International Meeting

September 4-9th 2017, Figueres, Catalonia

ORDOVICIAN GEODYNAMICS:

The Sardic Phase in the Pyrenees, Mouthoumet and Montagne Noire massifs

J. Javier Álvaro, Josep Maria Casas and Sébastien Clausen (eds.)

(cover photo: panorama of the Canigó massif from the south)

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FOREWORD

The presence of a Middle-Ordovician gap punctuating the stratigraphy of southwestern Sardinia (Teichmüller, 1931; Stille, 1939) has represented during the last decades a standard for correlation in southwestern Europe and northwest Africa. Any “similar” gap stratigraphic marking the absence of Furongian (“late Cambrian”)-to-Late Ordovician fossils has been used to invoke the correlation of the Sardinian event in the Alps (e.g., Stampfli *et al*., 2002), the eastern Pyrenees and the Montagne Noire (this vol.). After increasing research, the correlated gaps have finally displayed different time spans, such as the Furongian virtual absence of sedimentary strata in the Moroccan Anti-Atlas (except in the El Graara massif) and the Ossa-Morena Zone of Spain (recently re-interpreted as a break-up or rift-drift unconformity and named Toledanian Phase in the Central-Iberian Zone of the Iberian Peninsula) or the lack of Dapingian (Middle Ordovician) fossil-bearing strata in the Anti-Atlas. Therefore, the “Sardic Phase” is in need of re-evaluation outside Sardinia: any correlation should be based on a multidisciplinary approach based on recognition of stratigraphic gaps, geometrical reconstruction of sedimentary bodies and onlapping features capping inherited palaeoreliefs, structural studies of deformation predating any distinct gap and geochemical analysis of contemporaneous volcanic activies.

The presence of the Sardic Phase and associated Middle Ordovician stratigraphic gaps has been reported in the Eastern Pyrenees, the Mouthoumet massif and the Cabrières klippes of the southern Montagne Noire (Catalonia and Occitanie, southwestern Europe). Late Ordovician fault-controlled subsidence and the record of rifting volcanism were coeval, in some areas, with the onset of the Hirnantian glaciation. As a result, the Upper Ordovician of SW Europe offers a complex mixture of erosive unconformities and intrusion of acidic plutons (Pyrenees), followed by the breakdown of platforms in horsts and grabens and the onset of rifting branches (Mouthoumet and Montagne Noire), onlapping patterns and final sealing of Sardic palaeotopographies during Silurian and Early Devonian times.

This meeting addresses the dynamics of Cambrian-Ordovician sedimentary basins in southwestern Europe and Northwest Africa and aims to bring together a wide range of studies focusing on geodynamics, tectonics, volcanism, sedimentary geometries, event stratigraphy and chronostratigraphic correlation. We aim to balance the study of geodynamic processes recorded throughout North Gondwana with worldwide analogues.

J. Javier Álvaro, Josep Maria Casas and Sébastien Clausen

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2. Meeting in Figueres (4-5th September)

2.1 Timetable for presentations (15'+5' using ppt or pdf files)

3rd September. Evening

20-21 h. Welcome to meeting participants and icebreaking-dinner in Hotel Travé, Figueres.

4th September. Morning

10.00-10.30 h. *Quesada*. Early Palaeozoic evolution of Gondwanan units in the Iberian Massif: from subduction through rifting and drift on the southern passive margin of the Rheic Ocean.
10.30-10.50 h. Álvaro & Vizcaïno. The Furongian break-up (rift/drift) unconformity in the central Anti-Atlas, Morocco.
10.50-11.10 h. *Eguiluz, Palacios, Sarrionandía & Jensen*. Post-Cadomian paleogeographic evolution of Iberia.
11.10-11.30 h. *Chichorro, Solá, Álvaro, Sánchez-García, Quesada & Díez Montes*. The Cantabro-Iberian Basin: Insights of the rift-stage revealed by detrital zircon age distributions.
11.30-11.50 h. *Sánchez-García, Díez-Montes, Álvaro & González-Clavijo*. Furongian-Lower Ordovician metavolcanites from the Toledo Mountains (Spain) and its relationship with the Toledanian phase.

11.50-12.10 h. Coffee break

12.10-12.30 h. *Chichorro, Solá, Quesada, Álvaro, Sánchez-García & Díez Montes*. Cryptic Early-to-Main rift-related Cambrian magmatism in Central Iberian and Galicia-Trás-os-Montes zones (Iberian Massif) revealed by U-Pb dating of inherited zircon.
12.30-12.50 h. *González-Clavijo, Díez-Montes, Dias da Silva & Sánchez-García.* An overview on the Ordovician volcanic rocks and unconformities in the Central Iberian and Galicia-Trás-os-Montes Zones of the Iberian Variscan Massif.
12.50-13.10 h. *Valverde-Vaquero, McNicoll, Beranoaguirre, Díez-Montes, Rubio-Pascual & Rodríguez-Fernández.* Lower Ordovician unconformities and the Ollo de Sapo magmatic event: coeval 478 Ma pyroclastic base-surge deposits and subvolcanic granite intrusion in the Hiendelaencina Antiform.
13.10-13.30 h. *Dias da Silva, González-Clavijo, Díez-Montes, Martínez-Catalán, Gómez-Barreiro, Hoffmann & Gärtner*. Furongian-Late Ordovician volcanism in the Upper Parautochthon of the Galicia-Trás-os-Montes Zone (NE Portugal): Paleogeographic meaning and geodynamic setting.
13.30-13.50 h. *Solá, Chichorro, Álvaro, Díez Montes, Sánchez-García & Quesada*. Provenance of Ordovician Armorican quartzite in Iberia: the linkage with West African Craton.

4th September. Afternoon

16.00-16.20 h. *Colmenar & Rasmussen.* The onset of the Great Ordovician Biodiversification Event (GOBE) in high-latitude peri-Gondwana.
16.20-16.40 h. *Cocco & Funedda.* The Sardic Phase in SE-Sardinia (Italy).
16.40-17.00 h. *Chichorro, Solá, Díez Montes, Álvaro, Sánchez-García & Quesada*. The evolution of the rift-drift transition in the Central Iberian and Galicia-Trás-os-Montes zones, Iberian Massif.
17.00-17.20 h. *Cocco, Casini, Funedda, Loi & Oggiano*. Ordovician tectonics in Sardinia (Italy).

17.20-18.00 h. Coffee break

18.00-19.30 h. Workshop: What on Earth originated the Sardic Phase? Subduction, rift/drift unconformity, rearrangement of drifting trends in neighbouring plates…

5th September, Morning

10.00-10.30 h. *Álvaro & Casas*. The Sardic Phase in the Occitan and Pyrenean Domains of southwestern Europe: an introduction to the fieldtrip.

10.30-10.50 h. *Puddu, Casas & Álvaro*. New knowledge on the Upper Ordovician rocks of El Baell, Ribes de Freser area, eastern Pyrenees.

10.50-11.10 h. *Capdevila, Martínez, Reche, Cirès & Peucat*. Lower-Paleozoic rifting-related magmatism in the Northeastern Iberian Variscan massifs of Nuria and Guilleries: geochemical approach and attempt of palaeogeodynamic interpretation.

11.10-11.30 h. *Margalef, Castiñeiras, Casas, Navidad & Liesa*. Comparison between detrital zircon populations from the Ordovician rocks of the Pyrenees and other Perigondwanan terrains: palaeogeographic implications.

11.30-11.50 h. *Clariana, Valverde-Vaquero, Rubio, Beranoaguirre & García-Sansegundo*. Upper Ordovician magmatism in the Central Pyrenees: First U-Pb zircon age from the Pallaresa Massif.

11.50-12.10 h. Coffee break

12.10-12.30 h. *Clausen, Padel, Poujol & Álvaro*. Cambrian-Lower Ordovician sedimentary provenance shifts in Northwest Gondwana.

12.30-12.50 h. *Puddu, Casas & Álvaro*. On the Upper Ordovician of the La Cerdanya area, Pyrenees.

12.50-13.10 h. *Casas, Sánchez-García, Álvaro, Puddu & Liesa*. Ordovician magmatism in the Pyrenees.

14.00-16.00. Meal

16.00. END OF MEETING

16.00-19.00 h. Visit to Dalí Museum for participants to fieldtrip.

6th September

Beginning of fieldtrip. Night in Bellver de Cerdagne

7th September

Night in Carcassonne

8th September

Night in Carcassonne

9th September

Arrival to Figueres before 18 h. Participants can arrange their night there or return from train station or airport.

2.2 Abstracts of presentations

The Sardic Phase in the Occitan and Pyrenean Domains of southwestern Europe: an introduction to the fieldtrip

J. Javier Álvaro[1] & Josep Maria Casas[2]

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The meaning of the Sardic Phase in the eastern branch of the Variscan Ibero-Armorican Arc is still a matter of debate. Although it is commonly associated with uplift, emersion and considerable erosion before the onset of Upper Ordovician deposition, it is related neither to metamorphism nor cleavage development. As a result, the Upper Ordovician rocks directly onlap an inherited palaeorelief formed by different formations of the pre-Sardic succession, ranging from the Ediacaran to the Lower Ordovician in the Central and Eastern Pyrenees, the lower Cambrian to the Lower Ordovician in the southern Montagne Noire and different units of the Lower Ordovician in the Mouthoumet massif.

A time gap of about 20 m.y. can be estimated for the Sardic Phase in the Pyrenees and Montagne Noire (Occitan Domain) and about 18 m.y. in SW Sardinia. The Sardic uplift (whatever its origin) was necessarily followed by a succession of Late Ordovician extensional pulsations, which preceded and were contemporaneous with the opening of grabens and half-grabens infilled with the alluvial-to-fluvial Rabassa Conglomerate Formation in the eastern Pyrenees and the volcanosedimentary Villerouge and Roque de Bandies formations in the Montagne Noire and Mouhoumet massifs.

A distinct Ordovician magmatism was contemporaneous with the pre-Late Ordovician episode of uplift and erosion that led to the formation of the Sardic unconformity. Early and Late Ordovician ages have been obtained for magmatic bodies from the Occitan Domain (e.g., the Axial Zone of the Montagne Noire) and the Pyrenees. The emplacement of Late Ordovician felsic granitic bodies is coeval in the Cabrières klippes (Montagne Noire) with a tholeiitic volcanic activity originated by melting of mantle and crustal lithosphere and sedimentary infilling of rifting branches.

The Furongian break-up (rift/drift) unconformity in the central Anti-Atlas, Morocco

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In the central Anti-Atlas of Morocco, the "middle Cambrian"-Furongian transition lies in the Jbel Lmgaysmat Formation. The latter is exclusively preserved in a synsedimentary graben that crops out in the El Graara massif, and was originally surrounded by tilted and uplifted areas of non-deposition and/or erosion. Its top marks a break-up unconformity that separates a Cambrian sedimentation primarily controlled by sharp (synsedimentary) modifications in accommodation space (rifting phase) from an Ordovician thermal-dominated (gradual)

subsidence (Álvaro *et al*., 2014a). The Cambrian rift, rich in tholeiitic and subordinate alkaline volcanism, aborted in the Furongian and no mid-ocean ridge basalt (MORB)-type lavas were set in Ordovician times. The true ocean opening probably took place farther to the NW, between the Moroccan Coastal Meseta and the rest of Avalonian blocks.

Lower-Paleozoic rifting-related magmatism in the Northeastern Iberian Variscan massifs of Nuria and Guilleries: geochemical approach and attempt of palaeogeodynamic interpretation

Ramon Capdevila[1], Francisco J. Martínez[2], Joel Reche[2], Jordi Cirès[3] & Jean-Jacques Peucat[4]

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The inner zones of the European Variscan Chain are characterized by the presence of varied volcanic and plutonic bodies metamorphosed at the Carboniferous. These bodies, mainly granitic, were considered of Precambrian age at a time where granite genesis and orogeny were closely linked, but it has been proved that these igneous rocks were emplaced at the Ordovician during a rifting episode. To further clarify the nature and origin of this magmatism, we studied two massifs located in the NE of Spain: the Núria Massif of the Pyrenean Axial Zone, where metagranits have around 457 Ma, and the Guilleries Massif of the Catalan Coastal Ranges, whose metaigneous protoliths have ages between 451 and 488 Ma.

In the Núria massif, three pre-Variscan igneous bodies have been studied. The Ribes of Freser granophyre, an undeformed syenogranite that we interpret as a partial melting product of a granitic basement with a long fractionation history. The Núria orthogneiss is a deformed alkali-feldspar granite with a major and trace element composition typical of S-type granites. We interpret the protoliths of these orthogneiss as metapelite fusion products. The compositionally-varied Carançà augen-gneiss forms a ring around the top of the Núria orthogneiss. We interpret it as a partial melt of an alkaline metaluminous igneous source which became highly peraluminous by hybridization with metapelites.

In the Guilleries massif, the pre-Variscan metaigneous rocks are orthogneisses, felsic metavolcanics and amphibolites. The orthogneisses have a composition of alkali feldspar WPG or POG granites. We interpret them as unfractionated low partial melting products of a metaluminous granitic source. The San Martí metavolcanics were submarine effusives: gray and white levels are distinguished in the field. The gray levels are alkali-feldspar rhyolites having the same chemical composition than the Guilleries orthogneiss of which they are the volcanic equivalents. The white levels are adakitic, being interpreted as melting products of crustal deep-seated garnet amphibolites. The amphibolites are cordierite- and orthoamphibole-bearing, indicating a pre-Variscan hydrothermal alteration of basaltic protoliths. Their immobile elements point to tholeiitic intraplate basalts as their protoliths.

During the Ordovician, the study area was the site of an important diversified magmatism because several fertile sources were activated at the same time: granitoid, metasediment, amphibolite and upper mantle. The intense Ordovician magmatism associated with a wide rifting could have been generated by a massive asthenospheric rise initiated by a mantle avalanche caused by a Panafrican ocean closure. This avalanche embraced the whole Cambrian Period, reached the mantle-core discontinuity and triggered a deep mantle upwelling that was the responsible for the widespread and intense magmatism.

Ordovician magmatism in the Pyrenees

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Successive Ordovician magmatic pulsations are well documented in the pre-Variscan basement of the Pyrenees. According to radiometric data, magmatism lasted about 30 m.y., from ca. 477 to 446 Ma, and although the magmatic activity seems to be continuous, two peaks can be distinguished at 473-472 Ma and 457 Ma. Based on geochronological and geochemical data, two different magmatic complexes can be distinguished: latest Early-Mid Ordovician and Late Ordovician magmatism.

(a) During Early to Mid Ordovician times, the magmatic activity gave rise to the intrusion of voluminous aluminous granites, about 500 to 3000 m in size, that constitute the protoliths of the large laccolithe-shaped, orthogneissic bodies that crop out at the core of the Aston, Hospitalet, Canigó, Roc de Frausa and Albera massifs that punctuate the backbone of the Pyrenees. Only a minor representation of basic coeval magmatic rocks is exposed and acidic volcanic equivalents have been only reported in the Albera massif. Granites are medium to coarse grained and exhibit porphyritic textures with rapakivi K-feldspars, they are peraluminous and subalkaline and the geochemical characteristics indicate that these rocks were mainly derived from a continental crustal source.

(b) A Late Ordovician magmatic pulse yielded a varied suite of magmatic rocks. Small granitic bodies are emplaced in the lowermost part of the successions of the Canigó massif and constitute the protoliths of the Cadí, Casemí and Núria gneisses. Moreover, metre-scale thick bodies of metadiorite are interlayered in the same area. Coeval calc-alkaline and explosive volcanic rocks (ignimbrites, andesites and volcaniclastic rocks) are interbedded in the Upper Ordovician of the Ribes de Freser and Bruguera units, together with a granophyre body that crops out at the base of the Upper Ordovician succession. Metadiorites are metaluminous with slightly negative εNd values (-0.8) and a TDM age of 1.18 Ga. Their protoliths were derived from mantle melts with heterogeneous crustal contamination. Acidic volcanic rocks are peraluminous and subalkaline and εNd values between -5.1 to -4.8 indicate a crustal origin. Similar isotopic values have been obtained for the Ribes granophyre (εNd -2.6) and similar geochemical characteristics and εNd negative values (-3.2 to -5.2), from the Cadí and Núria gneiss, indicate that this assemblage was also derived from different magmas of continental crustal source. Finally, the εNd values of the Casemí gneiss (-1.9 to -1.3) suggests that their protoliths were derived from mantle melts with heterogeneous crustal contamination.

It should be noted that the latest Early-Mid Ordovician magmatism is coincident with the episode of uplift and erosion that led to the formation of the Sardic unconformity. This uplift was followed by an extensional pulse that developed normal faults, directly affecting the onset of the basal unconformity and controlling deposition of the (post-Sardic) Upper Ordovician strata, which were coeval with the Late Ordovician magmatic activity.

Cryptic Early-to-Main rift-related Cambrian magmatism in the Central Iberian and Galiza-Trás-os-Montes zones (Iberian Massif) revealed by U-Pb dating of inherited zircon

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The Early Palaeozoic rift-related igneous activity recorded in the Iberian Massif is subdivisable into three main events separated by relatively quiescent episodes. The so-called Early, Main and late events represent space-temporal diachronic pulses characterized by a progressive variation in the nature of the rift-related magmatism. The first two, Early and Main rift-magmatic events, were defined in detail by Sánchez García *et al*. (2003) and are almost exclusive of the Ossa-Morena Zone (OMZ), while the late magmatic event mainly occurs in the Central Iberian (CIZ) and Galiza-Trás-os-Montes (GTMZ) zones (Ollo de Sapo Formation and associated plutons).

The xenocrystic/inherited zircon component recorded in Early and Main felsic igneous rocks, mostly represented in the OMZ, is scarce in most samples studied until now. The scarcity of inherited cores or older zircons in Cambrian orthogneisses was explained by melting from zircon-undersaturated source rocks which led to the production of zircon-undersaturated magmas (Chichorro *et al.,* 2008). This hypothesis implies a limited participation of sedimentary protoliths at least in some sectors of the OMZ. However, this apparently contradicts the high grade metamorphic rocks and the related products of crustal melting, apparently derived from sedimentary rock protoliths (Ediacaran Série Negra), located in the central OMZ (Monesterio antiform) and interpreted as extensional migmatitic domes developed during the Cambrian rifting (e.g., Expósito *et al*., 2003; Sánchez-García *et al*., 2003). However, the evidence provided by these upper-crustal level domes does not preclude the participation of deeper mafic zircon-undersaturated source rocks (representative of lower crustal sections) on other Cambrian magmas. Melting at the highest grade of metamorphism can also explain the zircon inheritance scarcity. Recent studies in Évora massif do not allow us to exclude this hypothesis. Nevertheless, this scenario implies the melting of a substantial portion of crustal material.

This fact contrasts with the much higher zircon inheritance found in a Late rift-related Cambrian-Lower Ordovician igneous units (ca. 490-460 Ma) exposed in the CIZ and in GTMZ (Bea *et al*., 2007). If the zircon inheritance reveals the history of older zircon-growth events, it is surprising the significant contribution of the Early and Main Cambrian inherited igneous zircons: 17% and 16% respectively. Those zircons match the ages of the first two events recorded in OMZ, which are not apparently represented in the CIZ and GTMZ. So, what is the source of those zircons? A possible scenario would consider the contemporaneous Furongian-Lower Ordovician detrital formations from the CIZ and GTMZ (themselves containing detrital zircon of those ages) involved in melting as source layer for those anatectic Late-rift stage magmas in higher upper crustal levels. However, high-grade melting at higher crustal levels, even admitting wet melting conditions, seems unlikely due to the large volume of magmas. Alternatively, a probable existence of Early and Main Rift magmas entrapped in the lower crust and crust-mantle transition can be admitted. Thus, the syn-rift Terreneuvian-Cambrian Series 3 zircons, incorporated in the Late-stage Rift Magmas ascending in the crust, are an undirected evidence of this Early Magmatism also in the CIZ. This scenario of underplating magmas is compatible with the onset of diachronous progressive stretching and thinning of the continental lithosphere proposed by Álvaro *et al*. (2014a).

The Cantabro-Iberian Basin: Insights of the rift-stage revealed by detrital zircon age distributions

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The purpose of this study is to compare the features of the Terreneuvian-Cambrian Series 3 siliciclastic rocks representative of rift stages in the Ossa-Morena (OMZ), Central Iberian (CIZ), West-Asturian Leonese (WALZ) and Cantabrian (CZ) zones of the Iberian Massif (the three latter representing the so-called Cantabro-Iberian Basin), based on zircon age distributions.

Taking the OMZ as standard reference, the detrital zircon ages distribution shows a remarkable similarity with the underlying Ediacaran Série Negra Succession). This means that over at least 40-45 m.y., the erosive and depositional system did not record substantial changes, with no significant palaeogeographic variations (i.e. a peri-West African craton palaeoposition persisted during the Cambrian as advocated for Neoproterozoic times; Fernández-Suárez *et al*., 2002). The Cambrian clastic rocks predominantly derived, both directly and by recycling, from sources dominated by Cadomian/Pan-African tectonothermal events. This implies that the Série Negra was partly emerged after the Cadomian collision and magmatic arc activity (Pereira and Quesada, 2006). In addition to the Série Negra sediments, an evident potential source for the detrital zircons is represented by Neoproterozoic igneous rocks, including some remnants of Cadomian magmatic arc, exposed mainly in northern and northwestern sectors of the OMZ (Eguiluz *et al*., 2000; Henriques *et al*., 2015).

Although in the OMZ, the Cambrian rift-related igneous rocks are volumetrically significant (Sánchez-Garcia *et al*., 2003), the scarcity/absence of Cambrian igneous zircons in the Cambrian sediments suggests that the magmatism was essentially submarine and associated with hypabyssal intrusive complexes. The dataset agrees with a rift system partitioned into fault-bounded half-grabens (Oliveira *et al*., 1992), some close to volcanic centres while the majority had no significant Cambrian volcanic contributions.

The same features exist in the coeval rift-stage detrital rocks of the CIZ, WALZ and CZ. This means that the main central magmatic complexes prevalent at the time in the OMZ are not exposed, neither supplying OMZ- or CIZ-related basins.

The zircon age distributios from the CIZ and OMZ are similar, although in the sandstones of the CIZ, there is a slight increase in the percentage of Stenian-Tonian zircons (Martínez Catalán *et al.* 2004) demonstrating that there were no significant differences in the paleogeography of both zones. Based on the Neoproterozoic curves and considering the scarcity to low representativeness of Stenian-Tonian zircons, the sources basically remained the same. However, it should be noted that in the WALZ and CZ, the age distribution clearly reveals a closer influence of Eburnian and Liberian sources, probably related to their proximity to the West African craton (Reguibat Shield and Western and Eastern Morrocan Meseta). In conclusion, the OMZ, CIZ and WALZ-CZ sectors of the Gondwanan margin were paleogeographically close during the syn-rift stage, although the WALZ-CZ depocenters appear to be in a more internal position iofn the WAC realm than the OMZ and CIZ.

The evolution of the rift-drift transition in the Central Iberian and Galicia-Trás-os-Montes zones, Iberian Massif

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In the Iberian Massif a voluminous igneous activity was recorded along its reputed rifting axial (i.e., the Ossa-Morena Zone, OMZ) where three phases, separated by relatively quiescent periods, are distinguished. The first two, Early and Main rift-stage magmatic events, were defined in detail by Sánchez García *et al*. (2003) and are almost exclusive of the OMZ, while the Late rift-stage magmatic event mainly occurs in the Central Iberian (CIZ) and Galicia-Trás-os-Montes (GTMZ) zones (Ollo de Sapo Formation and associated plutons).

The age distribution of detrital zircon from the Furongian to Lower Ordovician arkoses and pelites (overlain by the Armorican Quartzite) that characterize the rift-drift transition in the CIZ and GTMZ reveals a substantial difference by comparison with Terreneuvian-Cambrian Series 3 samples (Chichorro *et al*., this vol.).

Despite the Late rift-stage magmatic event being the most representative in the CIZ and GTMZ, the CIZ and GTMZ Furongian to Lower Ordovician sediments show a relatively high percentage (~18%) of Cambrian zircons with the age spectra covering all the Early, Main and Late rift-stages magmatic events. This fact suggests that the main source of Early to middle Cambrian zircons was probably derived from the OMZ Terreneuvian-Cambrian Epoch 3 rift-stage igneous rocks and their host sedimentary successions, including also the Cadomian basement (Série Negra). This scenario is also compatible with the prevalent Ediacaran peaks, the relative low representativeness of Stenian-Tonian zircons and also the residual Eburnian-Liberian events displayed by the distribution of detrital zircon ages from the OMZ, typical of a West African Craton provenance (Pereira *et al*., 2012 and references therein).

However, the most prominent characteristic is the simultaneity between the sedimentary processes of this interval and the Late Rift-Stage igneous processes. The presence of Furongian–Early Ordovician detrital zircons in coeval sediments is the evidence of partial erosion of the Late rift-related igneous rocks. Significantly, both the high percentage of those zircons and its remarkable decrease in the overlying Armorican Quartzite (Linnemann *et al*., 2008; Pereira *et al*., 2012; Shaw *et al*., 2014) suggest that these grains resulted from erosion and incorporation of a syn-volcanic component to the sediments during the rift-drift transition. Therefore, these units may be considered essentially epiclastites, sometimes pointing for a strong proximal character.

Upper Ordovician magmatism in the Central Pyrenees: First U-Pb zircon age from the Pallaresa Massif

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The Late Ordovician magmatism is well represented in all massifs of the Eastern Pyrenees and the Catalan Coastal Ranges, which contain Upper Ordovician rocks. Nevertheless, data about this magmatic episode in the Central Pyrenees are scarce. This work shows the first geochronological data for Upper Ordovician magmatism in the Pallaresa massif. This massif is a large E-W trend antiformal structure included in the metamorphic structural units of the Pyrenean Axial Zone, where Cambro-Ordovician and Upper Ordovician rocks crop out. The Cambro-Ordovician rocks consist of a low-grade monotonous alternation of quartzites and slates with some limestone and conglomerate intercalations, which show evidence of a pre-Variscan penetrative deformation. The Upper Ordovician succession lies unconformably on the older Cambro-Ordovician beds and is represented by a siliciclastic succession with an intermediate limestone level. Volcanic rock levels have been observed interbedded within what was considered a Cambro-Ordovician succession close to outcrops of Upper Ordovician rocks in the eastern part of Pallaresa massif. Both metasedimentary and volcanic rocks were deformed and metamorphosed during the Variscan Orogeny.The volcanic rocks are rhyodacitic to dacitic crystal-rich meta-tuffs with a pervasive foliation. Their mineralogical composition is mainly volcanic quartz, feldspar and biotite. The matrix consists of fine grained of biotite, muscovite, quartz and clinozoisite, the latter mainly developed in highly deformed bands. In addition to these mineral phases, idiomorphic crystals of zircon are recognized.

Zircon from one these volcanic levels were dated by U-Pb CA-ID-TIMS, and provide an age of 453.6 ± 1.5 Ma (Sandbian). The occurrence of Late Ordovician magmatism in the Central Pyrenees may be related to the angular unconformity between Upper Ordovician rocks and the underlying beds, as well as to the pre-Variscan deformation affecting the Cambro-Ordovician sequence. This new data, coupled with the occurrence of Zn-Pb deposits of SEDEX o Mississippi Valley type in this sector of the Axial Zone, support the evidence of extensional tectonic activity during Late Ordovician times.

