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CALEDONIAN AND VARISCAN STRUCTURES IN THE ROCROI-ARDENNE LOWER PALAEOZOIC BASEMENT (Belgium and adjacent countries)

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(19 figures)

ABSTRACT. - This paper presents a synthesis of the relevant structural evidence for a Caledonian tectonic event in the Ardenne, with special reference to the Rocroi Massif. The importance of the Variscan imprint in the basement is also discussed.

The Lower Palaeozoic basement of Belgium and adjacent countries outcrops in the Brabant Massif, the Condroz Ridge and in the High-Ardenne Massifs. In the Ardenne, the basement was submitted to two successive compressive tectonic events (Caledonian and North-Variscan), separated by a period of tectonic relaxation or distension. Recent microstructural evidence, mainly from the Rocroi and Stavelot-Venn Massifs allows some insight into evolution of these tectonic events.

The Caledonian compressive period is subdivided into by three deformation phases, successively characterized by the formation of a penetrative cleavage, associated with minor N-S folds and extension fractures (D1), by major folding along a regional E-W to ENE-WSW axis (D2) and by late upright folding of lesser importance (D3). The relationship between the first two types of structure has still to be clarified. However, observations indicate that the original cleavage formation might be older than the major E-W to ENE-WSW folding.

The period of tectonic relaxation between the Caledonian and Variscan compression is marked by magmatic intrusions, in extensional setting (D4).

The Variscan deformation in the basement occurred mainly by reactivation of pre-existing Caledonian anisotropies and by the development of new structures of minor importance, like boudinage, folds and mullions (D5). The existing anisotropy was determined by cleavage as well as bedding planes, which were probably already subparallel to each other, as a consequence of the intensity and the isoclinal style of the Caledonian folding. Late and minor effects of the Variscan event (D6) are expressed mainly by kink bands in the Southern zone of the Rocroi Massif, and by open, upright, hectometric-scale, E-W folds in the Northern zone.

The existence of Caledonian deformation, as witnessed by cleavage development and isoclinal folding, was also demonstrated by several authors in the other High Ardenne's Caledonian Massifs.

I. INTRODUCTION

In Belgium, the Lower Palaeozoic basement outcrops in several massifs of varied stratigraphic and structural settings (fig.1) : the Brabant Massif to the North, the Condroz Ridge and the High-Ardenne and Eifel massifs to the South. These massifs were all affected by the Caledonian orogeny, and later by the North-Variscan fold-and-thrust belt.

The Brabant Massif consists of Cambro-Silurian rocks, probably derived from the destruction of a proximal Proterozoic crystalline basement. The latter supposedly forms the Eastern extension of the 'Armorica Craton' (Andre & Deutsch, 1984; Vander Auwera & Andre, 1985). The sedimentary sequence was intruded by calc-alkaline magmas along an arcuate belt during Caradocian to Wenlockian times (Andre & al., 1986). The "Caledonian" deformation in the Brabant Massif is probably Early Devonian. It comprises folding, cleavage development, thrusting, greenschist regional metamorphism (Mortelmans, 1955; Legrand, 1967; Andre *& al.*, 1986).

The Condroz Ridge is composed of rocks of Tremadoc to Ludlow age. Conglomerate and flysch

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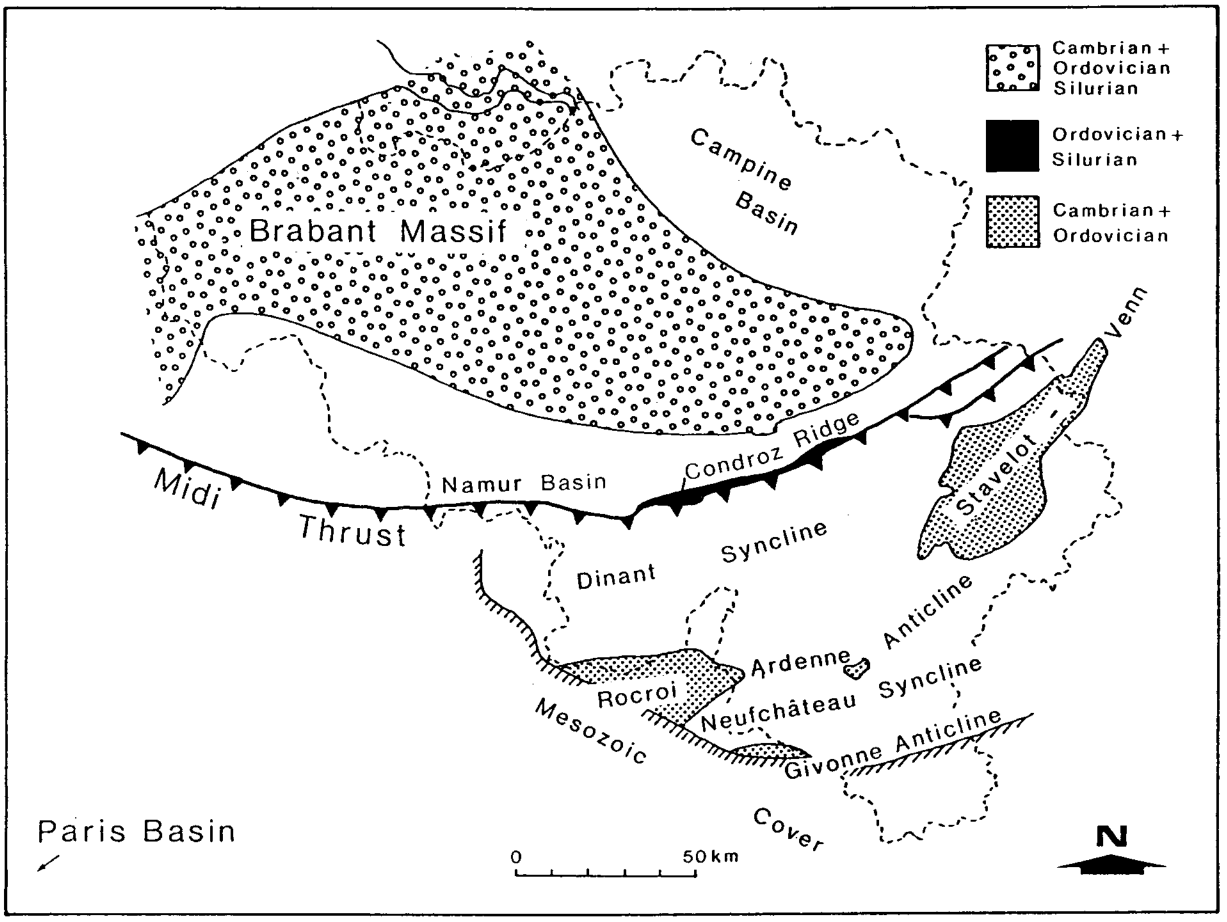

Fig. 1 - Major structural units of Belgium. The Lower Palaeozoic is outcropping in the Brabant Massif, the narrow elongate Condroz Ridge and the High-Ardenne Massifs of Rocroi, Serpont, Stavelot-Venn and Givonne. During the North-Variscan event, the Brabant Massif remained in a parautochthonous situation, while the Dinant Synclinorium was folded and thrust over the Namur Synclinorium, along the Midi fault. The High-Ardenne Variscan structure includes the Rocroi-Ardenne Anticlinorium, the Neufchâteau Synclinorium and the Givonne Anticlinorium. This structural framework is unconformably overlain by the Mesozoic cover of the Paris Basin.

deposits of Caradoc age probably results from the uplift related to Caledonian deformation in the Ardenne region. During Silurian, the Brabant and Condroz regions were supplied with siliciclastics bearing reworked Ordovician acritarch. These probably originated from the Southern Ardenne, in rapid uplift (Steemans, 1989). Tholeiitic intrusions of Silurian age are also known.

In the High-Ardenne the Lower Cambrian to Ordovician rocks are composed of sandy to shaly series, turbiditic in part (Vanguestaine, 1986). They were folded in a regional scale during the Ardenne "Caledonian" tectonic event, probably Caradocian (Fourmarier, 1954). Magmatism broadly occurred between the end of Caledonian and the beginning of Variscan deformation (Goffette *& al.*, 1991). Intrusions are of basic tholeiitic composition (Andre *& al.*, 1986).

In summary, the Caledonian orogenic event in Belgium occurred first in the Ardenne massifs and then migrated northward to the Brabant Massif. The Condroz Ridge seems to have been in an intermediate situation between the Ardenne and the Brabant Massifs. In these three Lower Palaeozoic areas, magmatism occurs during a long time interval. It is mainly post-kinematic in the High-Ardenne and pre-kinematic in the Brabant Massif.

The Lower Palaeozoic basement was later subjected to the Variscan orogeny. The geological history of the North-Variscan Front has recently been interpreted to have occurred in a two- step model of progressive deformation, in which the context evolved from a tensional one to a compressional one (Bless *& al.*, 1989; Meilliez *& al.*, 1990). During the Lower Devonian, the rate of downwarping and the thickness of the deposits was greatly variable, being the greatest in the Neufchâteau area (South of the Rocroi Massif) and the least near the Condroz area, indicating an active tensional context. From Middle Devonian to Middle Namurian times, the slower subsidence rate indicates a period of relative tectonic relaxation (Bless *& al.*, 1989). Renewed activity in Namurian times is

