui affleurent dans le domaine para-autochtone cévenol du sud du Massif Central français. Ces gneiss forment des massifs concordants dans la schistosité régionale, d'une épaisseur variant de la dizaine de mètres à plusieurs kilomètres, intercalés dans une épaisse séquence de schistes. Les deux types pétrographiques principaux sont un gneiss oeillé et un gneiss albitique, tous deux caractérisés par la présence de « phénocristaux » de quartz bleus d'origine

magmatique, les gneiss oeillés se distinguant par leurs pseudomorphes centimétriques de feldspath potassique. Les compositions chimiques en roche totale démontrent que de nombreux gneiss ont des compositions magmatiques et se rattachent aux séries ferrosiliceuses (acides, anormalement riches en FeO_t et pauvres en CaO). A contrario, certains gneiss albitiques ont des compositions qui les rapprochent de grauwackes. L'ensemble de ces observations suggèrent que les gneiss de Cézarenque–Joyeuse représentent d'anciennes coulées (pyroclastiques ou laviques) rhyodacitiques et leurs produits de remaniements proximaux. Un rare faciès de leucogneiss, subordonné en volume, correspondrait à d'anciens microgranites originellement intrusifs dans les édifices rhyodacitiques. Les datations U–Pb par LA–ICP–MS de grains de zircons extraits d'un gneiss oeillé et d'un leucogneiss indiquent que les magmas ont cristallisé entre 486.1±5.5 et 483.0±5.5 Ma, affiliant de fait les gneiss de Joyeuse–Cézarenque à la chaine volcanique d'âge Furongien–Ordovicien Inférieur d'Europe de l'Ouest. La gamme d'âge des zircons hérités ainsi que les thermomètres calibrés autour du zircon indiquent que les liquides proviennent de la fusion à 750–820°C de sédiments d'âge Ediacarien. Les ratios Eu/Eu* et Ce/Ce* des grains de zircon démontrent que ces liquides étaient fortement réduits (fO_2 proche du tampon fer–wustite), potentiellement à la suite de leur interaction en route vers la surface avec les sédiments riches en matière organique du Cambrien Inférieur. Ce caractère réduit a pu augmenter la solubilité du fer dans les liquides et expliquerait de fait le caractère ferrosiliceux. Cet évènement de fusion partielle de la croute continentale est vraisemblablement lié au sous-placage de magmas basiques dans un contexte de rift intracontinental et à la production de chaleur radiogénique anormalement élevée des séquences sédimentaires édiacariennes.

1. Introduction

The basement of south-western Europe consists of Gondwana-derived crustal blocks amalgamated during the Devonian–Carboniferous Variscan orogeny (e.g. Kroner and Romer, 2013; and references therein). A wealth of studies have documented the common occurrence within the Variscan metamorphic series of late Cambrian to early Ordovician (Furongian–Tremadocian–Floian: ~497–470.0±1.4 Ma, Cohen et al., 2013) felsic (meta)igneous rocks (Parga Pondal et al., 1964; Weisbrod and Marignac, 1968; Weisbrod, 1969; Guérangé-Lozes and Burg, 1990; Bea et al., 2007; Montero et al., 2007; Díez Montes et al., 2010; Melleton et al., 2010; Ballèvre et al., 2012; Díez Fernández et al., 2012; Talavera et al., 2013; Lopez-Sanchez et al., 2015; Gutiérrez-Alonso et al., 2016; Pouclet et al., 2017; Álvaro et al., 2020), the largest and most iconic example being the Ollo de Sapo formation in the Iberian Massif (see review in García-Arias et al., 2018). Considered together, the Furongian–Lower Ordovician igneous rocks delineate a > 2000 km-long arcuate volcanic-plutonic belt (Fig. 1) and shed light on a major magmatic event at the northern Gondwana margin by that time. Unravelling the spatial extent, duration, origin, and magma volumes emitted during this event is of marked importance since it has recently been suggested that the Furongian–Lower Ordovician belt would classify as a silicic Large Igneous Province (Díez Montes et al. 2010; Gutiérrez-Alonso et al. 2016; García-Arias et al. 2018) and hence would have had a strong influence on the early Ordovician climate and the living (Bryan, 2021). Besides, the acidic magmas (SiO_2: 65–70 wt.%) emitted during this magmatic flare-up have unusual compositions termed *ferrosilicic* (Fernández et al. 2008; Castro et al. 2009) as they show anomalously high FeO_t (> 2.5 wt.%) and low CaO (< 2.0 wt.%) contents. Even though melting of pre-existing crustal materials is generally advocated (Bea et al., 2007; Ballèvre et al., 2012; Montero et al., 2017), the origin of this peculiar signature remains debated (Castro et al., 2009; Díaz-Alvarado et al., 2016; Fiannacca et al., 2019). Finally, it is well-established that this magmatic event is coeval to the opening of the Rheic ocean (Murphy et al. 2008; Nance et al. 2010), yet, the actual trigger of crustal melting is disputed with models invoking extensional (Bea et al. 2007; Ballèvre et al. 2012; Pouclet et al. 2017; Álvaro et al. 2020) or contractional (Villaseca et al. 2016) settings, near or far from the influence of a wanning Cadomian subduction zone (Fernández et al. 2008; Castro et al. 2009; Villaseca et al. 2016).

This study focuses on the so-called Cézarenque–Joyeuse gneiss complex exposed in the Cévennes domain of the south-eastern French Massif Central (Fig. 1). Ballèvre et al. (2012) postulated that they could represent felsic igneous rocks of the Furongian–Lower Ordovician belt having experienced greenschist- to lower amphibolite-facies metamorphism during the Variscan orogeny. We explore this possibility by reviewing the field relationships, petrography and geochemistry of these gneisses and providing the first LA–ICP–MS zircon U–Pb and trace element data on such lithologies. Collectively, our results do tie the Cézarenque gneisses to the Furongian–Lower Ordovician magmatic event, provide novel constraints on the petrogenesis of ferrosilicic magmas and offer the opportunity to address the tectonic evolution of this segment of the northern Gondwana margin.

2. Geological setting

The French Massif Central (FMC) is built up by igneous and metamorphic rocks formed during the Variscan orogeny (Matte, 2007; Faure et al., 2009a) which itself resulted from the closure of one or several oceanic domains in the early Devonian followed by the late Devonian to Carboniferous collision of Gondwana, Laurussia and several microblocks (Vanderhaeghe et al. 2020, and references therein). At the scale of the eastern Massif Central, several tectonic–metamorphic nappes were identified (Burg and Matte, 1978; Ledru et al., 1989; Faure et al., 2009a), all of which being exposed in the Cevennes area. At the top, the Upper Gneiss Unit experienced a Devonian (c. 380–360 Ma, Chelle-Michou et al., 2017; Lotout et al., 2020, 2018) prograde high-pressure metamorphism, with peak conditions of ~800 °C and 10–14 kbar at Marvejols (Fig. 2, Bodinier et al., 1988). During its retrograde evolution, it was thrust at 360–340 Ma (Chelle-Michou et al., 2017; Costa, 1989) over a set of nappes devoid of HP record (and experiencing prograde metamorphism) which were locally subdivided into a Lower Gneiss Unit and a Parautochthonous Unit (Faure et al., 2009a). Those experienced a pervasive and protracted melting episode at 335–300 Ma as a result of crust thickening and marked heat advection from the mantle (Laurent et al. 2017). At the bottom of the nappe pile lies the barely metamorphic “Fold-and-Thrust Belt” (Faure et al., 2009a), which corresponds to the so-called “Viganais series” in the Cévennes area (Fig. 2). The whole sequence was unconformably overlain by Upper Carboniferous to Lower Permian coal-bearing sedimentary rocks.