Cambrian-Lower Ordovician sedimentary provenance shifts in Northwest Gondwana

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Detrital zircon from Cambrian-Lower Ordovician sediments of Northwest Gondwana is studied herein to assess the influence, across space and time, of different craton sources. Age distribution curves from new data obtained through LA-ICPMS U-Pb zircon dating from Cambrian sandstones sampled in the Pyrenees is compared to other available data from Morocco, the Iberian Peninsula, South France and Sardinia. Kolmogorov–Smirnov (K–S) test and crystallization age (CA) - depositional age (CA-DA) diagrams are applied herein to compare the zircon source populations and assess their possible correlation with the arc/rift geodynamic evolution recorded throughout this margin of Gondwana. During the Terreneuvian, zircon population allows distinction of (i) a southwestermost edge (Atlas and Ossa-Morena Rifts) mostly influenced by Panafrican and Atlasian sources (ca. 0.63-0.54), (ii) a northeasternmost edge (Sardinia) recording the influence of the Saharan Metacraton and the Arabian Nubian Shield, with an important Stenian-Tonian cluster (ca 1.2-0.9 Ga), and (iii) an intermediate palaeogeographic transect, where lies the Central-Iberian Zone, the Montagne Noire and the Pyrenees, sharing parent populations and a progressive influence of both sources. According to this gradual modification of zircon population percentages, the Cambrian Pyrenean Basin should be located between the Montagne Noire (Occitan Domain) and Sardinia. This trend of zircon compositions gradually disappeared from Cambrian Epoch 2 to Early Ordovician times, reflecting a distinct geodynamic evolution in Northwest Gondwana. The Atlas and Ossa-Morena Rifts show a rapid post-Panafrican/Cadomian shift to extensional conditions, with an arc/rift turnover across the Ediacaran-Cambrian transition. In this context, sediment supply of recently built Panafrican, Cadomian and Atlasian orogenic sources influenced the zircon age distribution curves of Terreneuvian deposits. During later Cambrian and Ordovician times, the relative influence of different cratons tended to balance, leading to a more spread age distribution curve, characteristic of extensional settings evolving from rift to drift (passive margin) conditions.

The Sardic Phase in SE Sardinia (Italy)

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The Paleozoic basement of SE Sardinia consists of four tectonic units stacked and metamorphosed under greenschist facies during the Variscan Orogeny. The stratigraphic succession in the nappes differs slightly from each other but all are characterised by a Middle Ordovician calc-alkaline volcanic complex that lies unconformable above a Cambrian-Lower Ordovician siliciclastic succession with embedded Furongian intermediate and felsic volcanic rocks. The angular unconformity, sometimes marked by the occurrence of discontinuous conglomerates, is more evident in the shallow unit (i.e. Sarrabus Unit) where the Variscan deformation is less penetrative. The volcanosedimentary succession sealing the angular unconformity is the main evidence of a plate margin active during the Ordovician. This Early Palaeozoic tectonic event is referred to the Sardic Phase and supposedly related to the emplacement of a volcanic arc on continental crust along the Northern Gondwana margin.

Geological mapping and structural analysis performed in the Sarrabus Unit show that the angular unconformity is due to a folding event affecting only the pre-Middle Ordovician succession. The Sardic folds have overturned limbs and lack axial planar foliation, demonstrating that their formation took place at shallow crustal structural levels. Furthermore, no older crust is involved in the Sardic deformation, indicating a thin-skinned tectonics.

These features do not fit completely with an Andean model because non-collisional orogens are tipically affected by strong uplift with exhumation of the lower crust and thick-skinned tectonics. Moreover, the volcanic arc should rest on the crystalline basement, while it lies on folded sediments in Southestern Sardinia. The lack of evidence of a back-arc basin rules out the Japanese model too.

These remarks led us to consider other geodynamic scenarios consistent with field observations from SE Sardinia. Overriding plates lacking large thrust sheets, strong uplift, lower crust deformation and exhumation could be representative of an oblique plate convergence, an erosive subduction zone or an Alaskan-type subduction-accretion orogeny. This last is the

model proposed to explain the evolution of other sectors of the Gondwana margin like the Lower Paleozoic basement of the western Alps and the Lachlan Fold Belt in Southeastern Australia.
The features recognized in the Sarrabus Unit do not allow so far to favour for one or the other model or to propose a new one, but we highlight that the Sardic shortening event did not occur in an Andean-like geodynamic setting. This should be taken into account in reconstructing the Early Palaeozoic plate-tectonic configuration also considering that other areas of the Variscan basement, mainly in the Eastern Pyrenees and Centro-Iberian Zone, show characteristics similar to those of SE Sardinia. This means that these sectors of the Variscan basement were affected by the same geodynamic processes and, albeit not necessarily, could be adjacent during Ordovician time. Other investigations on the geochemical signature of the magmatic products or the sedimentary sources areas of the succession involved in the Sardic Phase can support the field evidences in order to properly locate in Early Palaeozoic paleogeographic reconstructions the terranes now incorporated in the Variscan basement.

Ordovician tectonics in Sardinia (Italy)

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The segment of the Variscan chain exposed in Sardinia consists of three main structural domains stacked (from NE to SW) from a medium to high grade metamorphic inner zone to a green-schist facies nappe zone divided into external and internal, to a anchimetamorphic thrust-and-fold belt foreland.

Each Variscan domain is composed of different Palaeozoic successions, which share only few tracts, were assembled during the Variscan Orogeny and show clear evidences of tectonic instability during Ordovician times. In the foreland zone, folds only affecting the Cambrian to Lower Ordovician succession cause an angular unconformity sealed by a thick syn-tectonic succession mainly made of Upper Ordovician conglomerates that lies on both the Cambrian and Lower Ordovician rocks, indicating a shortening event followed by an important erosional phase. In the external nappe zone, folds in the pre-Middle Ordovician succession give rise to an angular unconformity sealed by Middle Ordovician volcanosedimentary complexes, and related intrusive rocks. In the internal nappe zone, Ordovician volcanic products became rare. In the inner zone, Ordovician magmatic rocks consist of calcalkaline orthogneisses, derived from granodioritic protoliths.

Ordovician tectonics in Sardinia is referred to the Sardic phase, as in other parts of the Variscan belt where similar and roughly coeval stratigraphic, tectonic and magmatic features have been found.

Several studies attempted to reconstruct the Ordovician paleogeography, arranging in the right position the Variscan crustal blocks based on their Ordovician features. In these paleogeographic reconstructions, Sardinia is often considered as a whole single crustal block located above a subduction zone where a volcanic arc developed. Some reconstructions interpret the foreland, the external nappes and the internal nappes as back-arc, arc and forearc, respectively. We want to point out that, being these structural domains amalgamated in Variscan time, it is a forcing to consider their provenance from zones adjacent during Ordovician times. Actually, there are several evidences supporting that the Palaeozoic blocks forming the Sardinian basement could be distant from each other before the Variscan Orogeny, including major differences in the sedimentary source, stratigraphic succession, bio-province of faunas and magmatism fingerprints between the foreland, nappe and inner zones. The pre-

Middle Ordovician succession in the nappe zone is siliciclastic and includes Furongian volcanites, absent in the foreland, which conversely exhibits a thick carbonate sequence. The sources of the Upper Ordovician sedimentary rocks, based on detrital zircon tipology, are different between nappe zone and foreland, which lacks detrital zircons related to the Middle Ordovician volcanism. Furthermore, the Silurian-Devonian succession in the nappe zone is characterized by a carbonate shelf succession that does not occur in the foreland. Also the Early Palaeozoic trilobite association found in the foreland belongs to a different bio-province then those collected in the nappe zone.

Taking into account all these features and using them to correlate each of the Sardinian blocks with the now-scattered Variscan terranes, it is essential to constrain the paleogeography and the geodynamic settings in the Ordovician, but also to better draw the Wilson cycle that ended in the Variscan orogeny.

The onset of the Great Ordovician Biodiversification Event (GOBE) in high-latitude peri-Gondwana

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After revising the Cambro-Ordovician brachiopod occurrences from Iberia, the onset of the GOBE have been investigated. Three brachiopod diversity pulses have been detected in this peri-Gondwanan margin through the study interval: (i) an initial increase in linguliformean diversity culminating during the Tremadocian (Tr3); (ii) a second pulse during the Darriwilian (Dw2) probably caused by a rise in orthid species; and (iii) a third pulse during Katian times (Ka3-4) characterized by the most diverse and phylogenetically complex faunas recorded in this study. The first pulse consists almost exclusively of remnants of the so-called Cambrian Evolutionary Fauna, whereas the second one shows a sudden rise of clades typical of the Palaeozoic Evolutionary Fauna. The fact that the otherwise typical Cambrian clades also radiated at the dawn of the Ordovician suggests that the triggers of the GOBE also benefitted somehow these ancient lineages. However, eventually the succeeding Paleozoic Evolutionary Fauna became dominant during the second diversity pulse. This pulse represents a high-latitude correlation to the main pulse of the Ordovician Radiation and may be correlated with the diversity increase observed in the lower-middle Darriwilian (*L. variabilis*-*Y. crassus* Zones) of Baltica and in the *H. holodentata* Zone of Laurentia. Although it coincides with a globally occurring transgression of eustatic nature, our analysis shows that the generic composition of the brachiopod associations of Iberia remained mostly endemic up through the Darriwilian-Sandbian interval. The third pulse instead coincides with the global warming Boda event, which also marks a time of global dispersal favoured by the widespread development of carbonate productivity, the warming of oceanic waters and coeval global sea-level rise. In Iberia, this is reflected by almost entirely cosmopolitan brachiopod faunas, suggesting that the phases of rapid speciation characterizing the Ordovician Radiation had ceased by the mid Katian.

Furongian-Late Ordovician volcanism in the Upper Parautochthon of the Galicia-Trás-os-Montes Zone (NE Portugal): Paleogeographic meaning and geodynamic setting

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In NE Portugal, the Upper Parautochthon (UP) of the Galicia-Trás-os-Montes Zone (GTMZ) displays one of the best exposures of the Furongian-Late Ordovician magmatic record, with a voluminous and temporally continuous volcanism, which is interbedded in a sedimentary sequence equivalent to the nearby autochthon, the Central Iberian Zone (CIZ) (Dias da Silva *et al*., 2014, 2015, 2016). This unit is structurally below the Morais Complex (Pereira *et al*., 2006); a thin-skinned tectonic pile composed, from bottom to top, by three units: the Lower, Middle (ophiolitic) and Upper Allochthons (Ballèvre *et al*., 2014). The UP structurally overlies an intervening Variscan synorogenic unit (the Lower Parautochthon, LP), which separates the GTMZ from the CIZ (Fig. 1). The magmatic rocks of the UP can be grouped -according to their age and geochemistry- in three clusters (Dias da Silva *et al*., 2015, 2016). The oldest, named Mora Volcanics, is a bimodal volcanic suite with 493 Ma. Stratigraphically above, the rhyolitic domes of the Saldanha Volcanics share chemical and age characteristics (483 Ma) with the Ollo de Sapo gneisses in Spain. The youngest and more voluminous volcanic suite is the bimodal Peso Formation. It is overlying the Algoso Formation (Armorican Quartzite) and shows an alkalinity increase when comparing with the underlying magmatic rocks. Their ages span from 470-455 Ma. Although there are three spikes of magmatic activity in the UP, it came out rather continuously, accompanying the entire stratigraphic record. The three magmatic blazes could be related to regional tectonic crisis in North Gondwana during a continuous extensional process along the margin. They are broadly contemporaneous to the formation of two of the main unconformities in the European Palaeozoic basement (Casas, 2010; Gutiérrez-Marco *et al*., 1990; Helbing & Tiepolo, 2005): the Toledanic (Furongian-Lower Ordovician) and the Sardic (Middle/Upper Ordovician); both formed by tilting of independent tectonic blocks that exposed substantial areas to erosion. Comparing the stratigraphic and magmatic records of the UP with the structurally adjacent units -the CIZ and the Lower Allochthon (LA)- we interpret that the UP was located paleogeographically in an intermediate position between the more continental domains (CIZ) and the marginal realms of Gondwana (LA). The increasing alkalinity of the magmatic rocks towards the Late Ordovician suggests an evolution towards oceanic rifting settings. It shows the evolution of a passive margin during the opening of the Palaeozoic seas (e.g. Rheic, Middle Allochthon) and the drifting of the peri-Gondwanan magmatic arc (Upper Allochthon), later reassembled during the Variscan continental collision of Laurussia and Gondwana (Gómez Barreiro *et al*., 2007; Martínez Catalán *et al*., 2007).

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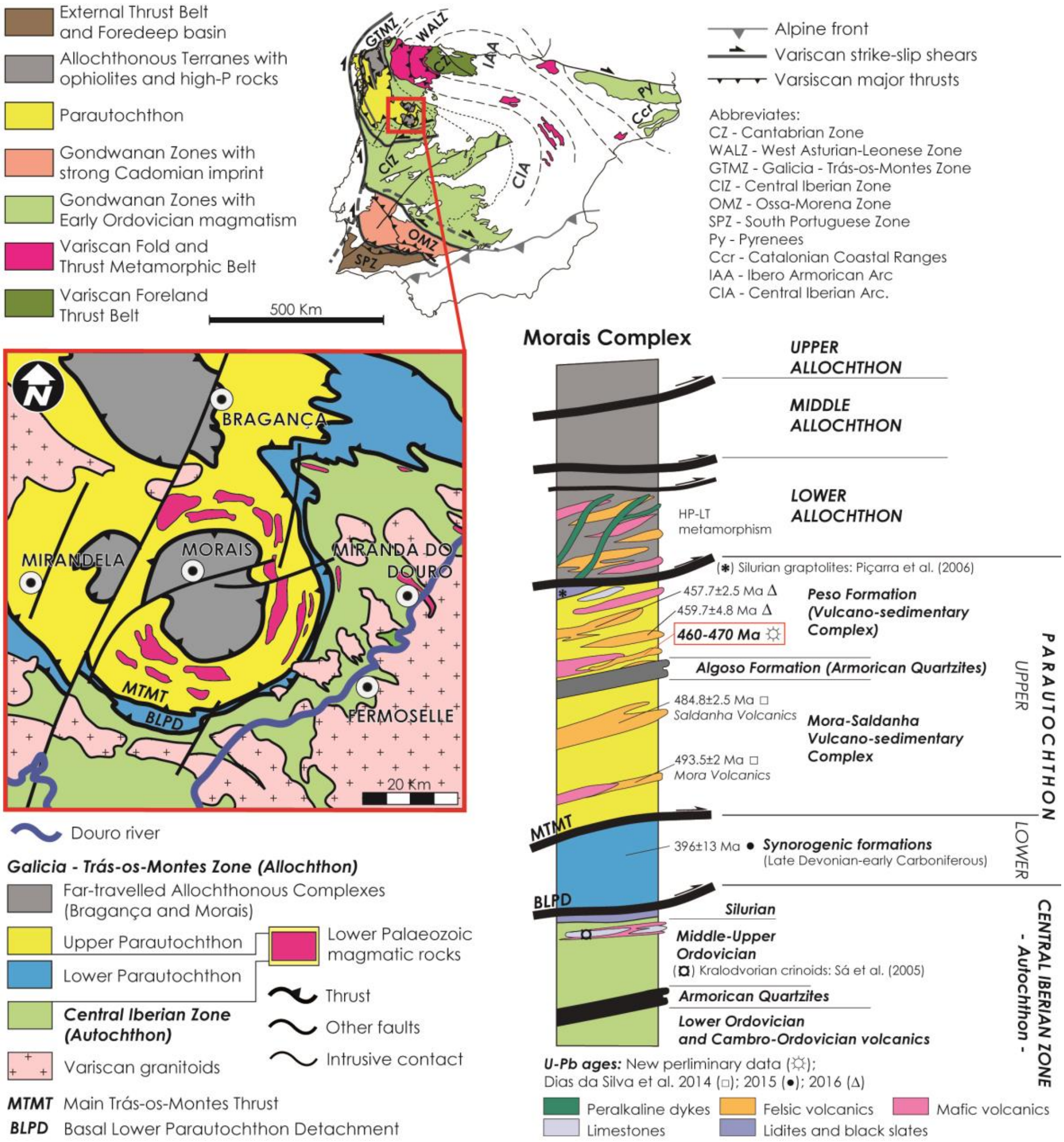

Figure 1. Location of the study area in NW Iberia and simplified logs for the Morais Complex.

Post-Cadomian paleogeographic evolution of Iberia

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The Iberian Massif represents a crustal section that records the convergence setting along the northern margin of Gondwana during the Ediacaran-early Cambrian times. During this interval, the Iberian Massif was an island-arc installed onto a thinned continental or oceanic crust, likely located close to the current Algeria region. The southwards (current coordinates) subduction was accompanied by strong left-lateral kinematics, whereas Amorican, Bohemian and several others Cadomian Massifs (Ibero-Armorican Belt) were located to the East (current coordinates).

Thus, we suggest that the Iberian Massif was originally rotated ~150º respect to the current position, and would be represented, from North to South, by (i) the Ossa-Morena Zone, which would correspond with an island-arc further from the Gondwana continental areas; (ii) to the south of this island-arc, a back-arc basin would be located, which would correspond with the southern areas of the Central-Iberian Zone (up to Valdelacasa anticline), and would contain turbiditic (Domo Group) and platform (Ibor Group) sequences, the latter with limestones included into olistostromic units; and (iii) the passive continental margin of Gondwana, which would correspond to the basement of the northern Iberian Massif areas. The continental margin of Gonwana would be separated from the back-arc basic by a narrow trough.

During early Cambrian times, the slowing down or stopping of the subduction process carried out the collapse of the island-arc (Ossa-Morena Zone). The pelitic materials generated by the island-arc and continental margin erosion filled the back-arc basin (Pusa and Villanueva Series), with the unconformable deposition of these sequences, which includes the Cambrian Lower Detrital, Carbonate and Upper Detrital series.

During mid Cambrian and Early Ordovician times, rifting conditions, which markedly affected the island-arc areas, carried out a widespread volcanism (final reactivation of the subduction?). This rifting event originated the opening of a basin that widened eastward and was filled with turbiditic series (Terena Formation). Other areas remained as a passive margin, implying that the Ordovician sequence was mainly controlled by the aforementioned troughs. The fore-arc, partially emerged or located at shallow marine conditions in the Furongian, accumulated Ordovician distal shales that fill both the Terena trough and the arc distal areas. Within both volcanic and back-arcs, largely emerged and subjected to erosion, sediments input to the south accumulated a shallow sequence rich in *Skolithos*, which changes to a clastic sequence that culminates with the 'Armorican Quartzite s.s.' restricted to this sector. In the continental passive margin an eminently clastic and thicker sequence, which has an important trough in the central zone, was deposited derived from the erosion of the African craton. Overlying the latter sequence, a pelitic level expanded, thicker along the trough axis. To the west (Iberia) the sequences are more diverse, complex and clastic in character. This scheme was modified in the end of the Silurian by the beginning of a new subduction episode with opposite directions to those previously existing.

An overview on the Ordovician volcanic rocks and unconformities in the Central Iberian and Galicia-Trás-os-Montes Zones of the Iberian Variscan Massif

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New detailed continuous digitalized maps (1:50.000 scale Geode program of the Spanish Geological Survey: IGME) facilitate an innovative regional approach to the Ordovician unconformities and volcanic rocks interbedded in the sequence (Clariana García et al., online; Villar Alonso et al., online). For this work, these maps have been actualized by up to date references from Spain and Portugal. At the selected areas of the Variscan Massif, two main Ordovician unconformities are generally described: (known ad the Toledanic and Sardic) but the regional approach unveils different stratigraphic positions at the diverse sectors. Besides, a new picture on the vulcanite abundance and variety arises. The general advances in geochronology and geochemistry of the interbedded volcanic bodies have been very useful to correlate the effusive events all over the vast area covered by this review (Coke *et al*., 2011; Dias da Silva *et al.*, 2014, 2015; Díez Montes, 2007; Talavera *et al*., 2013).

The Toledanic unconformity appears regionally mapped, but its position fluctuates through the different areas, changing from the base of the Cambro-Ordovician age acid volcanic rocks, to the base of the Tremadocian beds, or to the base of the Armorican type quartzite (Floian). This changing stratigraphic position may suggest the existence of several separate unconformities, but the cartographic compilation shows only one unconformity at each stratigraphic section. This event could be correlated to the "Ollo de Sapo" type vulcanites and associated plutonic rocks with ages spanning from Furongian to Floian (Cambro-Ordovician boundary).

Meanwhile, the Sardic unconformity, stated in some works and stratigraphic sequences (Martínez Poyatos *et al*., 2004; Sá, 2005), is not reflected on the maps but at some local exceptions. Furthermore, its stratigraphic position varies on the different proposals, but is always placed at the Middle-Upper Ordovician boundary or higher. According to the temporal range of the diverse schemes, the Sardic unconformity may be related to the bi-modal volcanic rocks interbedded around the Middle-Upper Ordovician boundary. The only mapped angular unconformity is located in NE Portugal, in the Central Iberian Zone, and placed at the base of the Kralodvorian limestones (Dias da Silva, 2014).

The Ordovician interbedded volcanites emerge as more widespread than previously considered and their chemistry is assorted, suggesting separate processes. Comparison between areas and tectono-metamorphic units allows us to determine an interesting zonation of major significance for better understanding the Ordovician evolution of the Gondwanan margin. The combination of the unconformities span and the volcanic rocks characteristics backs the previously stated extensional continental margin and block tilting structure for this part of Gondwana (Martínez Catalán *et al*., 2007); complementarily, it enables a more precise knowledge of the foremost extensional events and their regional distribution. This improved evolutionary Ordovician palaeogeographical constraints may be valuable for the understanding of the North Gondwana break-up and their importance on the following Pangea amalgamation during the Variscan times.

This work constitutes an example of the usefulness of the Geode type maps in the conception of regional hypothesis and the development of the subsequent research lines. The research has been financed by the Spanish national projects: CGL2011-22728 and CGL2015-64341-P; by the IGME project 531-GEODE-CXG; by SYNTHESIS project DE-TAF-5798; and by FCT postdoctoral grant SFRH/BPD/99550/2014.

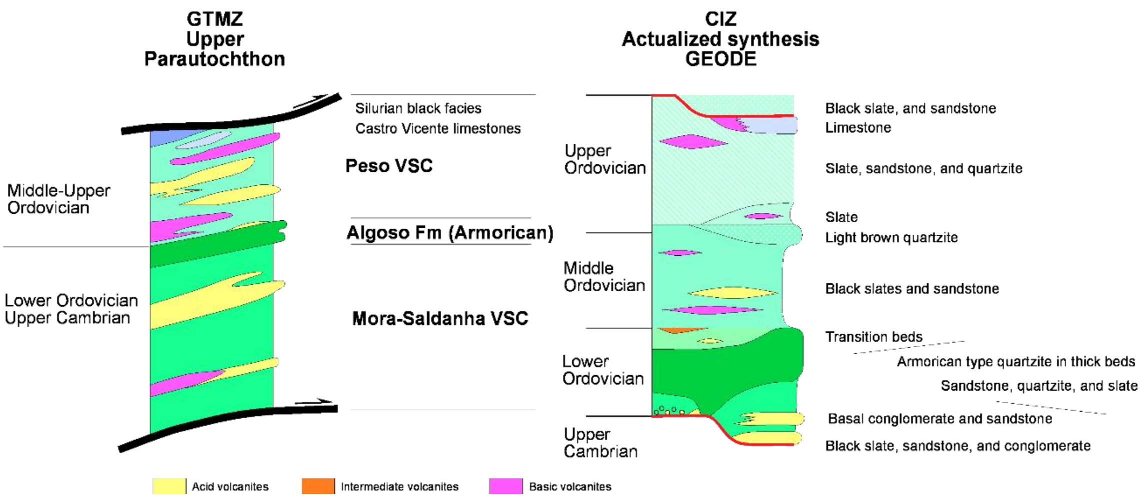

Figure 1. Sketch displaying the main Ordovician units at the central and NW Iberia.

Comparison between detrital zircon populations from the Ordovician rocks of the Pyrenees and other Perigondwanan terrains: palaeogeographic implications.

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The first LA-ICP-MS U-Pb detrital zircon ages from quartzites located below (three samples) and above (one sample) the Upper Ordovician unconformity of the Central Pyrenees (the Rabassa Dome, Andorra) are investigated. The maximum depositional age for the Jujols Group, below the unconformity, based on the youngest detrital zircon population, is around 475 Ma (Early Ordovician), whereas for the Bar Quartzite Formation, above the unconformity, the presence of only two zircons of 442 and 443 Ma precludes obtaining a precise maximum sedimentation age. A time gap of ~20 million years for the Upper Ordovician unconformity in the Pyrenees can be proposed, similar to that of the Sardic unconformity in Sardinia. The similar age patterns obtained on both sides of the Upper Ordovician unconformity suggest that there was no change in the source area of these series, while the absence of a Mid Ordovician age population may be due to a lack of sedimentation at that time. The four study samples present very similar U-Pb age patterns: the main age populations correspond to Neoproterozoic (Ediacarian-Cryogenian, ca. 550-750 Ma); Grenvillian (Tonian-Stenian, ca. 850-1100 Ma); Palaeoproterozoic (Orosirian, ca.1900-2100 Ma) and Neoarchean (ca. 2500-2650 Ma). The similarity with the Sardinian age distribution suggests that these two terranes could share the same source area and that they were paleogeographically close in Ordovician times in front of the Arabian-Nubian Shield.

New knowledge on the Upper Ordovician rocks of El Baell, Ribes de Freser area, eastern Pyrenees

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The Ribes de Freser area (Eastern Pyrenees), investigated by Robert (1980) and Muñoz (1985), is characterized by an Alpine antiformal stack, where several structural units are recognized. The stratigraphic succession exposed in these units and its thickness change considerably throughout the tectonic units. From bottom to top, three structural units can be distinguished: Ribes de Freser, El Baell, and Bruguera units. The aim of this work is to provide new structural and stratigraphic data about the rocks cropping out in Bruguera and El Baell units.

The Ribes de Freser unit is made up of a 200-600 m-thick succession composed of Upper Ordovician volcanic and volcanosedimentary rocks interbedded in Katian sediments (Muñoz, 1985; Martí *et al*., 1986), affected by two Variscan fold systems: a NW-SE and a E-W one.

The El Baell unit shows a fossiliferous carbonate succession referred to Upper Ordovician (Robert, 1980; Muñoz, 1985) that differs considerably from the classic one described by Hartevelt (1970), and used as reference for the main part of the Pyrenees. It lies on the Paleocene-Upper Cretaceous series of the Ribes de Freser unit, which crops out in a tectonic window, and it is topped by the Bruguera unit. The Upper Ordovician succession is made up of about 500 m of shales bearing centimetre-thick carbonate nodules (“*schistes troués*”), limestones, marlstones and siltstones, with three metre-thick levels of limestone rich in crinoids, brachiopods, echinoderms and conodonts, referred to a Katian age, and unconformably overlain by the Hirnantian Ansovell blackish shales. From bottom to top, the succession consists of siltstone/marlstone alternations, followed by the first limestone bar, which exhibits quartz veins and a synsedimentary breccia. Then, a thick pack of *schistes troués* and the second fossiliferous (echinoderm-rich) limestone bed are followed by *schistes troués*, siltstones and dark shales alternations, often intercalated with centimetre-thick carbonate levels, and marlstones. The last limestone bar shows fossiliferous and massive limestone, often recrystallized, and a *karst* on the top contact, which is capped by the Hirnantian black slates that contain scattered limestone dropstones, quartz veins and slumps.

Unfortunately, the base of the Upper Ordovician succession and the contact with Silurian beds is never exposed in this tectonic unit. The rocks of this unit are affected by a Variscan ENE-WSW-trending fold system, with tight folds, and N-plunging faults.

The Bruguera unit is composed of a slate-dominated succession of pre-Variscan age (Muñoz, 1985) overlain by rhyolitic ignimbrites and andesitic lavas, recently dated at ~455 Ma (Martí *et al*., 2014). Two Variscan fold systems, with E-W- and NW-SE oriented open folds and no penetrative foliation affect the Palaeozoic rocks of this unit (Muñoz, 1985). A Pre-Sardic fold system N-S oriented affects only the rocks of this unit.

On the Upper Ordovician of the La Cerdanya area, Pyrenees

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The aim of this work is to provide stratigraphic and structural features of the Upper Ordovician from the La Cerdanya area (Canigó unit), which exhibits a succession similar to that described by Hartevelt (1970) in the Central and Eastern Pyrenees. In the latter areas, the Upper Ordovician is represented mainly by a broad fining-upward package with some limestone key levels that ranges from 100 to 1000 m (Hartevelt, 1970).