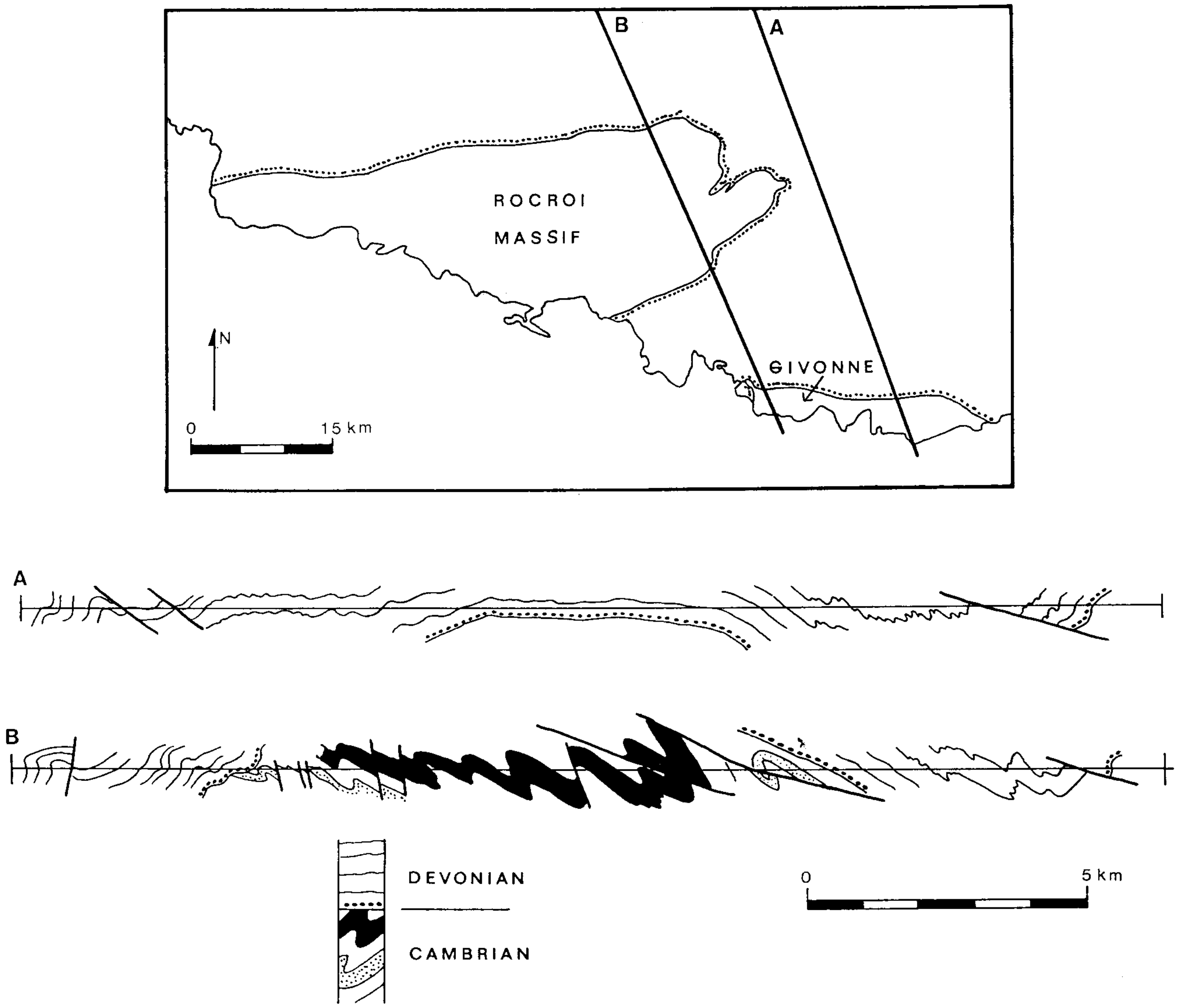

Fig. 2. - Schematized N-S section of the Rocroi Massif and its Devonian cover, illustrating the tightly folded Lower Palaeozoic unconformably overlain by the gently folded Lower Devonian sequence, after Asselberghs (1946), for the Lower Devonian and Beugnies (1963), for the Lower Palaeozoic.
a: Lower Devonian profile, a few kilometres East of the unconformity;
b: Meuse Valley profile.

shown by rapid increase in the rate of subsidence in the Dinant area and by the change of facies from stable platform carbonates in Middle Devonian to flysh sediments in Upper Namurian. By the time, the Ardenne Area was probably uplifted and being eroded. Evidence reported by Paproth (1987) and Bless *& al.* (1990) show that the Variscan deformation migrated northward, from Namurian to Westphalian.

During the North-Variscan tectonic event, the whole area was separated into two structural domains, bounded by the Midi thrust faults and the Condroz Ridge (Fig.2; Raoult, 1986; Raoult & Meilliez, 1986, 1987; Michot, 1989). The Northern domain was parautochthonous and comprised the Lower Palaeozoic Brabant Massif and the Devonian-Carboniferous Namur Synclinorium. The Southern domain comprised the Dinant Synclinorium, the High-Ardenne Anticlinorium with the Caledonian massifs of Rocroi, Serpont and Stavelot-Venn, the Neufchâteau-Eifel Synclinorium and the Givonne Anticlinorium. Evidence from boreholes, seismic and field data have let to several proposed tectonic models (Bless *& al.*, 1977; Raoult & Meilliez, 1986; Michot, 1989; Paproth, 1987; Bless *& al.*, 1989). There is general agreement that the Dinant Synclinorium may be a fold-and-thrust sheet (Dinant nappe), but the High-Ardenne Anticlinorium is interpreted either as allochthonous (Raoult & Meilliez, 1986, 1987) or as subautochthonous (Paproth, 1987; Bless *& al.*, 1989). These proposed models are all based on the interpretation of a seismic reflector, that supposedly extends southwards, below the High-Ardenne, from the outcrop trace of the Midi thrust faults.

The intensity of the North-Variscan tectono-metamorphic event is known to decrease northward, from the High-Ardenne, to the Brabant Massif (Hugon & Le Corre, 1979, Dandois 1981, Piqué *& al.*, 1984; Beugnies, 1986). The occurrence of a Caledonian regional folding event in the Ardenne-Eifel is generally accepted (Beugnies, 1963; Geukens, 1969, 1978). However, others aspects of Caledonian deformation are still controversial, e.g. the thermal history, the existence of axial-plane cleavage and the possible reactivation of Caledonian structures during the Variscan orogeny.

In the Rocroi Massif, both the Caledonian and Variscan major folds have similar E-W to ENE-WSW axial trend, with southward dipping axial planes. The Variscan deformation in the basement is generally superimposed on the Caledonian structures. As a result, the structures are often difficult to interpret without detailed microstructural analysis. In the Stavelot Massif, exposure is poorer than in the Rocroi Massif, but the E-W trending Caledonian belt is overprinted more obliquely, by the SW-NE Variscan trend. The Variscan metamorphism was most intense, of a very low grade to low grade greenschist facies in the High-Ardenne (Dandois, 1981 & ms.1984; De Béthune, 1986; Beugnies, 1986). This is probably a higher degree of metamorphism than the Cambro-Ordovician sequence suffered throughout the Caledonian event. In the Ardenne, the Variscan metamorphism developed mainly in static conditions, before the onset of Variscan deformation.

All these conflicting data have let to many controversial hypothesis on the respective importance of the Caledonian and Variscan deformations in the Lower Palaeozoic basement (Gosselet, 1888; Fourmarier, 1931; Waterlot, 1937, 1945; Beugnies, 1963; Hugon & Le Corre, 1979; Klein, 1980; Hugon, 1983; Delvaux de Fenffe & Laduron, 1984, 1987). This paper presents a synthesis of the most relevant structural evidence for a Caledonian event in the Ardenne, and especially in the Rocroi Massif. The importance of the Variscan imprint in the Lower Palaeozoic basement is also discussed. New field data are mainly from the Rocroi Massif; reference to literature is made for the other massifs. A general model of structural evolution is proposed for the Rocroi Massif.

II. SOUTHERN ZONE OF THE ROCROI MASSIF

1) Regional setting

The structure of the Rocroi Massif is known mainly from the lithostratigraphic maps of Waterlot (1937) and Beugnies (1963). However, they are relatively speculative, based mainly on the reconnaissance of the slaty series from place to place. Recent micropalaeontological data have demonstrated that, in places, the lithostratigraphy and biostratigraphy do not correspond (Vanguestaine, 1986). In absence of new geological maps based on the micropalaeontological data, the structure of the Rocroi Massif is assumed to be the one described by Beugnies, in 1963 (fig. 2 & 3).

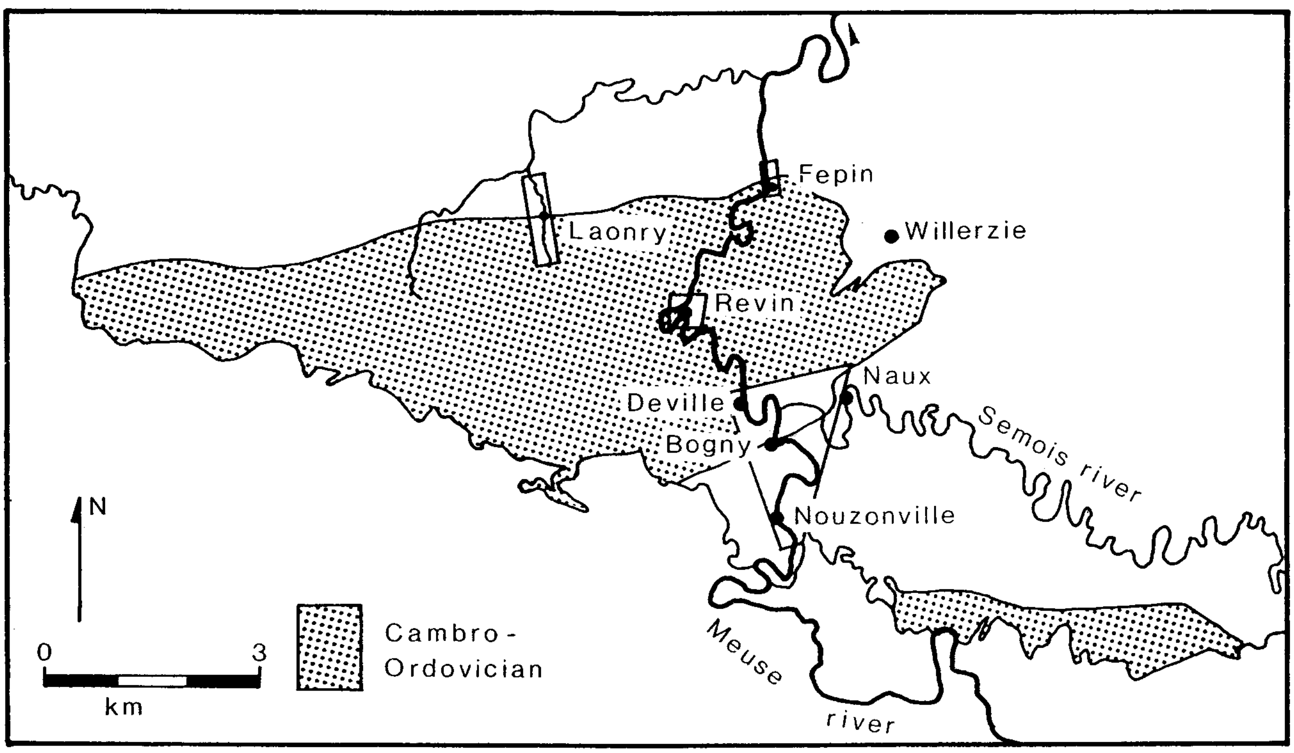

Fig. 3 - Location of the regions considered, in the Rocroi Massif.

