Low-grade metamorphism in the Montagne Noire (S-France) : Conodont Alteration Index (CAI) in Palaeozoic carbonates and implications for the exhumation of a hot metamorphic core complex

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Low-grade metamorphism in the Montagne Noire (S-France): Conodont Alteration Index (CAI) in Palaeozoic carbonates and implications for the exhumation of a hot metamorphic core complex

by *Ulrich Wiederer*[1], *Peter Königshof*[1], *Raimund Feist*[2], *Wolfgang Franke*[3] and *Michael P. Doublier*[4]

Abstract

The Montagne Noire is a metamorphic core complex situated in the southern foreland of the Variscan belt in South France. The core of the structure (Zone Axiale) exposes LP/HT gneisses metamorphosed in Carboniferous time. Exhumation of the hot gneisses has imposed amphibolite facies to low-grade metamorphism on the Palaeozoic rocks in the mantle of the Zone Axiale. Individual stratigraphic units can be traced without major changes in sedimentary facies across the metamorphic zonation, from diagenetic to epizonal grades. Therefore, the area is ideally suited for a comparison between different methods of assessing lower metamorphic grades. The conodont alteration index (CAI) analysed in c. 300 samples shows that CAI is not dependant on stratigraphic age, and is not significantly altered by secondary dolomitization. Locally increased CAI values are attributed to alteration by fluids. In samples with CAI ≥ 5, conodont elements show ductile deformation. Comparisons with an earlier set of crystallinity (IC) data (uncalibrated values of Engel et al., 1981) reveal a non-linear correlation between CAI and IC, with IC rapidly decreasing in samples with CAI ≤ 3. IC values between 0.23 and 0.46, which grossly correspond to the boundaries of the anchizone for Kübler indices, correlate with CAI of c. 4 to 3.

Although the earlier IC do not permit a satisfactory correlation with CAI, they clearly reflect the same metamorphic zonation and help to reveal the relationships between deformation and metamorphism. Metamorphism decreases away from the Zone Axiale and the metamorphic zonation cuts across the inverted limbs of recumbent fold nappes. Hence, metamorphism must have been acquired after crustal stacking, and represent dynamic contact metamorphism imposed by the rising, hot crystalline core. Accordingly, tectonic structures in areas of high CAI reveal the same extensional kinematics as in the Zone Axiale. The relatively narrow, present-day contours of the Zone Axiale are probably due to cooling and constriction of the extensional window.

Keywords: conodont alteration index (CAI), illite crystallinity (IC), metamorphic core complex, Variscides, Montagne Noire.

1. Introduction

The Montagne Noire is situated on the southern flank of the Variscan Belt in South France (e.g., Matte, 1991). It contains a low pressure metamorphic core with anatectic gneisses and granites, overlain and flanked by Palaeozoic metasediments (Fig. 1). The low metamorphic grade of the Palaeozoic cover and the presence of Carboniferous flysch sediments indicate an external position within the orogen, comparable with, e.g., the external massifs of the western Alps. Astonishingly, the high-grade core has yielded Carboniferous metamorphic and magmatic ages. The geodynamic context of this "hot external massif" is still a matter of controversy.

[1] Forschungsinstitut und Naturmuseum Senckenberg, Senckenberganlage 25, D-60325 Frankfurt a. Main. <againhe@web.de> and <pkoenigs@sng.uni-frankfurt.de>

[2] Institut des Sciences de l'Évolution, Lab. de Paléontologie, Université de Montpellier II, Place E. Bataillon, F-34095 Montpellier Cedex 05. <rfeist@isem.univ.montp2.fr>

[3] Institut für Geowissenschaften der Universität, Senckenbergstr. 3, D-35390 Gießen. <wolfgang.franke@geolo.uni-giessen.de>

[4] Institut für Geologie und Paläontologie der Universität, Senckenberg-Anlage 32, D-60325 Frankfurt a. Main. <Michael.P.Doublier@geolo.uni-giessen.de>

Apart from this geodynamic problem, the Montagne Noire offers ideal conditions for methodical metamorphic studies, since a laterally consistent sequence of Palaeozoic sediments (quartzites, greywackes, pelites, radiolarian cherts and carbonates) may be traced across the metamorphic zonation, from diagenesis into amphibolite facies. Earlier petrological studies have revealed a concentric zonation of low pressure metamorphism centred around the gneissic core (DEMANGE, 1985). We have started a multidisciplinary metamorphic study in the Palaeozoic mantle, in order to compare the records of different methods, and to provide additional constraints on the tectono-metamorphic evolution. This paper presents a survey of the conodont alteration index (CAI) in Devonian limestones, comparisons with an earlier study on illite crystallinity (IC), and tectonic consequences.

2. Geological setting

2.1. STATIGRAPHIC RECORD

The Palaeozoic sequence of the Montagne Noire starts with a thick pile of Early Cambrian carbonates and Middle Cambrian through Early Ordovician clastic sediments deposited at the N-Gondwana margin (e.g., PIQUÉ et al., 1994). In most of the area, Late Ordovician and Silurian deposits are missing. The Devonian oversteps the early Palaeozoic sequences with an angular unconformity of 10–20°, which might be interpreted to reflect an orogenic event. In fact, GEBAUER and GRÜNENFELDER (1976) have reported U–Pb ages on detrital zircon fractions and Rb–Sr whole rock ages from metamorphic rocks of the Zone Axiale spanning between 445 ± 10 Ma and 417 ± c. 35 Ma, which they take to indicate a "Caledonian" metamorphic event. However, the pre-Devonian rocks on the flanks of the Zone Axiale do not show a separate folding or cleavage. Besides, the "Écailles de Cabrières" of the southeastern Montagne Noire (see below) contain a stratigraphic sequence without any angular unconformity. There is biostratigraphic evidence for Early Ordovician, Late Ordovician and Silurian rocks (DREYFUSS, 1948), the latter being conformably overlain by the Devonian (FEIST and SCHÖNLAUB, 1974). On the northern flank of the Montagne Noire, the sedimentary sequence is restricted to the Cambrian through Early Ordovician with the exception of one isolated occurrence of fossiliferous Silurian rocks (CENTENE, 1977). These features rule out a major "Caledonian" orogenic event. The isotopic data cited above require confirmation by more reliable methods.