The Upper Ordovician succession lies on the Cambrian and Lower Ordovician Jujols Group (Laumonier *et al*., 2004; Casas & Palacios, 2012) by the angular Sardic unconformity, which marks the base of the post-Sardic succession.

In the study area the lowest part of the Upper Ordovician succession is made up of 0-

100 m thick of reddish-purple polygenic and heterometric conglomerates (Rabassa Conglomerate Formation), with clasts composed of vein quartz, quartzite and slate derived from underlying rocks. Locally, the Cambro-Ordovician is directly overlain by the Cava Formation (0-850 m thick), made up of conglomerates, sandstones and shales with volcanic intercalations, with fossiliferous levels of Katian age (Hartevelt, 1970). This formation is overlain by a 5-200 m-thick limestone and marly limestone, the fossiliferous Estana Formation of late Katian age (Gil-Peña *et al*., 2004). The top of the carbonate succession is capped by the black-grey shales of the Ansovell Formation, 20-320 m thick, and the Bar Quartzite, 2-20 m thick, referred to Hirnantian-Silurian age.

Both successions are affected by two Variscan fold systems with a penetrative foliation, while the Cambrian-Lower Ordovician succession is affected by a pre-Variscan fold system that produced open folds without related foliation sealed by the conglomerates of the Rabassa Formation (Casas, 2010; Casas *et al*., 2012).

The pre-Sardic succession, the Sardic Unconformity and the lower part of the post-Sardic succession are cut and offset by some N-S-trending syn-sedimentary faults, which sharply affect the thickness of the Rabassa and Cava formations (Casas & Fernández, 2007; Casas, 2010). The displacement of the faults, which show a throws of about 0.2 to 0.9 km, progressively decreases upward and peters out in the upper part of the Cava Formation (Fig. 1). A synsedimentary hydrotermal activity is associated with these faults lined with quartz veins, which is present in the contact of Cambro-Ordovician sediments with synsedimentary faults (Figs. 3-4), and as their clast counterparts in the Rabassa conglomerate (Figs. 2-3).

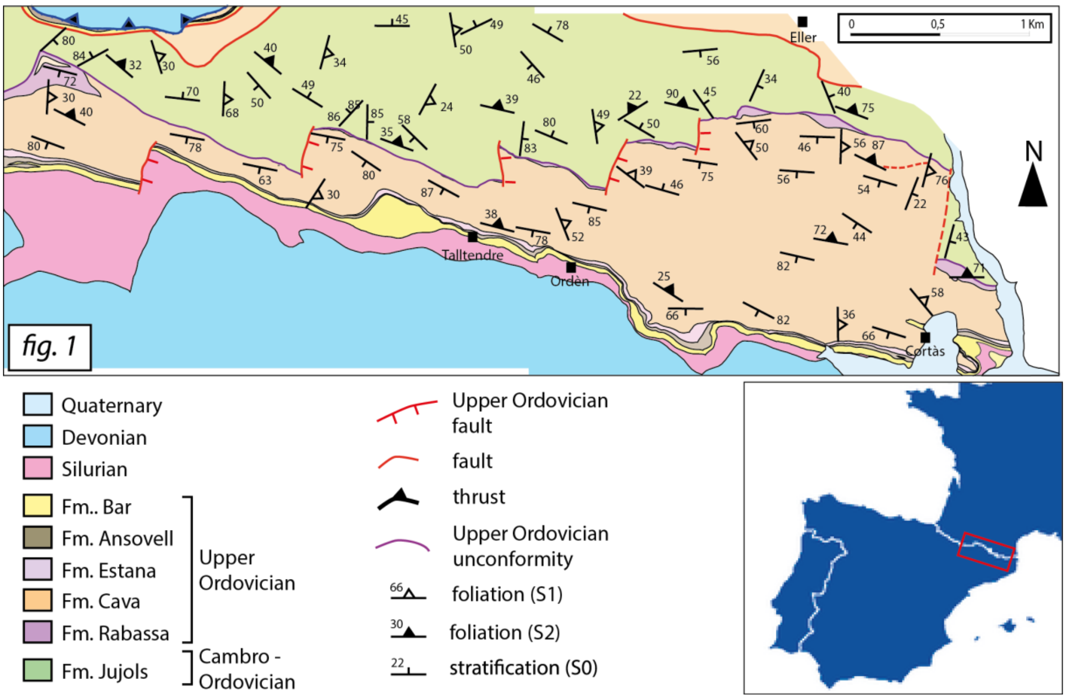
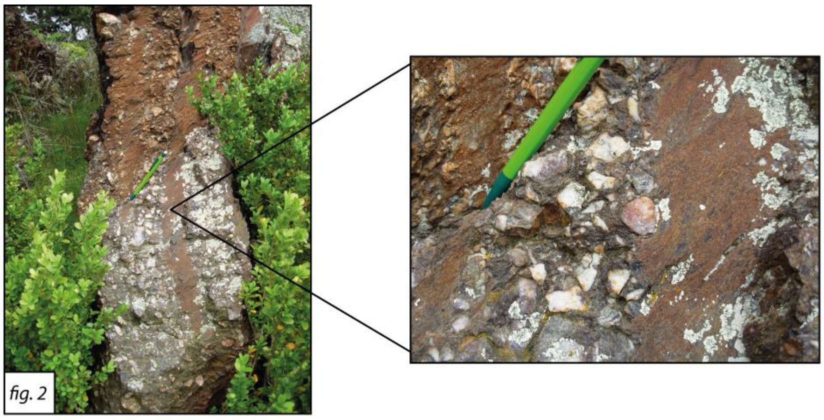
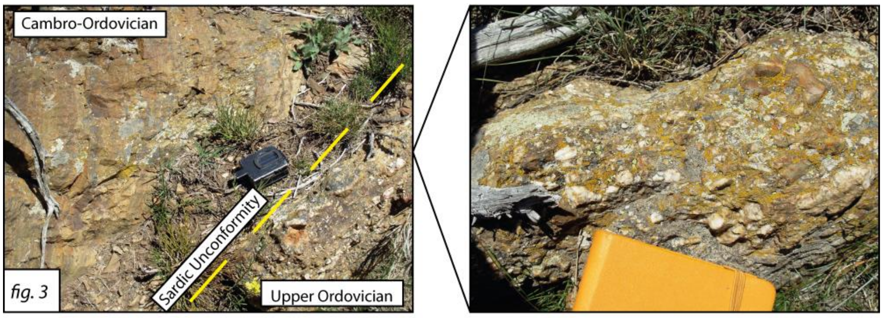


Figure 1. Geological map of the surroundings of Talltendre in the eastern Pyrenees.





Figures 2-4. Macroscopic aspect of the Jujols/Rabassa contact close to Talltendre, showing common incorporation of hydrothermal quartz clasts derived from synsedimentary-related faults fringed by hydrothermal dykes.

Early Palaeozoic evolution of Gondwanan units in the Iberian Massif: from subduction through rifting and drift on the southern passive margin of the Rheic Ocean

Cecilio Quesada

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Currently it is broadly accepted that the Iberian Massif consists of various tectonostratigraphic terranes, accreted to one another during the Variscan collision of Gondwana and Laurussia. These include: 1) units that occupied the northern margin of Gondwana prior to the Variscan orogeny (now exposed in the Cantabrian, West Asturian-Leonese, Central Iberian, lower allochthons of the Galicia-Tras os Montes, and Ossa-Morena zones, from inner to outer margin, respectively); 2) oceanic affinity units (ophiolites) exposed in the intermediate allochthons of the Galicia-Tras os Montes Zone and in the Pulo do Lobo Zone; and 3) exotic units thought to belong to Laurussia (upper allochthons of the Galicia-Tras os Montes Zone and the South Portuguese Zone). In this contribution focus is placed on the evolution recorded in the Iberian Massif's Gondwanan units during Cambrian through Ordovician times.

Breakup of the Neoproterozoic continent Rodinia started ca. 750 Ma ago and produced significant global plate reorganizations, among which the amalgamation of Gondwana that culminated near the Precambrian-Cambrian boundary is particularly relevant in our case. The future Iberian Massif was then a part of the northern active margin of Gondwana, under which subduction of the surrounding Mirovoi Ocean gave rise to the so-called Cadomian Arc and orogeny during the Ediacaran. Subduction until ca. 535 Ma (Lower Cambrian) is documented in the Ossa Morena Zone but it came to a sudden halt being replaced by extensional deformation at ca. 530 Ma. Oblique collision of a mid-ocean ridge with the outboard trench is held responsible for this abrupt change in the geodynamic scenario and for the progressive transformation of the previous active margin into a largely transcurrent one, of a kind similar to the San Andreas system in western N-America. Ridge-trench collision started earlier in the Moroccan Anti-Atlas (late Ediacaran) and propagated along the margin towards Ossa Morena (Early Cambrian). A combination of slab break-off and continued motion of the Gondwanan upper plate over the ridge-trench collision zone is envisaged as a likely scenario for generation of a slab window and subsequent propagation of rifting processes towards more inner parts of the Gondwanan margin, in a manner similar to the presently ongoing opening of the Gulf of California. Rift development started in Ossa Morena ca. 530 Ma ago, reaching the innermost West Asturian-Leonese and Cantabrian zones later in the Cambrian. By ca. 497 Ma (late Cambrian) a breakup unconformity marks the rift-drift transition in the Ossa Morena Zone, i.e. close to the rift axis. This is taken as evidence of opening of a brand-new oceanic tract (Rheic Ocean) and eventual drift-away of a ribbon of the previous Gondwana margin (Avalonia *sensu lato*). The break-up unconformity propagated inwards reaching the innermost Iberian Massif zones in Lower Ordovician times. It was succeeded by a rapid transgression that established relatively stable passive margin conditions throughout the Gondwanan continental shelf facing the Rheic Ocean, which prevailed until the onset of the Variscan collision in Devonian times.

In addition to structural and sedimentological evidence the Cambrian rift stage is characterized by coeval igneous activity. This took place in three main successive events, *Early* (ca. 530-525 Ma), *Main* (ca. 517-500 Ma) and *Late* (ca.490-470 Ma), with contrasting characteristics. Only in the Ossa Morena Zone the three events are expressed, whereas in the other zones the most important expression corresponds to the *Late* event. In fact, most of the igneous rocks of this *Late* event were emplaced after the breakup unconformity, suggesting persistence and propagation of the slab window beneath inner parts of the continent until well in the Ordovician. Assessing whether such propagation may have also been the cause of the younger Ordovician igneous events recorded in innermost parts of Gondwana (e.g. Pyrenees, Montagne Noire, Sardinia) or if they reflect far-effects of a brand new rift event (Paleotethys) still requires further research.

Furongian-Lower Ordovician metavolcanites from the Toledo Mountains (Spain) and its relationship with the Toledanian phase

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In the Toledo Mountains of the Central-Iberian Zone (CIZ), some scattered outcrops with volcanic and metavolcanic rocks occur interbedded across Cambro-Ordovician strata (Martínez Escorza, 1976; Ramírez-Merino *et al*., 2000). The authors differentiated two volcanic suites: (i) some volcanic rocks underlying the Cambrian Cortijos Sandstone Formation (VR1), and (ii) a volcanosedimentary complex (VSC) sandwiched between the Valdehierro and the Purple Series. Thirteen samples of both volcanic suites were selected for geochemical analysis. According to geochemical characteristics, VSC can be subdivided into two sets: (i) VSC1 displays similar geochemical characters than suite VR1 and is composed of andesite to subalcaline basaltic andesites in the Pearce (1996) diagram; whereas (ii) VSC2 exhibits a dacitic to rhyolitic composition (Fig. 1). Because of their stratigraphic position and age, VR1 and VSC are associated with the volcanic events of the Toledanian Phase, a widespread Furongian-Lower Ordovician tectono-magmatic episode related to generalize tilting, uplift and subaerial erosion in the Central-Iberian Zone.

The volcanism of the Toledo Mountains is compared with contemporaneous volcanic rocks of the CIZ and GTMZ, such as the Furongian Mora volcanic complex (Días da Silva *et al*., 2014) and the Carraxo, Covelo-Saldanha and Sanabria units (Díez-Montes *et al*., 2015), all belonging to the Lower-Ordovician Ollo de Sapo Formation. They are peraluminous, with minor differences: VR1 and VSC1 are similar to the Covelo-Saldanha and Sanabria complexes, whereas VSC2 presents more evolved features. REE patterns show more fractionated values for LREE than HREE and all groups display negative anomalies in Nb, Ti and Sr in the spider-diagram. In the Zr vs. TiO_2 tectonic discrimination diagram of Syme (1998), VSC2 plots in the extensional rhyolites field and the remaining data in the arc-association rhyolites field.

This contribution identified in the Toledo Mountains several groups of volcanites with similar geochemical characteristics supporting an extensive regime as previously stated for the area.

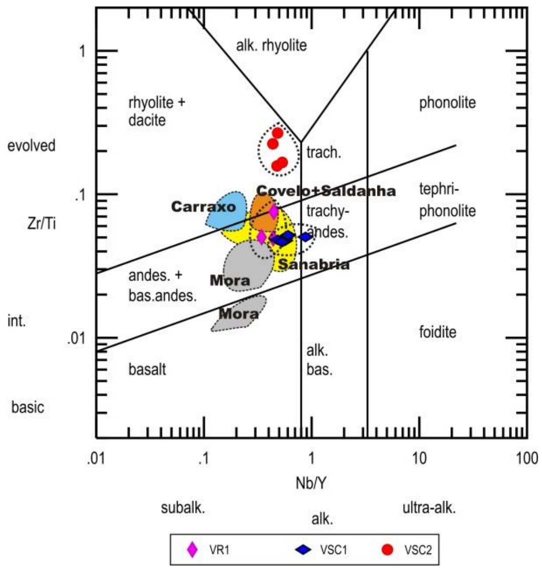

Figure 1. Geochemical features of the different groups in the Zr/Ti versus Nb/Y diagram (Pearce, 1996).

Provenance of Ordovician Armorican Quartzite in Iberia: the linkage with the West African Craton

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This study discusses the differences in detrital zircon populations of the Lower Ordovician Armorican quartzite. In the Iberian Massif, this proximal deltaic to shallow-water nearshore depositional facies (Gutiérrez-Marco *et al*., 2002) occur widespread in the Cantabrian (CZ,) West Asturian-Leonese (WALZ) and Central Iberian (CIZ) zones. The Armorican quartzite represents a passive margin sequence of the Gondwana margin facing the Rheic Ocean, related to transgression and strong subsidence. At this time, in the OMZ, the equivalent sedimentary units from Lower-Upper Ordovician are represented by more distal shales (Barrancos Formation).

There are two marked differences with major palaeogeographic implications between the Floian units of the CZ, WALZ and northern-central CIZ compared to those observed in southwestern CIZ (Luso -Alcudian southern subdivision, Lotze, 1945): (i) the former zones show a significant percentage of Stenian-Tonian detrital zircons, which are absent in the SW-CIZ; (ii) there is a significant discrepancy between these two domains in the percentage of (Rift-Stage) Cambrian zircons (30-10% in SW-CIZ and just 1.7% in northern-central CIZ).

The Cambrian zircons found in the Lower Ordovician quartzites from Southern CIZ fits the interval of Early and Main stages of magmatic activity in the OMZ (absent in the CIZ, WALZ and CZ) representing the onset of rifting in North Gondwana. This suggests that the main source of detritus was the OMZ), which represents the West-African margin. By contrast, the quartzites representative of the northern-central Iberian Massif were deposited far from the influence of peri-West African craton. According to Shaw *et al*. (2014), the continental margin of Gondwana represented by the CZ, WALZ and Central-Northern CIZ, Floian quartzite maintained a linkage with the Saharan Metacraton and Arabian-Nubian Shield. The differences between both groups demonstrate that the deposition of thick sequences of sandstones (Armorican Quartzite Formation) along the northern continental margin of Gondwana is the response to gradual source variations along the margin or, alternatively, that the differences underline a palaeogeographic limit separating two independent sectors inside the Luso-Alcudian Southern zone (Lotze, 1945). In fact, this sector is marked by the Late-Rift Stage Oledo-Zarza Lineament which may probably highlight an important Furongian detachment zone separating independent coastal areas. As Shaw *et al*. (2014) demonstrated, the depositional realm of the Armorican Quartzite in NW Iberia was linked to Central-East Africa but the data presented herein show that the SW-CIZ quartzites were undoubtedly linked to OMZ (Pereira *et al*., 2012).

Moreover, the difference between the U-Pb age distributions of detrital zircon populations from the Lower Ordovician Armorican quartzite and those found on Furongian-Lower Ordovician strata (absence *vs* presence of Furongian zircons, Chichorro *et al*., this vol.) argues against recycling from immediately underlying arkoses. This fact is in agreement with the assumed continuous subsidence and marine transgression (Paris *et al*., 1982) that led to the deposition of Armorican quartzite in passive margin settings.

Lower Ordovician unconformities and the Ollo de Sapo magmatic event: coeval 478 Ma pyroclastic base-surge deposits and subvolcanic granite intrusion in the Hiendelaencina Antiform

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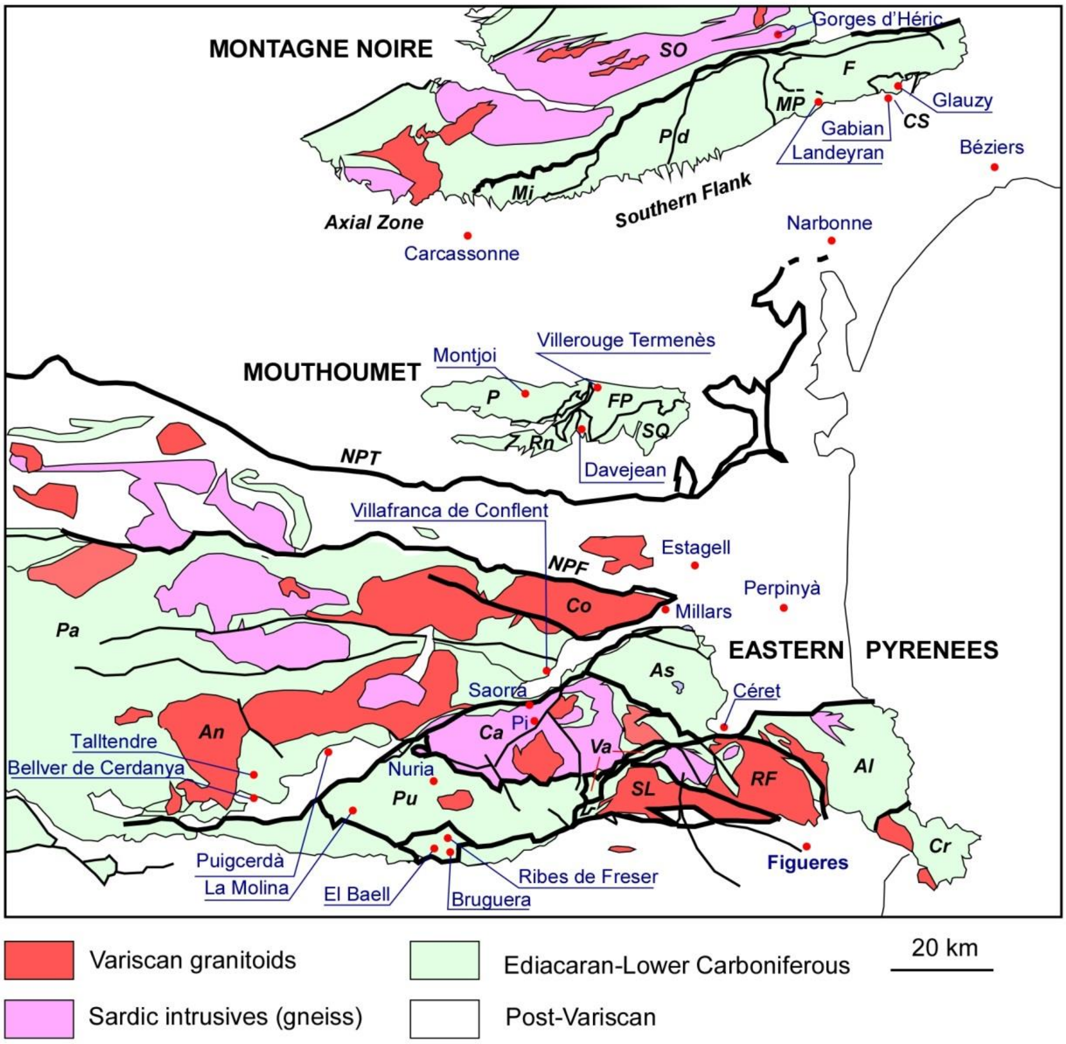
The Cambro-Ordovician Ollo de Sapo volcanic event in the Central Iberian Zone has different stratigraphic unconformities associated with it. In the Ollo de Sapo antiform of northwest Iberia and its continuation in the Hiendelaencina antiform of Central Spain, there is not clear evidence of an angular unconformity at its base. However, an erosive unconformity is mentioned in the literature separating the coarse Ollo de Sapo from the overlying Ordovician sequence in the Hiendelaencina Antiform. The Bornova conglomerate of Soers (1972) was interpreted as a product of the erosion of the underlaying felsic tuffs and ignimbrites of the Hiendelaencina gneiss of Schäfer (1969). Closer examination of this unit, along new clean road cuts near the village of La Bodera indicates that these “microconglomerates” are part of metric-scale pyroclastic base surge deposits. These base surge deposits are approximately 1 m thick with a felsic volcanic breccia at the base, a coarse unit with well-developed cross bedding and abundant rounded volcanic quartz and thin 1-2 cm thick siltstone at the top. The deposits are locally truncated by successive surges and intercalated with quartzite beds, suggesting that they might be the product of phreatomagmatic eruptions in a near-shore environment.

Zircon dating from one of these pyroclastic surges has provided a concordant age of 478±1.5 Ma (U-Pb ID-TIMS). In the Hiendelaencina antiform we have also dated the Bustares granite (Schäfer, 1969). This is a subvolcanic granite emplaced in the ignimbritic rocks (feinkörnige /grobkörnige gneiss) of the Hiendelaencina Series of Schäfer (1969). A concordant U-Pb ID-TIMS zircon age of 478±2 Ma obtained for this rock dates the intrusion and provides a minimum age for the coarse grain Ollo de Sapo in this area. The εHf isotopic values of zircon from the Bustares granite determined by LA-MC-ICPMS (UPV, Bilbao) have a range of between -3,5 to -15 with a bimodal distribution with a major peak at -5 and another peak at -10. These 478 Ma ages mark the final stages of the Ollo de Sapo volcanism in this area, suggesting that the Bustares granite might be a plug associated with the final phreatomagmatic eruptions. These data point to a paraconformity rather than an unconformity at the contact of the Ollo de Sapo with the overlying Ordovician rocks.

3. Guide to Field Excursion (6-9th September)



Al: Albères, An: Andorra, As: Aspres, Ca: Canigou/Canigó, Co: Conflent, Cr: Creus, CS: Cabrières ; F: Faugères, FP: Félines-Palairac, Mi: Minervois, MP: Mont Peyroux, NPT: North-Pyrenean Thrust, NPF: North-Pyrenean Fault, P: parauthocthon, Pd: Pardailhan, Pa: Pallaresa, Pu: Puigmal, RF: Roc de France/Frausa, Rn: Roc de nitable, SL: St.-Laurent-de-Cerdans, SO: Somail, SQ: Serr de Quintillan, Va: Vallespir

Figure 1. Geological sketch of the eastern Pyrenees and Mouhoumet and Montagne Noire massifs with setting of stops described below.

3.1 – 6th September – Fieldtrip to eastern Pyrenees, Catalonia

Upper Ordovician rocks of the Ribes de Freser antiformal stack

In the Ribes de Freser area of the southern Canigó massif, an Alpine antiformal stack can be recognized. In this antiformal stack, bounded to the north by the out-of-sequence Ribes-Camprodon thrust, three Alpine units (from bottom to top, the so-called Ribes de Freser, El Baell and Bruguera units) exhibit characteristic Upper Ordovician successions (Muñoz, 1985; Fig. 2) that are different from the classic one that can be recognized, with some lithologic variations, all across most part of the cordillera (Hartevelt, 1970). Preliminary restoration of the Alpine deformation allows us to situate the uppermost Bruguera unit in a pre-Alpine northernmost position, the intermediate El Baell unit in an intermediate setting, and the lowermost Ribes de Freser unit would lie originally in a southernmost position.

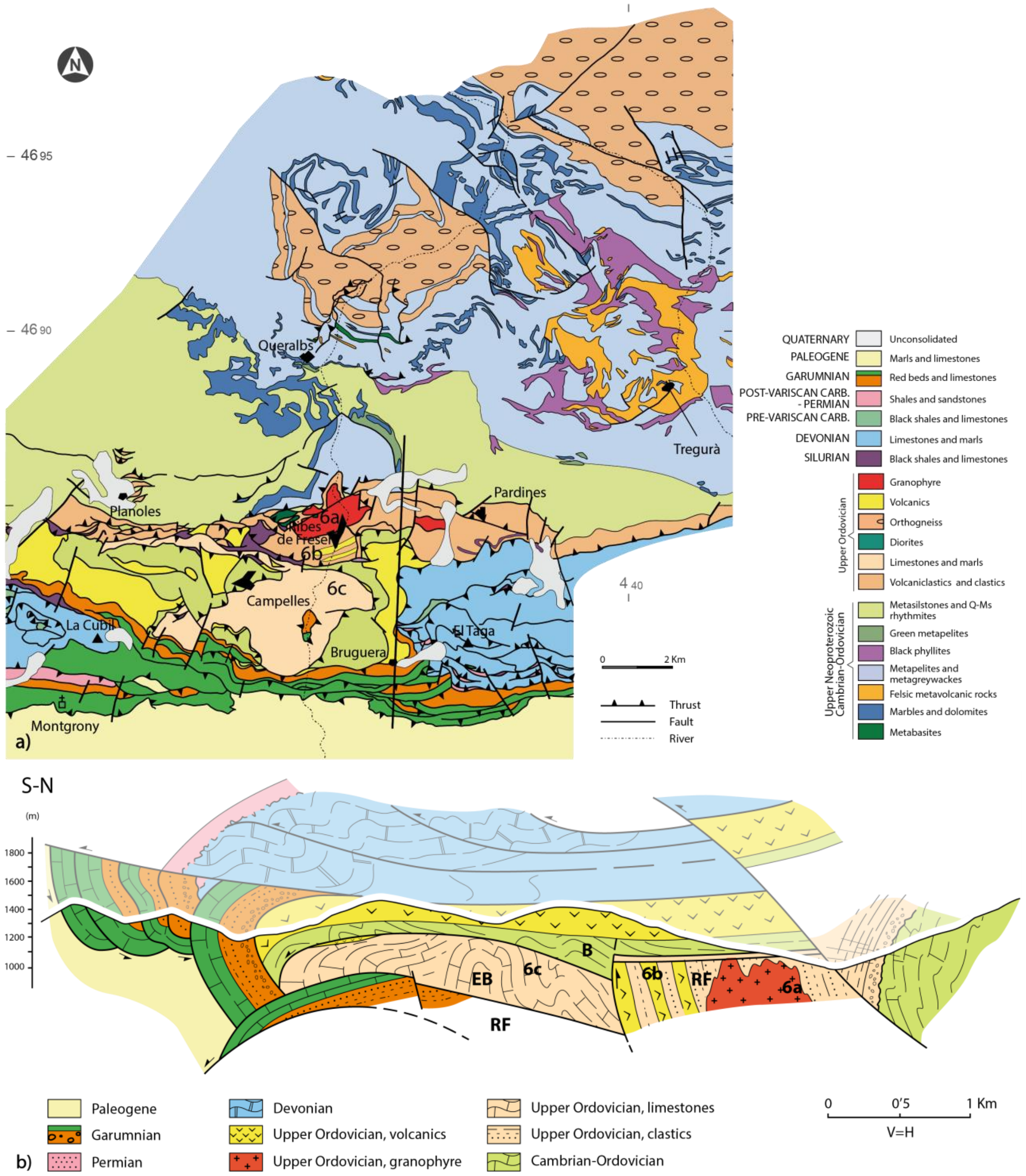


Figure 2. Geological map of the Ribes de Freser area (a) with location of the cross-section along the Freser river antiformal stack (b), with the Ribes de Freser (RF), El Baell (EB) and Bruguera (B) units; modified from Muñoz (1985) and Casas *et al*. (2015).