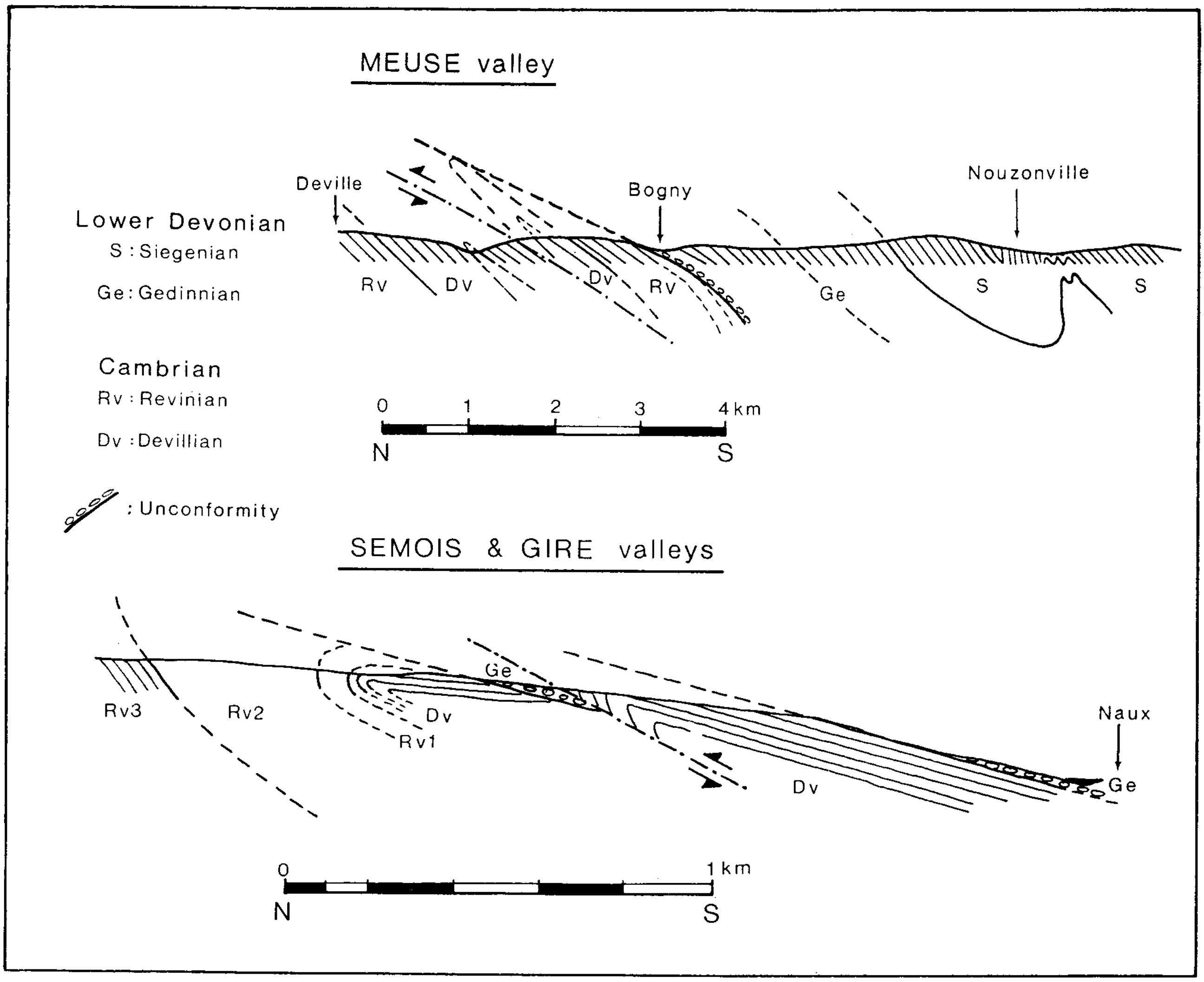

Fig. 4 - Detailed N-S sections across the Southern zone of the Rocroi Massif, showing the different style of folding, the unconformity overlain by the basal Fépin formation, and the effects of Variscan thrusting.

As a result of the good exposure, the structure of the Southern zone of the Rocroi Massif is relatively well known (fig.4). It is situated in a slightly inclined, normal flank of a large Variscan E-W fold with axial plane dipping to the South. The base of the Lower Devonian (Fépin series) unconformably overlies the Cambrian and its Caledonian E-W major folds (fig.4). The Caledonian and Variscan major folds are nearly coaxial and both face northward.

The metamorphism in the Southern zone of the Rocroi Massif is characterised by the occurrence of magnetite, chloritoid, ilmenite and pseudomorphosed andalusite. It is undoubtedly of Variscan age, because chloritoid porphyroblasts are found both in the matrix and in the pebbles of the basal "Gedinnian" conglomerate. The metamorphic peak is considered to be pre-pre-kinematic with respect to the Variscan cleavage development (Beugnies, 1963; Dandois, ms.1984; Delvaux de Fenffe & Laduron, 1984).

2) Caledonian event

Two deformation phases, in the Southern zone of the Rocroi Massif, were first established from microstructural evidence in the "Rocher de l'Hermitage" at Bogny-sur-Meuse (Delvaux de Fenffe & Laduron, 1984). The first deformation (D1) is described as a sub-isoclinal, N-S oriented, westward-facing folding with subhorizontal, axial-plane slaty cleavage. The second deformation phase (D2) comprises upright N-S folds, without axial-plane cleavage. The microstructural investigation was later extended to a larger area between Deville, Naux and Bogny-sur-Meuse (Duchâteau, ms.1983; Delvaux de Fenffe & Laduron, 1986). Hugon & Le Corre (1979); Hugon (1983) and Le Corre & Hugon (1987) that no penetrative deformation with cleavage development occurred in Caledonian times and that all the structures within the basement were solely the result of the Variscan deformation. However, as already discussed (Delvaux de Fenffe & Laduron, 1987), the occurrence of a Caledonian major synkinematic deformation is assumed here.

a. Pre-kinematic structures

In the Cambro-Ordovician sequence, turbidites are common. Sedimentary structures like graded bedding, cross bedding, flute cast and slumping are frequently observed.

A preferred orientation of very fined micas, parallel to the bedding, is observed locally in the more slaty layers. It is possibly related to compaction during diagenesis. No evidence of Caledonian metamorphism has been distinguished.

b. Major deformation phase

The major Caledonian deformation phase in the Southern zone of the Rocroi Massif is relatively complex and it is still incompletely understood. Two type of structures occur independently. They differ either by their orientation and by their style of folding. Their chronological relationship is not yet understood.

The most evident Caledonian structures are the regional E-W folds, overturned to the North (Waterlot, 1937; Beugnies, 1963). Here (fig.4), two N-S cross sections along the Meuse and Semois-Gire Valleys, were constructed, representing the real inclinations of bedding. The regional basal Devonian unconformity lies across major Caledonian folds. The folds are tight to isoclinal, but it is possible that, initially, they may have been of a more open profile and subsequently tightened during the Variscan event. Minor folds associated with the Caledonian E-W folding are difficult to distinguish from the E-W Variscan folds.

The other structures, also Caledonian, are minor NNW-SSE recumbent folds, facing WSW and associated with an axial plane slaty cleavage. These were described as P1 folds and S1 cleavage by Delvaux de Fenffe & Laduron (1984). These folds are locally associated with quartz-filled N-S extension fractures. The latter are common in most of the quartzitic layers and display a constant N-S direction over the area (fig.5). Some fractures are cut and displaced by tension veins associated with the Variscan E-W boudinage (fig.6a), while others are wrapped around later Caledonian folds, or also around Variscan folds (fig.6b). Extension fractures in some pebbles of the Fépin conglomerate (fig.6c) suggest a Caledonian origin.

In summary, the Caledonian major deformation, consists in two types of structures, with different structural style:

- subhorizontal slaty cleavage development, N-S extension fractures in quartzitic layers and occasional NNW-SSE folds, facing westward (net fig.8a);
- regional northward facing E-W folding and possible cleavage development (and/or reworking of the older one, if this is the latest event).

In the Southern zone of the Rocroi massif, there is no clear evidence for the relative chronology between the two types of structures, but it is postulated that the N-S structures are the first (D1) and the E-W ones are the latest (D2). In a previous work (Delvaux de Fenffe & Laduron, 1984), these two different structures were associated in a single deformation stage (D1). Here, they are split into two stages (D1 and D2). Consequently, the second stage of the previous work has now to be named D3.

c. Minor N-S folding

Near Bogny and Naux, in the Meuse and Semois Valleys, there is evidence for a late Caledonian (D3)

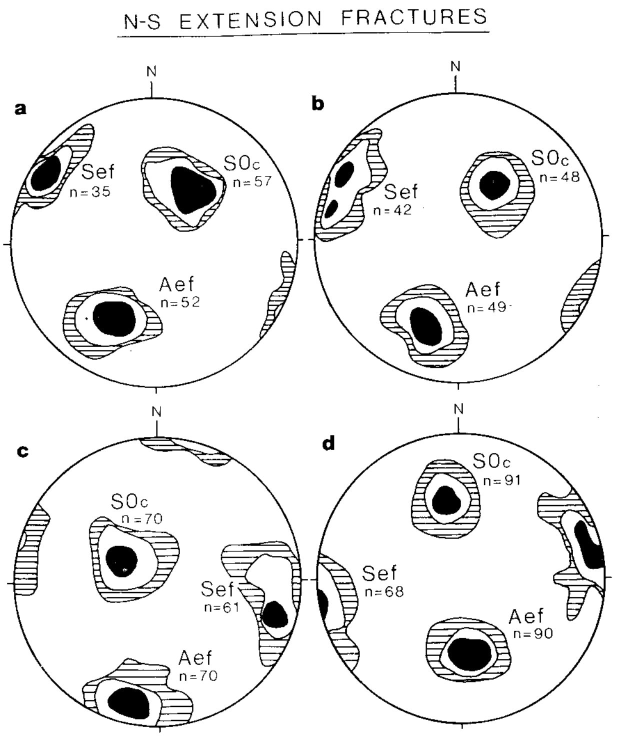

Fig. 5 - Schmidt nets, lower hemisphere projection, for quartz-filled extension fractures cutting quartzite layers at high angle to the bedding, in the Southern zone of the Rocroi Massif. S0c: poles of bedding planes of Cambrian quartzite layers, Sef: poles of extension fractures, Aef: axis of intersection between extension fractures and bedding planes. Density surfaces at 1, 15 and 30% per 1% area.

a: Deville-Monthermé area.

b: Bogny, left side of Meuse Valley.

c: Bogny, right side of Meuse Valley

d. Naux area.

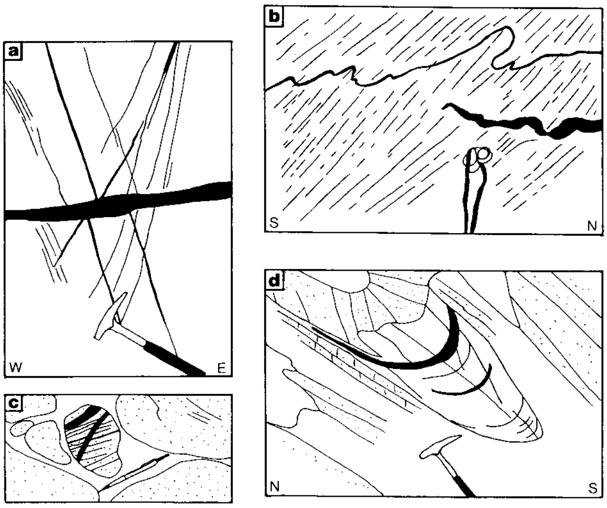

Fig. 6 - Structural relationships between N-S extension fractures in quartzite layers and later structures, in the Southern zone of the Rocroi Massif (drawing from photograph).

a: Bedding plane of a Lower Cambrian quartzite layer in which two sets of N-S extension fractures are intersected by E-W interboudinal quartz veins. Hammer is 32 cm long.

b: Bedding plane of a subvertical N-S quartzite layer, situated in the steep flank of a N-S upright fold. It shows the traces of folded quartz-filled extension fractures, and of cleavage planes of variscan orientation. This indicates that relatively intense deformation occurred in the quartzite layer already affected by extension fractures, along the cleavage planes. Lens is 5 cm long.

c: Quartzite pebble in Lower Devonian basal conglomerate, with two sets of quartz-filled extension fractures. Pen is 15 cm long.

d. Lower Cambrian quartzite layer with pre-existing quartz-filled extension fractures wrapped, together with the bedding, in a fold of probable Variscan origin. Hammer is 32 cm long.

folding event. Upright folds of metric to hectometric scale, with N-S axis were described first in Bogny-sur-Meuse (P2 folds of Delvaux de Fenffe & Laduron, 1984). No axial plane cleavage is associated with them, but the preexisting slaty cleavage is folded together with the bedding planes. A detailed mapping has shown that these folds are relatively frequent in the Southern zone of the Rocroi Massif. They are of various amplitude and their profile range from light undulations to well developed concentric folds with steeply inclined limbs (fig. 6b and 7b). In several places, they are intersected by the «Gedinnian» basal unconformity. Consequently, this folding event is of Caledonian origin, but is posterior to the major deformations D1 & D2, and is hence named D3.