2.1 Tectonic-metamorphic evolution of the Cévennes domain

The Cévennes domain is primarily composed of detrital metasediments including a wide variety of petrographic types (mica and quartz schists, locally feldspar-rich, quartzite, calc-silicate gneisses, paragneisses) commonly gathered under the term "Cévennes schists" (Brouder, 1963; Barbey et al., 2015). It also encompasses several gneiss massifs (Peyrolles, Cézarenque–Joyeuse, Fig. 3), all of which forming decametre- to (several) kilometre-thick concordant bodies within schists (Brouder, 1963; Weisbrod and Marignac, 1968; Roger, 1969). At the regional scale, four main tectonic-metamorphic events were described. The main deformation phase D_1 produced a near-horizontal schistosity S_1 with a N–S stretching and mineral lineation, a fabric resulting from a combination of pure and simple shear with top-to-the S kinematics (Mattauer and Etchecopar, 1976; Lacassin and van den Driessche, 1983). Ar–Ar dating of micas and amphibole yielded 340–330 Ma dates for D_1 (Caron 1994) which took place under metamorphic conditions of c. 500°C and 5.2 kbar (Arnaud, 1997) and would attest to an early crust thickening and nappe stacking stage (Toteu and Macaudière, 1984; Faure et al., 2009b). In the eastern Cévennes, Faure et al. (2009b) advocated that the Cézarenque–Joyeuse gneisses and overlying micaschists represent an allochthonous complex (klippen) stacked over schists during D_1, putting forward the existence of a high-strain zone at the base of the main gneiss massif (Mattauer and Etchecopar, 1976). Such complex was thus regarded as an equivalent of the Lower Gneiss Unit, and the underlying schists attributed to the Parautochthonous Unit sensu Faure et al. (2009a). Alternative views consider that all metamorphic rocks from the Cévennes (including the Cézarenque–Joyeuse gneisses) belong to the Parautochthonous Unit (Matte 2007).

A D_2 event affecting the Cézarenque–Joyeuse gneisses and abutting schists was identified by Bouilhol et al. (2006). It is typified by asymmetric folding reworking S_1, a NW–SE trending stretching lineation and local C/S structures collectively indicative of top-to-the NW displacements. The P–T conditions of this event are ill-defined. The D_3 event took place in the time period 320–310 Ma (Caron, 1994; Couzinié et al., 2021) while a thermal anomaly developed beneath the Cévennes and resulted in a LP–HT metamorphic field gradient (Chenevoy and Ravier, 1968; Weisbrod, 1968; Tobschall, 1971; Montel et al., 1992; Rakib, 1996; Bouilhol et al, 2006) with peak temperatures reaching 700–800 °C at 3–6 kbar in the southern Velay dome (Ledru et al., 2001; Barbey et al. 2015; Villaros et al. 2018; Couzinié et al., 2021). Evidence for top-to-the-NE shearing during D_3 are conspicuous in the northern Cévennes (Faure et al., 2001; Bouilhol et al. 2006). The D_4 event (310–295 Ma) features the widespread crystallization of retrograde, greenschist-facies mineral assemblages associated with (brittle-)ductile top-to-the-ESE shearing (Faure et al., 2001) and the emplacement of the main granite plutons of the Cévennes (Borne, Mont-Lozère, Aigoual–S^{t}-Guiral–Liron; Brichau et al., 2007; Couzinié et al., 2014; Laurent et al., 2017). This event stamps the transition to an extensional regime which led to the opening of the Alès (Westphalian D–Autunian) coal basin (Allemand et al., 1997).

2.2 Nature and age of the metamorphic protoliths

In the northern Cévennes, where the deepest structural levels are exposed (Faure et al. 2001; Arnaud et al. 2004), amphibolite-facies metasediments yielding Ediacaran zircon U–Pb maximum depositional ages between 592.4±5.5 and 568.3±4.9 Ma (Couzinié et al., 2019) are interlayered with augen gneiss and “leptynites” (a local term somewhat equivalent to fine-grained leucogneisses) of the Velay Orthogneiss Formation, a former c. 540 Ma-old S-type granitic complex which was originally intrusive within the sedimentary protoliths (Couzinié et al. 2017).

In the central Cévennes, decametric gneiss layers (stars on Fig. 2) within mica and quartz schists, interpreted as felsic metavolcanics, yielded eruption ages of 482±8 and 476±6 Ma (zircon Pb evaporation, Faure et al., 2009b). Further south, the Peyrolles orthogneiss is a metagranite which emplacement age was estimated at 433±4 Ma (zircon U–Pb LA–ICP–MS, Faure et al., 2009b) or 465±12 Ma (Rb–Sr whole-rock isochron; Sabourdy, 1975). Collectively, these data demonstrate that the sedimentary protoliths of the "Cévennes schists" were deposited from the late Ediacaran to the Ordovician.

In the eastern Cévennes (Fig. 2,3), the origin of the "Cézarenque–Joyeuse" gneiss complex (Elmi et al., 1974, 1989) is disputed and its age unknown. The gneisses form decametre- to kilometre-thick concordant bodies within mica schists (Brouder 1963; Weisbrod and Marignac, 1968; Roger, 1969) and were variously interpreted as former volcano(-sedimentary) rocks including felsic lavas and ignimbrites (Chenevoy 1968a, 1968b; Faure et al., 2001), intrusive granitic bodies (Crevola et al., 1983; Elmi et al. 1989) or detrital sedimentary rocks reworking igneous material (conglomerates, Weisbrod and Marignac, 1968).

3. Lithological components of the eastern Cévennes

3.1 The "Cézarenque–Joyeuse" gneiss complex

Based on available descriptions from the literature (Brouder, 1963; Elmi et al. 1974, 1989; Magontier, 1988) and our own field observations, three lithological facies were identified: augen gneisses, albite gneisses, and leucogneisses. Their descriptions (presented below) are complemented by a review of their whole-rock major element compositions, based on a compilation of the data presented by Chenevoy (1968b), Weisbrod (1970), Magontier (1988) and Elmi et al. (1989) supplemented by two new analyses of an augen gneiss and a tourmaline-bearing leucogneiss (dated samples, see section 4). Their whole-rock chemical compositions were analysed at the Service d'Analyse des Roches et Minéraux (SARM, CNRS, Nancy) from powdered samples using a Thermo Fischer ICap 6500 ICP-OES for major elements and a Thermo Fisher X7 Q-ICP-MS for Zr. The analytical procedures, reproducibility and limits of detection are detailed in Carignan et al. (2001). The compilation and the newly obtained data are provided in the Supplementary Table 1.

In the literature, the term “leptynite” was traditionally used to describe any quartz–feldspar rich (and mica poor) granofelsic to gneissic lithology and thus encompasses both felsic meta-igneous rocks and impure metasandstones and arkoses. Hence, samples referred to as “leptynites” were screened to retain only those of igneous compositions by: (i) a careful examination of the field descriptions (notably evidence for original intrusive relationships) and, (ii) applying the procedure of Davoine (1969), i.e. selecting the high SiO_2 analyses that met the criteria: (K_2O+Na_2O) > 7% and CaO < 2%. This way, a total of 71 analyses were retained for the three main gneiss facies. In the following, we assume that the whole-rock compositions were not significantly modified during the Variscan metamorphic evolution, as demonstrated for other gneiss massifs of the eastern Massif Central (Couzinié et al. 2017). Geochemical diagrams were plotted using GCDkit (Janoušek et al., 2006).

3.1.1 The augen gneisses

The augen gneisses are strongly to poorly foliated rocks typified by their mm- to cm-sized bluish quartz porphyroclasts generally making up 5-10% of the rock (Fig. 4a,b,c). Such grains show undulose extinction and are very commonly “corroded” and engulfed (Fig. 5a). Near rectangular (euhedral) to ovoid (lenticular) 1 to 10 cm-large polycrystalline aggregates (Fig. 4a,b,c; 5-20% of the rock) dominantly comprise albite plus quartz and were demonstrated to represent pseudomorphs after K-feldspar (Chenevoy, 1968b), in agreement with the local preservation of microcline in their inner part (Fig. 5b). Bluish quartz and feldspar aggregates are embedded in a fine-grained (av. 0.5 mm) foliated matrix of quartz, albite, muscovite, and biotite (often chloritized).