The Ribes de Freser unit is predominantly made up of volcanic and volcanosedimentary rocks (Robert & Thiebaut, 1976; Ayora, 1980; Robert, 1980; Muñoz, 1985, 1992; Martí *et al*., 1986; Muñoz & Casas, 1996), with a variable thickness ranging from 600 to 1200 m. Its lower part consists of diorite bodies and volcanosedimentary rocks, whereas rhyolitic lava flows and ignimbrites predominate in the central part, and ash levels, ignimbrites and volcaniclastic rocks in its upper part. A granophyre body is located into the lower part of the succession (Fig. 3). The volcanic activity was mainly explosive and displays a calc-alkaline affinity reflecting crustal melting (Martí *et al*., 1986).

The El Baell unit comprises a 300 m-thick succession entirely composed of limestones, marly limestones ("schistes troués") and shales (Robert, 1980; Muñoz, 1985). Three limestone-dominated thickening-upward parasequences, up to 30 m thick, can be distinguished. Conodonts and crinoids allowed Robert (1980) to attribute an early Katian (former Caradoc) age to the beds forming this unit.

The Bruguera unit lies on the top of the El Baell unit and is composed of a 200 m-thick slate-dominated succession, pre-Variscan (Cambrian-Ordovician?; Muñoz, 1985) in age, overlain by a volcanic complex (Robert, 1980). The latter consists of rhyolitic ignimbrites and andesitic lava flows. The age of this volcanic complex is controversial and based on a polen assemblage found in continental breccia located at the base of the volcanic complex, Robert (1980) attributed these rocks to a late Variscan (Carboniferous) volcanism. However, recent U-Pb data suggest a radiometric age at ca. 455 Ma (Martí *et al*., 2014) that is similar to that of the Ribes granophyre and volcanic rocks cropping out in the Ribes de Freser unit.

In the stops 6a, 6b and 6c, we will examine the Upper Ordovician rocks of the Ribes de Freser and El Baell units.

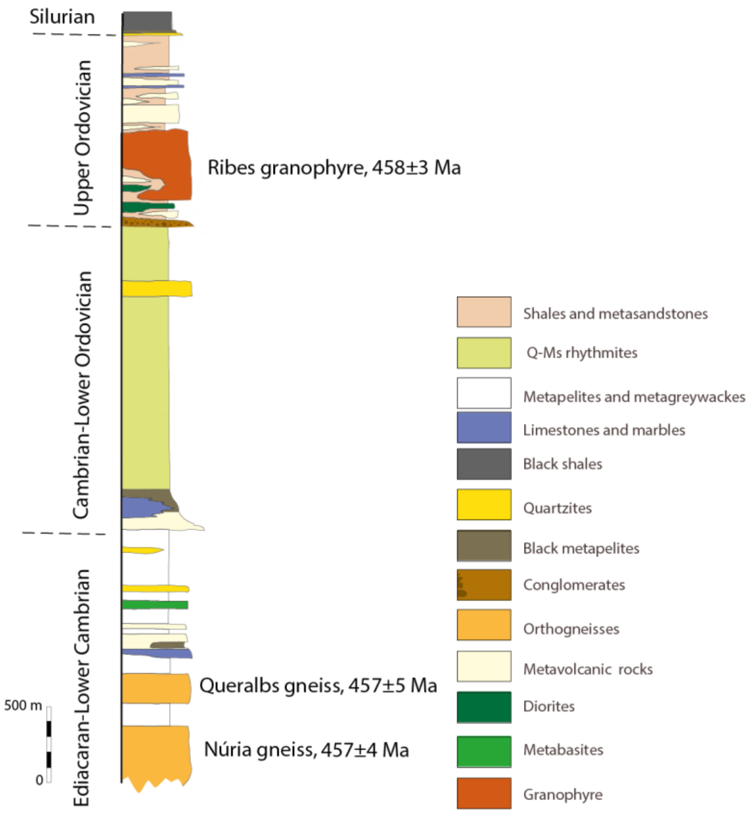


Figure 3. Synthetic stratigraphic log of the pre-Silurian rocks from the Ribes de Freser area. Geochronological data of the protoliths of the Ordovician gneisses and the Ribes granophyre after Martínez *et al*. (2011).

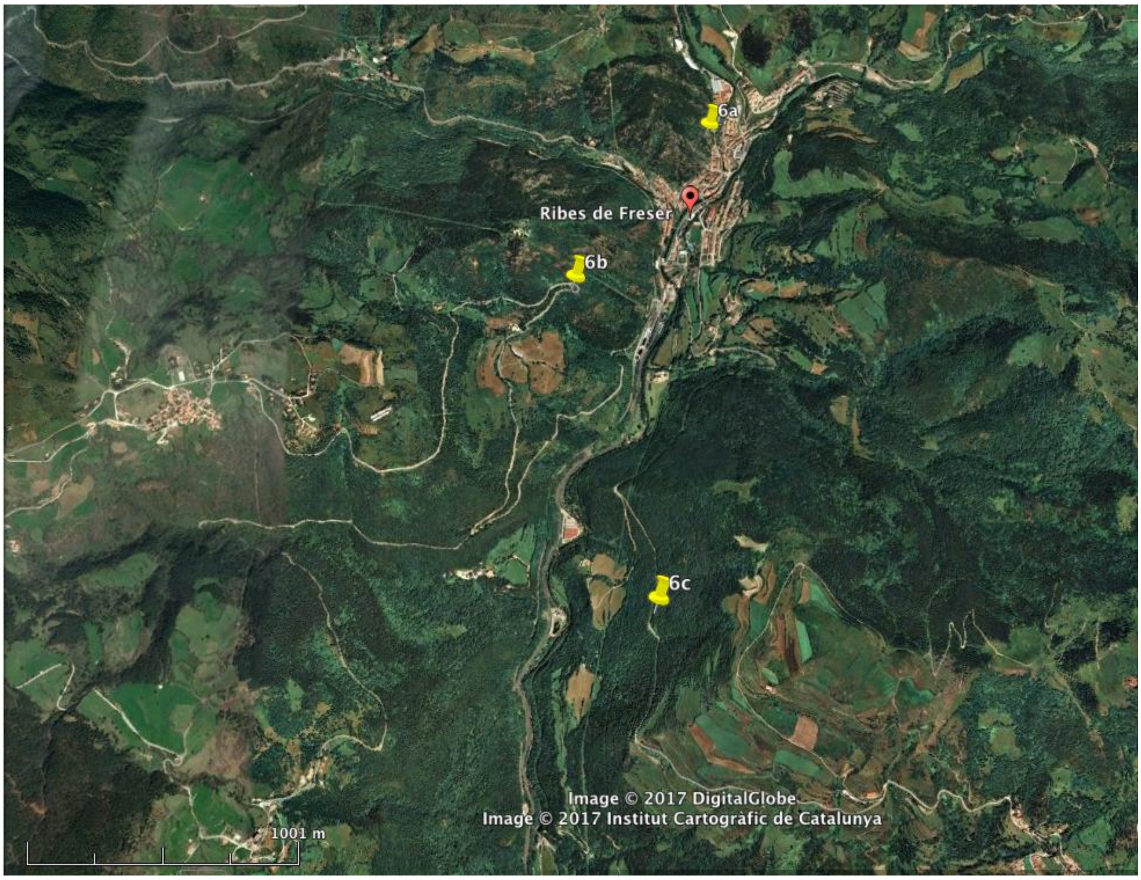


Figure 4. Location of stops 6a, 6b and 6c.

STOP 6a

Ribes de Freser granophyre

Josep Maria Casas-Joan Martí

The outcrop 6a is located on the north end of the RIbes de Freser village, at the beginning of the road GIV-5217 to Queralbs (Fig. 4). The Ribes de Freser granophyre is located in the lower and middle parts of the Upper Ordovician metasedimentary succession of the Ribes de Freser unit (Fig. 3). It is a non-foliated intrusive body of leucogranitic composition, mainly consisting of quartz and potassic (orthotic) feldspar, sometimes perthitic, with some crystals of sodium plagioclase and biotite. In general, quartz and potassium feldspar present graphic associations (Fig. 5). The contacts of the granophyre with the surrounding rock tend to be sub-parallel to the bedding planes. The grain-size increases substantially from the edges toward the inside of the granophyric body (from 0.3 to 1 cm), as well as the development of the graphic texture, although in the inner zone the granular texture gives the rock an appearance of leucogranite (ss). The granophyre has an alkaline leucogranite composition with high K and low Na and Fe contents, which implies high content of normative orthoclase and low in hypersthene. The most remarkable fact, however, is its Sr (7 ppm) poverty (Martí *et al*., 1986). The granophyre has been firstly considered as a Variscan magmatic body (Fontboté, 1949; Santanach, 1972a; Robert, 1980), although Ayora (1980) and Muñoz (1985) suggested a Late Ordovician age on the basis of field relationship. A Late Ordovician (Sandbian) age has been confirmed by

Martínez *et al.* (2011), who obtained a SHRIMP U-Pb zircon radiometric age of 458±3 Ma. The granophyre body constitutes a characteristic element of the Ribes landscape and its image has been used as a geologic identifier of the village (Fig. 6).

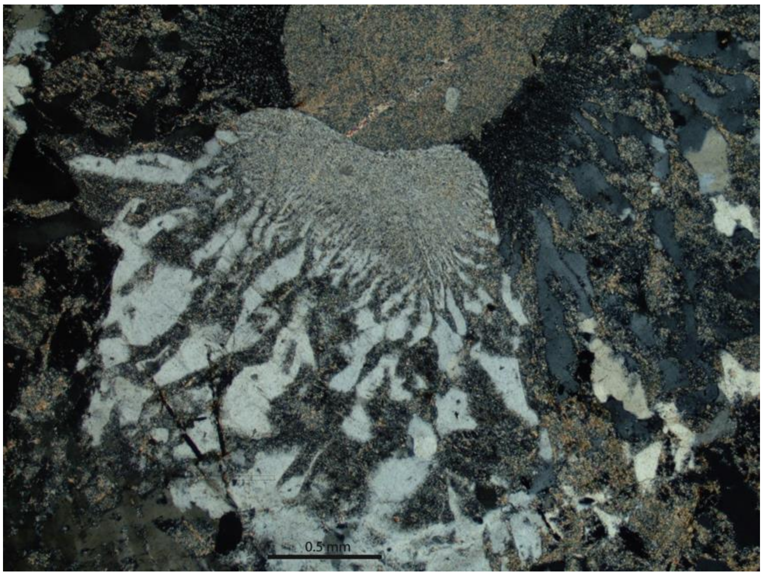


Figure 5. Graphic textures of the Ribes de Freser granopyhre.

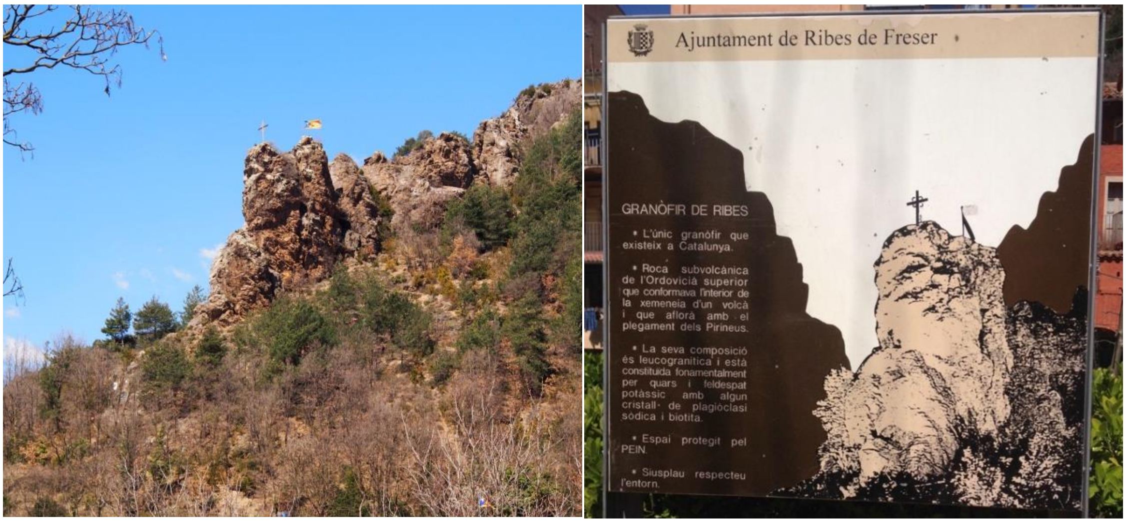


Figure 6. The granophyre as a geological identifier of the Ribes de Freser village.

STOP 6b

Upper Ordovician volcanism in Ribes de Freser

Joan Martí & Josep Maria Casas

The Upper Ordovician volcanism of the Ribes de Freser area is represented by abundant pyroclastic rocks indicating the predominantly explosive character of this volcanic episode, while associated lavas and subvolcanic intrusive rocks are scarce. The composition of the volcanic rocks evolved in time, ranging from intermediate (Rocabruna andesites, Ribes de Freser diorites) in the oldest rocks to becoming progressively more silicic (Ribes de Freser rhyolites and ignimbrites). Volcanic rocks are practically present throughout the Upper Ordovician succession, which suggests that the volcanic event spread throughout the Sandbian and Katian. In this field trip (stop 6b), we will visit the partially welded ignimbrites at the upper part of the succession. The spatial association between the granophyre, the rhyolites and the ignimbrites, and their similar age and composition, suggest a certain genetic relationship between them, being this subject for discussion in this field stop.

One of the main ignimbrite units crops out at a sharp bend on the GIV-4011 road that goes from Ribes de Freser to Campelles (Stop 6b, Fig. 4). This is a several metres thick, partially welded ignimbrite, rich in lithic fragments that is now totally devitrified and partially recrystallized (Fig. 7). The macroscopic texture of the ignimbrite is characterised by fiamme type pumice relicts, now totally transformed into phyllosilicates and chlorite, which indicate that the deposit emplaced at certain temperature sufficient to deform plastically the pumice fragments. Microscopically, these fiammes have a porphyritic texture, with albite and quartz phenocrysts. The matrix is normally heterogeneous and correspond to devitrified glass shards transformed into phyllosilicates and aggregates of quartz and feldspar. At some levels, the fluidic nature of these deposits is clearly expressed by the elongation of the original pumice fragments.

The presence of ignimbrites (and associated ashfall beds) interbedded with the Upper Ordovician metasediments indicate the existence of contemporaneous explosive volcanism, the source regions (volcanic vents) of which is not known. Compositionally, these rocks have a calc-alkaline nature, although their K content varies significantly from one to the other of the two streams. This variation does not significantly affect the content of trace elements, which suggest orogenic continental character (Martí *et al*., 1986).

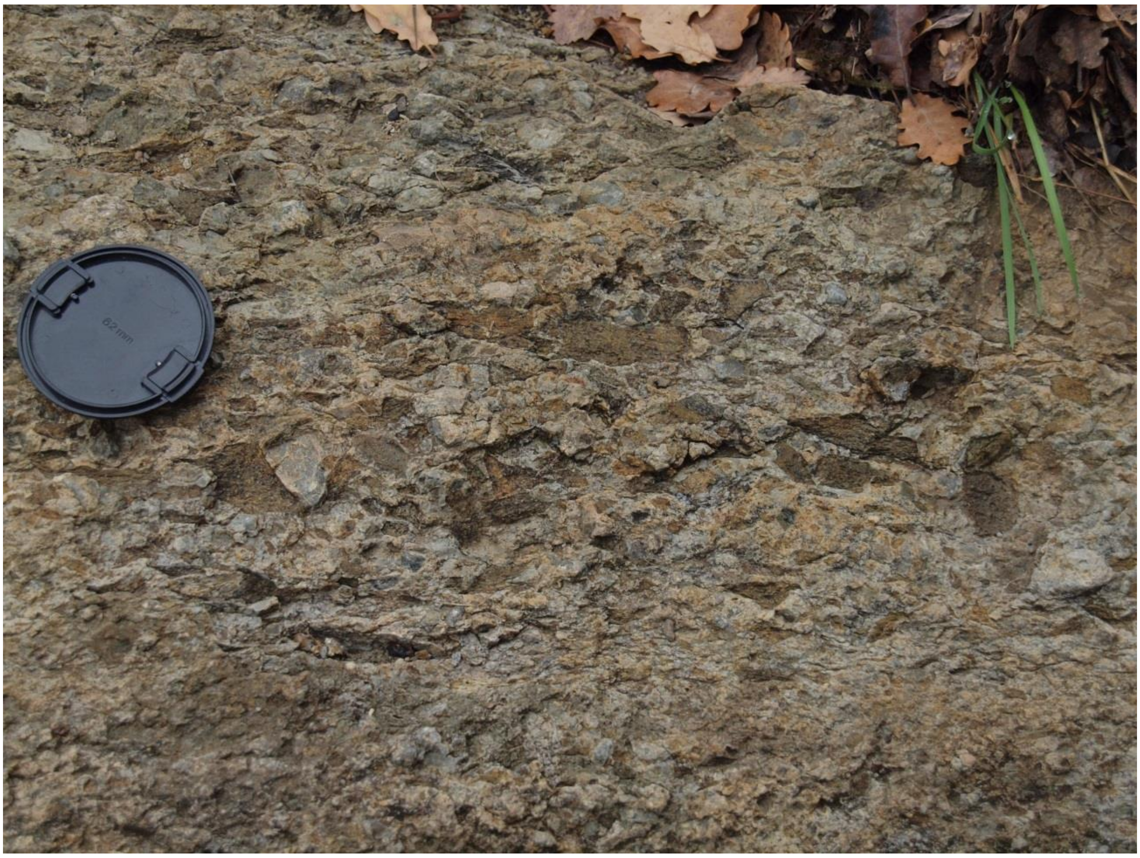


Figure 7. Partially welded ignimbrite cropping out at the GIV-4011.

STOP 6c

Upper Ordovician limestones along Ribes-Bruguera road

Josep Maria Casas, Claudia Puddu and J. Javier Álvaro

In the GIV-5263 road from Ribes de Freser to Bruguera (stop 6c, Fig. 4), we can examine a 500 m-thick succession entirely composed of limestones, marly limestones ("schistes troués") and shales (Figs. 8a-b). Its age is constrained by conodonts and crinoids that allowed Robert (1980) to propose an early Katian (former Caradoc) age. In this stop we can examine the uppermost part of this succession that exhibit an erosive contact with the overlying dark shales attributed to the Hirnantian. The succession is affected by Variscan south verging E-W oriented folds related to the formation of a cleavage, especially well developed in the marly limestones ("schistes troués"). Several normal faults cut the folds (Fig. 8c).

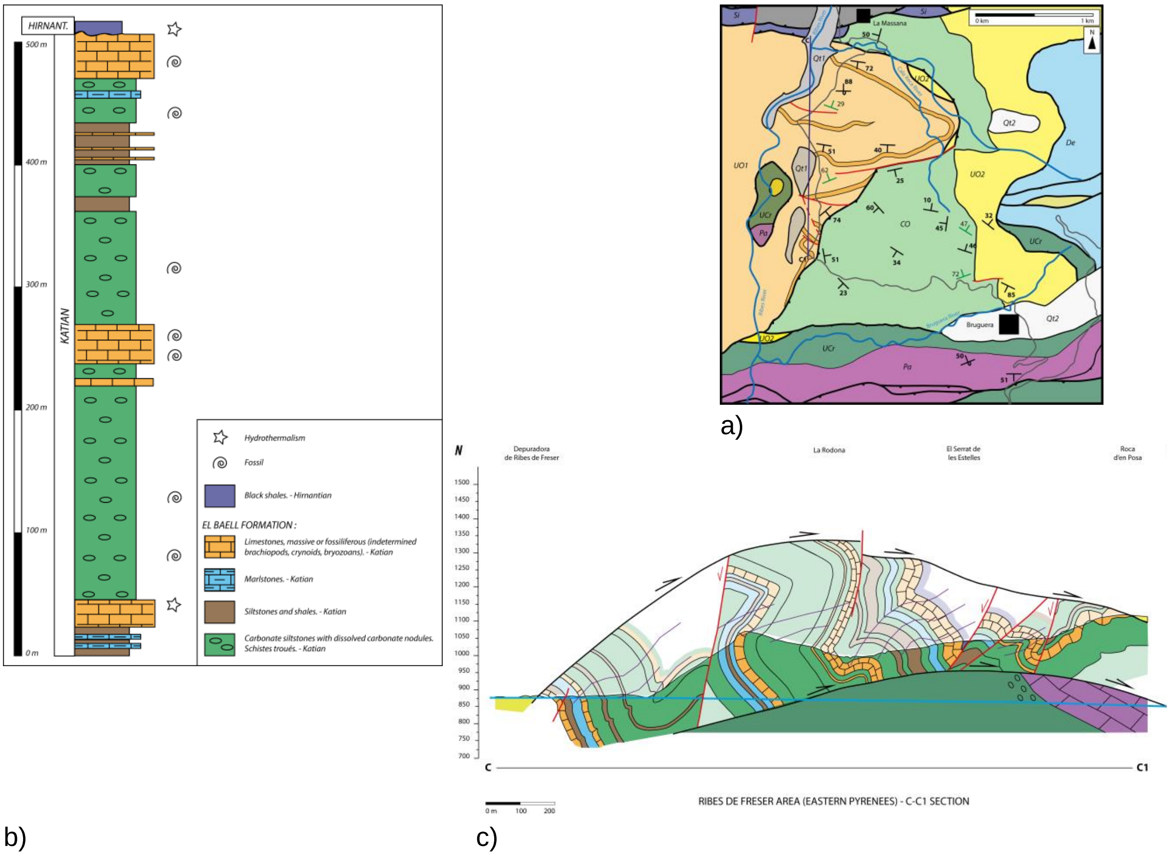


Figure 8. Geological map (a), stratigraphic log (b) and cross-section of the El Baell unit along the Ribes de Freser-Bruguera road, after Puddu *et al*. (this vol.).

The initial position of this unit cannot be pinpointed because, until now, similar successions have not been described in the Pyrenees. A preliminary restoration allows us to situate this unit in a pre-Alpine northernmost position, between the Ribes-Camprodon thrust and the Tet valley, where Upper Ordovician rocks crop out at the southern limb slope of the Vilafranca del Conflent syncline.

STOP 6d

Sardic unconformity at La Molina station

Josep Maria Casas

In this stop we will visit the la Molina area, where Santanach (1972b) described the Upper Ordovician unconformity in the Pyrenees (Fig. 9). After his work, it is widely accepted that the Upper Ordovician succession unconformably overlies either the Jujols or Canaveilles groups (García-Sansegundo & Alonso, 1989; Den Brok, 1989; Kriegsman *et al*., 1989; García-Sansegundo *et al*., 2004; Casas & Fernández, 2007). However, the origin of this unconformity has been object of several interpretations. Santanach (1972b) in this zone and García-Sansegundo *et al*. (2004) in the Garona dome, attributed the Sardic unconformity to basement tilting, related to of a Late Ordovician faulting episode and subsequent erosion. To the west, in the Lys-Caillaouas massif, Den Brok (1989) and Kriegsman *et al*. (1989) proposed the existence of a pre-Variscan deformation event. A pre-Late Ordovician folding episode has been also suggested as related to the unconformity in this area (Casas, 2010; Casas *et al*., 2012). However, the meaning of this deformation episode is unclear: it is related neither to metamorphism nor cleavage development, although it seems related to uplift, widespread emersion and considerable erosion before the onset of Upper Ordovician deposition. As a result, the Upper Ordovician rocks directly onlap different formations of the pre-Sardic succession in the Central and Eastern Pyrenees.

Acritarchs recovered from the uppermost part of the Jujols Group near this stop point to a broad Furongian-Early Ordovician microphytoplancton assemblage (Casas & Palacios, 2012), which is coincident with a maximum depositional age of ca. 475 Ma for the uppermost part of the Jujols Group in the La Rabassa dome, on the basis on the youngest detrital zircon population (Margalef *et al*., 2016). In the Albera massif, metapelites and metapsammites from the uppermost part of a metasedimentary succession that can be correlated with the Jujols Group are crosscut by acidic subvolcanic dykes, which constrain its minimum depositional age to 465-472 Ma (Liesa *et al*., 2011). All these data suggest a depositional age for the uppermost part of the Jujols Group at ca. 475 Ma. On the other hand, a ca. 455 Ma U-Pb age for the Upper Ordovician volcanic rocks directly overlying the Sardic unconformity has been proposed in the Bruguera unit (Martí *et al*., 2014) and in the Les Gavarres area (455±1.8 Ma, Navidad *et al*., 2010). Thus, a time gap of about 20 m.y. can be estimated for the Sardic Phase in the Pyrenees. Similar gaps are found in SW Sardinia (ca. 18 m.y.), the type area where the original unconformity was described, where the discontinuity is constrained by well-dated Upper Ordovician metasediments overlying upper Tremadoc-lower Floian(?) strata (Barca *et al*., 1987; Pillola *et al*., 2008).

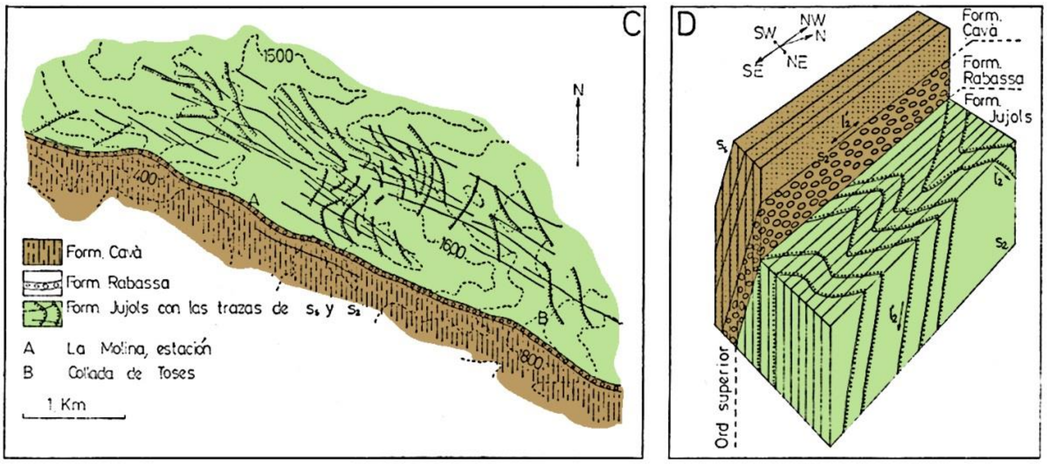


Figure 9. The Upper Ordovician unconformity at la Molina area after Santanach (1972b)

In this stop we will examine the Rabassa Conglomerate Formation, which constitutes the lowermost part of the Upper Ordovician succession (see stop 7a). It is mainly made up of subrounded to well-rounded clasts of vein quartz (Fig. 10). In this area, Santanach (1972b) reported different attitudes of the bedding planes in both the Upper Ordovician and the underlying Cambro-Ordovician successions. In the Upper Ordovician sequence, bedding is regularly oriented NW-SE and dips to the SW with minor variations in strike, whereas bedding of the Cambro-Ordovician succession presents a marked dispersion. This author attributed this difference to a pre-Upper Ordovician tilting affecting only the Cambro-Ordovician succession. Later on, Casas (2010) suggested that this different attitude is due to a folding event that gave rise to NW-SE to N-S oriented, metric to hectometric sized folds, without cleavage formation or related metamorphism. These folds, which do not affect the Upper Ordovician series, can account for the deformation and uplift of the pre-Upper Ordovician (Cambro-Ordovician) sequence and for the formation of the Upper Ordovician unconformity. These folds control the orientation of the Variscan main-folding phase minor structures, fold axes and intersection lineation in the Cambro-Ordovician sediments (Casas *et al*., 2012) (Fig. 11).