3) Variscan event

After an intermediate period of tectonic relaxation and distension (D4), the effects of the Variscan compressive event (D5) in the Rocroi Massif is relatively intense at its Southern zone. The thermal effect is expressed in the formation of magnetite, chloritoid, ilmenite and, probably, andalusite. The major structures are kilometric scale E-W folds with axial-plane cleavage, E-W boudinage in quartzitic layers and N-S stretching lineation in slates. Late Variscan deformations (D6), as minor kink bands and other folds which affect the slaty cleavage, are also known in the Southern zone of the Rocroi Massif and in the Neufchâteau Synclinorium (Fourmarier, 1944; Fourmarier *& al.*, 1962; Delvaux, 1991).

These structures occur both in the Cambrian basement and in the Devonian cover. However, metamorphism and deformation was not strictly contemporaneous. In the Cambrian and Lower Devonian slates, cleavage planes are deflected around the metamorphic blasts, indicating a pre-kinematic crystallisation (Beugnies, 1967, 1986; Delvaux de Fenffe & Laduron, 1984). This show that the peak of metamorphism occurs prior to cleavage development and folding. Also in the Cambrian at the Southern zone of the Rocroi Massif (Delvaux de Fenffe & Laduron, 1986), and in the Devonian of Bastogne (Lambert & Bellière, 1976), boudinage developed prior to folding and cleavage. This idea is supported by quartz veins between individual boudins, intersected by cleavage planes.

Although the whole Palaeozoic sequence was affected by the same tectonic event, the Cambro-Ordovician and the Devonian were deformed in a significantly different manner. In the Lower Devonian, the main Variscan deformation is characterized by kilometric-scale E-W folds, axial-plane slaty cleavage, N-S stretching lineation, local E-W boudinage and minor folds, facing North. The dip of the cleavage planes gradually changes, from 70-65°S in the cover, to 40-25°S in the basement.

In the Lower Palaeozoic, the Variscan microstructures have a greatly variable orientation, in contrast with the situation known in the Devonian. However, most minor folds in the basement face northward (NW, N or NE). This great variability of orientation can be easily explained, by the superimposition of neoformed Variscan minor folds upon preexisting Caledonian medium-scale structures. With this regard, the decametric to hectometric scale upright N-S folds of late Caledonian origin (D3) probably had a great influence on the orientation of the neoformed Variscan minor folds. For the development of the latter, two cases can be distinguished, depending on the attitude of the plane of anisotropy in which they developed (bedding and sub-parallel slaty cleavage):

a) In the limbs of the late Caledonian N-S upright folds (D3), the plane of anisotropy was oriented at high angle to the Variscan finite deformation X-Y

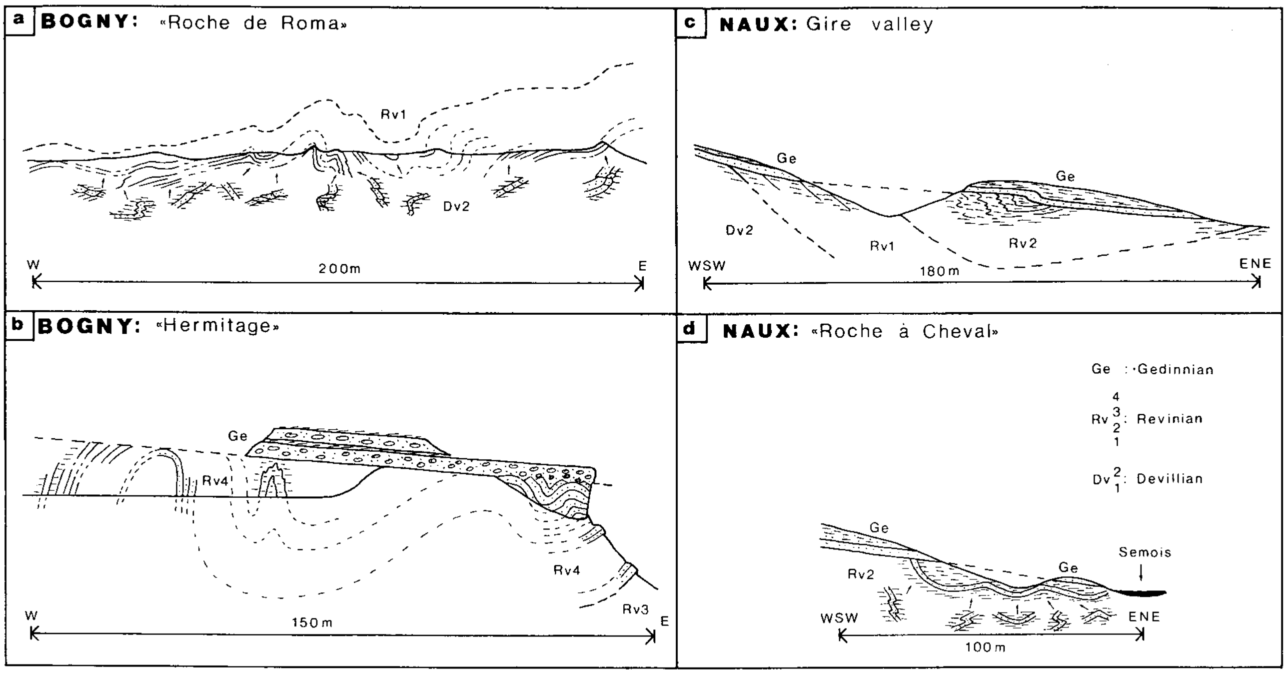

Fig. 7 - Selected E-W sections in the Southern zone of the Rocroi Massif, showing hectometric-scale upright N-S P3 folds. In E-W section, the Variscan crenulation cleavage (S4) appear subhorizontal. The Variscan cleavage and associated microfolds are clearly superimposed in the limbs of the upright D3 folds (late Caledonian). The early Caledonian S1-S2 cleavage (not shown) is subparallel to the bedding, around the P3 folds. Dv1-2: Devillian (Lower Cambrian), Rv1-4: Revinian (Upper Cambrian), Ge: "Gedinnian" basal conglomerate and sandstones (Lower Devonian).
a: Bogny (Roche de Roma).
b: Bogny (Roche de l'Hermitage).
c: Naux (Gire Valley).
d: Naux (Roche à Cheval).

plane (maximum flattening). Such a setting facilitates microstructures like folds, crenulation cleavage and mullions, with axial orientation depending on the initial attitude of the host bed: SW in one flank of the N-S folds and SE in the other (fig.8c).

b) Where existing anisotropy was at a low angle to the Variscan X-Y plane, boudinage, non-cylindrical folding and slickensides on bedding planes are dominant (fig.8d). The older slaty cleavage is no more crenulated, but local perturbations around porphyroblasts reflects its renewed flattening.

This latter situation is the most frequent, because of the limited development of the upright D3 folds. As a consequence, well-developed crenulation cleavage and mullions are relatively scarce and limited to specific areas. This explains why, generally, only a simple slaty cleavage is observed throughout the Rocroi Massif.

In the Cambrian sequence, boudinage commonly affects the most competent quartzite beds. A careful geometric study was made for 83 boudins in quartzite layers. Both symmetric and asymmetric boudins are present and some show rotation of one boudin relative to the other, in a backward sense (individual boudins dip less steeply than the general bedding: fig.9a).

Investigations by Ramsay (1967) and Goldstein (1988) on boudinage development have shown that symmetric boudinage forms when bedding is parallel to the principal direction of extension, and rotated asymmetric boudinage develop when bedding is oblique to the principal direction of extension. In the present case of back-rotated boudins, the principal direction of extension should be less inclined then the general bedding plane (or the direction of compression should be less inclined than the normal to the bedding). In this situation, an inverse shear component should exist in the bedding plane. The observation of slickensides on the bedding planes confirms that sliding may have occurred between individual strata, and their N-S direction coincides with the theoretical slip direction on the bedding plane.

Careful observation of the sense of asymmetry and rotation of individual boudins in relation to the major Caledonian E-W reveals that the backwards sense of boudin rotation is always preserved, in both normal and reverse flanks of the Caledonian folds (fig.9c). This is inconsistent with the idea that rotation of boudins generally occur in opposite sense in the two

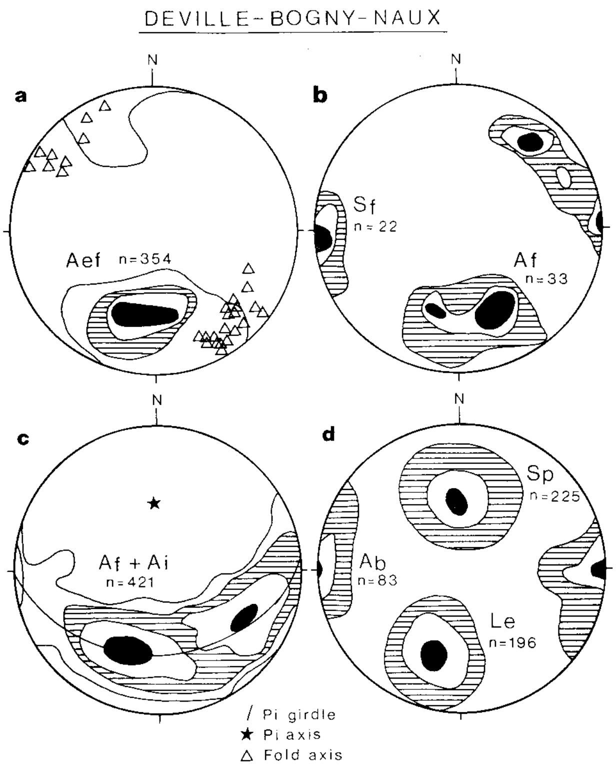

Fig. 8 - Schmidt nets, lower hemisphere projection, for the Southern zone of the Rocroi Massif. Density surfaces at 1, 15 and 30% per 1% area.

a: Caledonian sub-isoclinal D1 folds with NNW-SSE axial direction. Southward-plunging intersections between extension fractures and bedding planes (Aef).

b: Late Caledonian N-S upright D3 folds (Af: fold axis; Sf: poles of axial-plane).

c: Variscan axis (Af) of D4 fold and intersection lineations (Ai) in the Cambrian series. Great circle whose PI-axis coincides with the poles of cleavage planes. The two maxima (SSW and SE) correspond to the two flanks of the upright P3 folds, dipping respectively W and E.

d: Variscan structures of constant orientation, both in the Lower Palaeozoic and in Lower Devonian rocks (Sp: poles of principal cleavage planes, Ab: axis of boudinage, Le: plunge of mineral stretching lineation contained in Sp).