The occurrence of scarce garnet in the matrix along with relictual oligoclase and microcline was reported by Elmi et al. (1989). From a chemical point of view, the augen gneisses show elevated SiO_2 (av. ~68.5 wt.%) and FeO_t (av. ~3.9 wt.%), but distinctively low CaO (always < 1.5 wt.%) contents. In the Al/3-Na vs. Al/3-K diagram of de La Roche (1968), designed to discriminate igneous and sedimentary rocks, their compositions largely overlap with those of silicic Variscan plutonic and volcanic rocks of similar SiO_2 contents (Fig. 7a) demonstrating that the protoliths of the augen gneisses were igneous rocks. They show subalkaline rhyodacite compositions (Fig. 7b) and are highly peraluminous (Fig. 7c) which, together with the high Fe and low Ca contents, indicates that the igneous protoliths of the augen gneisses share a ferrosilicic signature (in the sense of Castro et al., 2009). Considering the alkalinity and Fe/Mg balance, the augen gneisses classify as calc-alkaline to alkali-calcic and magnesian (both sensu Frost et al., 2001, Fig. 7d,e) with $FeO_t/(FeO_t+MgO)$ clustering around 0.7–0.75. Their K/(K+Na) ratios are always close to 1 (Fig. 7f).

3.1.2 The albite gneisses

The albite gneisses (Fig. 4d) are composed of a foliated fine-grained (< 0.2 mm) matrix of quartz, albite, biotite (often chloritized) and muscovite, with some albite porphyroblasts locally reaching 1–2 mm (Fig. 5c). Muscovite and albite are overall more abundant than in the augen facies. Fragments of ovoid strongly stretched (dismantled) feldspar aggregates along with mm-sized bluish quartz grains locally occur in the foliated groundmass but remain scarce and of smaller size than in the augen gneisses. From a chemical point of view, the compositions of albite gneisses are more varied than those of their augen counterparts (e.g. SiO_2: 57–72%). A clear overlap exists as illustrated by most geochemical plots but two thirds of the albite gneisses are typified by higher Al–Fe–Mg (Fig. 7a,c) together with lower SiO_2 and alkali contents (Fig. 7b) with respect to the augen facies. Such samples plot in the fields of greywackes (60%) and shales (40%) in the Al/3-Na vs. Al/3-K diagram of de La Roche (1968) and their compositions resemble those of the Cévennes mica schists, albeit differing by more elevated alkali (Na_2O+K_2O) vs. CaO ratios (Fig. 7).

3.1.3 The leucogneisses

The leucogneisses are leucocratic fine-grained (< 0.2 mm) rocks (Fig. 4e,f) typified by their foliated (or lineated) "granitic" assemblage of albite (20-25 vol.%), quartz (30-40%) and K-feldspar (microcline, 15-25%) and a lower mica content with respect to other gneiss varieties (< 10-15%), with muscovite always more abundant than biotite (Fig. 5d). Locally, subhedral cm-sized grains of tourmaline, partly corroded, recrystallized, and truncated (Fig. 5d), underline the rock stretching lineation. The leucogneisses are strongly silicic (SiO_2: 72–76 wt.%) and plot in the field of meta-igneous rocks in the diagram of de La Roche (1968). They have rhyolitic compositions (Fig. 7b), are felsic peraluminous, calc-alkaline and slightly potassic (K/Na+K: 0.5–0.6, Fig. 7c to e) and show variable Fe/Mg ratios (Fig. 7f).

3.2 The "Cévennes schists"

This term encompasses a variety of metamorphic rocks derived from a range of pelite, semi-pelite, greywacke, sandstone, and arkose protoliths. The most representative petrographic type is a mica schist showing a well-developed foliation (Fig. 6a) defined by alternating mm-scale quartz and chlorite–muscovite (± graphite ± biotite ± garnet) layers. The mode of quartz can significantly increase and the rocks grade to quartz schists (Fig. 6b). Albite porphyroblasts (0.5–1 mm) are generally scarce (< 15% of the rock) but locally concentrated along decimetre-thick layers (Fig. 6c). Cordierite is restricted to the northern part of the study area (corresponding to the deepest structural levels) and the local occurrence of staurolite was reported in garnet–biotite schists near Ribes (Fig. 3, Bouilhol et al., 2006). A peculiar dark-coloured quartz schist type, composed of fine-grained (sub-mm) quartz (> 80% vol.) and biotite (Fig. 6d), was identified as 50 to 200 m-thick elongated bodies within the Cézarenque–Joyeuse gneisses.

3.3 Field relationships between the gneisses and the “Cévennes schists”

At the map scale, the contact between gneisses and the “Cévennes schists” is systematically parallel to the regional foliation and both lithologies appear interlayered. The gneisses define lenticular 10 to 250 m-thick bodies within the schists to the NE (near S^{t} André and Ribes) and a larger massif to the S, reaching a thickness of ~2 km (Fig. 3). At the outcrop scale, contrasting relationships were observed. The contact is very often gradational, with albite gneisses progressively losing their feldspar load and grading to mica schists. In contrast, SE of Malons (Fig. 3), gneisses and schists are juxtaposed via a shear zone marked by the concentration of quartz lenses, intense folding, and the occurrence of mylonites. Asymmetric folds, interpreted as drag folds, consistently indicate a top-to-the NW transport, i.e. thrusting of the gneisses over the schists (Fig. 4h), in agreement with the observations of Elmi et al. (1989) and Bouilhol et al. (2006). Finally, the eastern contact of the main gneiss massif is underlined by a distinctive decametric white quartzite (quartz > 95% vol.) layer (referred to as the “Peyremale” quartzite; Fig. 3, 4g). The contact is sharp, parallel to the bedding in the quartzite (evidenced by grain-size variations) and can be followed along several kms.

Within the gneiss bodies, the transition between “augen” and “albite” types is very often gradational and intermediate facies do exist (Elmi et al. 1989). Mapping of the augen gneiss-rich zones indicates that those are broadly concordant with the outline of the massif and the regional foliation. Conversely, the transposed contact between the leucogneisses and the augen gneisses, well-exposed in the Baume Valley (Fig. 3), is sharp and discordant (Fig. 4f), in agreement with the observations of Crevola et al. (1983), pointing to an intrusive relationship between the leucogneiss protolith and the augen gneiss protolith.

4 Zircon U–Pb dating and trace element compositions of the gneisses

Two representative Cézarenque–Joyeuse gneisses were selected for zircon U–Pb dating to clarify the age and origin of their protoliths. Both samples (augen gneiss 18CEZ01, and a tourmaline-bearing leucogneiss 19CEZ54, Fig. 4c,e) were collected from the Baume river cross-section (see Fig. 3).

4.1 Analytical techniques

Zircon grains were separated using standard techniques (jaw crusher, panning, heavy liquids), cast in epoxy resin and polished down to a near-equatorial grain section. Cathodoluminescence imaging was performed at the CRPG (Nancy, France) using a Jeol SM-6510 SEM equipped with a Gatan CL detector. Zircon U–Pb isotope and trace element analyses were carried out at ETH Zürich, Switzerland, by laser ablation–inductively coupled plasma–sector field–mass spectrometry using a RESOlution (ASI, Australia) 193 nm ArF excimer laser system attached to an Element XR (Thermo Scientific, Germany) mass spectrometer. Further details on the analytical procedures and the results of secondary standard and sample measurements are available in the Supplementary text and Supplementary tables 2, 3, 4 and 5.