Figure 10. Field aspect of the Rabassa conglomerate rich in subangular hydrothermal quartz clasts at La Molina.

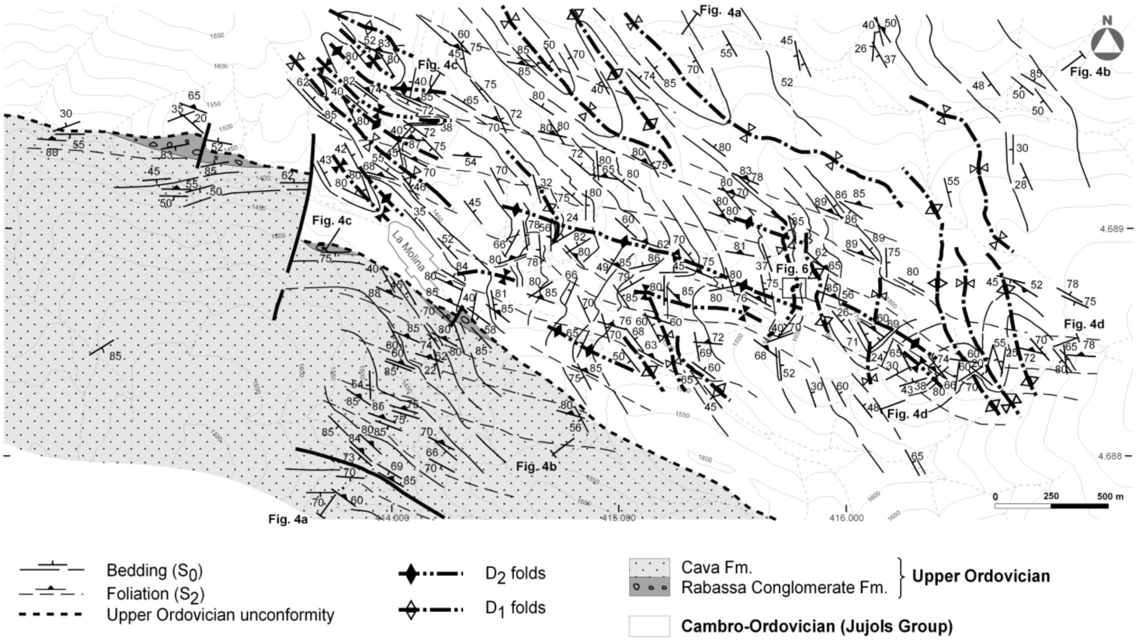


Figure 11. Geological map of the contact between the Upper Ordovician and the Cambro-Ordovician successions in La Molina showing (UTM coordinates); numbers indicate dip value (Casas *et al*., 2012).

3.2 – 7th September – Fieldtrip to eastern Pyrenees, Catalonia and Occitanie

STOP 7a

Upper Ordovician succession in Talltendre

Josep Maria Casas, Claudia Puddu and J. Javier Álvaro

The aim of this stop is to recognize stratigraphic and structural data from the Upper Ordovician rocks of the La Cerdanya area (Canigó unit). This area exhibits an Upper Ordovician succession similar to that described by Hartevelt (1970) in the Central and Eastern Pyrenees, which constitutes a broad fining-upward megasequence bearing a key limestone-marlstone interbed and marked thickness variations, ranging between 100 and 1000 m. Hartevelt (1970) defined five formations, which can be recognized with some lithologic variations all across most part of the cordillera (Fig. 12). Furthermore, as can be seen in the Ribes de Freres area (stops 6a and 6b), various volcanic and volcanosedimentary complexes crop out in different areas.

Unconformably overlying the Sardic-related palaeotopography (see stop 6d), the Rabassa Conglomerate Formation is made up of reddish-purple, unfossiliferous conglomerates with sharp lateral variations in thickness, from zero to 200 m. Conglomerates are composed of subrounded to well-rounded clasts rich in slates, quartzites and vein quartz, up to 50 cm in diameter, embedded in a green-purple granule-sized matrix. Their massive-to-channelized sets are interpreted as alluvial-to-fluvial deposits (Hartevelt, 1970). Due to its stratigraphic position, this author attributed the Rabassa conglomerates to the Sandbian-Early Katian (former Caradoc).

The overlying Cava Formation, 100-800 m thick, which either cover the Sardic unconformity or the Rabassa Conglomerate Formation, consists of feldspathic conglomerates and sandstones in the lower part, grading upward into variegated shales and fine-grained sandstones, with strongly burrowed quartzites in the uppermost part (Belaustegui *et al*., 2016). A contemporaneous volcanic influence is distinct in the southwestern part of the Canigó massif, where ash levels, andesites and metavolcanic rocks are embedded (e.g., in Ribes de Freser). Brachiopods, bryozoans and echinoderms are locally abundant, concentrated in fine-grained sandstones of the middle part of the formation, based on which, Gil Peña *et al*. (2004) attributed a Katian (former late Caradoc-early Asghill) age to this formation.

The Estana Formation, which lies above the Cava Formation, consists of limestones and marly limestones, up to 10 m thick. The unit constitutes a good stratigraphic marker bed, the so-called “schistes troués”, “Grauwacke à *Orthis*” and “Caradoc limestones” of French and Dutch geologists. Conodonts, brachiopods, bryozoans and echinoderms are abundant, yielding a Katian (former mid Ashgill; Gil Peña *et al*., 2004) age for the development of echinoderm-bryozoan meadows on shelly, offshore-dominated substrates.

The “Ansovell” Formation (Ansobell *sensu* Hartevelt, 1970) unconformably overlies the Estana limestone and consists of blackish shales with common slumping and convoluted layers, close to the base, and minor quartzite interlayers in the uppermost part. Where the Estana Formation tapers off, the Ansovell shales directly overlie the Cava sandstones. Finally, the Bar Quartzite Formation marks the top of the Upper Ordovician as a quartzitic layer, 5-10 m thick. An Hirnantian age (former late Ashgill) was proposed for the Ansovell and Bar formations by Hartevelt (1970), and confirmed by Roqué *et al*. (2017). Westward, in the Orri, Pallaresa and Garona domes, Gil-Peña *et al*. (2000, 2004) reported a calcareous conglomerate, up to 8 m thick, directly capping the erosive unconformity that marks the Estana/Ansovell contact, and attributed it to a Hirnantian glacial event.

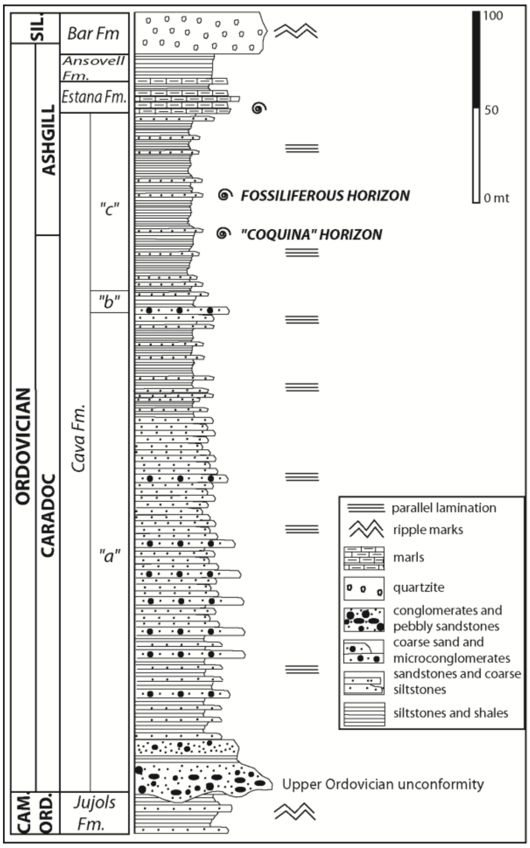


Figure 12. Synthetic stratigraphic column of the Upper Ordovician succession in the Ordèn-Talltendre area, after Puddu & Casas (2011).

In the Talltendre area, we can recognize the Upper Ordovician succession, the basal unconformity and several normal faults affecting the Upper Ordovician succession, the Cambro-Ordovician sediments and the basal unconformity (Fig. 13). Talltendre town is located on the quartzite of the Bar Quartzite Fm and, following the path from Talltendre to stop 7a, we can recognize, from top to bottom, all the Upper Ordovician succession (Fig. 13).

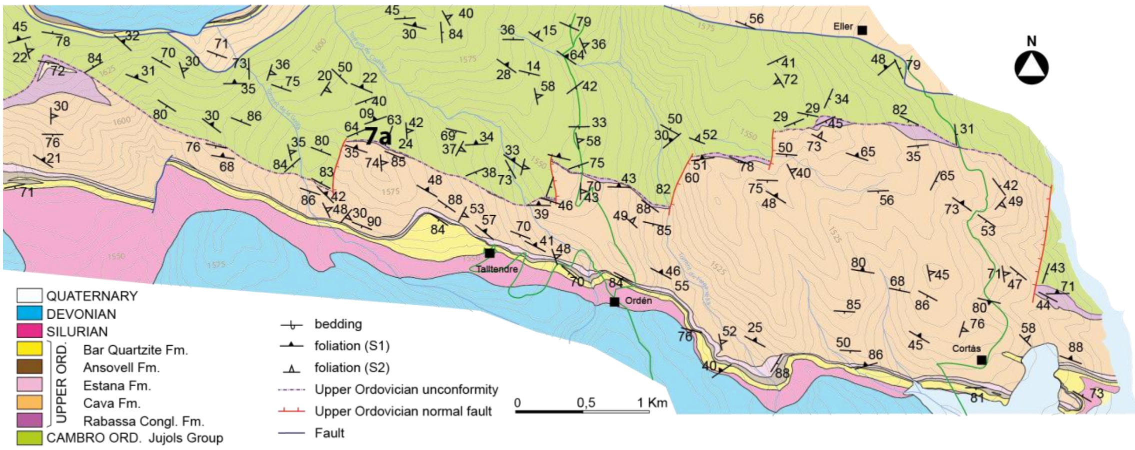


Figure 13. Geological map of the Talltendre area, north of Bellver de Cerdanya; modified from Puddu & Casas (2011).

It should be noted that near this area three brachiopod genera (*Porambonites* sp., *Eoanastrophia* sp., and *Dolerorthis* sp.) were collected, which have not yet been described in the Cava Formation (Fig. 14). The new brachiopod fauna comes from the uppermost part of the "c" member of the Cava Formation, above the “coquina” horizon described by Hartevelt (1970). Unfortunately, the state of preservation of the fossils only allowed a generic assignment, which represents an intermediate fauna between the “late Caradoc-early Ashgill” brachiopods collected in the “coquina” horizon located in the upper part of this member (*Svobodaina havliceki*, *Rostricellula* sp., *Rafinesquina* sp.; Gil Peña *et al*., 2004) and the “mid Ashgill” brachiopods of the Estana Formation (*Dolerorthis* sp., *Eoanastrophia pentamera*, *Iberomena sardoa, Leangella anaclyta*, *Longvillia mediterranea*, *Nicolella actoniae*, *Porambonites* (*Porambonites*) *magnus*, *Ptychopleurella villasi*; Gil Peña *et al*., 2004).

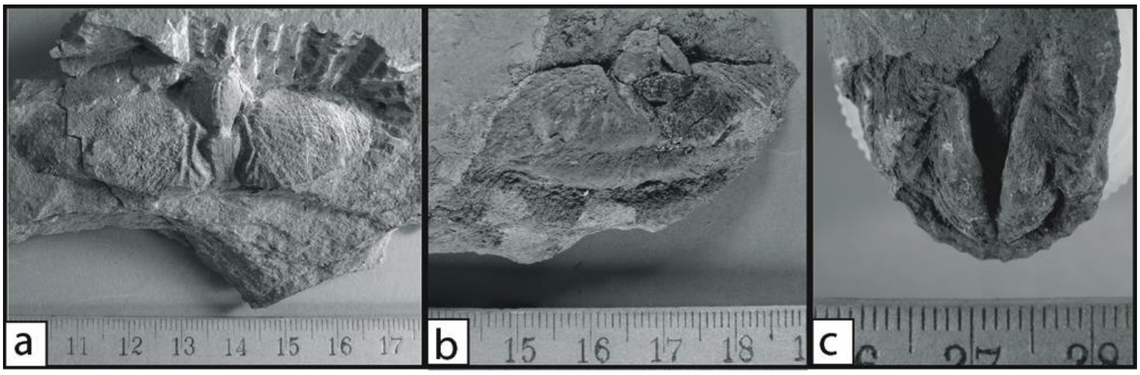


Figure 14. Brachiopods collected in the upper part of the “c” member from the Cava Formation: (a) *Porambonites* sp. (internal mould of ventral valve), (b) *Dolerorthis* sp. (internal mould of ventral valve), and (c) *Eoanastrophia* sp. (internal mould of dorsal valve).

Moreover, in this area, several normal faults affect the Upper Ordovician succession, the Cambro-Ordovician sediments and the unconformity. The faults are steep and currently exhibit broad N-S to NNE-SSW trending. In most cases, their hangingwall is the eastern block despite the presence of some antithetic faults; maximum throws of about 0.2 to 0.9 km can be recognized. Displacement progressively diminishes upward and peters out in the Cava rocks (Fig. 13). Based on these orientations, an E-W extension (in present day coordinates) can be proposed. The original orientation of the faults cannot be pinpointed owing to subsequent deformation events, although an original N-S orientation can be proposed. This orientation probably prevented the faults from being inverted during subsequent Variscan and Alpine deformation events, although the faults probably suffered rotations on horizontal E-W axes during these deformation episodes. On the other hand, sharp variations in the thickness of the Upper Ordovician succession have been reported by several authors; Llopis Lladó, 1965; Hartevelt, 1970; Speksnijder, 1986). Hartevelt (1970) documented variations from 200 to more than 850 m in the thickness of the Cava Formation: e.g., eastward from La Seu d’Urgell, the thickness of the Rabassa and Cava formations attain more than 800 m before sharply diminishing to some tens of metres within a few kilometres (Casas & Fernández, 2007). There, the maximum observed thickness occurs associated with the maximum grain size of the conglomerates (pebbles exceeding 50 cm in diameter are common). Sharp variations in thickness and grain size can be attributed to palaeorelief formation controlled by fault activity and subsequent erosion of uplifted palaeotopographies, with subsequent infill controlled by alluvial-fan and fluvial deposition.

In stop 7a, we can recognize the basal Upper Ordovician unconformity that separates the Upper Ordovician sediments from the underlying Cambro-Ordovician ones. This unconformity can be identified from detailed mapping in several neighbouring areas. In stop 7a, the unconformity has a NW-SE trend and cuts the bedding of the pre-unconformity deposits at different angles, ranging from a few to 90º (Fig. 15). As in the La Molina area, a synsedimentary hydrotermal activity gave rise to quartz veins and dykes, which subsequently feed the Rabassa conglomerates as vein quartz pebbles (Fig. 15).

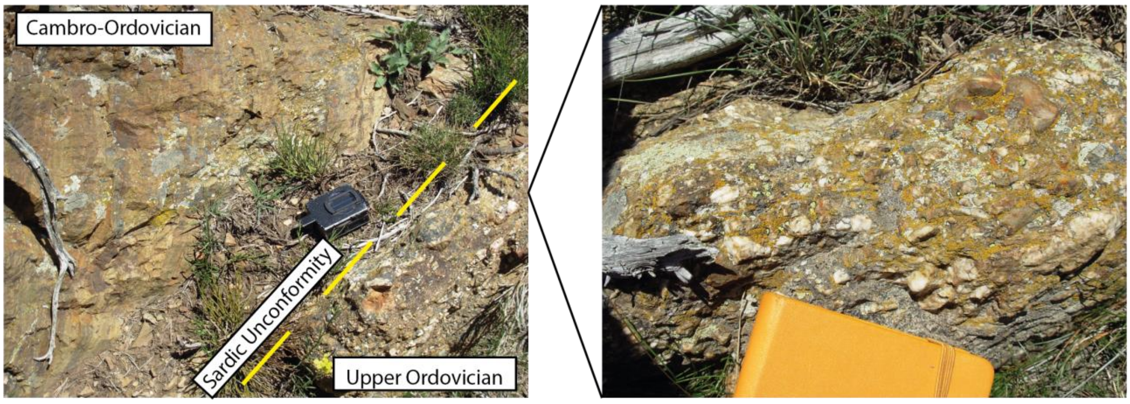


Figure 15. Sardic unconformity at the Taltendre area (stop 7a)

STOP 7b

Graus de Canaveilles along Conflent valley

Bernard Laumonier

The so-called Relais de l'Infante (Bains de Canaveilles) cross-section is certainly the best place in the Pyrenees to present and discuss the long-lasting question about the significance of the Canigou gneisses and the relationships between these gneisses and the micaschists of the surincombant metasedimentary Canaveilles (Nyer) Formation.

In the 1950s, in accordance with the ideas prevaling then, the quartzo-feldspatic Canigou gneisses were thought to be metasomatic rocks ("migmatites") derived from grauwackes, and the gneiss/micaschist contact was considered as an upper metasomatic front (e.g. Raguin, 1938; Guitard, 1955). However, this metasomatic model ("feldspathisation régionale") was gradually and rapidly abandoned:

(i) In 1958, G. Guitard concluded that the upper part of the gneisses, especially the leptynitic ones and the La Preste gneisses of the southeast of the Canigou massif, were derived by simple *in situ* ("topochimique") recrystallisation of rhyolites of variable thickness, unlike the underlying metasomatic biotite augen gneisses.

(ii) A few years later (Guitard,1963a, b), the author suggested that most of the Canigou gneisses are in fact orthogneisses, either metavolcanic (leptynitic gneisses) or metagranitic (augen gneisses). As a result, their large K-feldspar grains ("eyes") should not be porphyroblasts but porphyroclasts derived from the deformation of K-feldspar phenocrysts of former porphyritic, rapakivi biotite granites. However, another question was under debate then: did these metagranites represent an old Precambrian basement or did they correspond to a younger laccolitic intrusion?

(iii) In the following years, Guitard (1958, 1970[1965], *in* Jaffrezo et al., 1977), Autran *et al.* (1966, 1977), Fonteilles & Guitard (1988), Guitard *et al.* (1996, 1998), Laumonier & Guitard (1986) and Laumonier (1988) definitely favoured the first option, suggesting a model involving a Cadomian granitic basement (the G2 augen gneisses), a major post-Cadomian unconformity, a discordant Cambrian sedimentary cover (the Canaveilles series *sensu* Cavet, 1957), the base of which (G1 gneisses) is rhyolitic, and a major recumbent fold resulting in the reappearence of the Canaveilles Series under the stratoid Canigou gneisses, forming the Balatg Series.

The strongest arguments for the basement model (see Guitard *et al.*, 1996), apart from the micrographic characteristics, are:

a, the presence, all around the Canigou massif, of a carbonate level at (or very near) the base of the Canaveilles series; this level, called "marbre de base", locally associated with arkoses and/or volcanic rocks ("transition gneisses"), represents the transgressive base of the marine Canaveilles Series upon a major post-Cadomian erosive surface;

b, the fact that the Canaveilles Series unconformably overlies different lithologies, such as orthogneisses (Canigou, Roc de France) and/or paragneisses (Aston, Albères);

c, the fact that the metagranites (now, orthogneisses) do not cut the Canaveilles series;

d, radiometric ages available at this time (between 630 and 535 Ma, whole rock Rb/Sr and U/Pb on zircons) pointed to a Cadomian age for the metagranites; and

e, the attribution of the Canaveilles Series to the Cambrian (*s.l.*).

Nowadays, these arguments are more or less re-evaluated, in the context of an Ordovician laccolitic model for the Canigou orthogneisses because:

(i) the existence of a carbonate level all around the Canigou massif is confirmed, but it is any more interpreted as a "marbre de base", but simply as the lowest carbonate level (M1) of the Canaveilles Group (Canaveilles/Nyer Formation); the latter represents the overlying country-rock to the Canigou laccolith;

(ii) the supposed unconformity is the upper contact of the intrusion; this highlights the remarkable stratiform character of the laccolith; the transition gneisses are derived from the most leucocratic facies of the laccolith;

(iii) the intrusive contacts of the metagranites across the Nyer Formation indeed occur but are very rare;

(iv) after some unclear radiometric ages, ranging from ~451 Ma to ~425 Ma, the first indisputable dating of the Canigou metagranites (*in situ* U/Pb zircon dating by SIMS) gave an Early Ordovician age (475±10 Ma, Deloule *et al*., 2002); this age was subsequently confirmed several times (Cocherie *et al.,* 2005; see also stops 6, *supra*, for other references);

(v) the lower part of the Canaveilles (Nyer) Formation is considered to be late Ediacaran (~580 Ma) (Padel, 2016), as a result of which, the Canigou metagranites (Ordovician) are necessarily intrusives in the Ediacaran series (Laumonier *et al*., 2004).

Other observations (Barbey *et al*., 2001) confirmed these conclusions based on the discovery of a relic contact metamorphic aureole (melted metapelites, relics of andalusite-cordierite porphyroblasts, pre-Variscan garnet-clinopyroxenes associations in calcsilicate rocks).

The Relais de l'Infante cross-section is illustrated in the Figure 1:

The Canigou metagranites are represented by: (i) fine-grained leucogneisses (leptynites) with thin augen gneiss intercalations (levels A and C); (ii) banded gneisses ("gneiss rubanés") containig alternating leucogneisses and augen gneisses (level B); and (iii) massive augen gneisses (porphyritic granite, level D).

The Canaveilles Formation of the French geological maps (now named Nyer Formation of the Canaveilles Group; Padel, 2016) mainly consists of: (i) metapelites (cordierite or cordierite-andalusite micaschists), (ii) pure calcareous marbles (levels A to D, F), (iii) impure marbles with detrital quartz, muscovite and microcline, and (iv) rare calcsilicate gneisses (level E). Separating the effects of Ordovician contact metamorphism and Variscan regional HT-BP metamorphism is not easy…

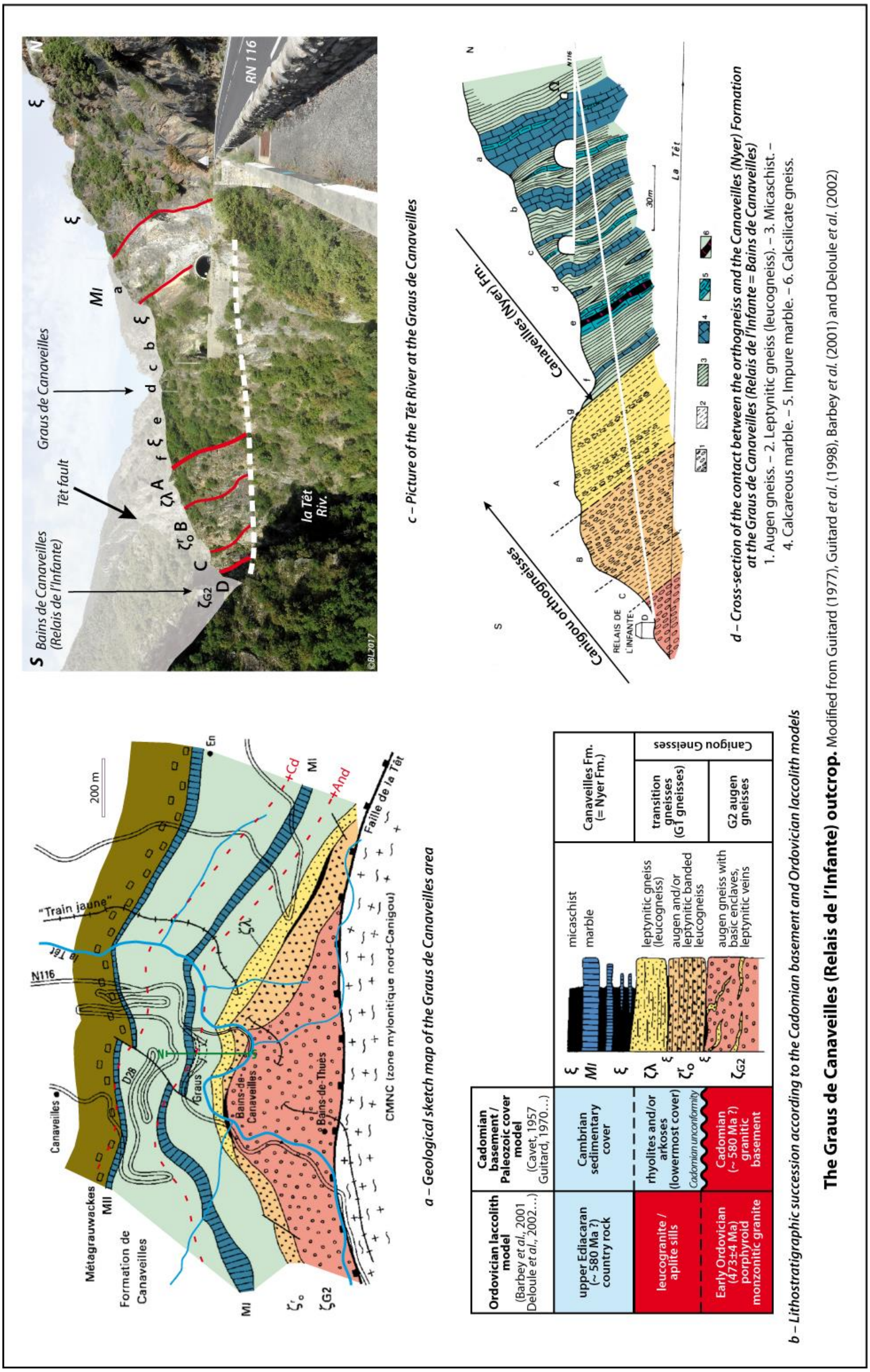


Figure 16. Relais de l'Infante cross-section (see also Jaffrezo, 1977).

Canigó granitic orthogneiss along Saorra-Pi road

Josep Maria Casas

This stop is located along the D6 road between Saorra and Pi and is devoted to the Canigó G-2 type augen gneiss (Guitard, 1970). They show a porphyroclastic texture with a thick-medium grain granoblastic matrix. Their metamorphic mineral composition is formed by quartz, plagioclase, K-feldspar, biotite and ilmenite and, as accessories minerals, apatite and zircon. The porphyroclasts are formed by K-Feldspar twined in Carlsbad with a rim formed by a cataclastic mosaic of grains; they include drops and veins pertites. Plagioclase mega-crystals are a polisynthetic and draughtboard twin and show patches of albite and muscovite spills along their crystallographic face.

Figure 17. Field aspect of the granitic Canigó G-2 type gneiss at the Saorra-Pi road.

We are in the northern slope of the Canigó massif, one of the gneissic core domes aligned in the backbone of the Pyrenees. The regional gneissic foliation dips to the N-NE (left hand in Fig. 17) and a related stretching lineation oriented NE-SW. The Canigó gneiss forms a 2000 m thick body with laccolithic morphology and, as discussed previously in stop 7b, derives from an Ordovician intrusive (Fig. 18). In the pre-Variscan basement of the Pyrenees successive Ordovician magmatic pulsations are well documented. According to radiometric data, magmatism lasted about 30 m.y., from ca. 477 to 446 Ma. Although the magmatic activity seems to be continuous, two peaks can be distinguished at 473-472 Ma and 457 Ma (Fig. 19). Based on geochronological and geochemical data, two different magmatic complexes can be distinguished: latest Early-Mid Ordovician and Late Ordovician magmatism.