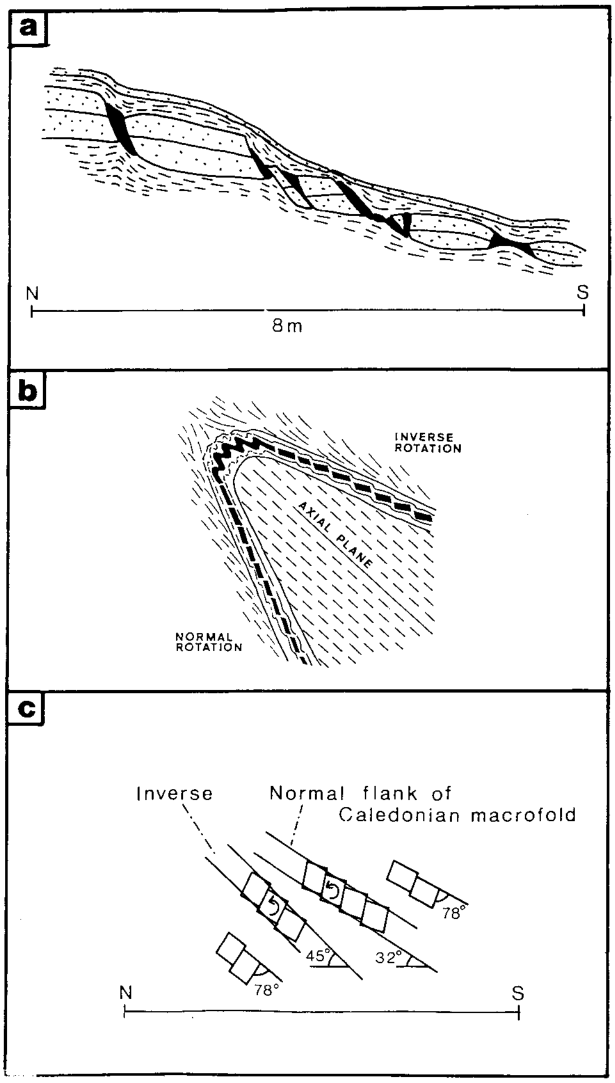

Fig. 9 - Occurrence and interpretation of boudinage in the Cambrian of the Southern zone of the Rocroi Massif.

a: Asymmetrical boudinage in a quartzite layer with back-rotation of each boudin relative to the other (Revinian of Naux).

b: Schematized occurrence of asymmetric rotated boudins in an inclined fold (after Rast, 1956). When boudinage and folding belong to the same deformation, the sense of asymmetry and rotation of isolated boudins is opposite to each other in the two flanks of the fold. The bedding of individual boudins tend to be less inclined (back-rotation) in the normal flank of the fold and more inclined in the overturned flank.

c: Schematized occurrence of asymmetric boudinage in relation with the major Caledonian folding. The sense of asymmetry and the back-rotation of isolated boudins is the same in both the normal and the overturned flanks of the major Caledonian folds. This suggests that the boudinage belongs to a distinct and later event than the folding.

flanks of a major fold, when belonging to the same event (fig.9b). However, the backward sense of rotation is consistent with a structural setting in a normal flank of a regional fold with axial plane dipping to the South. This is the case for the Variscan structure of the whole Southern zone of the Rocroi Massif. The geometrical characteristics of the back-rotated boudins thus confirms that they are not related to the Caledonian major folding, but to the Variscan major deformation.

The disposition of cleavage and microfolds in the two flanks of the nearly recumbent Caledonian E-W fold, near Naux (fig.10), also evidences the superposition of Variscan minor folding, in the overturned flank a Caledonian structure.

These latter observations have three important consequences: (1) they confirm the existence of Variscan boudinage and minor folds in the Lower-Palaeozoic basement, (2) they provide new evidence for superimposed deformation, and (3) they indicate

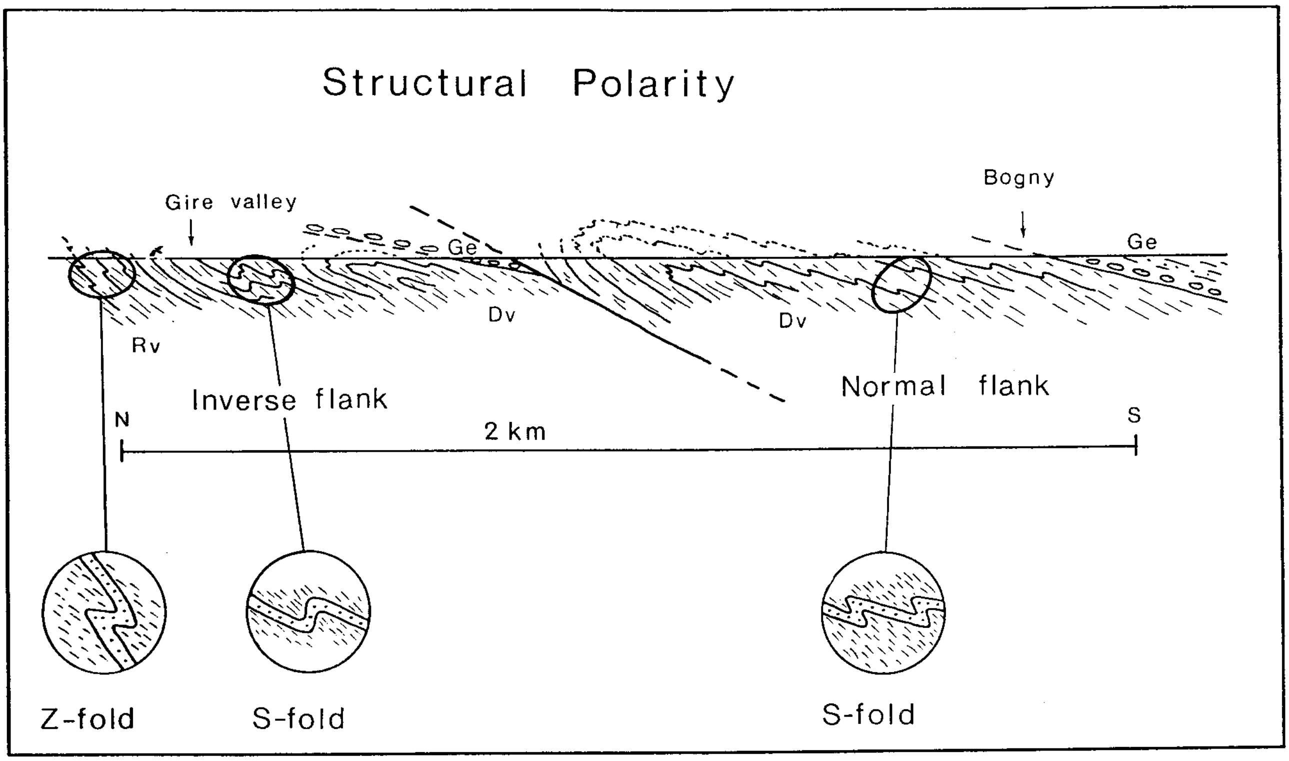

Fig. 10 - Disposition of principal cleavage and Variscan microfolds in the two flanks of the Caledonian major E-W fold, near Naux (Southern zone of Rocroi Massif). S folds occur in the two flanks of the Caledonian major structure, but Z folds occur only in the overturned flank. The Z folds in the overturned flank are compatible with a simultaneous development of the major host structure, during Caledonian E-W folding. However, the development of S folds in the overturned flank of the major fold is incompatible with the formation of this fold and must have a younger (Variscan) origin.

that all of the Caledonian basement of the Southern zone of the Rocroi Massif was deformed during the Variscan event, in the normal flank of a major Variscan inclined fold.

III. CENTRE AND EASTERN ZONE OF THE ROCROI MASSIF

In the East of the Rocroi Massif, the basal Devonian unconformity is poorly exposed. Meilliez (1989) shows, near Willerzie, that the basal Devonian is affected by a system of very open folds of limited amplitude, facing NNW and plunging slightly to the ENE (N060°E/10°NE), and that the whole sequence is affected by a weakly developed cleavage, dipping 15 to 50° SSE.

In the centre of the Rocroi Massif, along the Meuse River and in adjacent valleys, numerous structures are well exposed, but the absence of Lower Devonian sequence inhibits discrimination between Caledonian and Variscan structures. However, magmatic intrusions and microstructural arguments will be used instead.

The regional structural pattern consists of kilometric-scale ENE-WSW (N076°E) folds with axial-planes parallel to the principal cleavage. These folds were later affected by transverse fractures and minor folding (Beugnies, 1963 & 1972).

1) Magmatic intrusions

In the centre of the massif, diabase and micro-granite intrusions are quite common (Beugnies, 1963; Meilliez, 1981; Goffette, ms. 1986) and are considered to reflect a tectonic relaxation or distension event (D4). Geochronological data indicate a Middle to Upper Devonian U-Pb zircon age for this bimodal magmatism (Goffette *& al.*, 1991). In the Eastern zone of the Massif, at Willerzie, a volcano-sedimentary complex was recorded at the base of the Lower Devonian (Beugnies, 1968; Hanon, ms.1972; Roche, ms.1985; Meilliez, 1989). Thus, magmatism in the Rocroi massif, occurred mainly during the Devonian and hence these intrusions can be used as field marker to differentiate Caledonian from Variscan tectonic structures.

Meilliez (1981) investigated some thin diabase dykes near Revin. The dykes intersect bedding and small folds, and they are themselves affected by fracture cleavage. Their geometry seem to have been influenced by a pre-existing planar anisotropy, parallel to the axial-plane of the folds. Further South, near

Deville, Goffette (ms.1986) also found that dykes at least post-date one folding event and were affected by cleavage during later deformation.

2) Extensional structures

In the centre of the massif as in the Southern zone, boudinage is common in quartzite beds. It has a similar symmetric to asymmetric shape, with the boudins axis constantly oriented E-W to ENE-WSW (fig.13a). In the slates, mineral stretching lineations are less common, but they have the same N-S trend as in the Southern zone. If these structures are related to the Variscan event, as we postulate, they can be used as structural markers of the Variscan major deformation D5.

Near Revin, extensional structures occur in both quartzite and in slate beds. In the slates, they simulate the foliation-boudinage and large-scale extensional shear bands of the types described by Platt & Vissers (1980) and Lacassin (1988), in gneiss of amphibolite facies. Here, the same kinds of structures are found in slates of greenschist facies, in which they have never been described. By analogy, extensional structures in slates, resembling the foliation-boudinage in gneiss, will be called "cleavage-boudinage".

Inside Revin, a road section exposes a thick series of black slates of the Revinian Rv2 unit (Upper Cambrian). The slates show well-defined extensional structures (fig.11). Symmetric to asymmetric cleavage-boudinage and low-angle shear bands affect the pre-existing cleavage planes.

In another outcrop near Revin, composed of interbedded slates and quartzites, extensional structures are present both in the slates and in the quartzites. They probably initiated in the more competent quartzite layers, and propagated into the adjacent slates as quartz-filled tension gashes, which intersect the cleavage planes (fig.12).

In place, slickensides are also observed on cleavage planes. They plunge to the South and indicate a N-S sliding direction (fig.13a).

For the interpretation of these structures, reference is made to Platt & Vissers (1980). They demonstrated that symmetric and asymmetric foliation-boudinage in gneiss, forms in homogeneous rocks, characterized by a pronounced planar anisotropy and submitted to shortening at high angle to the anisotropy. Strong anisotropy limits the rate of ductile extension in the directions parallel to this plane. Extension then occurs by the opening of extensional fractures normal to the anisotropy (symmetrical foliation-boudinage) or oblique to it (asymmetrical foliation-boudinage). Also, low-angle shear bands develop, with relative displacement in a sense that causes extension along the anisotropy. The asymmetric structures are generally synthetic to the bulk sense of shearing (Lacassin, 1988).