4.2 Zircon U–Pb dates

For the augen gneiss sample 18CEZ01, a total of 202 analyses were performed on 148 grains, most of which were euhedral, showed pyramidal tips and ranged in length between 150 and 300 µm with aspect ratios from 1:1.8 to 1:3.7. Cathodoluminescence imaging revealed that half of the grains were composed of a (often CL-bright) core wrapped around by a rim showing concentric oscillatory zoning (Fig. 8). Such rims yielded concordant $^{206}Pb/^{238}U$ dates comprised (n=38) between 475±7 (#164) and 498±6 (#140) Ma. Three rims yielded older concordant dates of 525±9 (#2), 575±8 (#124), 621±11 (#198) Ma. The discordant data (n=17) showed a similar range of $^{206}Pb/^{238}U$ dates. For 10 of them, elevated signals on mass 204 indicated that the discordance was at least partly related to common Pb incorporation. The core analyses (n=64) yielded dominantly Neoproterozoic (40 concordant plus 4 discordant spots, $^{206}Pb/^{238}U$ dates from 542±10 Ma, #55, up to 999±13 Ma, #202) and Paleoproterozoic to Neoarchean dates (14 concordant and 7 discordant spots, with $^{207}Pb/^{206}Pb$ between 1870±47, #50, and 2718±25 Ma, #218). One core yielded a concordant Paleoarchean date of 3282±17 Ma (#20), making it the oldest concordant analysis so far obtained in the eastern Massif Central (see Chelle-Michou et al., 2017; Couzinié et al., 2019). The remaining 73 grains lacked any core–rim relationship and showed well-developed concentric magmatic oscillatory zoning. Thirty-seven analyses yielded concordant to slightly discordant $^{206}Pb/^{238}U$ dates clustered between 479±5 Ma (#213) and 499±7 Ma (#170). In contrast, 34 grains yielded concordant to slightly discordant Lower Cambrian to Neoproterozoic dates between 526±8 Ma (#167) and 753±15 Ma (#92) and 4 grains

(2 discordant) showed $^{207}Pb/^{206}Pb$ dates of 1898±30 (#106) to 2174±22 Ma (#179). Seven analyses were strongly discordant and not considered further.

For the tourmaline-bearing leucogneiss sample 19CEZ54, 100 analyses were performed on 68 grains which appeared often broken, slightly corroded and of smaller size (130–170 µm, rarely up to 250 µm) than in sample 18CEZ01. Aspect ratios of euhedral unbroken grains were also more variable and generally comprised between 1:2.2 and 1:2.8 but up to 1:5 in some cases. Fifty-two analyses were performed on grains devoid of core-rim relationships but showing prominent concentric oscillatory zoning (striped zoning in the most elongated grains). The concordant analyses clustered between 468±6 (#68) and 489±6 (#90) Ma (n=24) with 2 grains showing younger dates of 428±12 (#75) and 433±9 (#94) Ma and 12 grains older dates from 545±6 (#44) to 611±9 (#86), up to 1212±6 (#76) Ma. Out of the 16 discordant $^{206}Pb/^{238}U$ dates obtained, 8 overlapped with the concordant data (490–480 and at c. 540 Ma) while 8 defined a continuous trend from 455±7 (#87) to 414±6 (#36) Ma. The remaining analyses were performed on grains comprising a relict core surrounded by oscillatory-zoned overgrowths. Eleven rims yielded concordant overlapping $^{206}Pb/^{238}U$ dates between 480±5 (#57) and 490±6 (#41) Ma and a discordant date of 356±6 (#96) Ma. The 24 analysed cores showed concordant to slightly discordant $^{206}Pb/^{238}U$ dates between 492±11 (#52, disc.) and 695±10 (#82) Ma (16 concordant, 3 discordant) with 3 cores having older slightly discordant $^{207}Pb/^{206}Pb$ dates of 2045±25 (#95) to 2119±20 (#37) Ma. Out of the 27 discordant analyses, 16 showed detectable signals on mass 204 suggesting that discordance was at least partly related to common Pb incorporation.

To sum up, both gneiss samples were typified by the occurrence of oscillatory-zoned grains and rims with $^{206}Pb/^{238}U$ dates in the range 475–500 Ma, i.e. Furongian–

Tremadocian. The cores and the remaining oscillatory-zoned grains generally displayed Lower Cambrian–Neoproterozoic (525–999 Ma, peak at 600 Ma, Fig. 10b) and Paleoproterozoic–Neoarchean dates (1.8–2.7 Ga). For sample 18CEZ01, the density distribution of the Furongian–Tremadocian dates (n=72) was clearly symmetrical, centred on 487.4 Ma but importantly, the results of a χ^2_{red} test for homogeneity indicated that the data were overdispersed given the estimated analytical uncertainties (MSWD=2.3, p-value of 10^{-9}). This overdispersion is due to the 5 youngest and 5 oldest dates of the population. Alternatively, the statistical procedure of Montel et al. (1996) shows that the data can be modelled as a mixture of two populations, one centred at 492.2±1.3 Ma (n=46) and the other at 484.2±1.0 Ma (n=27, with p-value of 0.94). Importantly, the grains belonging to each population are undistinguishable based on texture or trace element compositions. For sample 19CEZ54, 34 analyses clustered in the range 490–468 Ma. Their distribution was asymmetrical and negatively skewed (towards younger ages). Excluding the three youngest analyses allowed to calculate a weighted average date of 483.0±1.2 Ma (±5.5 after propagation of systematic uncertainties) with an MSWD of 1.46 (p-value = 0.05) indicating that the considered grains may constitute a single population.

4.3 Zircon trace element compositions

Only analyses obtained from concordant U–Pb spots will be presented and further discussed, summing up to 159 spots for the augen gneiss 18CEZ01 and 64 for the leucogneiss 19CEZ54. Eu/Eu* were calculated as $Eu_N/(Sm_N \times Gd_N)^{0.5}$ but Ce/Ce* as $(Nd_N)^2/Sm_N$ following the approach of Loader et al. (2017). Since this method does not rely on the La and Pr contents, it is less sensitive to presence of minute inclusions of apatite, monazite or xenotime which would flaw the LREE budget of the analysed zircon grain and lead to inaccurate estimates of Ce/Ce* (e.g. Ni et al., 2020). In the formulas, the $_N$ stands for "normalized to the chondritic values", here of Boynton (1984). Ranges given in the description below correspond to the 1st and 3rd quartiles of the distribution unless stated otherwise.

For the augen gneiss 18CEZ01, all grains no matter their age or texture showed relatively steep HREE patterns with $(Dy/Yb)_N$ mostly between 0.15–0.25 (Fig. 11a,c), and a variable range of Ti contents (4.4–9.5 ppm). Furongian–Tremadocian grains and rims displayed very consistent trace element signatures that strongly contrasted with that of the Lower Cambrian/Neoproterozoic and Paleoproterozoic/Neoarchean grains and cores (Fig. 11). They were typified by lower Th/U ratios (0.03–0.13 vs. 0.30–0.74 and 0.53–0.92, respectively), lower LREE contents (Ce: 0.3–1.3 vs. 5.6–19.1 and 10.8–21.4 ppm; Nd: 0.4–1.1 vs. 0.8–2.4 and 1.3–5.2 ppm), deeper Eu negative anomalies (Eu/Eu*: 0.04–0.08 vs. 0.13–0.37 and 0.06–0.15) and less pronounced Ce positive anomalies (Ce/Ce*: 5–10 vs. 21–100 and 11–37).

A similar situation was observed in leucogneiss sample 19CEZ54 (Fig. 11) with all grains exhibiting steep HREE patterns (with $(Dy/Yb)_N$ centred around 0.16–0.25) and variable Ti contents (5.4–10.3 ppm). The Furongian–Tremadocian grains and rims had nearly identical trace element compositions than those extracted from 18CEZ01 which, here again, markedly contrasted with the signature of Neoproterozoic grains and cores. Their Th/U and LREE contents were systematically lower (Th/U: 0.08–0.32 vs. 0.49–0.85; Ce: 0.5–2.1 vs. 10.2–27.1 ppm; Nd: 0.6–1.7 vs. 1.3–5.5 ppm), the Eu negative anomalies were more pronounced (Eu/Eu*: 0.03–0.06 vs. 0.13–0.33) and the Ce positive anomalies weaker (Ce/Ce*: 5–9 vs. 16-71). Collectively, this indicates that the Furongian–Tremadocian and the older-than-Lower-Cambrian zircon grains crystallized in magmas of contrasted chemical compositions.