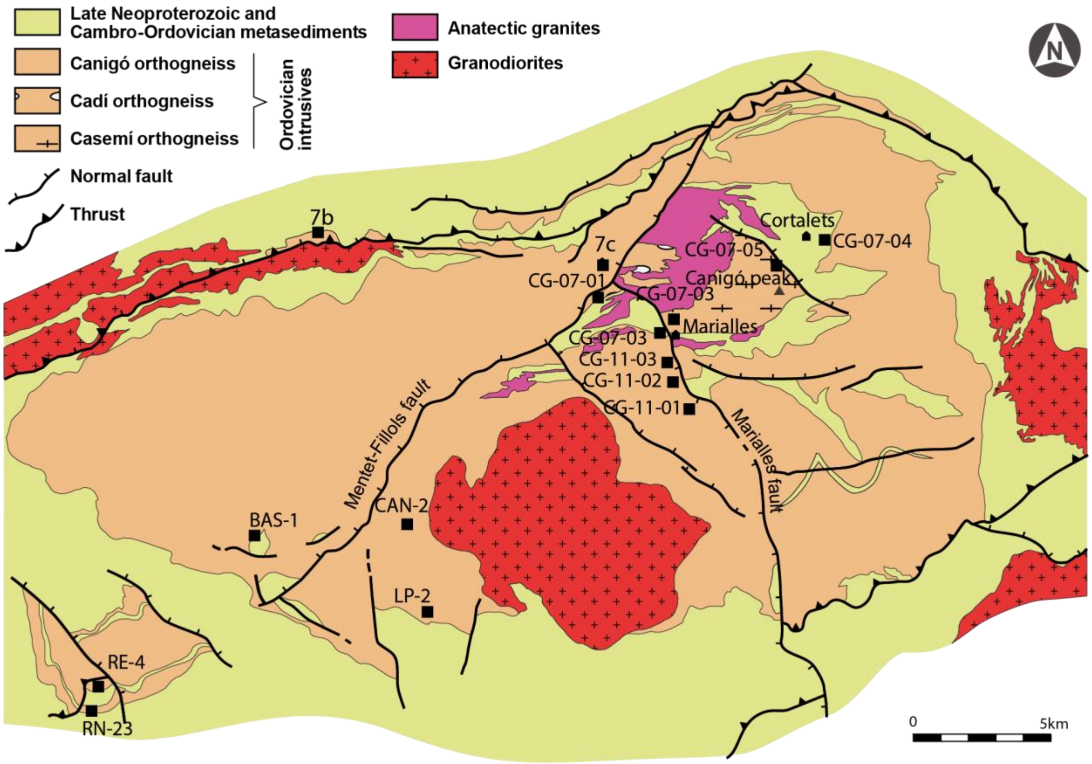


Figure 18. Schematic geological map of the Canigó massif with location of stops 7b and 7c and the geochronological samples referred in the Canigó log of the Fig. 20; modified from Guitard (1970) and Santanach (1972a).

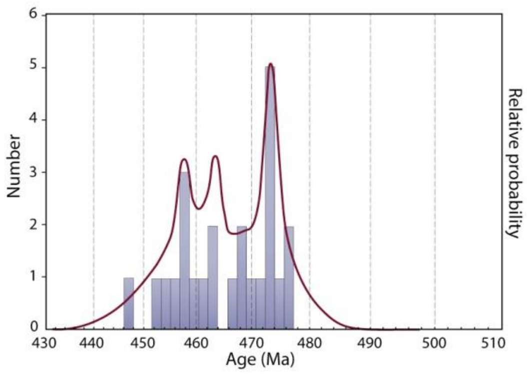


Figure 19. Relative probability plot of the geochronological ages for the Ordovician magmatism of the Pyrenees; data after Deloule *et al*. (2002), Cocherie *et al.* (2005), Castiñeiras *et al.* (2008), Denele *et al*. (2009), Casas *et al*. (2010), Liesa *et al.* (2011), Martínez *et al*. (2011), Mezger & Gerdes (2016) and Liesa *et al*. (unpublsh.).

During Early to Mid Ordovician times, the magmatic activity gave rise to the intrusion of voluminous aluminous granites, about 500 to 3000 m in size, which constitute the protoliths of the large laccolithe-shaped orthogneissic bodies that crop out at the core of the domal massifs that punctuate the backbone of the Pyrenees. These are, from west to east, the Aston (470±6 Ma, Denele *et al*., 2009; 467±2 Ma, Mezger & Gerdes, 2016), Hospitalet (472±2 Ma, Denele *et al.,* 2009), Canigó (472±6 to 467±7 Ma, Cocherie *et al*., 2005), Roc de Frausa (477±4 Ma, Cocherie *et al*., 2005; 476±5 Ma, Castiñeiras *et al*., 2008) and Albera (470±3 Ma, Liesa *et al*.,

2011) massifs (Fig. 20), which exhibit a dominant Floian-Dapingian age. It should be noted that only a minor representation of basic coeval magmatic rocks are exposed (e.g., Cortalet metabasite). The acidic volcanic equivalents have been reported in the Albera massif, where subvolcanic rhyolitic porphyroid rocks yielded similar ages than those of the main gneissic bodies: 465±4, 472±3, 473±2 and 474±3 Ma (Liesa *et* al., 2011; Liesa unpubl.).

The Late Ordovician magmatic pulse yielded a varied suite of magmatic rocks. Small granitic bodies are emplaced in the Canaveilles and Jujols strata of the Canigó massif and constitute the protoliths of the Cadí (456±5 Ma, Casas *et al*., 2010), Casemí (446±5 and 452±5 Ma, Casas *et al*., 2010) and Núria and Queralbs gneiss (457±4 and 457±5 Ma, Martínez *et al*., 2011). Metre-scale thick bodies of metadiorite interlayered in the micaschists of the Balaig series have also yielded a Late Ordovician age for the formation of its protolith (453±4 Ma, Casas *et al*., 2010). As we have seen in stops 6a and 6b, these rocks are coeval with calc-alkaline volcanic rocks (ignimbrites, andesites and volcaniclastic rocks) interbedded in the Upper Ordovician of the Ribes de Freser and Bruguera units.

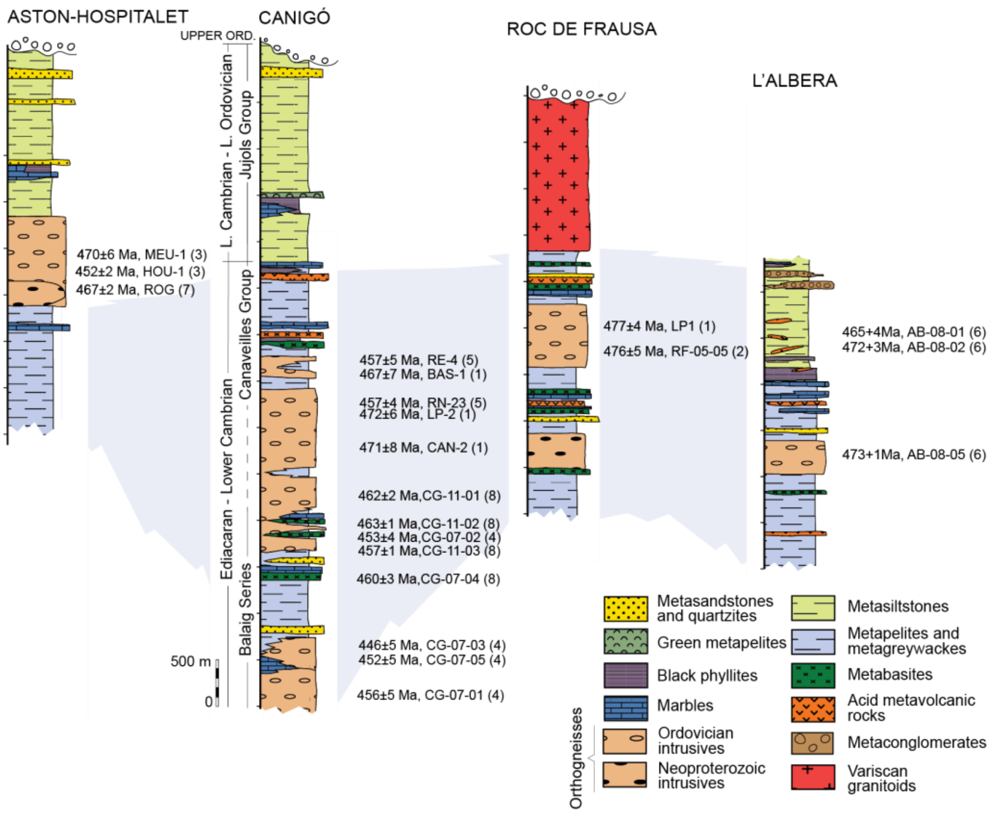


Figure 20. Stratigraphic logs of the pre-Upper Ordovician rocks from the Aston-Hospitalet, Canigó, Roc de Frausa and Albera massifs with the geochronological data of the protoliths of the Ordovician gneisses and metabasites: (1) Cocherie *et al*. (2005); (2) Castiñeiras *et al.* (2008); (3) Denele *et al*. (2009); (4) Casas *et al.* (2010); (5) Martínez *et al.* (2011); (6) Liesa *et al.* (2011) and (7) Mezger & Gerdes (2016); (8) Navidad *el al.* (en prensa). Stratigraphic data from Guitard (1970), Santanach (1972b), Ayora & Casas (1986), Liesa & Carreras (1989) and Liesa *et al.* (2011).

3.3- 8th September – Fieldtrip to Mouthoumet massif, Occitanie

The Mouthoumet massif lies south of Montagne Noire and north of the North Pyrenean frontal thrust (*sensu* Laumonier, 2015), as a result of which the massif belongs to the Occitan Domain (Álvaro *et al*., 2016; Pouclet *et al*., 2017). The Mouthoumet massif contains four tectonostratigraphic units, from east (tectonically top) to west (bottom), the Serre de Quintillan, Félines-Palairac and Roc de Nitable thrust slices, and an unnamed parautochthon (Berger *et al*., 1997; Bessière & Baudelot, 1988; Bessière & Schulze, 1984; Bessière *et al*., 1989; Cornet, 1980). The Ordovician has been traditionally reported as informal lithological units compared to the neighbouring and fossiliferous formations from the southern Montagne Noire (Cabrières klippes) and eastern Pyrenees (e.g., Gèze, 1949; Bessière *et al*., 1989; Berger *et al*., 1997). As in the case of the Pyrenees and Montagne Noire, the Middle Ordovician is absent and its gap allows differentiation between a Lower Ordovician sedimentary succession with subsidiary volcanic episodes and an unconformably overlying Upper Ordovician-Devonian sedimentary package that begins with a distinct volcanic episode.

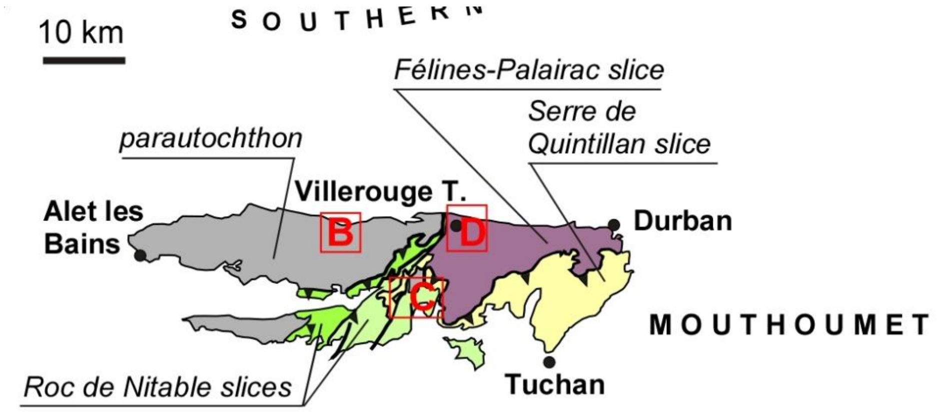


Figure 21. Variscan thrust slices of the Mouthoumet massif with setting of Ordovician formation stratotypes; B: Montjoi stratotype, C: Gascagne stratotype and D: Villerouge and Marmairane stratotypes; modified from Berger (1982), Bessière & Schulze (1984), Bessière (1987), Bessière & Baudelot (1988) and Berger *et al*. (1990, 1993, 1997).

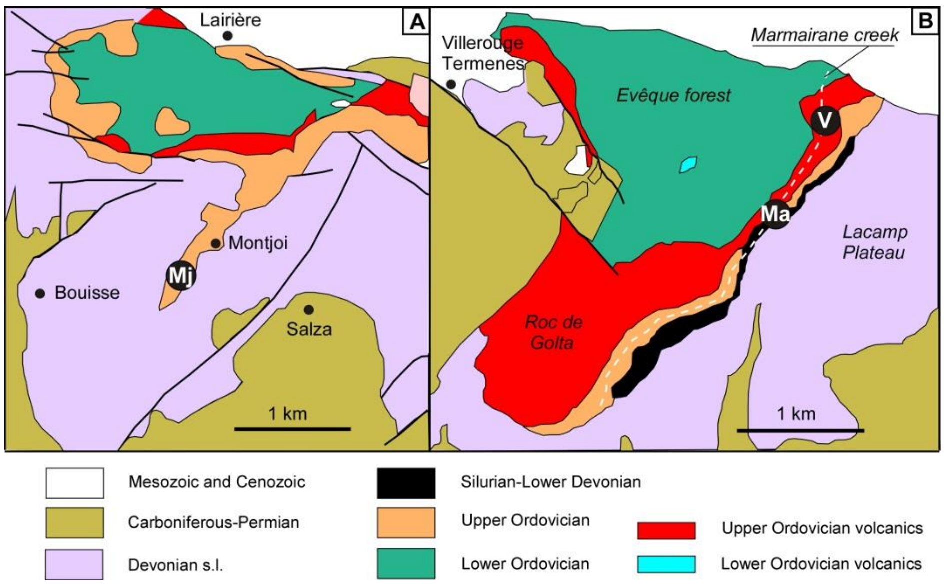


Figure 22. Geological sketch of the Mouthoumet study area. A. Surroundings of Montjoi, Mouthoumet parautochthon. B. Villerouge-Termenès, Félines-Palayrac thrust slice; Mj: Montjoi stratotype (STOP 8a), V-Ma:Villerouge and Marmairane stratotypes, respectively (STOP 8b); modified from Berger (1982), Bessière & Schulze (1984), Bessière & Baudelot (1988) and Berger *et al.* (1990, 1993, 1997).

The Lower Ordovician consists of greenish and brownish shales and sandstones, dated to Tremadocian-Floian(?) (former Arenig) by acritarchs (Baudelot & Bessière, 1975, 1977; Berger, 1982; Cocchio, 1981, 1982). This siliciclastic unit contains interbedded thick flows (up to 100 m) of porphyritic metarhyolites in the parautochthon, and metarhyolitic flows or sills overlain by metabasaltic flows in the lower part of the Serre de Quintillan slice and in the Davejean tectonic window (Berger *et al*., 1997; Bessière & Baudelot, 1988; Bessière & Schulze, 1984; Bessière *et al*., 1989). Both the lithostratigraphic units and volcanogenic products were recently revised in Álvaro *et al*. (2016).

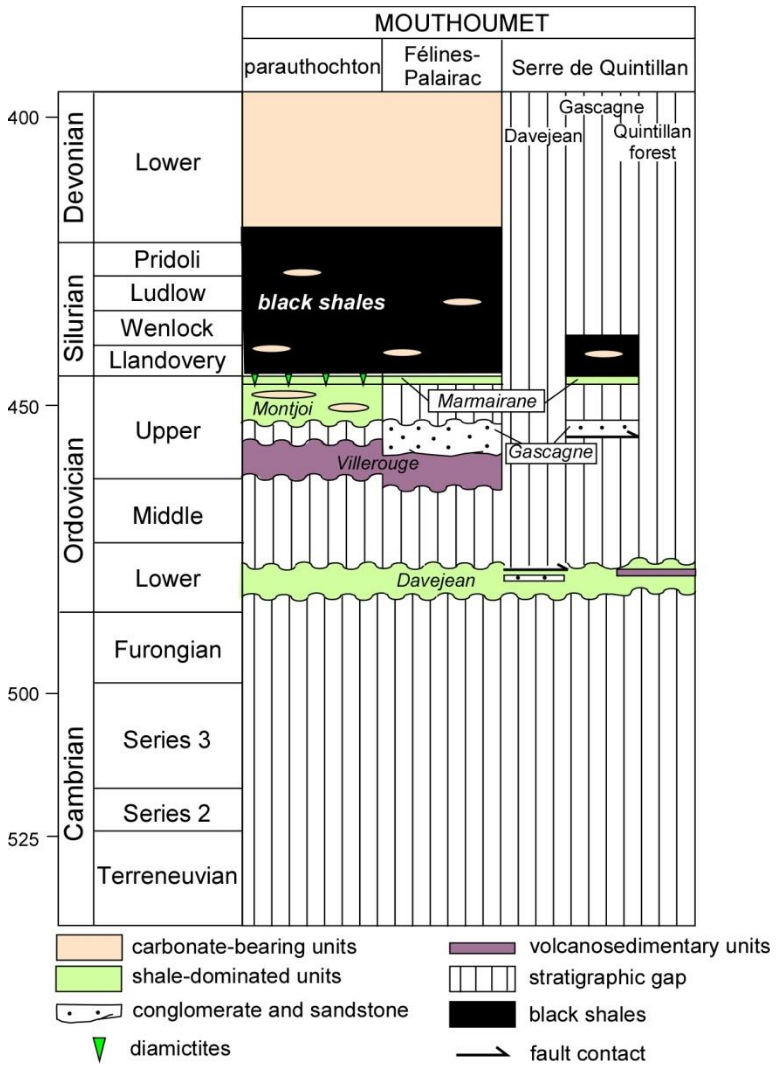


Figure 23. Stratigraphic framework of the Cambrian-Devonian strata and volcanosedimentary complexes in the Mouthoumet massif; modified from Álvaro *et al*. (2016).

STOP 8a

Lower Ordovician Davejean Volcanic Complex and Montjoi and Marmairane formations in Montjoi, Mouthoumet parautochthon

J. Javier Álvaro, Eric Monceret and Daniel Vizcaïno

Davejean Volcanic Complex

The most significant Lower Ordovician volcanics consist of (i) a thick flow (ca. 100 m) of porphyritic metarhyolites embedded in the sandstones of the parautochthon, dated to Tremadocian by acritachs (Cocchio, 1982); and (ii) flows or sills of metarhyolites overlain by flows of metabasalts encased in the Lower Ordovician shales and sandstones of the lower part of the Serre de Quintillan slice and in the Davejean tectonic window. These Lower Ordovician volcanics are referred to the Davejean Volcanic Group.

The Tremadocian metarhyodacite flow of the the Lairière volcanic dome (along the road D212), in the vicinity of Montjoi (parauthocthon) was studied by Pouclet *et al*. (2017), together with a metarhyolite flow interbedded between Lower Ordovician shales and sandstones from the Serre de Quintillan slice, and metabasaltic flows from the Davejean window and in the Maisons village, west of the Serre de Quintillan slice. The rhyolites are suspected to predate the subsequent basic products. Composition ranges from sodo-potassic dacite to rhyolite and display an alkali-calcic character. The A/CNK ratio is peraluminous, the rare earth elements are enriched and fractionated, and the lithophile elements are enriched.

The overlying basaltic lava flows are microlitic porphyritic and range from mafic composition with abundant phenocrysts of pyroxene and plagioclase to intermediate composition enriched in plagioclase and alkaline feldspar. They were metamorphosed with complete recrystallization of magmatic minerals and matrix in the greenschist facies paragenesis. In addition, many flows were spilitized and enriched in sodium (albite). The chemical composition is basaltic to moderately evolved. The norm calculation, though poorly constrained due to post-magmatic processes, suggests an oversaturated tholeiitic magmatic composition. The N-MORB normalized diagram shows parallel profiles of the mafic and evolved lavas belonging to the same magmatic batch. Compared to the CT average composition, the lavas can be defined as continental tholeiites, also shown by the Ti-Nb-Th ratios. In summary, basaltic rocks are evolved, REE-rich, and LREE-fractionated. They display a typical initial rift tholeiite (IRT) magmatic signature indicating contributions from both asthenospheric and lithospheric mantle sources.

By comparison with the remaining Occitan Domain, the Davejean and Peyrebrune Volcanic Complexes of the Mouthoumet and northern Montagne Noire massifs, respectively, represent a Tremadocian pulsation related to underplated basic magma rising. Basaltic rocks are evolved, REE-rich, LREE-fractionated and display a typical initial rift tholeiite (IRT) magmatic signature indicating contributions of both astenospheric and lithospheric mantle sources (Pouclet *et al*., 2017).

Montjoi Formation

The stratotype of theformation lies in the parautochthon, along the road D212, 1 km SW of Montjoi (N42°59′11.19″, E02°28′28.65″). This fossiliferous mixed (carbonate-shale) unit was traditionally known as the Ashgill “schistes troués” and “limestone layers” and contains a rich and diversified shelly fauna of bryozoans, brachiopods and echinoderms. It is a shale-dominated formation, up to 90 m thick, which displays common limestone nodules parallel to stratification and lenticular beds, up to 1.4 m thick.

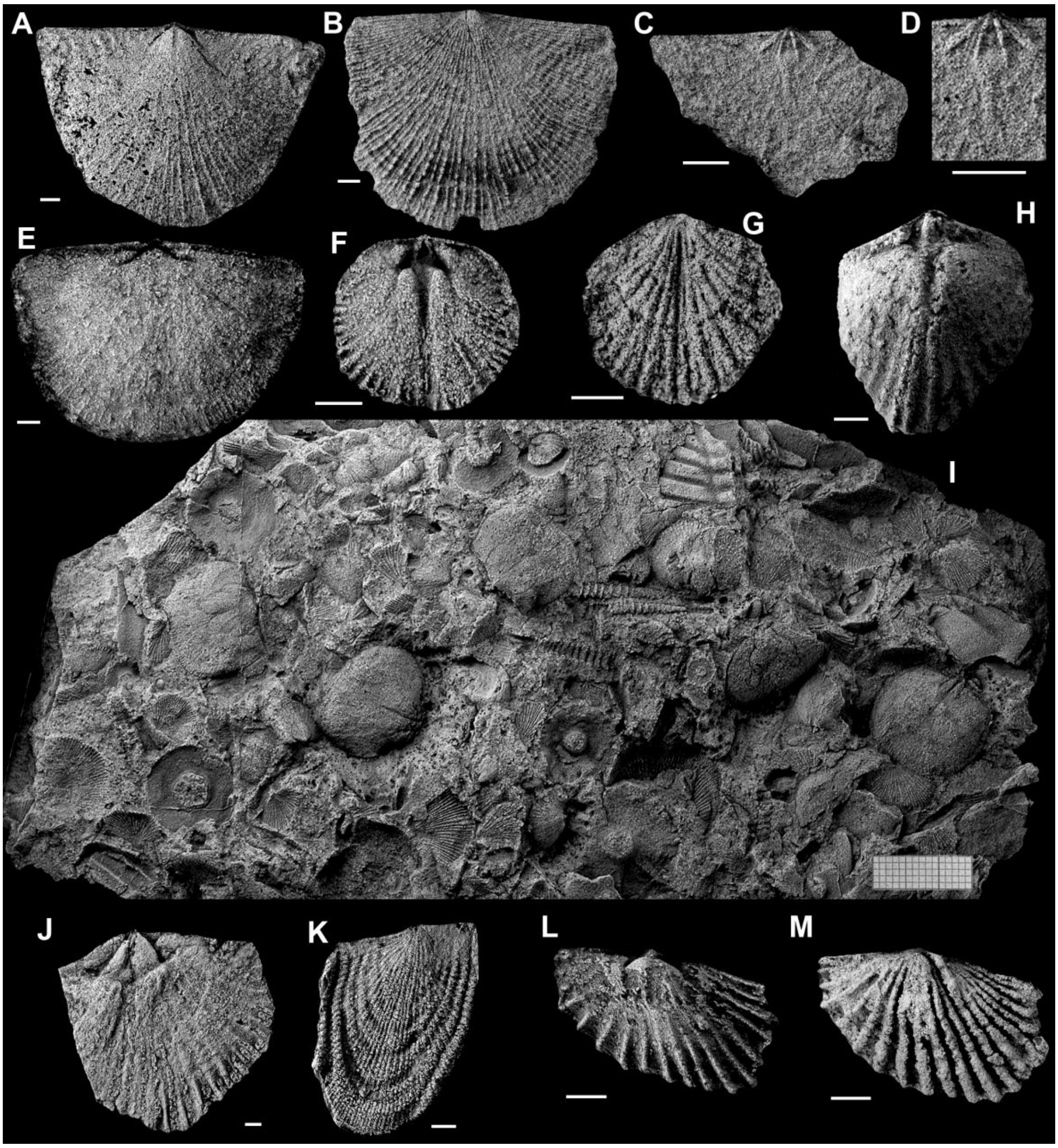


Figure 24. Representative fossils from the Gascagne Formation (A-I), Katian (Ka1-Ka2) in age, and from the Montjoi Formation (J-M), Katian (Ka2-Ka4) in age. A-D. *Kjaerina* (*Kjaerina*) *gondwanensis*, (A) internal mould of a ventral valve, (B) latex cast of exterior of a ventral valve. C-D. Latex cast of internal mould of a dorsal valve (C) with detail of cardinalia (D) of an incomplete specimen. E. *Iberomena sardoa*, internal mould of a dorsal valve. F-G. *Portranella exornata*, (F) internal mould of a dorsal valve, (G) latex cast of exterior of a dorsal valve. H. *Rostricellula termieri*, internal mould of a dorsal valve. I. General aspect of a fossiliferous level showing several valves of the brachiopods *Svobodaina*? sp., *Drabovia* sp., a pygidium of the trilobite *Calymenella* cf. *boisseli*, "*Cornulites*" sp. and cystoid columnar plates. J. *Dolerorthis* sp., internal mould of an incomplete dorsal valve. K. *Kjaerina* (*Villasina*) sp., latex cast of exterior of a ventral valve. L-M. *Nicolella actoniae*, internal mould (L) and latex cast of exterior of a ventral valve (M) of a small specimen. Scale bars = 2 mm.

The Montjoi Formation represents the widespread development of bryozoan-echinoderm meadows in mid-latitude settings of northern Gondwana. It has yielded a rich echinoderm association composed of *Aonodiscus spinosus*, *Caryocrinites crassus*, *C. elongatus*, *C. major*, *C. rugatus*, *Conspectocrinus celticus*, *Corylocrinus europaeus,* *Cyclocharax paucicrenellatus*, *Heliocrinites helmackeri*, *H. rouvillei*, *Mespilocystites* cf. *lemenni*, *M*. cf. *tregervanicus*, *Ristnacrinus cirrifer* and *Trigonocyclicus* cf. *vajgatschensis* (Touzeau *et al*., 2012).

Two brachiopod associations have been identified by Álvaro *et al.* (2016) (Fig. 24J-M): the Montjoi-LeMoulin (this stop) and Termes associations. The former includes *Nicolella actoniae* and *Dolerorthis* sp., and the latter *Kjaerina* (*Villasina*) sp., *Dolerorthis* sp., and Rafinesquinidae indet. The content of both associations, including some characteristic elements of the *Nicolella* Community, allows a correlation with laterally equivalent Mediterranean bioclastic limestones and dolostones, such as the Gabian Formation in the Montagne Noire (Colmenar *et al*., 2013), the Uggwa Limestone and Wolayer formations in the Carnic Alps of Austria (Schönlaub, 1998), the Estana Formation in the Spanish Central Pyrenees (Gil-Peña *et al*., 2004), the Cystoid Limestone in the Iberian Chains (Villas, 1985), the Portixeddu and Domusnovas formations in Sardinia (Leone *et al*., 1991), the Rosan Formation in the Armorican Massif (Mélou, 1987), the Upper Djeffara Formation in Libya (Buttler & Massa, 1996), and the upper part of the Porto de Santa Anna Formation in the Portuguese Central Iberian Zone (Colmenar, 2015). According to the correlation of Mediterranean formations bearing the *Nicolella* Community with the Pushgillian-to-Rawtheyan stages of the British scale (for a discussion, see Villas *et al*., 2002), the Montjoi Formation can be dated as Katian, stage slices Ka2-Ka4. 3.2.4.