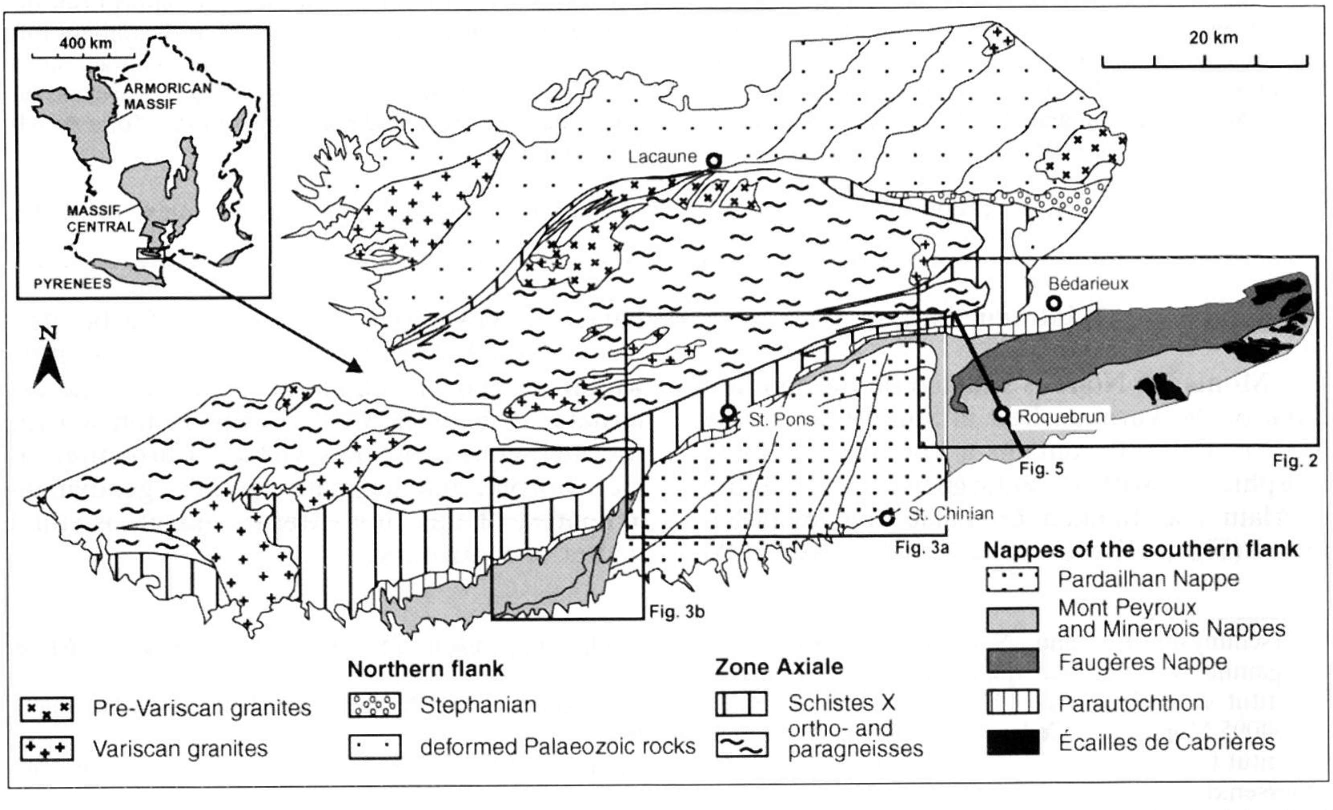

Fig. 1 Simplified geological map of the Montagne Noire, with insets showing the position of areas sampled for CAI (see Figs. 2 and 3).

Apart from a basal clastic member, the Devonian sequence is entirely composed of carbonates. Depositional environments show a clear trend of upward deepening, from sabkha-type early diagenetic dolomites in the Pragian towards hemipelagic nodular limestones in the Late Eifelian through to the basal Carboniferous (FEIST, 1985). These limestones are overlain by radiolarian cherts (Tournaisian to Mid-Viséan) and a thick sequence of synorogenic, deep water clastic sediments (flysch) of Late Viséan to Early Namurian age (ENGEL et al., 1981; FEIST and GALTIER, 1985).

2.2. TECTONO-METAMORPHIC UNITS

There are no constraints on the protolith ages of the paragneisses of the Zone Axiale. The widespread orthogneiss bodies have yielded Late Proterozoic (LÉVÊQUE, 1990) and Cambrian isotopic ages (HAMET and ALLÈGRE, 1972, 1973; DUCROT et al., 1979).

The main metamorphism in the Zone Axiale is characterized by low pressures and high temperatures up to anatectic grade, which overprint an earlier stage with higher pressures. The metamorphic zonation exhibits a concentric pattern surrounding the contours of the Zone Axiale (BARD and RAMBELOSON, 1973; THOMPSON and BARD, 1982; DEMANGE, 1985; OURZIK et al., 1991). Earlier, pressure dominated stages of the metamorphic evolution are indicated by local relics of eclogites and kyanite bearing metamorphic assemblages (DEMANGE, 1985). Ar–Ar dating on metamorphic muscovite and biotite has yielded ages of 316 ± 4, 308 ± 3 and 297 ± 3 Ma. The youngest ages occur at the northern and southern margins of the Zone Axiale (MALUSKI et al., 1991).

Late- to post-tectonic granites were intruded into the Zone Axiale between c. 330–320 Ma and into its northern flank around 290–280 Ma (Rb–Sr whole rock data of HAMET and ALLÈGRE, 1976). While U–Pb dating appears to confirm the earlier ages of intrusion into the Zone Axiale (327 ± 5 Ma, MATTE et al., 1998), they indicate younger ages for the granites on the northern flank (314 ± 8 Ma, LÉVÊQUE, 1986; 310 ± 2 Ma, MONIÉ unpubl., cited in SIMIEN et al., 1999).

The gneisses of the Zone Axiale are overlain by the "Schistes X", a variegated sequence of metasediments and few volcanic rocks, metamorphosed in amphibolite to greenschist facies. Subdivision of the Schistes X is based upon different protolith lithologies. U–Pb datings indicate the presence of Late Proterozoic meta-magmatic rocks (LESCUYER and COCHERIE, 1992). However, the Schistes X also contain lenses of fine-grained marbles (X_6 , X_9, , X_{10} and X_{15}), partly associated with black meta-cherts (X_9 and X_{15}, see BOGDANOFF et al., 1984), which remind the Tournaisian and Early Viséan lithologies of the less metamorphosed units further south (e.g., ENGEL et al., 1981).

The "Par-Autochthon", structurally above and to the S of the Schistes X, contains clearly identifiable Ordovician, Devonian and Carboniferous sequences. The repetition of conspicuous protolith lithologies in the Schistes X and in the Par-Autochthon suggests that these units do not simply represent basement and cover, but structurally deeper members of the nappe pile, which contain stratigraphic equivalents of the fossiliferous sediments in the overlying units.

The higher nappes consist of very low-grade metamorphic rocks and comprise (in order from bottom to top) the Faugères Nappe (Devonian and Carboniferous), the Mont Peyroux Nappe (Ordovician to Carboniferous), and the Pardailhan Nappe (Cambrian to Devonian, Figs. 1 and 2). The "Écailles de Cabrières" in the eastern part of the Mont Peyroux Nappe are a chaotic mass of olistoliths in a Carboniferous flysch matrix. The olistoliths consist of Early Ordovician through to Viséan rocks which are derived from the normal limb of the Mont Peyroux Nappe. They range up to 10 km^2 in size and several hundreds of metres in thickness. These miniature nappes were detached and emplaced by gravity slide during a late stage of the southward advance of the tectonic front (DE SITTER and TRÜMPY in GÉZE et al., 1952; ENGEL et al., 1978, 1981). The Minervois Nappe in the western Montagne Noire (Figs. 1 and 3b) is probably a structural equivalent of the Mont Peyroux Nappe.

2.3. TECTONO-METAMORPHIC EVOLUTION

During a first stage of the tectonometamorphic evolution, Palaeozoic sediments were deformed into large recumbent fold nappes, which were stacked by grossly southward-directed thrusting (ARTHAUD, 1970). It is possible that the kyanite-bearing assemblages reported by DEMANGE (1985) relate to this phase of crustal thickening. The formation of eclogites in a foreland position is very unlikely. Since there is no solid evidence for a major Caledonian event in the Montagne Noire, the eclogite relics (and possibly also the kyanite) probably belong to a pre-Cambrian (?Cadomian) orogenic cycle.

During the second stage, the present-day Zone Axiale was exhumed in the centre of the Montagne Noire. The dominant low pressure meta-