4.4 Ti-in-zircon and zircon saturation thermometry

The crystallization temperatures of Furongian–Tremadocian zircon grains and rims were calculated from their Ti contents using the equation of Ferry and Watson (2007) with a_{SiO2} of 1 and a_{TiO2} of 0.5 as suggested by Schiller and Finger (2019) for S-type felsic magmas. In both samples, estimated values spread over a large range of temperatures (between 670 and 900 °C) but with a main cluster at 730–770 °C (Fig. 12).

Zircon saturation temperatures were obtained based on the major element and Zr contents of the dated samples using the equation of Watson and Harrison (1983). A melt having a composition akin to augen gneiss sample 18CEZ01 would be saturated in zircon below 844±42 °C (considering a 5% uncertainty). A lower temperature of 721±36 °C was retrieved for the leucogneiss sample. Given the common preservation of zircon cores in both samples, these values should be regarded as maximum magma temperatures.

5 Discussion

5.1 Nature and age of the protoliths of the Cézarenque–Joyeuse gneisses

5.1.1 Interpretation of the field relationships, petrography, and whole-rock compositions

Three models exist regarding the origin of the Cézarenque–Joyeuse gneisses: (a) they represent metasediments (conglomerates and arkoses) reworking materials of igneous origin (granite pebbles, microcline clasts, Weisbrod and Marignac, 1968); (b) they are metagranites (Crevola et al., 1983); (c) they correspond to volcanic and volcano-sedimentary rocks (Chenevoy 1968a, 1968b; Faure et al., 2001). Model (a) should be discounted as a significant proportion of the gneisses (the augen facies and at least a third of the albite gneisses) show acidic *igneous* compositions (see section 3.1). Model (c) is preferred against model (b) as the latter cannot explain the gradual transition at the outcrop scale from rocks of igneous vs. sedimentary (two thirds of the albite gneisses) origin. The volcanic model is also supported by the elevated normative Qz/(Ab+Or) ratios (average of 0.66) of the igneous samples consistent with a low pressure of crystallization of their parental magmas (Wilke et al., 2017) and the sharp, stratigraphic, contact between the gneisses and the overlying Peyremale quartzite (see section 3.3), pointing to a near-surface emplacement.

In this frame, based on the SiO_2–alkalis systematics (Fig. 5b), we infer that the augen and albite gneisses with igneous compositions represent now-metamorphosed rhyodacites. Their blue quartz grains and pseudomorphs after K-feldspar (for the augen facies) embedded in a finer grained matrix are regarded as former igneous phenocrysts, a view consistent with the fact that the blue colour of quartz is commonly interpreted as resulting from rutile exsolution from grains crystallized at high temperature (T > 700°C, Seifert et al., 2011). The persistence of blue quartz grains in the albite gneisses of sedimentary origin and their composition akin to greywackes indicate that they represent the proximal remobilization products of the adjacent rhyodacites, i.e. epiclastic tuffs. Finally, the leucogneisses are strongly silicic and show a large range of Fe/Mg ratios (Fig. 7) which are typical of fractionated granitic melts. Given their cross-cutting field relationships and their elevated normative Qz/(Ab+Or) ratios (again pointing to crystallization at low pressure), they likely represent former microgranite dykes intrusive within the rhyodacites and associated epiclastic tuffs.

5.1.2 Interpretation of U–Pb results

In line with the abovementioned arguments, the augen gneiss sample 18CEZ01 should be regarded as a metamorphosed porphyritic (K-feldspar-bearing) rhyodacite and the leucogneiss sample 19CEZ54 as a metamicrogranite. In the following, only concordant zircon analyses will be discussed, discordant results being ascribed to a combination of common Pb incorporation and Pb loss.

The augen gneiss (metarhyodacite) 18CEZ01 contains many magmatic grains and rims yielding Furongian–Tremadocian $^{206}Pb/^{238}U$ dates, the distribution of which is well-centred around 487 Ma and would correspond to the crystallization age of the magma. Importantly, considering the estimated analytical uncertainties, the dates are statistically overdispersed. The lack of correlation between U contents and $^{206}Pb/^{238}U$ dates and the symmetrical shape of the date distribution collectively suggest that this overdispersion is of geologic significance (Spencer et al. 2016) and should not be ascribed to limited Pb loss. In line with the volcanic origin of 18CEZ01, it is likely that the analysed dataset comprises syn-eruptive zircon grains and rims plus antecrysts which formed in deep-seated magma chambers (e.g. Matzel et al., 2006) and were scavenged during the eruption. For such cases (mixture of pre-eruptive and syn-eruptive grains, undistinguishable on textural and chemical grounds), Vermeesch (2021) argued that the “Maximum Likelihood Age” (MLA) model designed by Galbraith (2005) has a high potential to unravel the actual crystallization age of the igneous rock. Indeed, the MLA model presumes that the antecrysts define a continuous range of pre-eruptive dates, which is arguably more realistic than the “two populations” (i.e. two age clusters) subdivision (see section 4.2) inferred based on the statistical procedure of Montel et al. (1996). Running the MLA algorithm yields 486.1±0.9 Ma (±5.5 Ma when systematic uncertainties are considered), which is the best estimate of the eruption age of the augen gneiss parental magma. We interpret the grains and cores showing older dates (from the Lower Cambrian to Paleoarchean) as inherited from the magma source or as xenocrysts incorporated from the country-rocks during magma ascent and emplacement.

For the metamicrogranite 19CEZ54, most oscillatory-zoned grains and rims also yielded $^{206}Pb/^{238}U$ dates in the range 475–500 Ma with two concordant analyses showing overlapping younger dates of 428±12 Ma (#75) and 433±9 Ma (#94). We do not ascribe any significance to these dates because in both cases the $^{206}Pb/^{238}U$ date / $^{207}Pb/^{206}Pb$ date ratios are < 90%, hence suggesting post-magmatic disturbance of the isotopic system (Spencer et al., 2016). Besides, the grain domain corresponding to analysis #94 showed elevated Ti (~160 ppm), Nb–Ta (9 and 10 ppm, respectively) and Pr contents (0.8 ppm), which are an order of magnitude higher than in other zircon grains and likely attests to the presence of Ti- (and possibly Nb–Ta) oxide and apatite inclusions. Excluding these two dates, we infer that the cluster at c. 483 Ma represent the crystallization age of the parental magma, in agreement with textural evidence. Importantly, the negatively skewed distribution argues for the occurrence of Pb loss from the 475–500 Ma magmatic population (Spencer et al., 2016) and, excluding the 3 youngest dates, we retain the weighted average date of 483.0±1.2 Ma (±5.5 Ma considering systematic uncertainties) as the best estimate of the crystallization age of the microgranite protolith.

The relative age difference of 3 Ma between the crystallization of the two felsic melts exceeds the internal uncertainties and, since both samples were dated during the same analytical session, we infer that such time interval is of geological significance. The slightly younger crystallization age of the microgranite 19CEZ54 is consistent with field observations supporting an intrusive relationship with the porphyritic metarhyodacite 18CEZ01.