Marmairane Formation

Although the above-reported Ashgill “schistes troués” were traditionally described and mapped as directly overlain by Silurian black shales, in some areas yielding graptolites of the Llandovery-Wenlock transition (Ovtracht, 1967; Centène & Sentou, 1975; Berger *et al*., 1997), the lower part of these supposed “Silurian black shales” is, in fact, represented by another distinct shaly unit. This formation, up to 10 m thick, consists of green shales with rare centimetric dolostone nodules and sandy interbeds. In the parautochthon, this unit becomes a diamictite (unsorted sandy shale with hematite cement). Its name derives from a creek linking the Lacamp plateau and the Evêque Forest. The formation conformably overlies the Montjoi Formation in the parautochthon and is either conformably or paraconformably overlain by the Silurian black shales in the parautochthon and the Félines-Palairac slice. Its stratotype lies along the homonymous creek (N43°0′2.38″, E2°40′8.91″).

In the parautochthon, some diamictitic beds are exposed along the road D212. They are formed by unsorted, sandy channel bodies that incised erosively into the underlying Montjoi Formation. Fossils have not been found in this bed and hence its age is constrained by lithostratigraphic correlation with laterally equivalent deposits of the Marmairane Formation.

STOP 8b

Villerouge, Gascagne and Marmairane formations along Marmairane creek, Félines-Palairac slice

J. Javier Álvaro, Jorge Colmenar, Eric Monceret and Daniel Vizcaïno

The Marmairane creek is accessible via two paths: from the north with a 4x4 and the south (after 20 minutes’ walk). We will visit these exposures from the south.

Villerouge Formation

The outcrops that contain the statotype of the Villerouge Formation are exposed in the vicinity of Villerouge-Termenès and cover part of the Lacamp Plateau and the Evêque Forest (Félines-Palairac slice). Its selected stratotype lies along the Marmairane creek (N43°00′14.13″, E02°40′07.28″), east of Villerouge-Termenès. The formation is ca. 50 m thick and consists of lava flows and volcanic breccias, interbedded with variegated shales and subsidiary litharenites. The base of the Villerouge Formation is marked by an erosive unconformity, and its top by the volcaniclastic conglomeratic lag that forms the basal part of the overlying Gascagne Formation. The top is characterized by a distinct level, about 4-8 m thick, composed of poorly sorted, gravel-to-boulder, highly weathered mafic and shaly fragments, subrounded to angular in shape and embedded in a heterolithic volcaniclastic matrix. Stratification is crude to absent, bases are inversely graded, and the fabric is dominantly matrix-supported . Silicification of clasts is pervasive. The presence of relic feldspar microlites and mafic clasts, altered and rich in iron oxyhydroxide cements, suggests the matrix was originally tuffaceous. This unit is interpreted as volcanic subaerial pyroclastic and debris flows (lahars) where the presence of conspicuous silicified clasts indicates cannibalization of the Lower Ordovician, shale-dominated, basement. Some acidic interbeds and dykes occur crosscutting the above-reported mafic assemblage.

Durand-Delga & Gèze (1956) included this volcanic unit in the Upper Ordovician-Silurian succession, whereas Berger *et al*. (1997) associated it with the Lower Ordovician acidic episodes of the area. A radiometric age is necessary to solve this stratigraphic puzzle, although we consider them as capping the Sardic unconformity: both the Villerouge and Roque de Bandies formations form the basal infill of the Upper Ordovician troughs recognized in the Mouthoumet massif and the Cabrières klippes, their bases are erosive unconformities, and their volcaniclastic content is exclusively recognized as reworked counterparts in the overlying Upper Ordovician strata.

The Roque de Bandies (Cabrières klippes) and Villerouge (Félines-Palairac slice) volcanosedimentary complexes are dominated by mafic lava and pyroclastic flows and laharic mudflows interbedded with shales. Geochemical affinity is tholeiitic and the volcanic products originated from melting of mantle and crust lithosphere. Volcanism is associated with fault-controlled subsidence and the generation of structurally controlled depocentres (grabens and half-grabens). After the end of volcanism, marine sedimentation was established leading to the sealing of previous (uplift and eroded) palaeoreliefs (Álvaro *et al*., 2016).

Gascagne Formation

This fossiliferous sandstone-dominated unit ranges from about 100 m thick in Laroque de Fa (Serre de Quintillan slice) to 5 m thick in the Félines-Palairac slice. It was previously known as "Caradoc sandstones" and consists of basal volcaniclastic conglomerates and lags covered by litharenite-to-arkose lenses and beds with shale interbeds increasing in thickness upsection. Its stratotype lies at the homonymous hill (N42°57′19.39″, E2°34′32.19″), east of Laroque de Fa. The lower part of the formation, directly overlying the Villerouge Formation, consists of polymictic (volcaniclastic-dominated) conglomerates and breccias, reaching up to 8 m in thickness. Clast size ranges from sand to boulder. They are chaotically oriented and display a clast-supported texture. Only locally, they are supported in a shale matrix suggesting local transport by debris flows.

A sampling from the upper sandstone/shale alternations of the formation in the Gascagne area has yielded, in order of abundance: the brachiopods *Portranella exornata*, *Kjaerina* (*K.*) *gondwanensis*, *Svobodaina*? sp., *Iberomena sardoa*, *Drabovia* sp., *Rostricellula termieri*, *Tafilaltia* sp., *Hirnantia* sp., and Strophomenidae indet; the trilobites *Calymenella* cf. *boisseli*, *Dalmanitina* sp., *Dreyfussina* sp., and *Onnia* sp.; and undetermined gastropods "*Cornulites*" sp. and cystoid columnar plates (Fig. 24A-I). Considering the brachiopod assemblage, the upper (volcanosedimentary-free) part of the Gascagne Formation can be correlated with the Glauzy Formation of the Montagne Noire (Colmenar *et al*., 2013), the lower part of the Porto de Santa Ana Formation in Buçaco, Portugal, the uppermost beds of the

“Bancos Mixtos” in Central Iberia (Villas, 1995), the upper half of the Fombuena Formation in the Iberian Chains, NE Spain (Villas, 1985), the Cava Formation in the Spanish Central Pyrenees (Gil-Peña *et al*., 2004), the base of the Portixeddu Formation in Sardinia (Leone *et al*., 1991), the Uggwa Shale Formation in the Carnic Alps, Austria (Havlíček *et al*., 1987), and the Upper Shale Member of the Bedinan Formation in Turkey (Villas *et al.*, 1995). Based on these correlations and the dating of the basal part of the Porto de Santa Anna Formation in Portugal, dated as Pusgillian by means of chitinozoans (Paris, 1979, 1981) and acritarchs (Elaouad-Debbaj, 1978), the Gascagne Formation can be dated as Katian, Ka1-Ka2 stage slices (see Bergström *et al*., 2009).

Marmairane Formation

The diamictite-free, offshore-dominated shales of the Marmairane Formation have yielded a new Hirnantian fauna of brachiopods and trilobites (Fig. 25). Five associations have been identified, four from Villerouge-Termenès (associations 1-4) and one from Gascagne (association 5). They are listed below, being the brachiopods ordered in descending abundance in each association: (1) Association 1 comprises *Hindella crassa incipiens*, *Plectothyrella crassicosta* ssp., and Dalmanellidae indet.; (2) Association 2 includes the brachiopods *Leptaena trifidum*, *Eostropheodonta hirnantensis*, Platyorthidae indet., *Glyptorthis sp*., Dalmanellidae indet, *Plectothyrella crassicosta* ssp., *Paucicrura* sp., *Rostricellula*? sp., *Orbiculoidea*? sp., and an indeterminate orthid; the trilobites *Mucronaspis* sp., *Flexycalymene* sp., and *Lichas* sp.; and undeterminated gastropods, dacryoconarids, and massive and ramose bryozoans; (3) Association 3 is represented by the brachiopods *Paucicrura* sp., Dalmanellidae indet, *Plectothyrella crassicosta* ssp., *Leptaena trifidum*, *Eostropheodonta* sp., Orthidae indet, and *Orbiculoidea*? sp.; the trilobites *Flexycalymene* sp. and Dalmanitidae gen. et sp. indet.; and undeterminated gastropods; dacryoconarids, and massive and ramose bryozoans; (4) Association 4 includes the brachiopods *Hindella crassa incipiens*, *Leptaena trifidum*, *Eostropheodonta* sp., Dalmanellidae indet, *Leangella* sp., and *Plectothyrella*? sp.; the trilobite *Flexycalymene* sp.; and undeterminated gastropods, dacryoconarids, and massive and ramose bryozoans; and (5) Association 5 comprises the brachiopods *Kinnella kielanae* (very abundant), *Hirnantia sagittifera*, *Eostropheodonta* sp., Dalmanellidae indet., *Dalmanella testudinaria*, *Plectothyrella crassicosta* ssp., Hindella crassa incipiens, *Leptaena trifidum*, *Drabovinella* sp., *Destombesium*? sp., and *Howellites*? sp.; the trilobite *Mucronaspis* cf. *mucronata*, and undetermined crinoid columnar plates and ramose bryozoans. The fifth association occurs in homogeneous black-to-grey shales representative of lower offshore clayey substrates, whereas the former ones were found in green shales with centimetre-thick sandy storm-induced interbeds and scattered carbonate nodules reflecting upper offshore clayey-dominated substrates.

Most of the recorded brachiopods are typical elements of the *Hirnantia* Fauna (Temple, 1965; Rong & Harper, 1988), such as *Hirnantia sagittifera*, *Hindella crassa incipiens*, *Leptaena trifidum*, *Eostropheodonta hirnantensis*, *Dalmanella testudinaria*, and *Plectothyrella crassicosta*. Based on this fossil content, the Marmairane Formation can be correlated with classical Hirnantian formations, such as the Hirnant Formation in Wales (Temple, 1965), the Kosov Formation in Bohemia (Marek & Havlíček, 1967), the Loka Formation in Sweden (Bergström, 1968), the Kuanyingchiao Beds (Rong, 1979) and the Langøyene Formation in Norway (Bergström & Cocks, 1982), among others. Based on this correlation, the Marmairane Formation can be dated as Hirnantian.

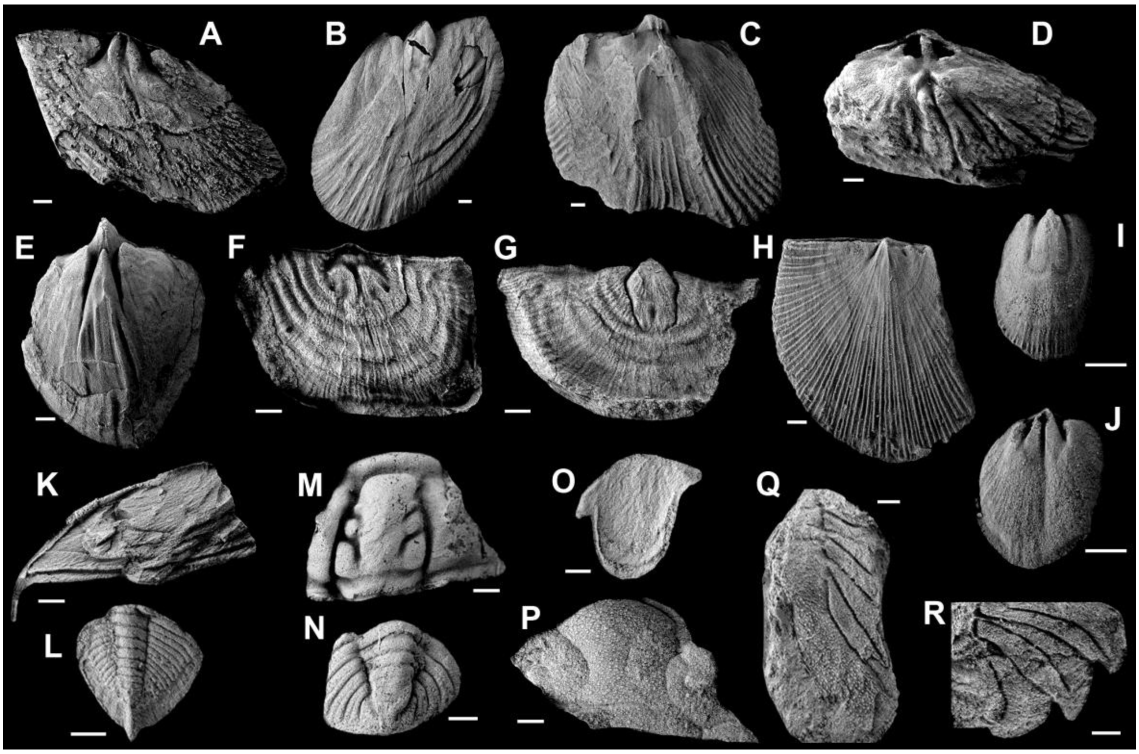


Figure 25. Brachiopods and trilobites from the Hirnantian Marmairane Formation. A-B. *Hirnantia sagittifera*, (A) internal mould of a dorsal valve and (B) internal mould of a ventral valve. C-D. *Plectothyrella crassicosta* ssp., (C) internal mould of a ventral valve and (D) internal mould of a dorsal valve. E. *Hindella crassa incipiens*, internal mould of a ventral valve. F-G. *Leptaena trifidum*, (F) internal mould of a dorsal valve and (G) internal mould of a ventral valve. H. *Eostropheodonta hirnantensis*, internal mould of a ventral valve. I-J. *Kinnella kielanae*, (I) internal mould of a ventral valve and (J) internal mould of a dorsal valve. K–L. *Mucronaspis* cf. *mucronata*, partial cephalon and pygidium, internal moulds. M-N. *Flexicalymene* sp., cranidium and pygidium, external moulds. O. Dalmanitid hypostome, internal mould. P-R. *Lichas* sp., cephalon (latext cast of external mould) and pygidia (internal moulds). Scale bars = 2 mm.

3.4- 9th September – Fieldtrip to Montagne Noire massif, Occitanie

The French Massif Central broadly consists of a stack of ductile and synmetamorphic nappes built during Devonian and Carboniferous times (Faure *et al*., 1997, 2005). Conversely to the northern areas, the eo-Variscan (Devonian) events are lacking in the southernmost part of the French Massif Central, or Montagne Noire. This area underwent its first deformation during the Mid Carboniferous (Visean to early Namurian) times.

Since Gèze (1949), the Montagne Noire is geologically subdivided, from south to north, into (i) a southern flank with imbricated Palaeozoic strata, including a Visean flysch; (ii) the Axial Zone made of metamorphic rocks, migmatites and plutonic rocks, globally arranged in a bulk domal shape; and (iii) a northern flank composed of Cambrian-to-Silurian strata. Both flanks contain low-grade metamorphic sedimentary sequences involved in kilometre-scale, late Visean south-verging recumbent folds and thrust sheets (Fig. 26).

Emplacement of the recumbent folds is stratigraphically dated of Late Visean to Namurian by the syntectonic foreland sedimentary basin in which they emplaced (Engel *et al*., 1980; Feist & Galtier, 1985).

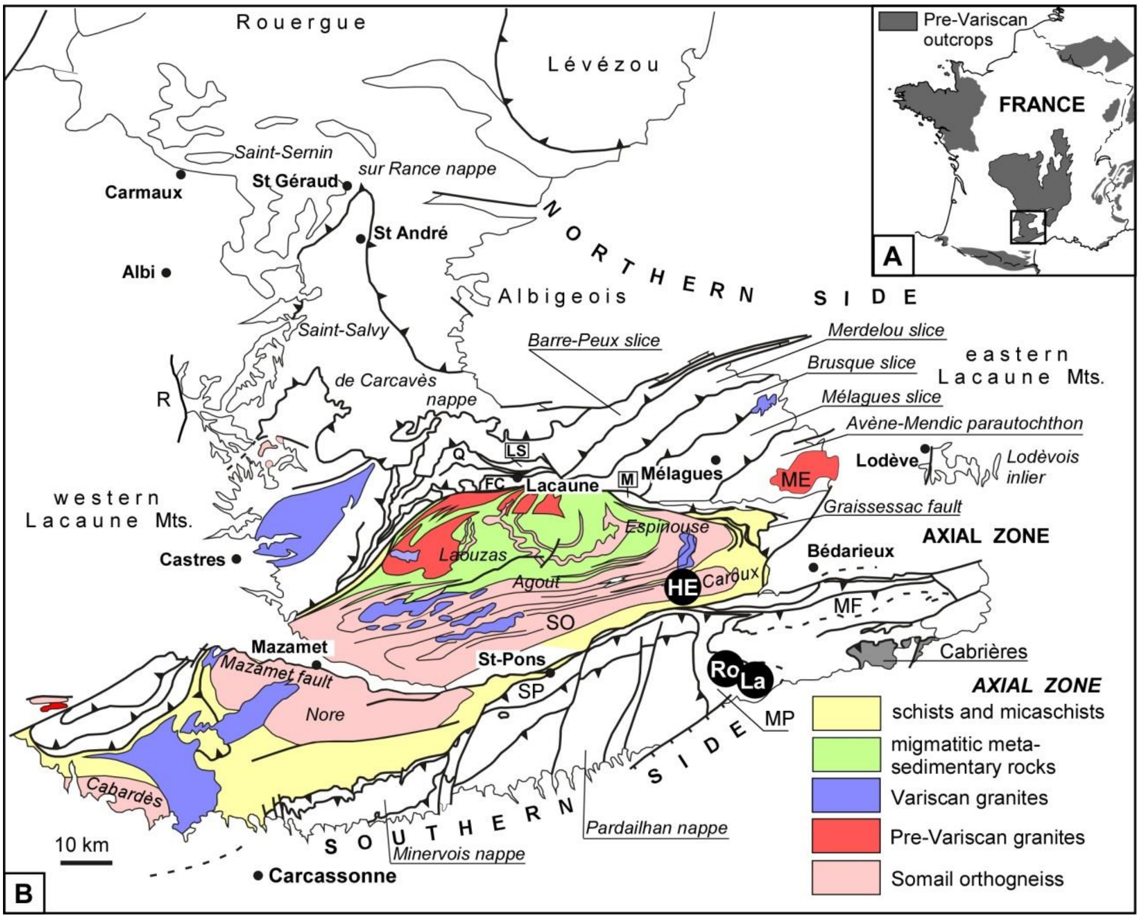


Figure 26. Geological sketch of the Montagne Noire with setting of main tectonostratigraphic units reported in the text; FC Farayruc-Calmejanne nappe, HE Gorges d'Héric (STOP 9c), La Landeyran stratotype (STOP 9b) LS La Sagne nappe, M Murat nappe, ME Mendic granite, MF Mount Faugères nappe, MP Mont Peyroux nappe, R Réalmont, Ro Roquebrun (STOP 9a), Q La Quille nappe, SP Saint-Ponais nappe; modified from Guérangé-Lozes (1987) and Álvaro *et al*. (2014b).

Axial Zone

The Axial Zone of the Montagne Noire (Gèze, 1949) is a ENE-WSW trending gneiss dome made of four principal lithological units: (i) schists and micaschists, (ii) migmatitic orthogneisses, (iii) metapelitic metatexites, and (iv) diatexites and granites (Rabin *et al*., 2015) (Fig. 27):

(i) The schist and micaschist units envelop the Axial Zone and are predominantly dominated by metapelitic packages showing a progressive increase in metamorphism from greenschist facies (sericite-chlorite-bearing schist) to amphibolite conditions and andalousite-cordierite-bearing micaschist toward the dome core (Thompson & Bard, 1982).

(ii) The migmatitic orthogneiss, also known as the Somail Orthogneiss, makes up the southern part of the Axial Zone, from the eastern Caroux to the western Cabardès domes (Fig. 26). The Somail orthogneiss is an Ordovician metagranite with large K-feldspar phenocrysts that was emplaced at 471 ± 4 Ma and 450 ± 6 Ma (U-Pb zircon ages) within a sedimentary pile, traditionally known as “Schistes X”. The latter, formally named St Pons-Cabardès Group, consists of schists, grauwackes, quartzites and subsidiary volcanic tuffs and carbonates (Demange, 1982, 1985, 1999; Demange *et al*., 1996; Alabouvette *et al*., 2003; Roger *et al*., 2004; Cocherie *et al*., 2005). The Somail orthogneiss is deformed and shows evidence of synkinematic partial melting.

(iii) In the deeper part of the dome, the so-called Laouzas metatexite was derived from a pelitic protolith.

(iv) In the Laouzas-Espinouse area, the metapelitic metatexites are accompanied by leucocratic diatexites corresponding to heterogeneous cordierite-bearing granites such as the Laouzas and Vialais granitoids. The age of their emplacement remains a matter of debate. U-Th-Pb EPMA chemical dating of monazite yielded an age for the Laouzas granitoid at 327 ± 7 Ma (Bé Mézème, 2005), while U-Pb LA-ICPMS gives an inherited age of 319 ± 2 Ma and an emplacement age at 299 ± 1 Ma (Poilvet *et al*., 2011). Two distinct monazite ages were also obtained in the Vialais granites: 327 ± 5 Ma (Matte *et al*., 1998) and 303 ± 4 Ma (Roger *et al*., 2015).

To the west, SW of Mazamet, the dome shape of the Axial Zone is well delineated by the geometry of the micaschist-gneiss contact, whereas, to the east, the Rosis micaschist synform subdivides the Axial Zone into the Espinouse and Caroux massifs (Figs. 26-27). The northeastern edge of the Axial Zone is bounded by the Graissessac fault, which represents the boundary of the Upper Carboniferous (Stephanian) Graissessac Basin (Bogdanoff *et al*., 1984; Echtler & Malavieille, 1990). East of Graissessac lies the Permian (Autunian) Lodève Basin. The southwestern edge of the Axial Zone, or Nore massif, is separated from the main part of the dome by the Mazamet Fault, a Neogene structure related to the Pyrenean Orogeny.

Despite the controversies on the Axial Zone dome emplacement, there is a large consensus for the relative chronology of the tectonic events, as follows: (i) recumbent folding and thrusting, (ii) doming, (iii) post-migmatitic granitoid emplacement, and (iv) opening and infill of the Graissessac and Lodève basins during widespread post-Variscan extensional pulses.

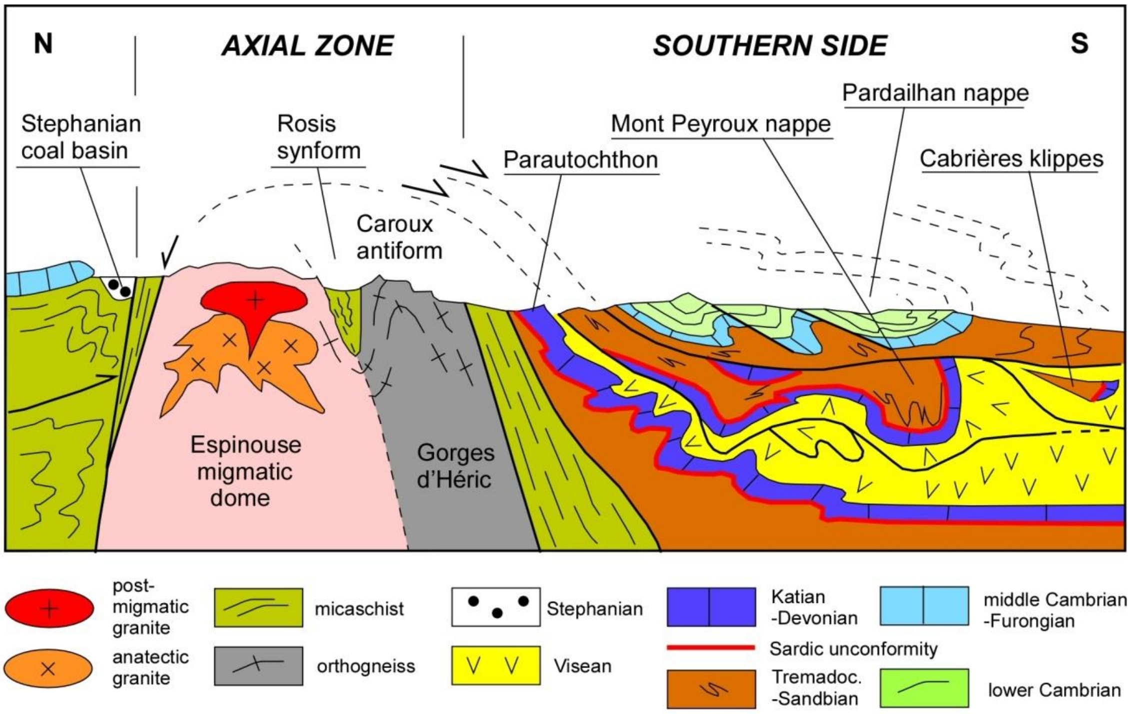


Figure 27. Synthetic cross-section of the Montagne Noire with location of the Sardic unconformity; modified from Engel *et al*. (1980), Faure *et al*. (2004) and Álvaro *et al*. (2016).

Southern Montagne Noire

The southern Montagne Noire, exposed as south-facing thrust slices and nappes, comprises a complete and fossiliferous, Cambrian to Carboniferous succession that represents the type area for many litho- and biostratigraphic units (Vizcaïno & Álvaro, 2001, 2003).

The Lower Ordovician sedimentary rocks of the southern Montagne Noire have been traditionally subdivided into lithostratigraphic units, alternating between sandstone- and shale-dominant intervals, whose contacts commonly coincided with biostratigraphic boundaries (Thoral, 1941; Dean, 1966; Courtessole *et al*., 1981, 1985). Vizcaïno *et al*. (2001) and Vizcaïno & Álvaro (2003) revised both scales and proposed two independent litho- and biostratigraphic subdivisions, which are followed below. Four Lower Ordovician formations are described below, from bottom to top, the La Maurerie, Cluse de l'Orb, Foulon and Landeyran formations. The Tremadocian-Floian boundary is tentatively located within the underlying St-Chinian Formation (Vizcaïno & Álvaro, 2003).

STOP 9a

Lower Ordovician La Maurerie, Cluse de l'Orb and Foulon formations, Mont Peyroux nappe

J. Javier Álvaro, Eric Monceret and Daniel Vizcaïno

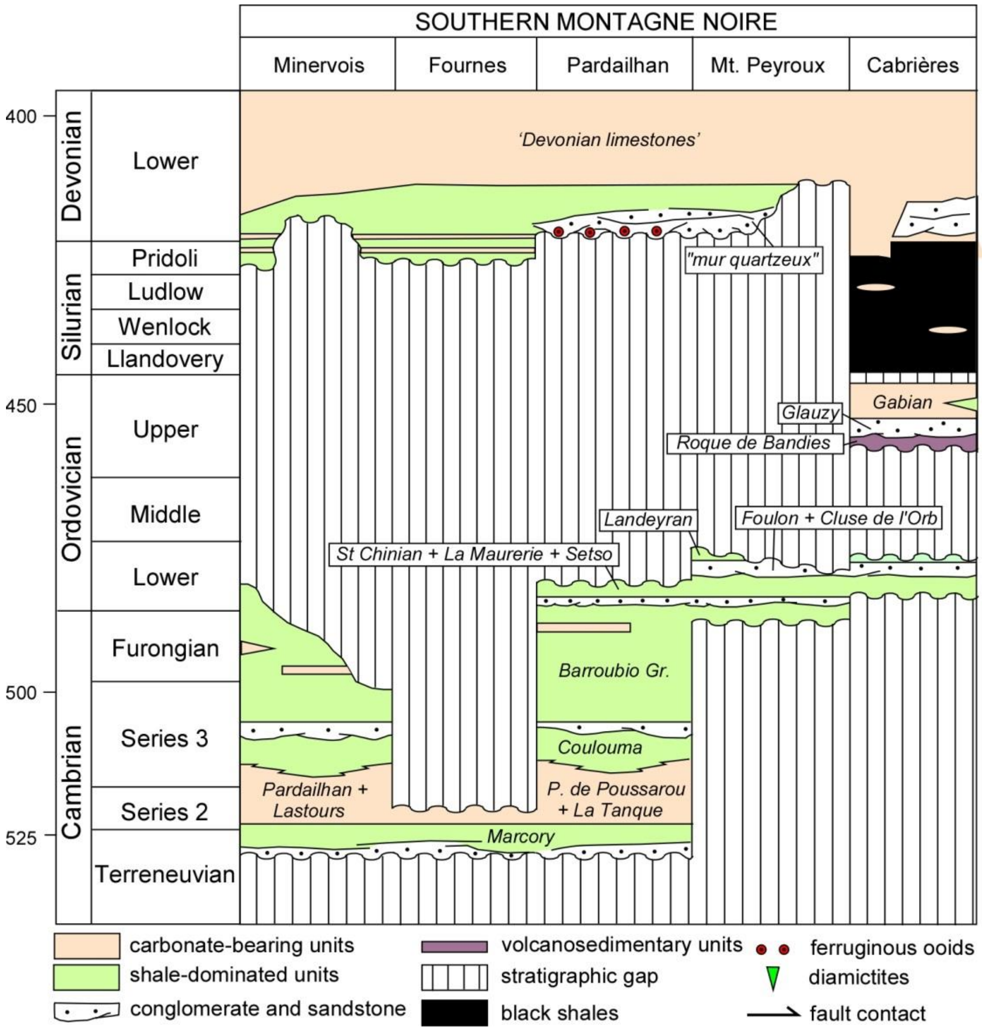


Figure 28. Stratigraphic framework of the Cambrian-Devonian strata and volcanic complexes in the southern Montagne Noire; modified from Álvaro *et al*. (2016).