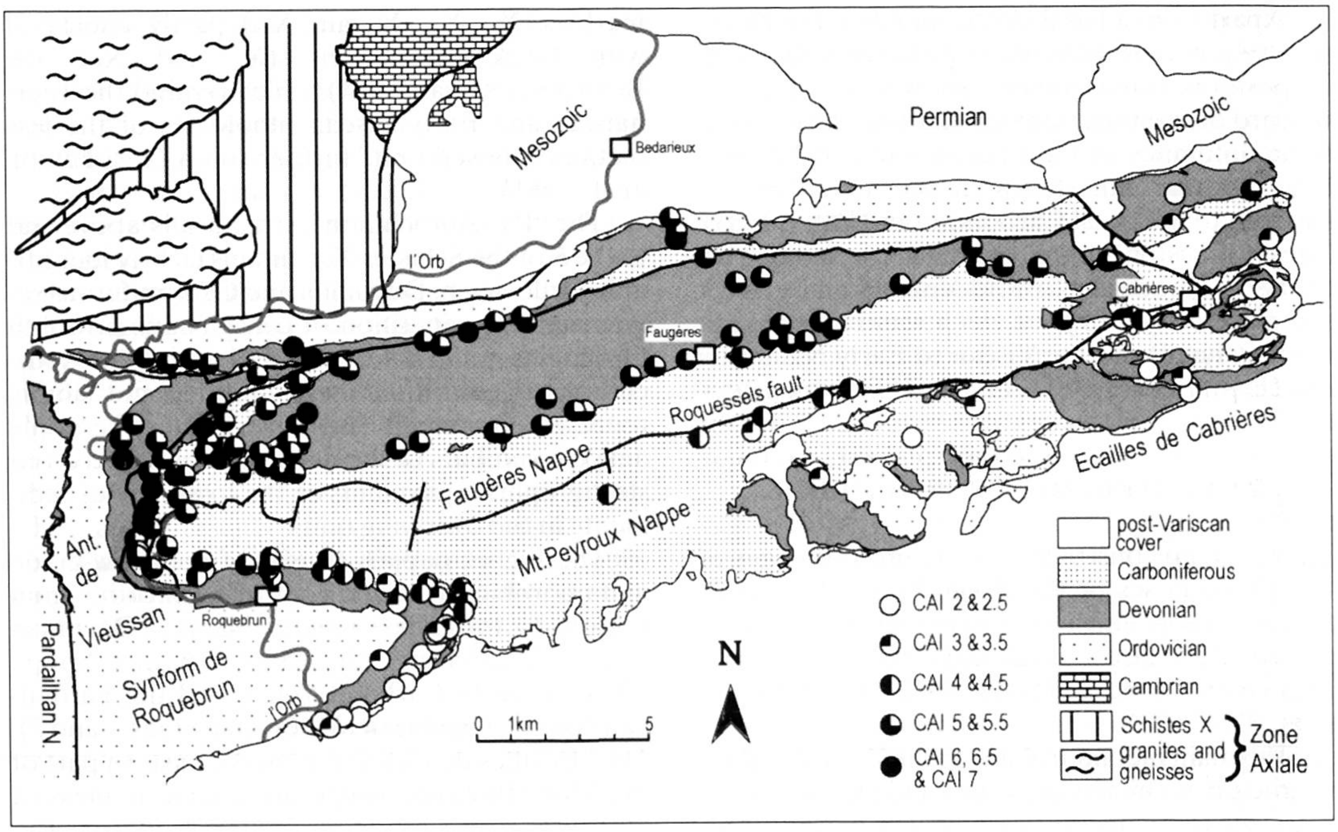

Fig. 2 Distribution of CAI in the southeastern Montagne Noire.

morphism occurred during this stage, and the metamorphic zonation was thinned by ongoing extension. Within only 1–2 km of tectonic thickness, gneisses with cordierite, sillimanite and K-feldspar were juxtaposed against identifiable Palaeozoic rocks of the Par-Autochthon. This dramatic reduction of the metamorphic sequence was accompanied by strike-slip faulting (e.g., ECHTLER, 1990). The southern flank of the Montagne Noire has undergone late refolding, which has created map-scale antiforms and synforms. Later still, dextral transtensional faulting juxtaposed the Zone Axiale against the Schistes X and the lower grade parts of the Palaeozoic cover. In this paper, we refer to the formation and stacking of recumbent fold nappes as D1, to the features relating to exhumation of the Zone Axiale as D2, and to the late refolding on the southern flank of the Montagne Noire as D3 (Roquebrun Synform, Vieussan Antiform, Figs. 2 and 5).

The primary thrust faults separating the fold nappes (D1) have nowhere survived. Segments of the former fold nappes are separated, today, by ductile to brittle extensional faults, partly with an important component of strike-slip. The contact between the Mont Peyroux and Faugères Nappes is an extensional detachment with displacement top to the E, refolded by F3 (DOUBLIER and FRANKE, unpublished observations). In the Vieussan Antiform, the fault separates the Devonian carbonates, and, further E, the Viséan flysch sequences of the Mont Peyroux and Faugères Nappes (Figs. 2 and 5). This eastern segment, the "Roquessels fault" (Figs. 2 and 5), was analyzed by ENGEL et al. (1981). It has been shown to reduce the zonation of IC, and must therefore postdate metamorphism. The contact between the Pardailhan and Mont Peyroux Nappe (Figs. 1, 2 and 5) is a westward dipping, brittle normal fault, which, toward the N, changes into a dextral transtensional fault parallel with the southern flank of the Zone Axiale. The fault postdates F3 (AERDEN, 1998; AERDEN and MALAVIEILLE, 1999).

The importance of extensional features has qualified the Montagne Noire, for several authors, as a prime example of "orogenic collapse", due to gravitational instability after tectonic thickening (e.g., VAN DEN DRIESSCHE and BRUN, 1992; AERDEN, 1998), possibly aided by strike-slip faulting (ECHTLER, 1990; ECHTLER and MALAVIEILLE, 1990). Alternatively, the Zone Axiale has been interpreted as a conical fold (MATTE et al., 1998), a transpressive feature (DEMANGE, 1999), as a diapiric uplift (e.g., FAURE and COTTEREAU, 1988; FAURE, 1995), or else by a combination of diapirism and extensional collapse (SOULA et al., 2001).

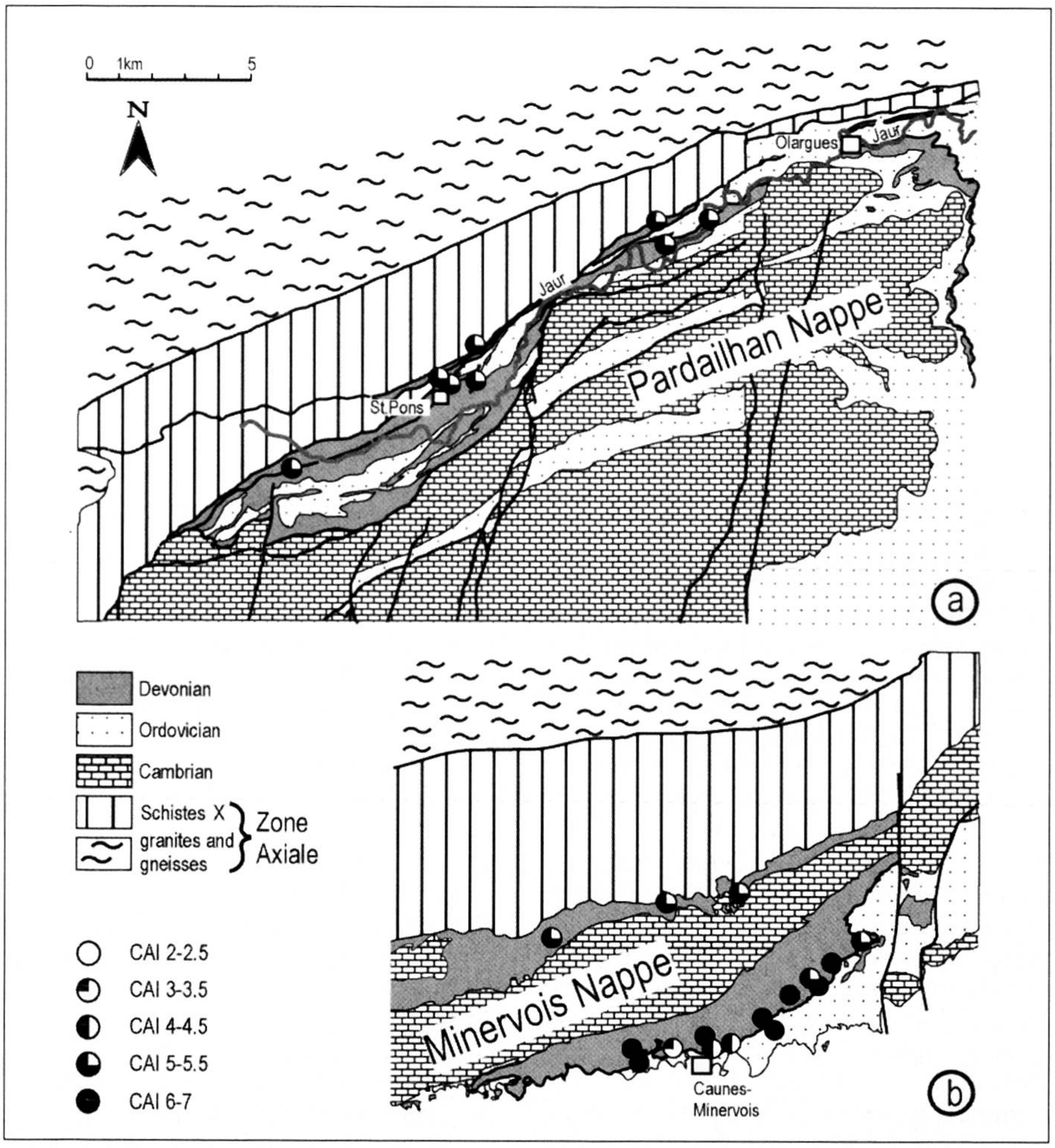

Fig. 3 (a) Distribution of CAI around St. Pons. (b) Distribution of CAI in the Minervois.