In Revin, the presence of extensional structures affecting the slaty cleavage indicate that the slates already possessed a strong planar anisotropy, before being submitted to a new deformation. Cleavage-boudinage in slates are assumed to be equivalent to

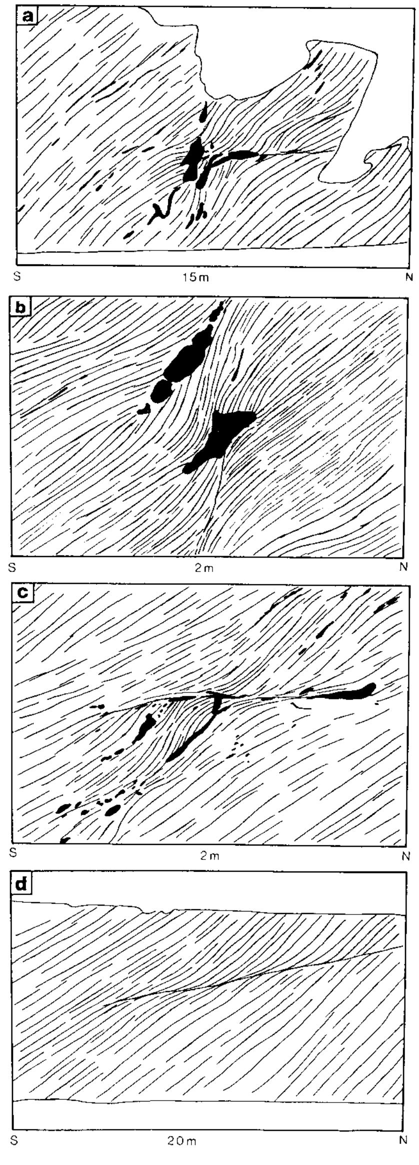

Fig. 11 - Extensional structures in a thick series of Rv2 black slates, near Revin (drawings from photographs). Observation in N-S cross sections, at right angle to the mean boudinage axis. Quartz-filled tension gashes are shown in black. (Definition and interpretation as in Platt & Vissers, 1980 and Lacassin, 1988).

a: Large-scale symmetric cleavage-boudinage, passing laterally to asymmetric cleavage-boudinage.

b: More evolved, metric-scale symmetric cleavage-boudinage.

c: Metric-scale asymmetric cleavage-boudinage, with a low-angle shear-band fracture. Inverse displacement along the fracture and reverse drag of the cleavage, particularly in the centre.

d: Large-scale shear band.

the bedding-boudinage in the quartzite layers and, thus, to be representative of the Variscan deformation. Because, during the Variscan deformation, the onset of boudinage occurs before the cleavage development (Lambert & Bellière, 1976 and herein), the pre-existing cleavage affected by extensional structures must be Pre-Variscan.

The occurrence of asymmetric extensional structures indicate that the new direction of maximum compression was not strictly perpendicular to the pre-existing cleavage, but was more inclined than the normal to the bedding or cleavage. This situation has let to inverse shearing movements parallel to the bedding or cleavage.

These characteristics are similar to those observed in the Southern zone of the Rocroi Massif, but, there, extensional structures are present only in the competent quartzite beds. This suggests that, in the Southern zone, the slates deform in a more ductile manner.

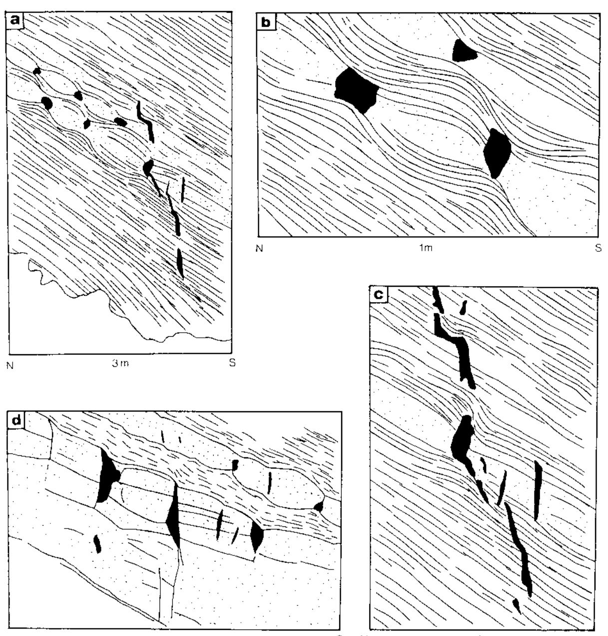

Fig. 12 - Extensional structures in a series of interbedded black slates and more quartzitic layers, near Revin (drawings from photographs). Observation in N-S cross sections, at high angle to the mean boudinage axis. Quartz-filled tension gashes are shown in black. (Definition and interpretation as in Platt & Vissers, 1980 and Lacassin, 1988).

a: Symmetric boudinage probably initiated in the more competent quartzitic layers. The quartz-filled tension veins propagate in the surrounding slates, cutting the cleavage planes obliquely.

b: Detail of symmetrical barrel-shaped boudinage of the more competent quartzitic layers.

c: Detail of tension gashes, of intermediate type between symmetric and asymmetric structures, suggesting asymmetric evolution of early symmetric structures.

d: Detail of asymmetric boudinage in quartzitic layer, with inverse rotation of isolated boudins.

3) N-S extension fractures

Numerous N-S quartz-filled extension fractures, similar to those in the Southern zone of the Rocroi Massif, are observed mainly in the quartzite layers. They intersect the beds at a high angle and the intersection lineations plunge 30-40°S (fig.13b). They usually occur in these competent beds bearing E-W boudinage structures. Some are clearly intersected and even displaced by the E-W interboudinal tension veins. Using the boudinage as a Variscan structural marker, the N-S extension fractures should be pre-Variscan.

4) Minor folds

Minor, tight to isoclinal folds are common near Revin, as elsewhere in the Rocroi massif. The orientations of their axes show considerable

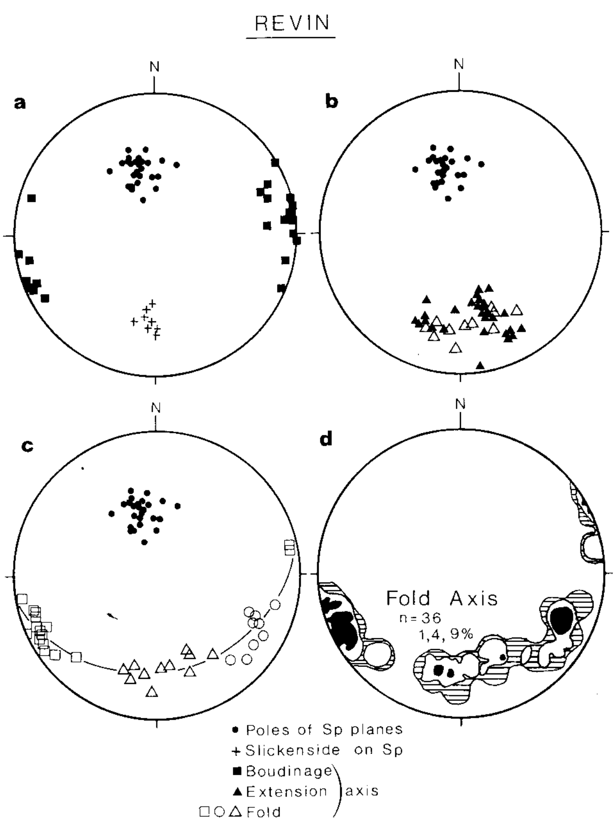

Fig. 13 - Schmidt nets, lower hemisphere projection, for Revin, in the centre of the Rocroi Massif (Revinian, Upper Cambrian).

a: Poles of cleavage planes (Sp), axes of boudinage in quartzite layers and orientation of slickensides on cleavage planes.

b: Poles of cleavage planes (Sp) and axes of N-S structures of probable Caledonian origin: sub-isoclinal folds and intersections between extension fractures and bedding planes.

c: Distribution of fold axes of various origin, along a great circle whose Pl-axis coincides with the mean pole of cleavage planes.

d: Density surfaces per 1% area for fold axis of net 13c. Clustering in three sets, with maxima towards ENE-WSW, S and SE directions.

dispersion, along a great circle corresponding to the mean cleavage plane (fig.13c). The data from Beugnies (1972), Meilliez (1981) and fig.13c are not sufficient to allow reliable statistical analysis, but a tentative classification of the folds into three sets can be made :

(1) SSE to SSW plunging folds, subparallel to the N-S extension fractures and associated with the axial-plane cleavage (probable Caledonian D1);

(2) ENE-WSW to E-W trending folds, parallel to the general trend of the boudinage and to the major folds, but at high angle to the mineral extension lineation (associated with the major E-W Caledonian D2 or Variscan D5 structures);

(3) intermediate SE plunging folds, most probably of Variscan origin (D5).

Transverse folding and faulting with simultaneous deformation of bedding and cleavage are described by Beugnies (1972, near Revin) and Baudet (1981, near Fumay). No direct field evidence is available to decide whether they are of late-Caledonian (D3) or late-Variscan age. Beugnies (1963) mapped transverse folds and faults throughout the Rocroi Massif and found them to affect also the Lower Devonian cover. But the N-S folds described in the centre of the massif are somewhat analogous with the N-S upright folds (D3) observed in the Southern zone (Delvaux de Fenffe & Laduron, 1984) which are intersected by the basal Fépin beds. Further investigation is needed to specify their relative ages.

IV. NORTHERN ZONE OF ROCROI MASSIF

The Devonian unconformity in the North of the Rocroi Massif is well exposed in the Meuse Valley, near Fépin, and in the Laonry quarry, along the Pernelle Valley. On a regional scale, the ENE-WSW to NE-SW major folds in the Cambrian basement are clearly intersected by the unconformity of the Lower Devonian. This upper sequence is affected by more open, E-W folds, which are very different from those underneath the unconformity (Waterlot, 1937, Beugnies, 1963).

1) Fépin unconformity (Meuse Valley)

Along the Meuse Valley, the Lower Devonian unconformity on the Cambrian sequence is well exposed in the cliffs of the "Signal de Fépin". The base of the Fépin formation (Lower Devonian) is folded in an asymmetrical syncline, facing to the NNW and of hectometric scale (Fépin Syncline: Gosselet, 1879, Anthoine, 1940). This structure is interpreted (Meilliez & Mansy, 1990) to be the result of progressive non-coaxial deformation during the major Variscan folding event (formation of a thrust sheet, followed by folding and cleavage development and ending with local rotation of both bedding and cleavage).

In the normal flank of the Fépin syncline, the Fépin Conglomerate lies unconformably on tight to isoclinal folds belonging to the Revinian (fig.14). The best example is a fold of 20-30 m scale, with axial plane dipping to the NNW and Pi-axis gently plunging to the ENE (fig.15a,b). Both limbs of this fold are overlain directly by the subhorizontal Fépin formation. This fold is associated with axial-plane slaty cleavage, minor folds and boudinage structures, of the same ENE trend as the Pi-axis (fig.15a).

In the steep flank of the Fépin syncline, cleavage planes with a constant dip of 10-30°S (fig.15c), lie at a high angle to the Fépin Conglomerate, in subvertical position. The main small-scale structures in the Cambrian rocks are SE to SW-plunging, tight to isoclinal folds with associated N-S extension fractures (fig.15c), and ENE-WSW boudinage (fig.15d). Boudinage structures are common and can be related to Variscan deformation, while the NE to SW folds and extension fractures are thought to reflect earlier Caledonian deformation. Where ENE-WSW boudinage and N-S folds or extension fractures are superimposed, the ENE-WSW boudinage intersects the earlier N-S structures.