5.2 Typology and petrogenesis of the igneous association

Nearly overlapping crystallization ages and key geochemical markers (including almost identical whole-rock Na/K ratios, Fig. 7d, and similar magmatic zircon trace element systematics, Fig. 11) collectively indicate that the metarhyodacites and the metamicrogranites were genetically related and constitute a single magmatic system. Considered together, they define a peraluminous, calc-alkalic to alkali-calcic, magnesian and sodi-potassic association (Fig. 7), which is the hallmark of crust-derived magmatic suites (Bonin et al. 2020). Further evidence for a crustal origin is provided by the marked zircon inheritance (Laurent et al. 2017) and zircon trace element systematics. Indeed, the low Th/U ratios displayed by the magmatic zircon grains, positively correlated with LREE contents (Fig. 11e,f), can be regarded as a consequence of coeval monazite precipitation from the melt upon cooling which is a typical feature of highly aluminous and P-rich “S-type” granitic melts formed by melting of sedimentary rocks (Cuney and Friedrich, 1987). Besides, as Ce^{4+} and Eu^{3+} are less incompatible with respect to zircon than Ce^{3+} and Eu^{2+} (e.g. Trail et al., 2012, and references therein), the low Ce/Ce* and Eu/Eu* ratios (weak Ce positive anomalies and marked Eu negative anomalies, Fig. 11g,h) displayed by magmatic zircon grains demonstrate that the parental magmas were notably reduced, as commonly observed in S-type granitic melts (Whalen and Chappell, 1988).

Examination of the inherited zircon date distributions and trace element systematics sheds light on the nature and origin of the crustal source. The main Neoproterozoic (Ediacaran–Cryogenian, 541–720 Ma) and subordinate Paleoproterozoic (1.9–2.2 Ga) clusters are remarkably matching those observed in Ediacaran metasediments from deeper structural levels of the Cévennes domain (Chelle-Michou et al., 2017; Couzinié et al., 2019) which strongly suggests that the Cézarenque–Joyeuse magmas formed by melting of Ediacaran sedimentary rocks. The occurrence of two inherited grains showing overlapping c. 525 Ma U–Pb dates indicates that the magmas interacted at some point of their evolution with a Lower Cambrian crustal component, possibly during ascent (in which case these zircon grains are xenocrysts).

Interestingly, the inherited Precambrian grains have a wider range of trace element compositions compared to magmatic grains and include a significant proportion of zircon with higher Th/U and LREE contents together with stronger positive Ce anomalies and weaker Eu negative anomalies. These latter signatures suggest that the Precambrian igneous rocks which erosion products fed the Ediacaran sedimentary basins largely originated from more oxidised (Fig. 11g,h), probably arc-derived magmas (Ishihara 2004). For the Neoproterozoic grains, the long-lived accretionary Cadomian orogeny, developed along the northern Gondwana margin (Garfunkel, 2015, and references therein), constitutes the most likely source. The Paleoproterozoic to Archean zircon grains would stem from the recycling of old Gondwanan crustal materials, possibly originating from the Saharan Metacraton (Couzinié et al., 2019).

The Cézarenque–Joyeuse felsic igneous rocks have whole-rock compositions matching those of the so-called *ferrosilicic* suites (Castro et al. 2009), i.e. they are anomalously rich in Fe and Mg and low in Ca. Such signatures have been thought to result from very high melting temperatures (> 1000 °C) of sedimentary rocks (mostly greywackes, Castro et al., 2009). However, several observations argue against this model for the Cézarenque–Joyeuse gneisses. First, a large pool of inherited zircon grains and cores were preserved throughout the magmatic evolution implying either moderate melting temperatures (< 850 °C, below the zircon saturation temperature of the augen gneiss) or very fast melt production and transfer to the upper crust (Watson, 1996; Bea et al., 2007). In the latter case, the magmas are expected to have followed a nearly adiabatic path from source to surface. This way, crystallization temperatures deduced from Ti-in-zircon thermometry (clustering at 730–770 °C, see section 4.4) should lie within 50 °C of the actual melting temperatures (Holtz and Johannes, 1994), which, therefore, could not have exceeded 820 °C. As an alternative to the very high melting temperature model, Fiannacca et al. (2019) suggested that the *ferrosilicic* signature should be explained by the selective incorporation in the melt phase of mafic materials (mostly peritectic garnet, Stevens et al., 2007) present in the sedimentary source. Yet appealing, this model cannot be directly tested in that any petrographic evidence for such entrainment was irremediably erased during the dissolution of the added crystal load.

Novel insights on the origin of the *ferrosilicic* signature may be gained from the examination of the Eu/Eu* and Ce/Ce* systematics of the Cézarenque–Joyeuse magmatic zircon grains and rims. Measured values (Fig. 11g,h) are significantly lower compared to zircon encountered in “typical” (not *ferrosilicic*) sediment-derived granites from the Lachlan Fold Belt (Burnham and Berry 2017) and southern Tibet (Wang et al. 2012; Gao et al. 2016), calling for anomalously reducing conditions. The fO_2 prevailing during zircon crystallization was most likely close to the value expected for the iron–wüstite buffer considering the calibration of Trail et al. (2012). Little is known on the influence of fO_2 on anatectic melt compositions, but Gaillard et al. (2001) showed that the Fe solubility of a subaluminous (A/CNK between 1 and 1.1) melt at 930 °C is negatively correlated to fO_2: at NNO+1.5, maximal FeO_t contents reach ~1.8 wt.% but are higher than 3 wt.% at fO_2 below FMQ. Since the alumina content of a granite melt at a given fO_2 has no influence on the Fe solubility (Holtz et al., 1992), this result should also be valid for peraluminous melt compositions. Hence, we posit that the high Fe contents displayed by the *ferrosilicic* rocks may result from an enhanced Fe solubility in the melt phase itself caused by strongly reducing conditions. Those may be inherited from the source level (melting in a graphite-buffered environment) or acquired during ascent via interaction with organic matter (possibly oil-bearing) sedimentary rocks, a phenomenon known to deeply affect the fO_2 of magmas (e.g. Iacono-Marziano et al., 2012). As a matter of fact, the Lower Cambrian sequences of the southern Massif Central *autochthon* host a > 200 m-thick black shale formation (Alvaro et al. 2014), to which graphite-bearing mica schists of the Cévennes parautochthon may be correlated. Hence, the low fO_2 and peculiar high Fe contents of the Cézarenque–Joyeuse parental magmas may result from their interaction with the local Lower Cambrian sediments prior to the eruption (Fig. 13).

5.3 The Cézarenque–Joyeuse magmatism in the frame of the northern Gondwana evolution

Felsic (sub)volcanic and volcanosedimentary associations coeval to the Cézarenque–Joyeuse gneisses are widespread over the northern Gondwanan terrains of SW Europe. In this section, we first provide a short review of correlative formations and then address the geodynamic setting and the trigger(s) of this magmatic flare-up.

The best-characterized Furongian–Lower Ordovician metavolcanic suite of the northern Gondwana realm is arguably the Ollo de Sapo Formation in the Iberian Massif (see review in García-Arias et al., 2018; von Raumer and Stampfli, 2018). The Ollo de Sapo gneisses strikingly resemble their Cezarenque–Joyeuse counterparts: lithological types (augen vs. fine-grained facies) and whole-rock chemical compositions are alike (Fig. 7), as already noted by Weisbrod (1969); their ages overlap within error (Montero et al., 2007, 2009, 2017; Díez Montes et al., 2010; Lopez-Sanchez et al., 2015); the same petrogenetic model is inferred, i.e. melting of Ediacaran sedimentary rocks (Montero et al. 2017). Additional correlations can be established with several massifs exposed throughout the Variscan nappe pile: (i) in the Massif Central (Fig. 1), the Cézarenque–Joyeuse gneisses recall the metarhyolites from the St-Salvi-de-Carcavès–St-Sernin-sur-Rance and Thiviers–Payzac units (Guérangé-Lozes and Burg, 1990; Melleton et al., 2010; Alvaro et al., 2014; Pouclet et al., 2017) and from the Indre group (Nouzier metavolcanics, Quesnel et al., 2009) ; (ii) in the Armorican Massif, they resemble the felsic rocks from the “Porphyroid nappe” and the Brétignolles, Île de Groix and La Châtaigneraie metavolcanics (Bouton and Branger, 2008; Ballèvre et al., 2012; El Korh et al., 2012); (iii) in the Iberian Massif, besides the Ollo de Sapo Formation, they are akin to several metarhyolites bodies exposed in the parautochthonous “Schistose Domain” (Valverde-Vaquero et al., 2005; Dias Da Silva et al., 2014; Farias et al., 2014); (iv) in Sardinia, they compare to the volcanics of the San Vito Formation and the Li Trumbetti and Mt. Geisgia rhyolites (Oggiano et al., 2010). Furthermore, many peraluminous crust-derived (meta)granites of similar Furongian–Lower Ordovician age have been described in the same areas and possibly represent former deep-seated magma reservoirs that fed the eruptions (Álvaro et al., 2020). Relevant examples include the Mendic pluton in the Massif Central (Demange, 1982; Alvaro et al., 2014; Pouclet et al., 2017), the Mervent orthogneiss in the Armorican massif (Diot et al., 2007) and numerous massifs in Iberia (Díez Fernández et al., 2012; Talavera et al., 2013).