3. Conodont alteration index (CAI)

3.1. METHOD

Conodont elements are tooth-shaped, microscopic remains of the food-processing apparatus (CLARK et al., 1981) of marine, Late Cambrian to Late Triassic microfossils. Conodonts probably belong to the early chordates (e.g., BRIGGS et al., 1983; ALDRIDGE et al., 1987, 1993). Conodont elements are composed of calcium phospate. They generally range from 0.1 to 1 mm in length. A variety of morphological shapes occur within one "conodont apparatus". Continuous growth of the conodont elements throughout the lifetime of the conodont animal has produced an internal structure consisting of apatite lamellae separated by thin organic layers. Conodont elements are often abundant in marine carbonates and are relatively resistent to corrosion and recrystallization under diagenetic and low-grade metamorphic conditions. Because of their preservation potential, their generally rapid evolution and their pandemic distribution, conodonts have proven extremely useful as biostratigraphic index fossils.

Conodont elements are also valuable indicators of metamorphic grade. Colour alteration of conodonts was first noted by ELLISON (1944). However, it was not until the pioneer work of EPSTEIN et al. (1977) that these colour differences were placed in a systematic context. Using experimental pyrolysis and samples from the field, EPSTEIN et al. (1977) demonstrated that conodont elements visibly change colour, from pale yellow to light brown, dark brown, and black (conodont colour alteration indices, CAI 1–5) which is assigned to different temperature regimes during metamorphism. The change of colour is the result of coalification of minute amounts of organic matter sealed within the conodont elements. REJEBIAN et al. (1987) extended the range of the method to higher temperatures with CAI values of 5–8, which are characterized by colour changes from black over grey to opaque white and, finally, glassy. These higher grade changes result from carbon loss, release of structurally bound water, and recrystallisation. In order to derive metamorphic temperatures for natural rocks, the authors used an Arrhenius plot to extrapolate their experimental data to geologic time spans.

During the last decades, detailed CAI studies have been used to assess organic maturation and hydrocarbon potential in diagenetic to very low-grade metamorphic regimes. Examples of prospective hydrocarbon areas identified using conodont colour alteration data include the Canning Basin (NICOLL and GORTER, 1984) or eastern Canada (LEGALL et al., 1982). CAI maps have also been produced for parts of the Variscan basement in Europe (e.g., KÖNIGSHOF, 1992; HELSEN and KÖNIGSHOF, 1994; GARCIA-LOPEZ et al., 1997; SARMIENTO et al., 1999). Furthermore, conodont elements have also been used to determine thermal aureoles related to igneous (NICOLL, 1981; KÖNIGSHOF, 1991) and (or) hydrothermal processes (e.g., HARRIS et al., 1990), and to test the mineralization potential (e.g., WARDLAW and HARRIS, 1984). NÖTH et al. (1991) found a good correlation between CAI and vitrinite reflectance.

Conodont elements which have suffered from short-term, relatively high temperature events (e.g. contact metamorphism) do not necessarily follow the same gradational colour alteration sequence. Many of these conodonts are not uniformly altered, but only show superficial alteration such as a grey patina. In these cases, CAI values may vary considerably within a small area, within a narrow stratigraphical interval, within one single sample, or even within one conodont element. Conodont elements may be etched and/or significantly corroded by pore waters, saline waters or hydrothermal solutions. The corroding fluids may also oxidize the organic matter within the conodont elements, thereby producing grey to white colours which must not be mistaken as evidence for high metamorphic temperatures.

Samples were processed in 7.5% formic acid during a maximum of 48 hours. The insoluble residue containing the conodonts was wet-sieved (125 μm mesh). In most of the samples, conodonts were hand-picked directly from the insoluble residue. Some of the samples required concentration of conodonts by gravity settling in bromoform.

Colours of conodont elements were determined optically by comparison with a CAI reference set produced in the laboratory or from naturally altered samples (EPSTEIN et al., 1977; REJEBIAN et al., 1987). This permits in most cases the determination of CAI to a precision of half CAI steps. Most populations of conodont elements consist of different morphs with differing thickness, and reveal some variation in colour. Therefore, we have evaluated, in most samples, the CAI of at least 20 specimens.

3.2. DERIVATION OF SAMPLES

We have analyzed 294 limestone samples, including 73 conodont concentrates from localities in the "Écailles de Cabrières". Since, in this area, several samples were taken from each locality, these data are represented by only 21 data points in Fig. 2.

We have tried to analyze samples of uniform lithology. Most of the samples are hemipelagic, grey, nodular carbonate mudstones of Late Devonian to Early Carboniferous age. These have usually yielded abundant conodont faunas (up to 300 specimens/kg). Some biodetrital shelf limestones of Middle Devonian age were sampled in areas, where the nodular limestones have been excised by faulting, or as part of a sample sequence taken to test possible interrelationships between CAI and stratigraphic age. This latter group of samples covers a stratigraphic thickness of c. 500 m. Some samples of secondary dolomite were taken to test the influence of dolomitization.

Most of the productive samples come from a belt of Devonian to earliest Carboniferous carbonates, which, due to refolding by F3, define westward plunging, kilometric syn- and antiforms (Synforme de Roquebrun, Antiforme de Vieussan, Fig. 2). Eastwards, this belt is continued in the Faugères Nappe and terminates NW of Cabrières.

Three samples from the Schistes X did not yield any conodonts. In the area around St. Pons (Fig. 3a), the narrow northwesterly tract of Devonian carbonates belongs to the Par-Autochthon (4 samples). To the SW of Olargues, the southeasterly tract of Devonian rocks represents the basal part of the Pardailhan Nappe. This applies probably also to the southeastern carbonates around St. Pons, although attribution to the Mont Peyroux Nappe cannot be entirely ruled out. In this part of the Montagne Noire, the Devonian of the Mont Peyroux Nappe has mostly been excised by the post-D3 transtensional fault at the base of the Pardailhan Nappe.