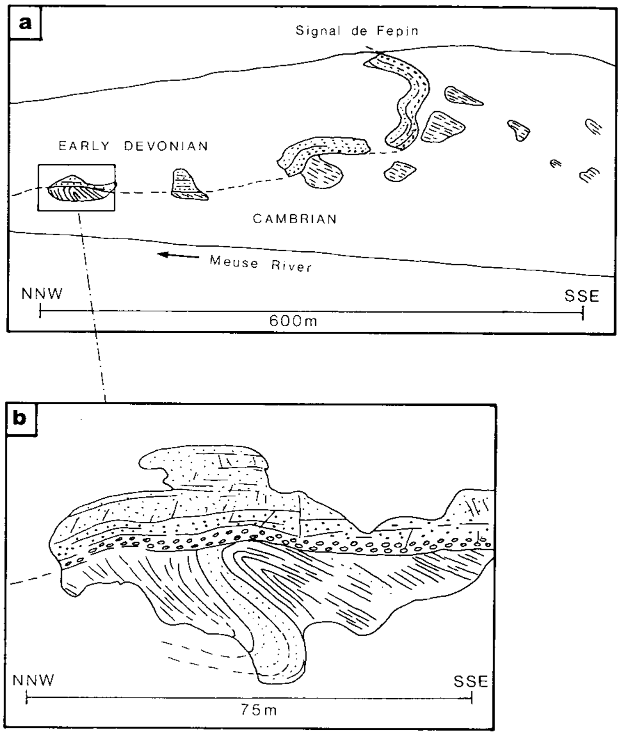

Fig. 14 - General structure of the "Signal de Fépin" unconformity, in the Northern zone of the Rocroi Massif (for more details: Anthoine, 1940 & Meilliez, 1989).

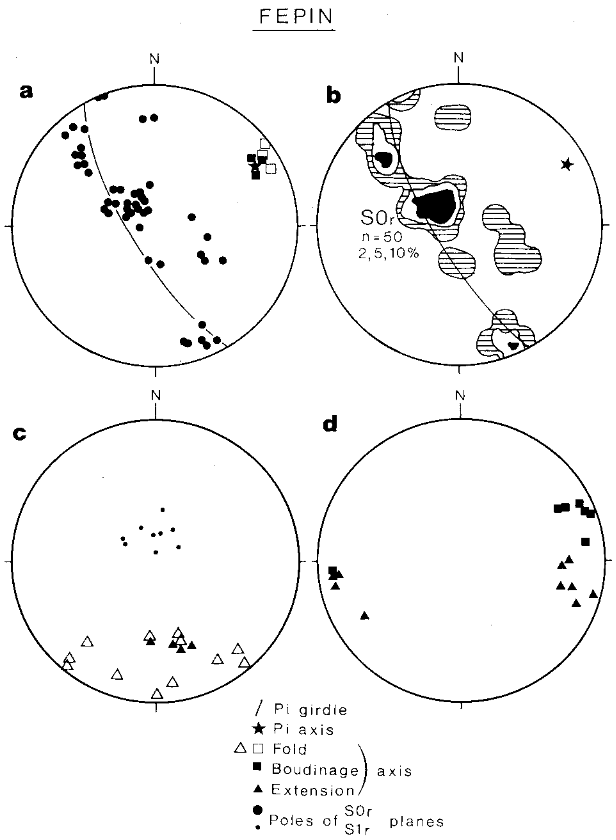

Fig. 15 - Schmidt nets, lower hemisphere projection for the Fépin unconformity, in the Northern zone of the Rocroi Massif (Revinian, Upper Cambrian).

a: Poles of bedding planes (S0r), along an inclined fold intersected by the Lower Devonian unconformity, in the normal flank of the Fépin Variscan syncline. Great circle whose PI-axis corresponds to the orientation of some minor folds and boudinage structures observed in the vicinity.

b: Density surfaces per 1% area for S0r poles of net 15a.

c: Poles of cleavage planes (S1r), fold axis and intersections of N-S extension fractures with bedding, in the steep flank of the Fépin Syncline (Revinian).

d: Axis of boudinage and E-W extension fractures in the Revinian rocks of the steep flank of the Fépin Syncline.

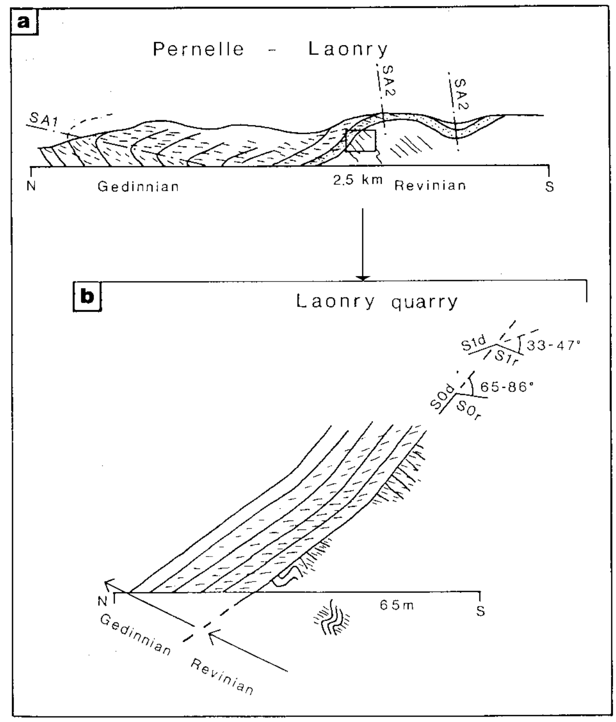

Fig. 16 - Structure of the Pernelle Valley and the Laonry quarry, in the Northern zone of the Rocroi Massif (adapted from Stasse, ms.1988).

a: General Variscan structure of the Lower Devonian (Lochkovian), showing a recumbent, northward facing and kilometric-scale anticline (D5), with gently southwards dipping axial-plane and cleavage. It also shows a later upright fold system (D6) with subvertical axial plane, affecting the normal flank of the recumbent D5 anticline. The Laonry quarry is situated in the North flank of this later anticline.

b: Schematized angular relationships in the Laonry quarry between bedding planes (S0) and cleavage planes (S1), in each side of the unconformity (d=Devonian, r=Revinian, Upper Cambrian).

From these relationships, it appears that, in Fépin, Variscan deformation in the basement occurs, at least partly, by passive reactivation of the Caledonian slaty cleavage. This deformation is mainly expressed by extensional structures (bedding-boudinage and cleavage-boudinage), and also by some minor folds and slickensides on cleavage planes. The cleavage in the Cambrian sequence has a constant dip throughout the area, and, apparently, was only slightly reoriented during the Variscan folding. Therefore, the deformation in the basement during the Variscan event could be modelled by a combination of (1) a renewed flattening and ENE-WSW extension along the planes of anisotropy and (2) shearing along the pre-existing planes of anisotropy. This anisotropy comprises both cleavage and bedding planes, which are often subparallel as a consequence of the tight to isoclinal Caledonian folding.

Therefore, in the Cambrian sequence, the main folding and the slaty cleavage should be Caledonian, and the boudinage, Variscan. This is a reaffirmation of the suggestion made by Gosselet (1879), that the cleavage of the Revinian slates at the "Signal de Fépin" was acquired before the deposition of the Lower Devonian Fépin formation.

2) Laonry unconformity (Pernelle Valley)

The Laonry unconformity in the Pernelle Valley has been much studied, (Fourmarier, 1909, Anthoine, 1940; Asselberghs, 1946; Stasse, ms.1988; Malengreau *& al.*, in preparation). The structure of this area consists of a kilometric-scale recumbent Variscan D5 anticline, facing northward, with subhorizontal axial plane and cleavage (fig.16a & 17a). Its normal flank was later deformed into an open hectometric-scale upright D6 anticline - syncline system. The unconformity is exposed mainly in the Laonry quarry (fig.16a), in the Northern flank of the upright D6 anticline. Cleavage and bedding measurements in the Lower Devonian from the two flanks of this fold indicate that the slaty cleavage has been rotated with the bedding, during this later upright D6 folding (stereonet 17c-d). Consequently, the Variscan structure in the Pernelle Valley can, as in the Meuse Valley, be regarded as the result of a progressive deformation.

Stasse (ms.1988) demonstrated that the bedding in the Revinian sequence dips generally towards the South while the basal Devonian series (Fépin Conglomerate) dips northwards (fig.16b). Along this unconformity, the mean angular discordance is 65-86° between the Revinian and Devonian bedding (S0) and 33-47° between associated cleavage planes.

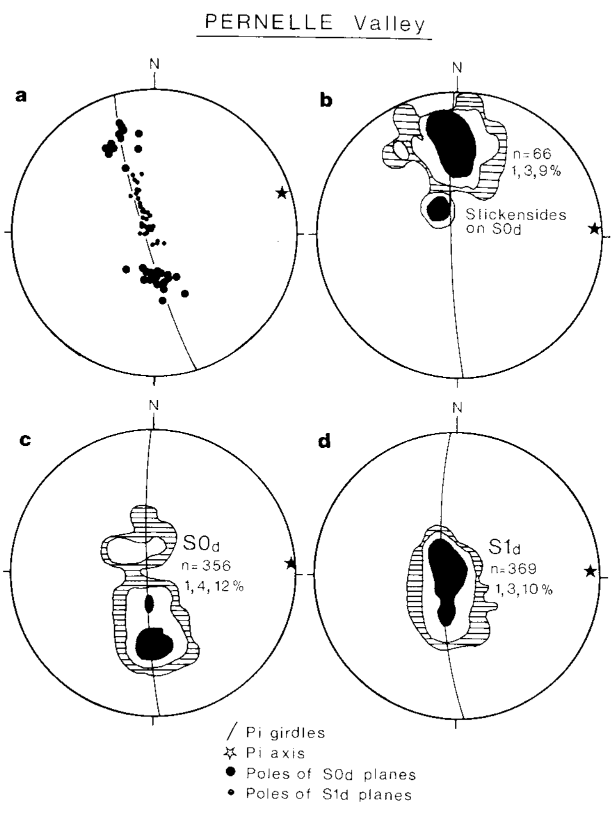

Fig. 17 - Schmidt nets, lower hemisphere projection for Variscan structure of the Lower Devonian rocks along the Pernelle Valley, Northern zone of the Rocroi Massif. Density surfaces per 1% area.

a: Poles of bedding planes (S0d) and cleavage planes (S1d) for the recumbent northward facing kilometric-scale D5 anticline. Great circle with PI-axis plunging slightly to the ENE.

b: Slickensides on bedding planes, observed in the normal flank of the D5 anticline and indicating a general N-S transport direction.

c-d: Poles of bedding planes (S0d) or cleavage planes (S1d) along the late Variscan D6 upright fold affecting the normal flank of the early Variscan D5 recumbent anticline. The great circles with PI-axis plunge slightly to the East.