Altogether, petrological and geochronological evidence substantiate that a pervasive crustal melting event took place along a > 2000 km segment of the northern Gondwana margin (Álvaro et al., 2020). From a geodynamic perspective, an intracontinental rift-setting (Fig. 13) should be retained based on: (i) the linear shape of the magmatic belt and the synchronous deposition of thick sedimentary sequences (Pouclet et al., 2017); (ii) the occurrence of “anorogenic” A-type magmatic rocks in NW Iberia (Díez Fernández et al., 2012, and references therein), in the Maures massif (Seyler, 1986; Briand et al., 2002), in the Armorican Massif (Ballèvre et al., 2002) and in the southern Massif Central (Albigeois area, Pin and Marini, 1993); (iii) the lack of regional metamorphism related to crust thickening (e.g. Montero et al., 2007); and (iv) the coeval opening of the Rheic ocean (Díez Montes et al., 2010; Nance et al., 2010). The ultimate origin of this rifting event remains debated. Many authors put forward the role of the southwards (in Ordovician coordinates) subducting Iapetus slab in controlling crustal extension (Fernández et al., 2008; Díez Montes et al., 2010; Díaz-Alvarado et al., 2016; Oriolo et al., 2021) while others retained a plume-induced origin (Briand et al. 1992, 2002). The geological record of the Cévennes parautochthon does not provide any evidence that would help clarify this point.

An atypical feature of this rifting event is the voluminous generation of peraluminous crust-derived magmas, which are more commonly encountered in syn- to post-collisional orogenic settings (Barbarin, 1999). Two factors seem to have played a key role in enabling the crust to produce granitic melts under such conditions. First, evidence for mantle-derived magma underplating and associated advective/conductive heat transfer are provided by the occurrence of coeval mafic magmatic rocks in the vicinity of the felsic volcanic centres (Bea et al., 2007; Montero et al., 2009; Díez Montes et al., 2010; Díaz-Alvarado et al., 2016; Pouclet et al., 2017). However, such rocks represent small volumes compared to the felsic suites and are not systematically exposed (e.g., in the Cézarenque–Joyeuse area) thus calling for an additional heat supply. As a matter of fact, the thick Ediacaran sedimentary sequences which constitute the source rocks of the Furongian–Lower Ordovician magmas were characterized by anomalously high radiogenic heat productions: > 2.7 $\mu W.m^{-3}$ at 550 Ma in the Iberian “Schist and Greywacke Complex” (Bea et al., 2003) and average of 3.0 $\mu W.m^{-3}$ in the Massif Central (Couzinié, 2017), i.e., 50% higher than the mean upper crustal composition of Rudnick and Gao (2003). This feature was explained by the recycling of Cadomian felsic igneous rocks which selectively enriched the Ediacaran detritus in K, Th and U (Bea et al., 2003). Numerical modelling indicates that the presence of layers with elevated radiogenic heat productions within the crust can significantly affect its thermal structure and provoke heating during subsidence (Sandiford et al., 1998). Therefore, we infer that the pre-rifting crust structure and composition (inherited from its late Neoproterozoic evolution) played an important role in enhancing melt production at Furongian–Lower Ordovician times.

6. Conclusion

Field relationships, petrography and whole-rock geochemistry collectively indicate that the Cézarenque–Joyeuse gneisses represent former rhyodacite edifices and their erosion products originally interlayered within detrital sedimentary sequences. Zircon U–Pb dating of the gneisses demonstrate that the felsic magmas erupted or were emplaced at very shallow crustal levels between 486.1±5.5 Ma and 483.0±5.5 Ma. In that regard, the Cézarenque–Joyeuse gneisses do represent a newly identified fragment of the Furongian–Lower Ordovician volcanic belt of SW Europe. Inherited zircon date distribution and magmatic zircon trace element systematics further substantiate that the source of the Furongian–Lower Ordovician magmas corresponded to Ediacaran sediments and suggest that the *ferrosilicic* signature they commonly exhibit may be explained by the strongly reduced character of the silicic melts (fO_2 close to the iron–wüstite buffer) acquired through interaction with Lower Cambrian organic matter-bearing sediments (Fig. 13). Such a model would be valid for other Furongian–Lower Ordovician ferrosilicic rocks of the northern Gondwana realm as Lower to Middle Cambrian black shales have been described in the Central Iberian Zone (Alvaro et al., 2020b) and in the southern Armorican massif (Pouclet et al. 2017). In light with available paleogeographic constraints, it is inferred that crustal melting took place in an intracontinental rift setting and was enhanced by mantle-derived magma underplating and the anomalously high radiogenic heat production of the Ediacaran sedimentary sequences.

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Figure captions

Figure 1: Sketch map of the Variscan terrains of SW Europe outlining the location of several felsic (sub)volcanic complexes of Furongian–Lower Ordovician age, adapted from Ballèvre et al. (2014). Zones in the Iberian Massif: CIZ, Central Iberian; CZ, Cantabrian; GTMZ, Galicia–Trás-os-Montes; OMZ, Ossa-Morena; SPZ, South-Portuguese; WALZ, West Asturian–Leonese. V: Vosges; BF: Black Forest; MM: Maures Massif; CO: Corsica. Abbreviations for the (sub)volcanic formations. In the

Iberian Massif, LQ: Loiba and Queiroga series (Valverde-Vaquero et al. 2005); MS: Mora and Saldanha volcanic complexes (Dias Da Silva et al. 2014); P: volcanics from the Paraño Group in the Bragança–Alcañices area (Farias et al. 2014). Armorican Massif, Pn: Porphyroid nappe and B: Brétignolles metavolcanics (Ballèvre et al. 2012); IG: Ile-de-Groix felsic gneisses (El Korh et al. 2012); LC: La Châtaigneraie metavolcanics (Bouton and Branger 2008). In the Massif Central, the Larroque rhyolitic formation in the S^{t}-Sernin-sur-Rance (R) and S^{t}-Salvi-de-Carcavès nappes (SSC) (Guérangé-Lozes and Burg 1990; Pouclet et al. 2017); TP: the Génis rhyolites in the Thiviers-Payzac Unit (Melleton et al. 2010; Pouclet et al. 2017) ; Nz : the Nouzier metavolcanics (Quesnel et al. 2009). In Sardinia, SV: San Vito Formation volcanites and TG: Li Trumbetti and Mt. Geisgia rhyolites (Oggiano et al. 2010).

Figure 2: Geological map of the Cévennes domain outlining the main lithotectonic formations. Redrawn and adapted from Chantraine et al. (2003).

Figure 3: Geological map and associated interpretative cross-sections of the eastern Cévennes highlighting the intricate association between the different gneiss facies and the “Cévennes schists”. The location of the dated samples is indicated in the inset and the cross-section centred on the Baume valley. Map redrawn and adapted from the works of Elmi et al. (1974, 1989) and Bouilhol et al. (2006).