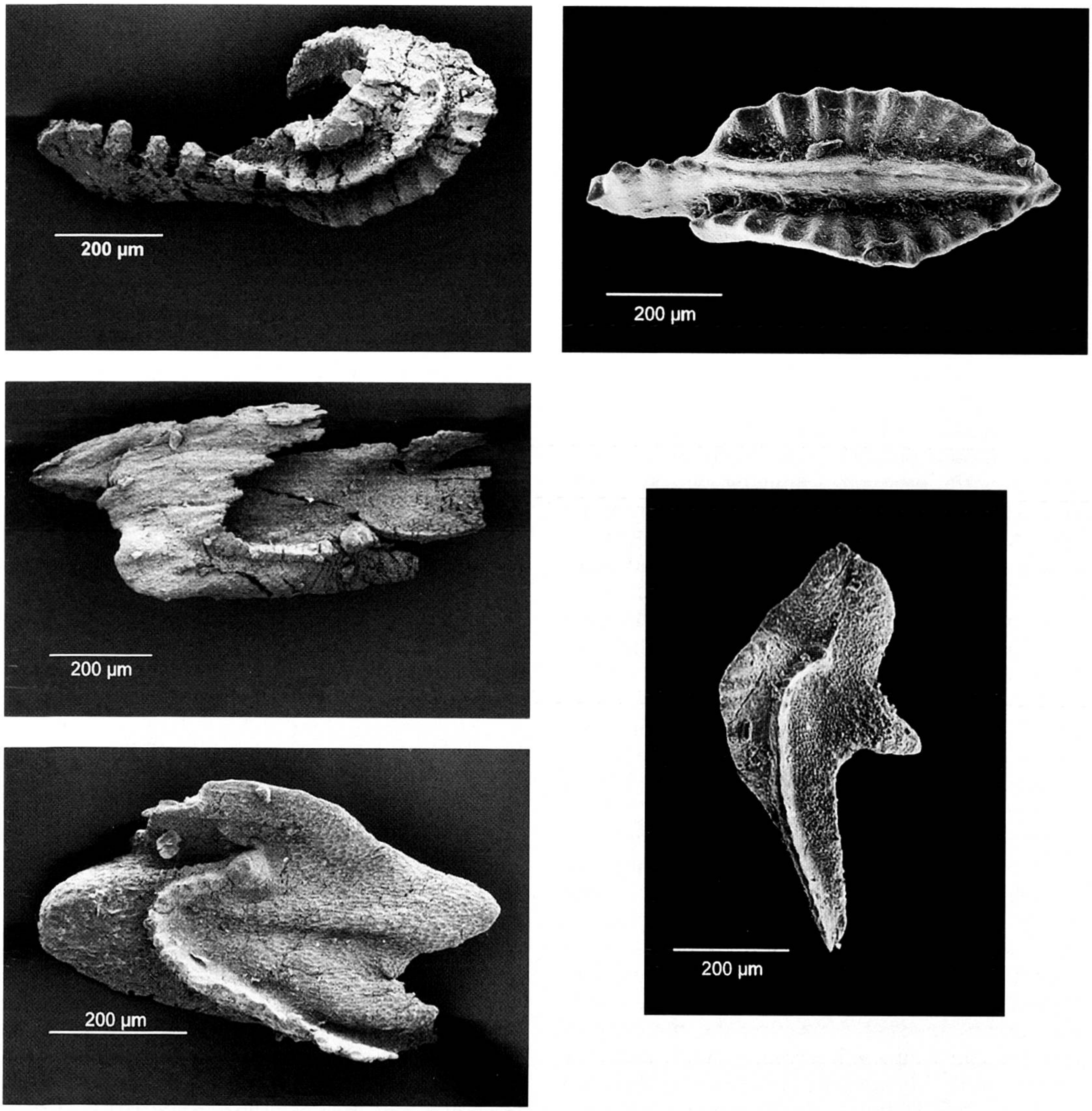

Fig. 4 Scanning electron micrographs of ductilely deformed conodonts from the Montagne Noire (sample nos. PMN...), compared with undeformed specimens of similar conodont elements from the Rhenish Massif (sample nos. Di...) Top left: *Polygnathus* sp. (CAI 5, PMN 109977). Top right: *Polygnathus pennatus* (CAI 4, Di-88). Left middle and bottom: *Palmatolepis* sp. (CAI 5, PMN 39953 and 1099110). Bottom right: *Palmatolepis perlobata schindewolfi* (CAI 3.5, Di-90).

In the southwestern Montagne Noire (Fig. 3b), outcrops of Devonian carbonates are limited. The northern carbonate belt in Fig. 3b probably belongs to the Par-Autochthon (ARTHAUD, 1970), although an assignment to one of the higher nappes can not be excluded (e.g., ALABOUVETTE et al., 1993). The southeastern group of samples in Fig. 3b belongs to the Minervois Nappe, which probably represents a westward continuation of the Mont Peyroux Nappe.

3.3. RESULTS

3.3.1. Areal distribution of CAI data

In order to test a possible influence of long-term, shallow burial under a growing sedimentary overburden, we have analyzed a series of 11 samples from the northern limb of the Roquebrun Synform. The samples span the Middle Devonian through to the Early Viséan and do not show any

Table 1 Metamorphic temperatures derived from CAI data for different durations of heating (after REJEBIAN et al., 1987).

CAI	temperature (°C) for 3 Ma	temperature (°C) for 10 Ma	temperature (°C) for 20 Ma	temperature (°C) for 30 Ma
2	80	70	70	65
3	150	145	140	135
4	225	215	212	210
5	345	330	323	325
5.5	395	365	362	360
6	420	405	402	400
6.5	500	470	475	478

significant differences.

The best areal coverage of CAI is achieved in the southeastern Montagne Noire (Fig. 2). Along the belt of outcrop of Devonian carbonates, which delineates the westward plunging syn- and antiforms (F3), CAI is seen to decrease away from the Zone Axiale, toward the S: in the Par-Autochthon as well as in the Faugères and Mont Peyroux Nappes in the Vieussan Antiform, CAI attains values between 5 and ≥ 6, and decreases in the Roquebrun Synform to 2.5–2. Toward the E, the Faugères Nappe retains CAI 5.5–5, whereas CAI in the Mont Peyroux-Cabrières Nappe decreases to 2.5–2. This eastward increasing difference of CAI between the adjacent nappes is caused by the Roquessels fault and a NW-trending fault NW of Cabrières.

In the more southwesterly parts of the Montagne Noire (Figs. 3a and b), samples adjacent to the Schistes X on the southern flank of the Zone Axiale show CAI of 5.5–5, like their structural equivalents to the NE (Fig. 2). CAI in samples from the southeastern part of the Minervois Nappe near Caunes-Minervois (Fig. 3b) partly goes down to 4.5–4. However, the majority of samples near Caunes show high CAI. Values may vary considerably not only between neighbouring localities (≥ 6–3), but also within one single sample (7–5).

Samples with exceptionally high CAI (7–5.5) also occur in the Vieussan Antiform (Fig. 2). In the southwestern as well as in the eastern Montage Noire, the conodonts with elevated CAI values display a sugary surface, are often corroded and show a grey patina. Apparently, these features do not affect the entire conodont elements, but are restricted to their surface. Some conodont elements reveal grey and white stripes parallel with brittle fractures, which suggests loss of organic matter with concomitant changes in colour.

The Devonian limestones have often undergone secondary, post-tectonic dolomitization, and minor amounts of dolomite in the samples may go unnoticed. Since MARCH-BENLLOCH and SANTISTEBAN (1993) and HELSEN (1995) have found abnormally high CAI values in dolomitized limestones, we have analyzed several limestone/dolomite sample couplets. In most dolomite samples, we did not find conodont elements at all. In one case, the CAI in the dolomite (sample no. 39904, CAI 5) turned out to be one grade higher than that of the neighbouring limestone (sample no. 39905, CAI 4). In four other sample couplets, dolomites and limestones yielded identical values (CAI 5 or 5.5). We have therefore included the dolomitized samples into the metamorphic maps.

3.3.2. Deformation of conodonts

It is a well known feature that conodont elements are deformed together with the encasing rock. In weakly metamorphosed areas which are usually favoured for biostratigraphic purposes, deformation of conodont elements is brittle, which leads to their fragmentation during the processing of the samples. Brittle fracturing is a frequent phenomenon also in our samples with CAI ≤ 5. However, there is a large number of samples in which the conodont elements show ductile deformation (Fig. 4). As it has been reported in earlier studies (KOVÁCS and ÁRKAI, 1987; KÖNIGSHOF, 1992), ductile behaviour in our samples occurs at CAI ≥ 5, although not as a regular phenomenon. We conclude that the temperature level attained in CAI 5 defines the brittle-ductile boundary of fine-grained apatite for the (unknown) lithostatic pressures and strain rates prevailing.