La Maurerie Formation

The La Maurerie Formation (after the village of the same name) is an alternation of sandstones and sandy shales, more than 900 m thick, bearing centimetric (fossiliferous) siliceous nodules. The uppermost part is distinguished as the Setso Shales (after the Setso stream, SE of Roquebrun), a succession of homogeneous black to grey shales, up to 30 m thick, proposed by Thoral (1935) as a member of the La Maurerie Formation.

The shale interbeds of the La Maurerie Formation have yielded graptolites (such as *Clonograptus*, *Dictyonema* and *Temnograptus*), trilobites (*Ampyx*, *Ampyxinella*, Asaphidae indet, *Asaphopsoides*, *Megistaspis*, *Ogygiocaris*?, *Paramegalaspis*?, *Platycalymene*, *Symphysurus*, and *Taihungshania*), echinoderms (*Anatifopsis*, *Balantiocystis*, *Balanocystites*, *Cheirocystella* and *Lingulocystis*), gastropods (*Carcassonnella*, *Lesueurilla*, *Pararaphistoma*, *Sinuites* and *Thoralispira*), hyoliths (*Circotheca*), cephalopods (six species of the unformally defined '*Orthoceras*'), conocardioids (*Eopteria*), monoplacophorans (*Archinacella*) and bivalves (*Babinka*, *Coxiconcha*, *Ekaterodonta* and *Redonia*) (Capéra *et al*., 1975, 1978; Courtessole *et al*., 1981, 1983, 1985; Babin *et al*., 1982).

One trilobite-based zone has been identified in the La Maurerie Formation, named the *Taihungshania miqueli* Zone (former Courtessole's G and H "faunizones"), subsequently subdivided into the *Asaphelina barroisi berardi* + *T. miqueli* and the *T. miqueli* acme subzones

(Fig. 29). The base of the zone is located at the top of the underlying St.-Chinian Formation and its top at the basal part of the overlying Setso Member (Vizcaïno & Álvaro, 2003).

Cluse de l'Orb Formation

The Cluse de l'Orb Formation (named after the Orb river) has been traditionally recognized as the lingulid-rich Armorican Quartzite counterpart. However, the assignation to the Armorican facies was rejected after evidencing the sharp diachroneity of both formations. The presence of the brachiopod species *Ectenoglossa lesueuri* was one of the best biostratigraphically significant levels of correlation with the British and Armorican "Arenig" at the beginning of the 20th century. The formation, 150-220 m thick, consists of an alternation of white quartzites/grey sandstones and shales, bearing some siliceous fossiliferous nodules and glauconitic remains (Dabard & Chauvel, 1991), commonly encrusted by a wide diversity of microbially induced sedimentary structures (Noffke & Nitsch, 1994) and representing deposition on a storm-dominated platform (Eschard *in* Courtessole *et al*., 1985). Both lithologies are relatively rich, respectively, in brachiopods (genera *Ectenoglossa*, *Lingulepis* and *Lingulobolus*) and trilobites (genera *Asaphellus*, *Colpocoryphe*, *Ogyginus*, *Paramegalaspis*, *Platycalymene*, *Pradoella* and *Taihungshania*). Other taxa found in the formation are echinoderms (*Lingulocystis*), gastropods (*Thoralispira*) and bivalves (*Coxiconcha* and *Ekaterodonta*) (Rauscher, 1968, 1971; Capéra *et al*., 1975, 1978; Havlíček, 1980; Courtessole *et al*., 1981, 1983, 1985; Babin *et al*., 1982).

Two trilobite-based zones are recognized in the Cluse de l'Orb Formation, the lower *Taihungshania shui landeyranensis* and the *Colpocoryphe maynardensis* zones (former Courtessole's "faunizones" I and J, respectively), the top of the latter marking the basal part of the overlying Foulon Formation (Vizcaïno & Álvaro, 2003).

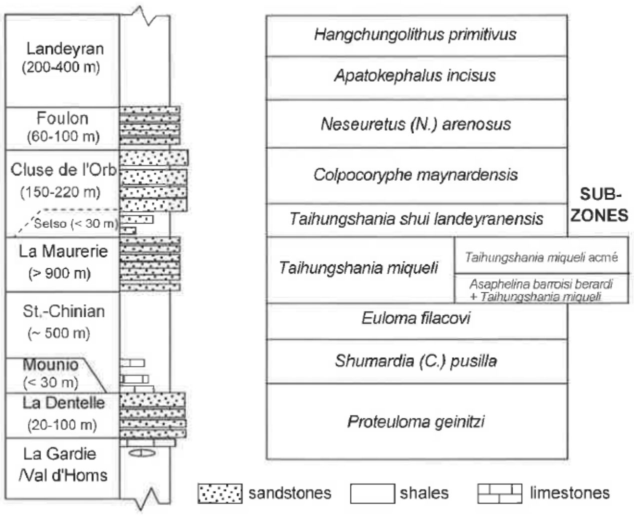


Figure 29. Lower Ordovician stratigraphic framework of the southern Montagne Noire; after Vizcaïno & Álvaro (2003).

Foulon Formation

The Foulon Formation (place-name found at the left bank of the Orb river) is 60-100 m thick and consists of fine-grained sandstones and sandy shales bearing limestone and siliceous nodules rich in fossil fauna and glauconitic remains. The fossil assemblage of the formation is rich, and comprises trilobites (genera *Asaphellus*, *Bathycheilus*, *Hungioides*, *Merlinia*, *Neseuretus*, *Niobella*, *Ogygiocaris* and *Salterocoryphe*?), brachiopods (*Rafanoglossa*), bivalves (*Coxiconcha*, *Cymatonota*?, *Ekaterodonta*, *Modiolopsis*, *Redonia* and *Thoralia*), rostroconchia (*Ribeira*), gastropods (*Thoralispira*), problematica (*Hanadirella* and *Solandangella*), echinoderms (*Lingulocystis*), and graptolites (*Didymograptus*) (Rauscher, 1968, 1971; Ubaghs, 1969, 1991, 1994; Capéra *et al*., 1975, 1978; Havlíček, 1980; Courtessole *et al*., 1981, 1983, 1985, 1991; Babin *et al*., 1982; Dean, 1986; Horný, 1994; Horný & Vizcaïno, 1995; Horný & Peel, 1996; Lefebvre & Vizcaïno, 1999; Vizcaïno & Lefebvre, 1999).

The base of the trilobite-based *Neseuretus* (*N*.) *arenosus* Zone (former Courtessole's "faunizone" K) is identified at the basal aprt of the Foulon Formation, whereas its top marks the basal part of the Landeyran Formation (Vizcaïno & Álvaro, 2003; Fig. 29).

STOP 9b

Lower Ordovician Landeyran Formation, Mont Peyroux nappe

J. Javier Álvaro, Sébastien Clausen, Eric Monceret and Daniel Vizcaïno

The Landeyran Formation (named after the homonymous valley) is 200-400 m thick and consists of brown, green and grey, homogeneous shales bearing siliceous nodules. It contains a rich and diverse fauna of trilobites (*Ampyx*, *Apatokephalus*, *Basiliella*, *Bathycheilus*, *Carolinites*, *Colpocoryphe*, *Courtessolium*, *Hoekaspis*?, *Euloma*, *Foulonia*, *Geragnostus*, *Hangchungolithus*, *Incisopyge*?, *Megistaspis*, *Neseuretus*, *Niobella*, *Ogyginus*, *Otarion*, *Platycalymene*, *Platycoryphe*, *Pliomerops*, *Pradoella*, *Prionocheilus*, *Proetidella*, *Selenopeltis*, *Taihungshania* and *Toletanaspis*), echinoderms (*Ampelocarpus*, *Anatifopsis*, *Balanocystites*, *Balantiocystis*, *Cheirocystella*, *Cothurnocystis*, *Lagynocystis*, *Lingulocystis*, *Lyricocarpus*, *Nanocarpus, Ovocarpus, Proscotiaecystis, Pyrgocystis*?, *Ramseyocrinus*, rhombiferans and *Thoralicystis*), gastropods (*Carcassonnella*, *Lesueurilla* and *Thoralispira*), brachiopods (*Aportophylla*, *Conotreta*, *Hesperonomia*, *Rafanoglossa*, *Ocorthis*, *Orthambonites*, *Paralenorthis*, *Paurorthis*, *Prantlina*, *Progonambonites*, *Ranorthis*, *Sinorthis* and *Spondyglossella*), hyoliths (*Elegantilites* and *Nephroteca*), machaeridians (*Lepidocoleus* and *Plumulites*), conularids (*Eoconularia*), bivalves (*Babinka*, *Coxiconcha*, *Ekaterodonta*, *Redonia*, *Synek*, *Thoralia*), rostroconchia (*Ribeiria* and *Tolmochovia*), graptolites (*Didymograptus*, *Phyllograptus* and *Tetragraptus*), and ostracodes (in study) (Rauscher, 1968, 1971; Ubaghs, 1969, 1991, 1994; Capéra *et al*., 1975, 1978; Havlíček, 1980; Courtessole *et al*., 1981, 1983, 1985, 1991; Babin *et al*., 1982; Dean, 1986; Horný, 1994; Horný & Vizcaïno, 1995; Horný & Peel, 1996; Lefebvre & Vizcaïno, 1999; Vizcaïno & Lefebvre, 1999).

The Landeyran Formation contains two trilobite-based zones, the *Apatokephalus incisus* and the *Hangchungolithus primitivus* zones (former Courtessole's "faunizones" L and M, respectively) (Vizcaïno & Álvaro, 2003; Fig. 29). The top of the Landeyran Formation is marked by a major stratigraphic gap, which is unconformably overlain, in the Pardailhan nappe by the Devonian "mur quartzeux" (Quémart *et al*., 1993).

STOP 9c

Somail augen gneisses at Gorges d'Héric, Caroux Dome

J. Javier Álvaro

Although the augen gneisses that form the Axial Zone were traditionally considered as Precambrian (Cadomian-related) in age, new geochronological dating has been recently available for the various generations of granitoids that form the core of the Axial Zone (Cocherie, 2003; Roger *et al*., 2004; Bé Mézème, 2005; Charles *et al*., 2009). Augen orthogneiss yield zircon U-Pb ages of 471 ± 4 Ma, 456 ± 3 Ma and 450 ± 6 Ma interpreted as the magmatic age (Roger *et al*., 2004; Cocherie *et al*., 2005). These Ordovician granitoids intruded a sedimentary succession of grauwacke, schist, quartzite and rare marble, acidic volcanosedimentary layers known as the St Pons-Cabardès Group or “Schistes X” (Demange *et al*., 1996; Alabouvette *et al*., 2003). The Sériès tuffs that cap the group were dated to 545 ± 15 Ma by Lescuyer & Cocherie (1992); as a result, the Sériès tuff is a lateral equivalent of the Rivernous metarhyolitic Formation of the southern Montagne Noire (Lodevois inlier), dated to 542.5 ± 2.4 Ma and 539.2 ± 2.7 Ma (Padel *et al*., in press), and the acidic volcanic episode recorded in the Pic de la Clape Formation of the eastern Pyrenees (ranging from 542.9+5/-1.23 Ma to 532.9±6.9 Ma; Padel, 2016).

The core of the Espinouse dome consists of cordierite migmatites developed at the expense of augen orthogneiss and their host rocks. Metre to kilometre-sized HT gneiss septa are preserved (e.g. La Salvetat area). Toward the dome core, migmatites progressively grade into the Laouzas cordierite anatectic granite (Gèze, 1949). Monazite from the Laouzas granite yields U-Th-Pb chemical age of 333 ± 6 Ma (Bé Mézème, 2005). The migmatite age obtained in several places by monazite U-Th-Pb chemical method ranges between 327 and 322 Ma (Bé Mézème, 2005). The northwestern part of the Axial Zone is occupied by the Montalet pluton: although this peraluminous granite was considered as an early body (Demange *et al*., 1996; Alabouvette *et al*., 2003), syn-magmatic monazites in this granite yielded U-Th-Pb chemical ages at 327 ± 7 Ma and 333 ± 4 Ma (Bé Mézème, 2005; Charles *et al*., 2008).

Several small-sized, late to post-migmatitic delimited plutons intrude the Axial Zone. They are either biotite-rich granite (e.g. Anglès granite) or muscovite-biotite ± garnet leucogranites (e.g. Soulié, Vialais granites). Age of these plutons, determined by U-Th- Pb chemical method on monazite, is comprised between 324 and 318 Ma (Bé Mézème, 2005; Charles *et al*., 2008). Lastly, several generations of garnet-tourmaline pegmatite and aplite dykes and sills intrude all the above-described rocks (Collot, 1980). Therefore, from rim to core, the Axial Zone appears as formed by concentric envelopes of micaschist, augen gneiss, migmatites, and anatectic granite.

STOP 9d

Roque de Bandies, Glauzy and Gabian formations, Cabrières klippes

Jorge Colmenar, Daniel Vizcaïno and J. Javier Álvaro

Roque de Bandies Formation

The volcanic activity of the Roque de Bandies Formation consists of lava and volcaniclastic flows and tephra deposits interbedded with volcanosedimentary sequences. Its geochemical composition and interpretation are documented in stop 8b.

Glauzy Formation

The formation, up to 50 m thick, consists of basal volcanosedimentary (litharenitic) conglomerates and conglomerate/shale alternations, grading upsection into subarkoses and greywackes rich in quartzite pebbles, bioclastic lags marking the base and laminae of trough cross-stratified sets, and iron-oxydroxide cements, conspicuous as coatings and impregnations. Glaucarenites are locally abundant. The lower part represents reworking of the underlying Roque de Bandies Formation, grading upsection into the onset of shoreface-dominated substrates.

Some basal shale interbeds have yielded a poorly preserved acritarch association with *Priscogalea striatula* and *Striatotheca principalis parva* (F. Martin op. cit. Nysæther *et al*., 2002) of probable Mid-Ordovician (pre-“Caradoc”) age. However, the probable reworking of the Lower Ordovician acritarch-bearing basement (well constrained in the conglomerates and litharenites that characterize the Roque de Bandies/Glauzy transition) should be taken into consideration.

The bioclastic sandstones of the upper part of the formation have yielded, in order of descending abundance, the brachiopods *Proclinorthis vailhanensis*, *Kjaerina* (*K.*) *gondwanensis*, *Svobodaina havliceki*, *Longvillia mediterranea* and *Triplesia*? sp.; indeterminate crinoid columnar plates; the trilobites *Dreyfussina exophthalma*, *Dalmanitina* (*Dalmanitina*) *proaeva*, *Calymenella boisseli* and *Onnia grenieri*; the annelid *Cornulites* sp.; and undetermined ramose bryozoans. All these brachiopods belong to the *Svobodaina havliceki* Community (Colmenar *et al*., 2013), a recurrent peri-Gondwanan benthic association characterized by a relatively low diversity and mainly composed by endemic genera.

Considering the brachiopod content (Fig. 30), the uppermost beds of the Glauzy Formation can be correlated with the base of the Porto de Santa Ana Formation in Buçaco, Portugal; the uppermost beds of the “Bancos Mixtos” in Central Iberia (Villas, 1995); the upper half of Fombuena Formation in the Iberian Chains, Spain (Villas, 1985); the Cava Formation in the Spanish Central Pyrenees (Gil-Peña *et al*., 2004); the base of the Portixeddu Formation in Sardinia (Leone *et al*. 1991); the Uggwa Shale Formation in the Carnic Alps of Austria (Havlíček *et al*., 1987); and the Upper Shale Member of the Bedinan Formation in Turkey (Villas *et al*., 1995). Except for the Portuguese and Austrian formations, these correlations are strengthened by the co-occurrence in the same units of the trilobites *Dreyfussina exophtalma* and *Calymenella boisseli* (see Hammann & Leone, 2007).

Based on the correlations described above and the dating of the basal horizons of the Porto de Santa Anna Formation in Portugal as Pusgillian, by means of chitinozoans (Paris, 1979, 1981) and acritarchs (Elaouad-Debbaj, 1978), the uppermost beds of the Glauzy Formation can be dated as Katian, Ka1-Ka2 stage slice (see Bergström *et al*., 2009). Because no fossils are known from the lower half of the Glauzy Formation, no correlation can be attempted for its lower part.

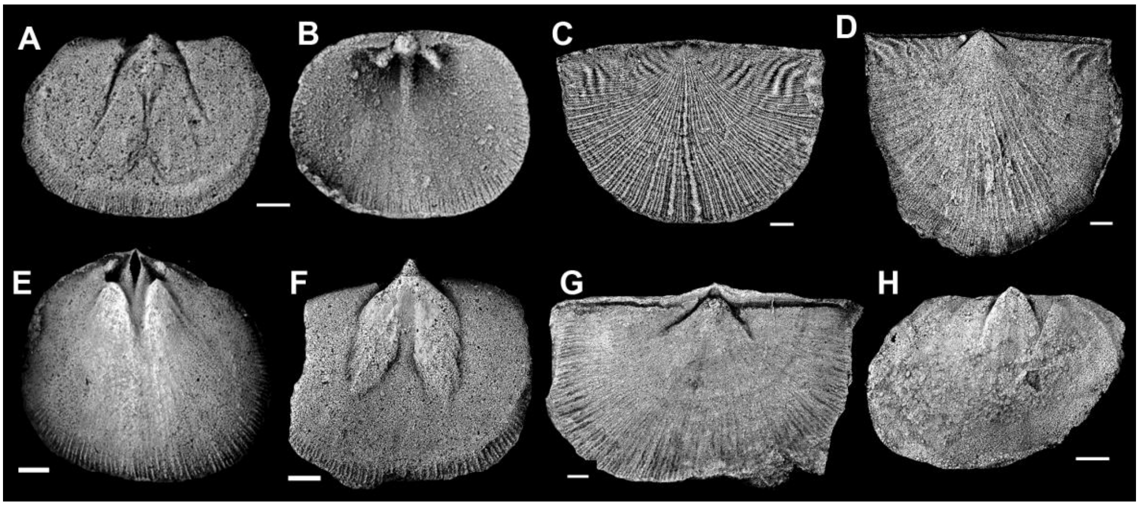


Figure 30. Representative brachiopod taxa from the Glauzy Formation: A-B. *Proclinorthis vailhanensis*, (A) internal mould of ventral valve and (B) latex cast of interior of dorsal valve. C-D. *Kjaerina* (*K.*) *gondwanensis*, (C) latex cast of exterior of ventral valve and (B) internal mould of ventral valve. E-F. *Svobodaina havliceki*, (E) internal mould of dorsal valve and (F) internal mould of ventral valve. G. *Longvillia mediterranea*, internal mould of a ventral valve. H. *Triplesia* sp., internal mould of a ventral valve. Scale bars = 2 mm.

Gabian Formation

This unit, 10-25 m thick, consists of floatstone-dominated limestones, marlstones and shales bearing carbonate nodules rich in bryozoans, cystoids and subsidiary brachiopods and trilobites. The formation represents the development of bryozoan-echinoderm meadows similar to other "lower-middle Asghill" carbonate units associated with the Boda event (Villas *et al*., 2002; Fortey & Cocks, 2005).

The Gabian Formation have provided an abundant number of brachiopod taxa (Fig. 31) such as *Nicolella actoniae*, *N. decemcostata*, *Nicolorthis instans*, *Eridorthis angusticulata*, *Dolerorthis* cf. *maxima*, *Tafilaltia trifida*, *Tissintia glauzensis*, *Heterorthis* cf. *alternata*, *Svobodaina feisti*, *S. barnumosa*, *Dalmanella rostrata*, *Portranella exornata, Drabovia* cf. *latior, Protomendacella profuga*, *Triplesia cymbula*, *Leptestiina meloui*, *Leangella anaclyta, Aegiromena meneghiniana*, *Kozlowskites*? sp., *Iberomena sardoa*, *Longvillia mediterranea*, *Rafinesquina* (*Mesogeina*) *gabianensis*, *R.* (*M.*)? sp., *Kjaerina* (*Kjaerina*) *gondwanensis, K.* (*Villasina*) *pedronaensis*, *Leptaena* sp. B, *Porambonites* (*Porambonites*) *magnus*, *P.* (*P.*) *dreyfussi* and *Eoanastrophia pentamera*. Dreyfuss (1948) described the trilobite *Dalmania socialis* var. *proaeva* and listed a total of 8 cystoid species. Babin *et al*. (1988) recorded some remains of the conodont *Amorphognathus ordovicicus*. Ernst & Key (2007) described 68 species of bryozoans from museum collections.

The fossil content of this formation rich in elements of the brachiopod *Nicolella* Community, allows a direct correlation with other Upper Ordovician Mediterranean bioclastic limestones and dolostones, such as the Uggwa Limestone and Wolayer formations from the Carnic Alps, Austria (Schönlaub, 1998), the Estana Formation from the Spanish Central Pyrenees (Gil-Peña *et al*., 2004), the Cystoid Limestone from the Iberian Chains (Villas, 1985), the Portixeddu and Domusnovas formations from Sardinia (Leone *et al*., 1991), the Rosan Formation from the Armorican Massif (Mélou, 1987), and the Upper Djeffara Formation from Libya (Buttler & Massa, 1996).

According to the correlation of the Mediterranean formations bearing the *Nicolella* Community with the Pushgillian to Rawtheyan stages of the British scale (see Villas *et al*., 2002), the new formation can be dated as Katian (stage slices Ka2-Ka4) on the global scale. The finding of *Amorphognatus ordovicicus* within the unit (Babin *et al*., 1988) strengthens the correlation with the upper Katian (see Bergström *et al*., 2009).

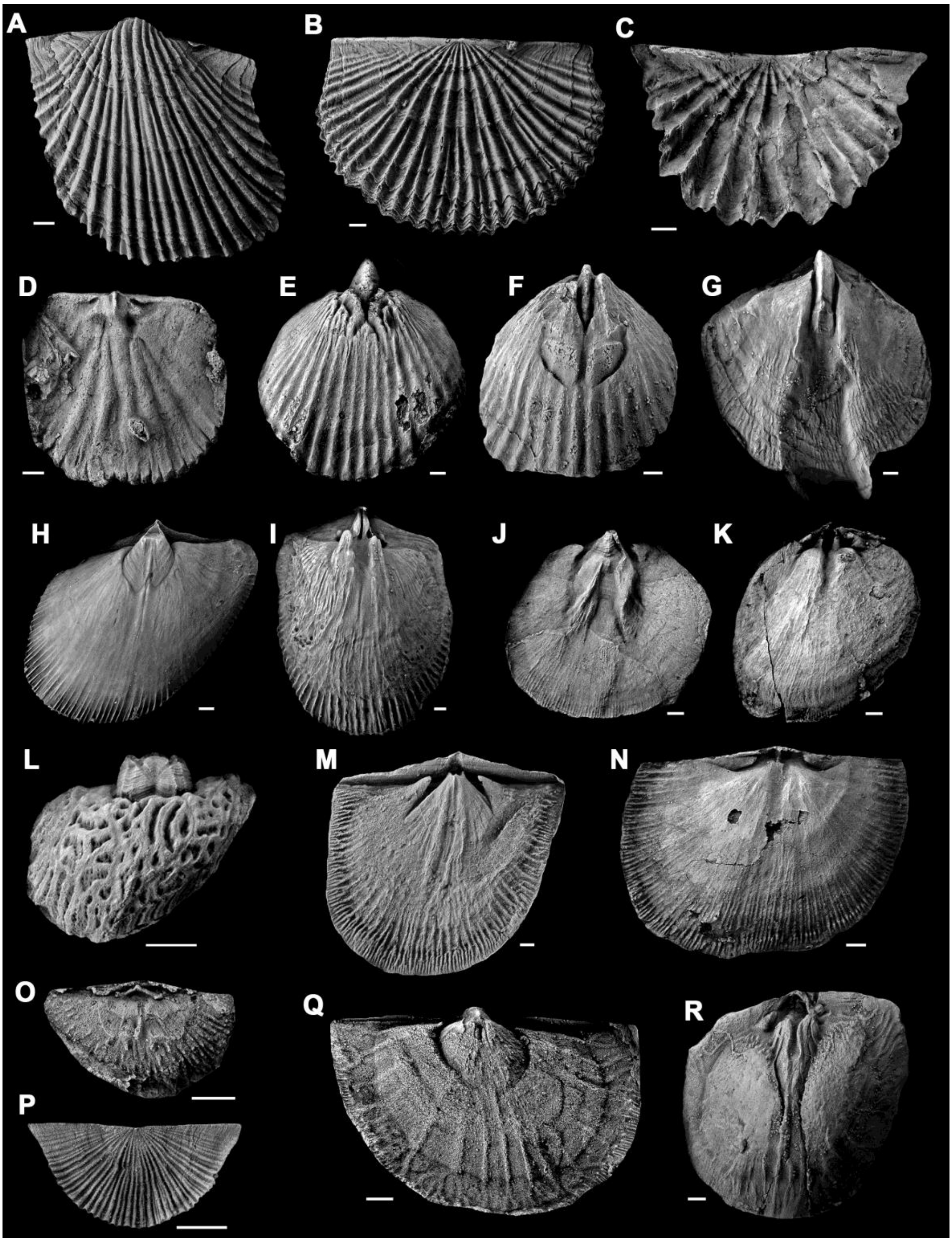


Figure 31. Representative brachiopod taxa from the Gabian Formation: A-B, D) *Nicolella actoniae*, (A) latex cast of exterior of a ventral, (B) latex cast of exterior of a dorsal valve and (D) latex cast of interior of a dorsal valve. C. *Nicolella decemcostata*, latex cast of exterior of dorsal valve. E-F. *Eoanastrophia pentamera*, (E) internal mould of ventral valve, (F) internal mould of dorsal valve. G. *Porambonites* (*Porambonites*) *magnus* (Meneghini, 1880), internal mould of ventral valve () from the Estana Formation (Pyrenees). H-I) *Dolerorthis maxima*, (H) internal mould of ventral valve and (I) internal mould of dorsal valve. J-K. *Svobodaina feisti*, (J) internal mould of ventral valve and (K) internal mould of dorsal valve. L. *Leangella anaclyta,* internal mould of ventral valve. M-N. *Longvillia mediterranea*, (M) internal mould of ventral valve and (N)

internal mould of dorsal valve. O-P. *Aegiromena meneghiniana,* (O) latex cast interior and (P) latex cast of exterior of dorsal valve. Q-R. *Iberomena sardoa*, (Q) internal mould of ventral valve and (R) internal mould of dorsal valve. Scale bars = 2 mm.

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