Figure 4: Representative photographs of the Cézarenque–Joyeuse gneisses and their relationships with adjacent lithologies: (a) augen gneiss with a well-developed foliation, cm-sized lenticular quartz-feldspar aggregates and bluish quartz grains (44.3884°N, 4.0876°E); (b) poorly-foliated augen gneiss facies with euhedral K-feldspar and blue quartz phenocrysts embedded in a finer-grained matrix (44.3935°N, 4.0787°E); (c) augen gneiss, dated sample 18CEZ01 (44.5130°N, 4.2134°E) ; (d) isoclinally folded albite gneiss (44.5023°N, 4.1224°E); (e) fine-grained tourmaline-bearing leucogneiss,

dated sample 19CEZ54 (44.5103°N, 4.2107°E); (f) oblique contact between a leucogneiss and the S_{0-1} foliation of the augen gneiss, interpreted as former intrusive relationships (44.5124°N, 4.2135°E); (g) concordant contact between the S_{0-1} foliation of the augen gneisses and the overlying “Peyremale” quartzite (44.3526°N, 4.0676°E); (h) asymmetric folds in schists, interpreted as evidence for local top-to-the-NW shearing at the gneiss–schist contact (44.4750°N, 4.1309°E).

Figure 5: Representative thin section photomicrographs of the Cézarenque–Joyeuse gneisses: (a) augen gneiss with 0.5 cm-sized engulfed quartz grains embedded in a finer-grained foliated matrix with alternating layers of quartz–albite and biotite (chloritized)–muscovite; (b) close-up on the inner part of a 2 cm-sized lenticular quartz–feldspar aggregate showing the preservation of relict K-feldspar (microcline) surrounded by a fine albite (sericitized)–quartz association; (c) albite gneiss with alternating fine-grained quartz–albite and biotite (chloritized)–muscovite layers overgrown by mm-sized albite porphyroblasts; (d) leucogneiss composed of a foliated quartz, feldspar and muscovite assemblage with truncated tourmaline grains (dated sample 19CEZ54, see text and Fig. 4).

Figure 6: Representative field photographs of the “Cévennes schists”: (a) metapelitic chlorite–muscovite schist (44.5028°N, 4.1213°E); (b) schist with fine alternations of quartz-rich (poorly foliated) and quartz-poor domains (44.5833°N, 4.0545°E), likely derived from a semi-pelite protolith; (c) schist typified by layers rich in mm-sized albite porphyroblasts (44.5924°N, 3.9745°E); (d) dark (biotite-bearing) quartz schist, a facies primarily encountered close to the Cézarenque–Joyeuse gneisses (44.3937°N, 4.0781°E).

Figure 7: Whole-rock geochemistry of the Cézarenque–Joyeuse gneisses. (a) Al/3-K vs. Al/3-Na cationic diagram of de La Roche (1968) aiming at discriminating meta-

igneous from metasedimentary rocks. (b) Total alkali vs. silica classification diagram of Le Bas et al. (1986). The dashed line is the subalkaline–mid-alkaline boundary following Rittman (1957). (c) B–A cationic classification diagram of Debon and Le Fort (1988) with the subdivisions of Villaseca et al. (1998). (d) SiO_2 vs. Na_2O + K_2O - CaO and (e) SiO_2 vs. FeO_t/(FeO_t+MgO) (in wt.%) diagrams of Frost et al. (2001). (f) K/(K + Na) vs. B. cationic classification diagram of Debon and Le Fort (1988). The dotted lines represent the contours encompassing 85% of the data for the following distributions: Variscan plutonic and volcanic rocks from the French Massif Central with SiO_2 between 63–72% (n=1350, Moyen et al., 2017), the Ollo de Sapo gneisses (n=141, based on the compilation of García-Arias et al., 2018), the Cévennes mica schists (n=48, data from Weisbrod, 1970, Harlaux, 2016, and Couzinié, 2017). The contour levels were drawn using the kde2d function of R (Venables and Ripley 2002). The B–A values for sediment-derived experimental melts at 2–15 kbar and 700–900 °C were taken from the compilation in Couzinié et al. (2017). The black arrow in (c) depicts the increase in Fe solubility observed by Gaillard et al. (2001) in a subaluminous 930 °C granitic melt when fO_2 decreases from NNO+1.5 down to NNO-0.7 (~QFM, NNO sanding for Ni–NiO buffer).

Figure 8: Representative cathodoluminescence images of zircon grains from the Cézarenque gneisses. The locations of laser spots (white circles) are indicated along with the spot name (#XX). The corresponding $^{206}Pb/^{238}U$ dates (if < 1.2 Ga) or $^{207}Pb/^{206}Pb$ dates (if > 1.2 Ga) are quoted with 2σ uncertainty, in Ma. All displayed analyses are concordant.

Figure 9: Zircon U–Pb results for the Cézarenque gneiss samples. (a,b) Tera–Wasserburg diagrams ($^{238}U/^{206}Pb$ vs. $^{207}Pb/^{206}Pb$). Error ellipses/ages are quoted at the 95% confidence level. Analyses were not corrected from common Pb. (c,d)

Individual Furongian–Lower Ordovician ^{206}Pb/^{238}U dates (grey bars) and their distributions represented as Kernel Density Estimates (with an adaptative bandwidth). Only concordant data were considered. (e) Radial plot and Maximum Likelihood Age estimate for sample 18CEZ01 (see Vermeesch, 2021).

Figure 10: Histograms and Kernel Density Estimates (plotted with an adaptive bandwidth) showing the U–Pb date distribution of older-than-Lower Cambrian zircon grains, rims, and cores from the Cézarenque–Joyeuse gneisses. Only concordant ^{206}Pb/^{238}U dates (if < 1.2 Ga) and concordant to slightly discordant (“^{206}Pb/^{238}U date”/”^{207}Pb/^{206}Pb date” > 92%, if > 1.2 Ga) were considered. The data are compared to the U–Pb date distribution of Ediacaran metasediments from the Cévennes parautochthon of the eastern Massif Central, also represented as Kernel Density Estimates (Chelle-Michou et al. 2017; Couzinié et al. 2019).

Figure 11: Zircon trace element data for the Cézarenque–Joyeuse gneisses. (a,b,c,d) Rare Earth Elements patterns normalized to the chondrite values of Boynton (1984); (e,f) LREE contents (Ce+Nd) vs. Th/U diagram for the different date populations; (g,h) Eu/Eu* vs. Ce/Ce* diagram. The background yellow shading mimics the contours of the distribution of 209 zircon analyses from S-type granites which whole-rock compositions are not ferrosilicic (data from Burnham and Berry, 2017; Gao et al., 2016; Wang et al., 2012), drawn using the kde2d function of R (Venables and Ripley 2002). For sake of consistency, Ce/Ce* of the literature zircon were recalculated using the same methodology as for the Cézarenque–Joyeuse gneisses, i.e. following the approach of Loader et al. (2017). Were also plotted the relationships between zircon Eu and Ce anomalies and melt temperature–oxygen fugacity, as estimated by Trail et al. (2012) for peraluminous melt compositions. The oxygen fugacity (fO_2) in this plot is expressed relative to the NNO (nickel–nickel oxide) buffer. At 800 °C, the fO_2 of the

quartz–fayalite–magnetite (QFM) buffer corresponds to NNO-0.8 and that of the iron–wüstite buffer is NNO-5.

Figure 12: Histograms and density distribution of temperatures for magmatic Furongian–Lower Ordovician grains and rims of the Cézarenque–Joyeuse gneisses, obtained via the Ti-in zircon thermometer using the equation of Ferry and Watson (2007) with a_{SiO2} of 1 and a_{TiO2} of 0.5 as suggested by Schiller and Finger (2019) for peraluminous felsic magmas.

Figure 13: Interpretative geodynamic sketch illustrating the Furongian–Lower Ordovician evolution of the northern Gondwana margin. The crustal column on the right panel summarises the petrogenetic model proposed for the protoliths of the Cézarenque–Joyeuse gneiss complex.