3.4. DISCUSSION

3.4.1. Assessment of metamorphic temperatures

The age of the tectono-metamorphic evolution is constrained by biostratigraphic findings. As documented by index fossils, sedimentation of flysch sediments continued into the latest Viséan (EN-

GEL et al., 1981) and locally into the earliest Namurian (FEIST and GALTIER, 1985), i.e., until 325–320 Ma (isotopic calibration of the time scale by MENNING et al., 2000). Deformation and metamorphism must have occurred later. Stefanian sediments unconformably rest upon the folded and metamorphosed Montagne Noire basement, but have still been thermally affected by the Zone Axiale (BECQ-GIRAUDON and GONZALEZ, 1986; ALABOUVETTE et al., 1993). Deformation and metamorphism were definitely terminated before the Permian, i.e., before c. 300 Ma. Therefore, we are left with a maximum time span of c. 25 Ma for the tectono-metamorphic evolution.

This consideration is important, since the metamorphic alteration of organic matter (CAI as well as vitrinite reflectance) is dependent not only on the maximum temperature achieved during metamorphism, but also on the duration of heating (e.g., TEICHMÜLLER, 1987). REJEBIAN et al. (1987) have published an Arrhenius diagram based on heating experiments, which models the evolution of CAI for different maximum temperatures and durations of heating (Table 1). For a heating time of 25 Ma, one arrives at a temperature of c. 210 °C for CAI 4, c. 325 °C for CAI 5, c. 360 °C for CAI 5.5 and c. 480 °C for CAI 6.5.

A comparative study of CAI and IC by GARCIA-LOPEZ et al. (1997) proposes that the transition between diagenesis and anchizone (180–230 °C, e.g. KISCH, 1990; FREY and ROBINSON, 1999) corresponds to CAI 4, and the transition between anchi- and epizone (280–320 °C after BUCHER and FREY, 1994) occurs at about CAI 5.5. KOVÁCS and ÁRKAI (1987) have correlated CAI and IC in a petrologically heterogeneous set of metamorphic rocks from Hungary. They correlate the transition from diagenesis to anchizone with CAI 5, i.e., with a higher CAI than GARCIA-LOPEZ et al. (1997: CAI 4). As already pointed out by KOVÁCS and ÁRKAI (1987), varying correlations between CAI and estimated temperatures may be expected because of different metamorphic conditions, such as geothermal gradient or fluid pressure.

It is also necessary to critically review the temperature modelling of REJEBIAN et al. (1987). Temperatures derived for CAI ≤ 5 (Table 1) are grossly in accord with the temperature estimates of GARCIA-LOPEZ et al. (1997). For CAI ≥ 5.5, temperatures calculated after REJEBIAN et al. (1987) appear to be too high. CAI 6.5 would yield, for a heating period of 25 Ma, a temperature of c. 475 °C, i.e., close to the greenschist/amphibolite facies boundary at low pressures. For shorter heating times, calculated temperatures are even higher (Table 1). However, CAI 6.5 occurs in parts of the Montagne Noire, where the neighbouring Carboniferous greywackes do not reveal quartz recrystallization. Therefore, syntectonic temperatures close to 500 °C appear unlikely. The estimation of REJEBIAN et al. (1987) lead to too high temperatures for higher CAI values. Lastly, REJEBIAN et al. (1987) have pointed out that the evolution of CAI may be impeded by high fluid pressures, on which we so far have no control in the Montagne Noire. For all these reasons, and in the absence of independent thermometers, we refrain from an assessment of absolute temperature and contend ourselves with the use of CAI as a relative temperature gauge.

3.4.2. Interpretation of the CAI pattern

In the southeastern part of the Montagne Noire, where the sampling coverage is the best, it is obvious that CAI decreases away from the Zone Axiale. This applies also to the area with low CAI values in a relatively northern position NW of Cabrières (Fig. 2): since the Zone Axiale and its lower grade mantle plunges towards the ENE, a decrease of metamorphic grade in that direction is to be expected.

Higher grades of CAI adjacent to the Zone Axiale are also observed further to the SW (Figs. 3a and b). A decrease of CAI towards the SE is not immediately apparent from Fig. 3b, because the area around Caunes-Minervois shows a wide range of CAI. However, the evolution of CAI is irreversible, so that we interpret the higher CAI values near Caunes as a secondary alteration. In fact, these samples are situated in a narrow corridor between a ductile reverse fault to the NNW and a brittle reverse fault to the SE (DOUBLIER and FRANKE, unpublished mapping), so that alteration of CAI by percolating fluids is plausible. By analogy with Fig. 2, we suppose that the lower values near Caunes reflect the lower metamorphic temperatures of areas more distant from the Zone Axiale.

The higher grades of CAI observed in some samples of the Vieussan Antiform and the Minervois Nappe only affect part of the conodonts within a single sample, and only parts of the individual conodont elements. A similar heterogeneity of CAI in areas with CAI ≥ 5 was also observed by KOVÁCS and ÁRKAI (1987), and explained by different pressure and temperature conditions of the anchizonal and epizonal rocks in the individual tectonometamorphic units. However, such a partial increase in CAI may also be due to alteration by percolating fluids, e.g., during contact metamorphism (see chapter 3.1.). Since our samples are derived from one major geological structure

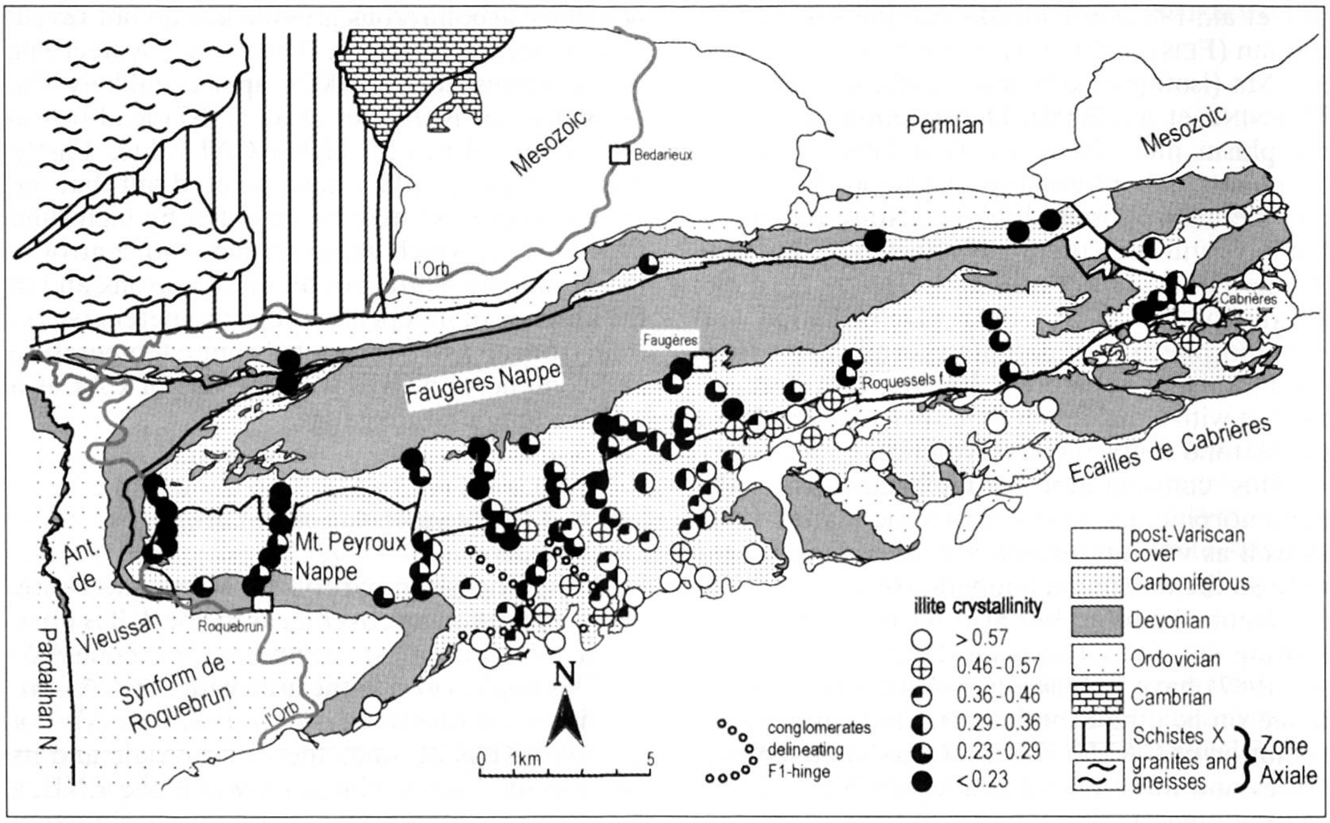

Fig. 5 Distribution of illite crystallinity in the Carboniferous flysch of the eastern Montagne Noire (after ENGEL et al., 1981).

with a more or less uniform metamorphic gradient, we prefer to interpret the spread of CAI values as due to local alteration by fluids.

This general array of the data – with CAI decreasing away from the core of the Zone Axiale – permits two interpretations:

– CAI is highest in rocks, which were most deeply buried during nappe stacking, and are therefore exposed close to the core of the later dome structure.

– CAI was acquired after nappe stacking, and imposed by the rising, hot core of the Zone Axiale.

This problem requires discussion of the relationships between metamorphism and deformation (chapter 5).

4. Comparison with illite crystallinity (IC)

A comprehensive study of IC and other indicators of low-grade metamorphism in the Montagne Noire is in progress. However, an earlier survey of IC is available for the mudstones and shales of the Early Carboniferous flysch, which overlies the Devonian carbonates (ENGEL et al., 1981). In this study, measurements were carried out according to the procedure described by WEBER (1972), i.e., without control on the density of the suspension from which the textured preparations were made, and without the use of external, inter-laboratory standards. Besides, the areal distributions of the Carboniferous shales and the Devonian carbonates are different, so that we can only compare a limited number of sample couplets taken from closely adjacent localities. For these reasons, IC data of ENGEL et al. (1981) only permit an approximate, preliminary correlation with our new CAI data.

The data set of ENGEL et al. (1981) is reproduced in Fig. 5. In order to account for the larger statistical variation in the lower anchizone and diagenetic zone, the authors choose a logarithmic subdivision of IC values, in which the Weber indices (Hb_{rel}) were subdivided into classes defined by $Hb_{rel} = 10^{2.n}$ $(2 \leq n \leq 6)$. In Fig. 5, the boundaries between these IC classes were retained, although the values are now given as $\Delta°2\theta$. As shown in the legend to Fig. 5, there are three classes between 0.46 and 0.23, which grossly correspond to the anchizone as defined by FREY and ROBINSON (1999, $\Delta°2\theta$ between 0.42 and 0.25).

The areal distribution of IC clearly reproduces the metamorphic zonation reflected in the CAI (Fig. 2). Like CAI, IC decreases away from the Zone Axiale toward the S and E. Where IC and CAI samples were taken closely adjacent to each other, such as in the Orb valley section and in the

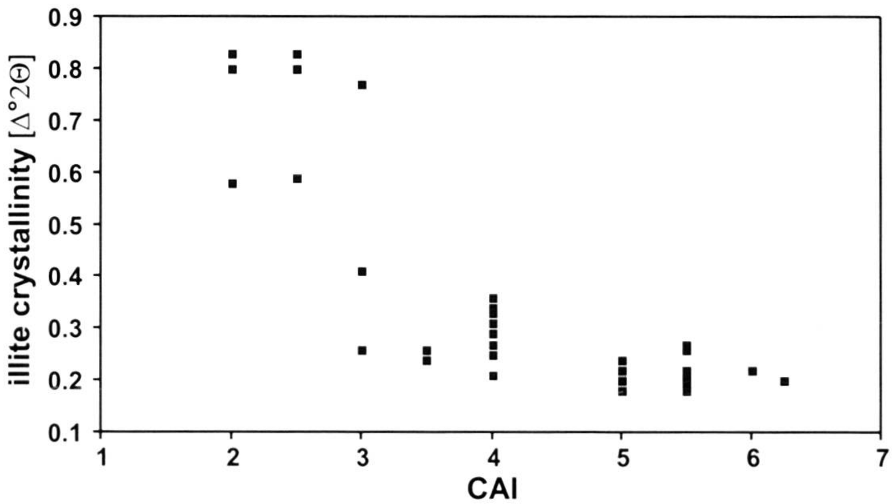

Fig. 6 Correlation diagram for CAI samples of Devonian carbonates and IC data of Carboniferous mudstones and shales (after ENGEL et al., 1981), from closely adjacent localities.

Écailles de Cabrières, it is possible to establish a very rough correlation (Fig. 6). In the range of CAI 6.5–3, IC decreases slowly. IC values around 0.25 (which, in Kübler indices, would correspond to the boundary epizone/anchizone) occur somewhere between CAI 5 and 3. This is backed up by the metamorphic maps (Figs. 2 and 5): the northern limb of the Roquebrun Synform is characterized by CAI 4–4.5, and by IC ranging in the group of 0.29–0.23. For CAI ≤ 3, IC rapidly decreases to values ≥ 0.5. Even with the relatively small and uncalibrated data set available it is possible to state that the correlation between IC and CAI is non-linear.

For comparisons with the literature, we want to emphasize again that the IC data obtained by ENGEL et al. (1981) may not be compared with Kübler indices obtained by correlation with external standards. FREY and ROBINSON (1999) correlate calibrated IC values of 0.25 with a CAI of 5.5. KOVÁCS and ÁRKAI (1987) have not proposed a precise correlation, but obtained CAI 5 (and sometimes higher) in the anchizone, and CAI 6–7 (sometimes lower) in the epizone, suggesting a transition from anchizone to epizone in the CAI range of 6–5. GARCIA-LOPEZ et al. (1997) report CAI up to 5.5 in the anchizone. Again, the anchi/epizone transition would occur around CAI 5.5, although the authors have no CAI data from the epizone. GARCIA-LOPEZ et al. (1997) correlate the lower limit of the anchizone with CAI 4. If one presumes that there is no major difference between IC and KI, our IC data would suggest that the limits of the anchizone are about one grade of CAI lower than in the publications cited above. Ongoing, precise measurements of IC will hopefully reveal whether this shift has methodological reasons (IC vs. KI), or else is due to the specific metamorphic environment of the Montagne Noire.

5. Relative timing of metamorphism and deformation

In the area around St. Pons and Olargues (Fig. 3a), Devonian limestones of the Pardailhan Nappe show the same elevated CAI (5.5–5) as the underlying Par-Autochthon. Apparently, the zonation of CAI cuts across the entire tectonic stack established during D1. In the Vieussan Antiform of the eastern Montagne Noire, both the zonations of CAI and IC clearly cut across the inverted limbs of F1 folds (Figs. 2 and 5). As shown in Fig. 5, the zonation of IC also cuts across the axial plane of a kilometric, originally recumbent F1 fold in the flysch. All these observations clearly indicate that the main metamorphic event recorded in the Palaeozoic sediments on the southern flank of the Montagne Noire post-dates the formation and emplacement of the recumbent fold nappes (D1).

In the area to the E of the large F3 syn- and antiforms (exposed in the Orb Valley), both CAI and IC are offset along the fault which separates the Faugères Nappe to the N from the Mont Peyroux Nappe and the Écailles de Cabrières to the S ("faille de Roquessels" in ENGEL et al., 1981; see Figs. 2 and 5). The metamorphic contrast is highest in the Cabrières area and decreases westwards along the fault. In the Vieussan Antiform, CAI in the Mont Peyroux is the same as in the underlying Faugères Nappe. It appears that the Roquessels fault, in its eastern segment, truncates the zonation of CAI and IC at a steeper angle, whereas it is more or less parallel with the zonation further W. Anyhow, the Roquessels fault has been refolded

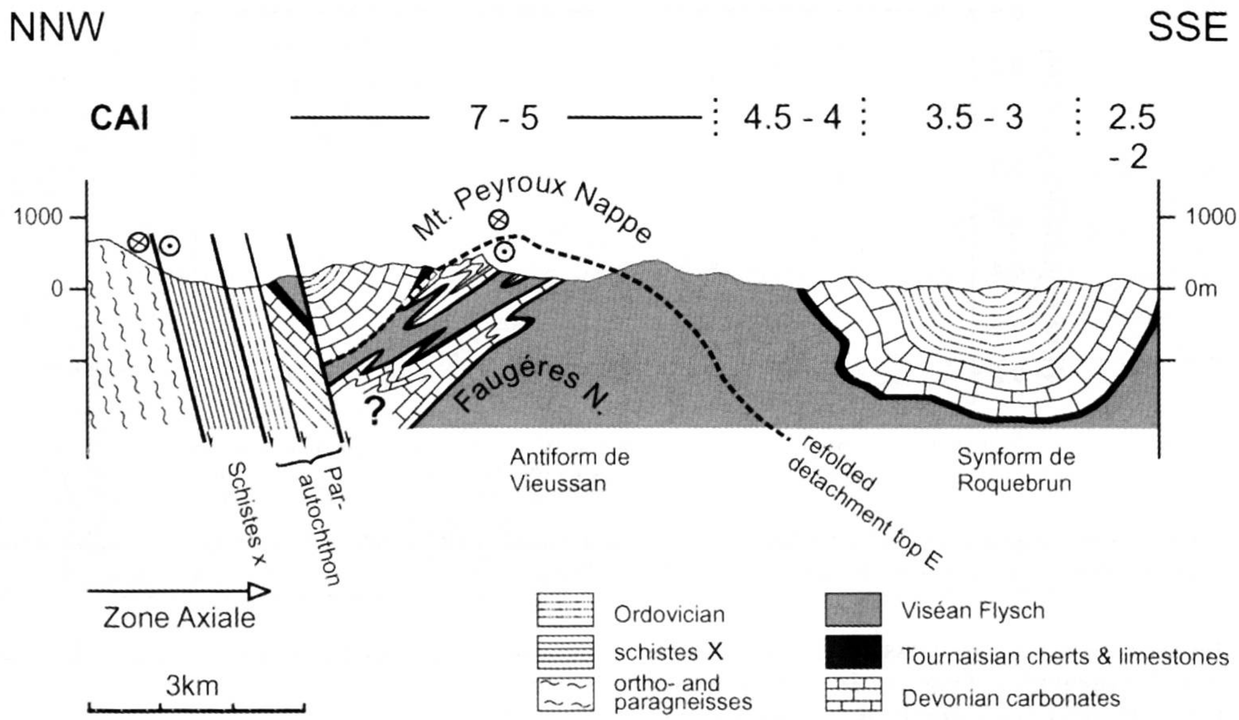

Fig. 7 Tectonic section through the Parautochthon, the Faugères- and Mont Peyroux Nappes in the Orb valley, with the zonation of CAI.

by F3 (Fig. 7). Taken altogether, these relationships suggest that the metamorphic alteration reflected in the CAI and IC post-dates F1 and predates F3, i.e., correlates with F2. Since the zonation of CAI has been refolded by F3, the present-day widths of the individual CAI zones must not be taken to derive the thickness of the individual CAI zones.

Post-nappe deformation on the southern flank of the Montagne Noire has already been noticed by LEE et al. (1988) and by MATTAUER et al. (1996). Our recent tectonic studies have confirmed LEE et al. (1988) in that the main deformation in the Par-Autochthon, the Faugères Nappe and northern parts of the Mont Peyroux Nappe occurred in a regime of ductile simple shear. Deformation is characterized by slaty cleavage, a pronounced stretching lineation in the Devonian limestones, and non-cylindroidal folds. Asymmetric clasts suggest transport towards the ESE (DOUBLIER and FRANKE, 2001; FRANKE, 2001). These features occur only in areas with CAI ≥ 5. Since CAI was established after D1, we conclude that the tectonic structures and fabrics associated with the higher CAI values also post-date D1, i.e. relate to the exhumation history of the gneissic core (D2). In fact, the style of deformation and kinematic indicators in the Devonian carbonates match those in the eastern part of the Zone Axiale (simple shear top to the ENE: e.g., MATTE et al., 1998). It appears that, during the time of ductile deformation, the Devonian limestones of the Mont Peyroux and Faugères Nappes were kinematically still coherent with the gneissic core.

Today, the core of the Zone Axiale is separated from the Palaeozoic rocks on its southern flank by a late, dextral strike-slip fault (ECHTLER and MALAVIEILLE, 1990). These relationships suggest that the area characterized by the extensional regime of the Zone Axiale originally was more extensive: it also comprised large, northerly parts of the southern flank, and later contracted into its present-day contours. This areal reduction of the kinematic regime was probably caused by the cooling of the structure.

The array of the metamorphic zonation indicates that metamorphism in the Palaeozoic cover was caused by the rising hot gneissic core. Ongoing isotopic studies suggest that high temperature/low pressure metamorphism in the gneisses was already active at 316 ± 1 Ma (FRANKE et al., 2002), or even before the intrusion of the post-tectonic Vialais granite at 327 ± 5 Ma (MATTE et al., 1998). Flysch sedimentation in the Montagne Noire continued at least into the late Viséan, which corresponds to an isotopic age of 325–320 Ma (MENNING et al., 2000). These constraints leave only a few Ma for deformation and metamorphism, probably too little for the usual "regional metamorphism", which is caused by tectonic stacking followed by thermal relaxation. We therefore suppose, that heat was advected by the intrusion of synkinematic granites, which are concealed in the subsurface of the Zone Axiale (dynamic contact metamorphism).

6. Conclusions

Our regional study of the conodont alteration index (CAI) in Devonian carbonates on the S flank of the Montagne Noire yields methodological results as well as new constraints upon the tectonic evolution of a Variscan metamorphic core complex.

Methodological aspects include the following points:

– There is no relationship between CAI and stratigraphic age (i.e., depth and duration of sedimentary burial).

– Secondary, post-tectonic dolomitization had no major impact on the development of CAI.

– The metamorphic zonation revealed by the CAI data correlates well with an earlier illite crystallinity (IC) map (ENGEL et al., 1981). Comparisons between CAI and IC show a non-linear correlation, with IC rapidly decreasing for CAI ≤ 3. The boundaries of the anchizone (0.25 – c. 0.40 $\Delta°2\theta$) range between CAI values around 4 and c. 3, i.e., at lower $\Delta°2\theta$ than in the previous literature. This difference may be due to methodological reasons (uncalibrated IC values in ENGEL et al., 1981), or else to the specific metamorphic regime of the study area.

– Some irregularities in the general pattern of CAI may be explained by secondary alteration of CAI by fluids circulating in fault zones.

Constraints upon the tectonometamorphic evolution of the Montagne Noire include the following items:

– The metamorphic zonation revealed by CAI and IC cuts across the inverted limbs of recumbent F1 fold nappes and across the axial plane of a large F1 fold in the Carboniferous flysch identified by ENGEL et al. (1981). On the other hand, an extensional fault which reduces the CAI and IC zonation, is refolded by F3. The pattern of CAI and IC must therefore reflect a second metamorphic event (M2), which postdates the emplacement of large recumbent fold nappes (F1) and has obliterated the M1 metamorphism associated with folding and nappe stacking.

– Tectonic structures and fabrics associated with CAI and IC include slaty cleavage, non-cylindroidal folds and shear criteria top to the ENE (here summarized as D2), which match the deformation within the crystalline Zone Axiale. It appears that – during D2 and M2 – an important, northern part of the Palaeozoic rocks on the southern flank of the Montagne Noire were kinematically coherent with the Zone Axiale. D2/M2 were caused by dynamic contact metamorphism during the extensional exhumation of the hot metamorphic core.

– Shrinking of the extensional regime into the present-day contours of the Montagne Noire probably occurred during cooling.

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