Some folds are also unconformably overlain by the basal conglomerate. They are relatively tight (interlimb angle of 60° to 120°), have WSW axial direction (N246°E/22-30°SW) and axial planes of variable inclination (fig.18c). The cleavage in the Revinian sequence is wrapped around the folds, keeping relative parallelism with bedding. This is clearly shown in the stereonet (fig.18d), where the poles of cleavage planes are dispersed along a great circle, with a PI-axis plunging SW (N244°E/23°). The latter is very similar to that defined by the poles of bedding planes whose PI-axis plunges SW (N246°E/25°) (fig.18c). Crenulation lineations are also found with the same orientation (fig.18d). These observations suggest the existence of superimposed deformation in the basement, beginning with cleavage development and followed by ENE-WSW tight folding. The age of the latter, however, may be questionable. The folds observed in the Cambrian sequence of the Laonry quarry are adjacent to the discordance surface and may have been induced by the Variscan tectonics. However, several arguments favour a Caledonian origin of these folds:

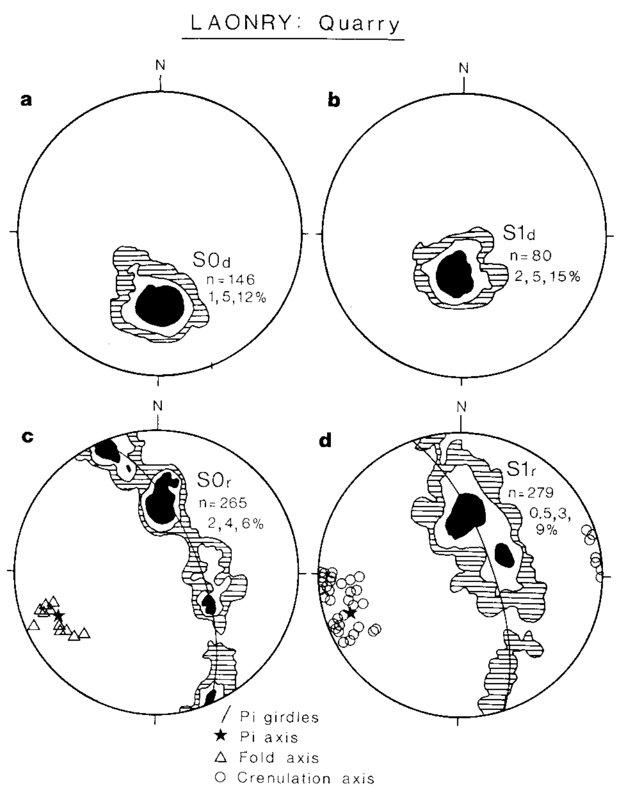

Fig. 18 - Schmidt nets, lower hemisphere projection for the unconformity in the Laonry quarry, Pernelle Valley, Northern zone of the Rocroi Massif. Density Surfaces per 1% area.

a-b: Poles of bedding planes (SOd) or cleavage planes (S1d), in the Lowermost Devonian (Fépin formation).

c-d: Poles of bedding planes (S0r) or cleavage planes (S1r) and axis of fold or crenulation cleavage, in the Revinian. The PI-axis of the great circle coincides with the mean fold or crenulation axes.

- The Devonian strata overlying the folded Revinian are nearly monoclinal at the scale of the quarry (fig.18a);

- There is no tectonic interface between the Lower Palaeozoic and the basal layer of the Fépin conglomerate. The Revinian beds which are generally at high angle to the Fépin beds are generally indented into the basal conglomerate. As a result, the basal Fépin layer is anchored to the underlying rocks, as in the Southern zone of the Rocroi Massif. The main movements necessary to accommodate the folding of the Lower Devonian cover probably took place in the thin pelitic layers between the thick competent beds of the Fépin sequence. Numerous slickensides on the Lower Devonian bedding planes indicate a N-S sliding direction, at right angle to the Variscan E-W trend (fig.17b).

- The Variscan folds in the cover plunge towards the ENE (N73°E/16°), while the fold axes in the basement plunge to the WNW (N250°E/20°).

- In the Cambrian outcrops surrounding the Laonry quarry, few folds are observed, but the dispersion of S0 poles in stereonet also occur along a great circle with PI-axis plunging to the West (N262°E/08° : fig.19a). The S1 poles also display similar dispersion (fig.19b).

In summary, the high angle unconformity of both bedding and cleavage planes and the absence of cleavage continuity through the unconformity suggest the existence of Caledonian S1 cleavage. This cleavage is little affected by the Variscan deformation. However, this cleavage was probably deformed later, by the ENE-WSW minor folds. These folds are overlain by the Lower Devonian unconformity and are probably related to the major Caledonian folding event (D2).

This is the first place where such a structural relationship between minor and major Caledonian structures is found. In the Southern zone and the centre of the Rocroi Massif, the Caledonian cleavage is related to N-S sub-isoclinal folds and N-S extension fractures. The ENE-WSW Caledonian folds observed in the Laonry quarry could be related to the major Caledonian folding event. Consequently, the relationships between the cleavage and the ENE-WSW folds suggest that the major cleavage development and its related N-S folds and tension fractures predates the major ENE-WSW folding event, both being of Caledonian age. This imply the first structures to be D1 and the latter, D2.

The stereograms for the Lower Palaeozoic surrounding the quarry show a slight dispersion of both

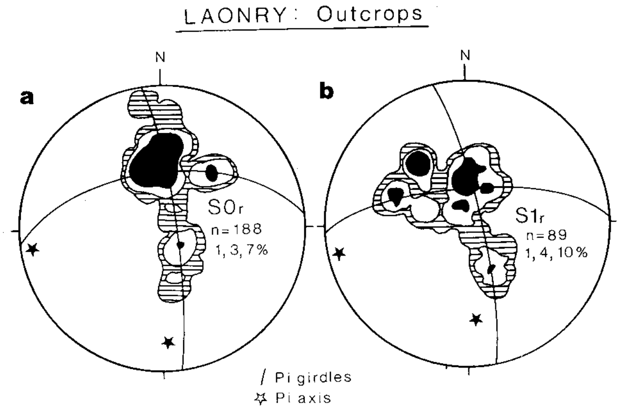

Fig. 19 - Schmidt nets, lower hemisphere projection for the outcrops surrounding the Laonry quarry, in the Northern zone of the Rocroi Massif (Revinian, Upper Cambrian). Surfaces per 1% area.
a-b: Poles of bedding planes (S0r) and cleavage planes (S1r). Principal distributions along great circles with PI-axis plunging to the West and secondary distributions along great circles with PI-axis plunging to the South.

S0 and S1 planes along a great circle with a PI-axis plunging South (fig.19a,b). This could indicate the presence of minor undulations which may also be of Caledonian origin (D3).

V. STRUCTURAL SYNTHESIS OF ROCROI MASSIF

From the field evidence summarized in this paper, the structural evolution of the Lower Palaeozoic basement in the Rocroi Massif may be subdivided into two periods of tectonic compression, separated by a period of tectonic relaxation or distension. The structural characters of each of the three events, are presented here, as a working hypothesis. However, it should be emphasized that several aspects of the proposed structural evolution are still debatable.

1) Caledonian compressional event (Caradocian)

The Caledonian major compressional event is incompletely known, but enough evidence exists to suggest that relatively intense and complex deformation occurred in the Cambro-Ordovician basement. Field observations indicate Caledonian microstructures with dominant N-S axial directions, but observations over a wider area indicate the existence of dominant ENE-WSW regional structures.

Caledonian microstructures comprise N-S tight to sub-isoclinal folds, axial-plane slaty cleavage and N-S extension fractures. Dominant are the axial-plane slaty cleavage and the N-S extension fractures. They were probably formed by deformation with its principal compressive axis at high angle to the bedding and its principal extension axis, E-W.

The regional structure is an E-W to ENE-WSW kilometric-scale fold, with axial-plane parallel to the regional cleavage. Some associated minor ENE-WSW folds are observed in Fépin and Laonry. Their age is undoubtedly Caledonian as they are intersected by the Lower Devonian unconformity.

The E-W to ENE-WSW folds are incompatible with the N-S folds and associated extension fractures, and this suggests that they reflect two distinct deformation events. However, it is difficult, to assign them a relative age. The relationship between the two events has only been observed in Laonry quarry, where the principal cleavage was deformed, like the bedding, in the ENE-WSW folds which are overlain and cut by the unconformity. This indicates that the original development of the slaty cleavage, and the related N-S microfolds and extension fractures, probably belong to a first deformation (D1), and that the E-N to ENE-WSW folding, to a second deformation (D2). For the latter, it is not know wether the D2 folding is associated with a new cleavage development (S2) or wether it has caused the reactivation of the older one (S1).

The transverse N-S upright folding of decametric to hectometric scale, mainly developed in the Southern zone of the Rocroi Massif, probably represents a later stage in the Caledonian event (D3). This folding has no associated axial-plane cleavage and affected the earlier Caledonian cleavage.

(2) Tectonic relaxation and distension

Between the end of the Caledonian deformation (Upper Ordovician) and the onset of the Variscan tectonic activity (Namurian, after Paproth, 1987), there was a long period of tectonic relaxation and distension (D4), with magmatic activity and sedimentation.

Two types of magmatic activity were recognized: a Lower Devonian volcano-sedimentary complex at the eastern border of the massif and some Middle to Upper Devonian diabase and microgranite intrusions in the centre.

The Lower Devonian sediments recorded an highly variable rate of subsidence in the whole area between the High-Ardenne and the Midi fault (Bless *& al.*, 1989 ; Meilliez, 1989). From Middle Devonian to Namurian, the rate of subsidence decreased rapidly and remained low until early Namurian (relaxation of tensional tectonics). During Namurian times, a progressive inversion from tensional to compressional deformation caused a rapid increase in the sedimentation rate, with strong clastic input (Bless *& al.*, 1989). This latter event marks the onset of the Variscan deformation.

(3) Variscan compressional event (Namurian to Westphalian)

The effects of the Variscan major compressional event (D5) are difficult to identify in the Lower Palaeozoic basement. This latter was highly anisotropic before the Variscan compression, while the Devonian cover was considerably less so. It is postulated that this physical contrast is the principal cause of the differences in style and intensity of deformation between the basement and its cover. In the Lower Palaeozoic basement, the Variscan deformation mainly reactivated the Caledonian structures and cleavage. The intensity of this reactivation across the Rocroi Massif decreases from the South to the North, in parallelism with the regional evolution of metamorphism and deformation in the whole Ardenne.

In the Southern zone of the Rocroi Massif, the main Variscan deformation causes (1) the reworking of the pre-existing cleavage in a mostly ductile process (synkinematic recrystallization of very fine micas, pressure shadow development, N-S mineral extension lineation, and crenulation cleavages), and (2) the development of boudinage, mullions and minor folds. The plane of principal flattening was more steeply inclined than the pre-existing Caledonian anisotropy. This is indicated by the structural polarity of rotated asymmetrical boudins and minor folds. It is also consistent with the Variscan cleavage planes, in the Devonian cover, a few kilometres South of the unconformity, being more steeply inclined (50-70°S) than the cleavage in the basement (25-40°S).

In the centre of the massif, cleavage-boudinage and extensional shear bands in slates indicates that the Variscan deformation resulted from flattening at high angle to a pre-existing cleavage. The cleavage in igneous rocks, E-W boudinage in quartzites and N-S striae on cleavage planes are all Variscan. However, no certified Variscan folds have been observed. The asymmetry and rotation sense of boudinage structures also indicates that the principal compression direction was more steeply inclined than the normal to the pre-existing cleavage.

In the Northern zone, the Variscan deformation also caused renewed flattening at high angle to the older slaty cleavage (cleavage-boudinage in slates, bedding-boudinage in quartzite). Variscan recumbent folding of the whole Lower Palaeozoic and Lower Devonian sequences possibly induced shear movements along cleavage and bedding. In the Lower Devonian, the mean attitude of cleavage planes is 20°N-30°S, depending on rotations associated with the later Variscan D6 deformations. In the basement, the cleavage planes near the unconformity have a relatively constant dip of 30-40°S at Laonry and 10-30°S at Fépin.

In conclusion, the cleavage planes have a relatively constant dip of 30-40°S throughout the Lower Palaeozoic of the Rocroi Massif. However, the cleavage planes in the Lower Devonian cover, dip more than 40°S in the Southern zone and less than 30°S in the Northern zone. In consequence, the Lower Palaeozoic acted as a relatively rigid body during the major Variscan deformation D5. It has imposed the general trend and inclination of bedding and cleavage planes, by reactivation of the older anisotropy. The present structure of the Lower Palaeozoic is then mainly inherited from the Caledonian orogeny.

The Variscan compressive event is also characterized by a second series of structures (D6 stage), which mark a second step in the course of the Variscan progressive deformation. The kink bands and deformation of the slaty cleavage observed in the Neufchâteau syncline (Fourmarier *& al.*, 1962; Delvaux de Fenffe, 1991) and the upright ENE-WSW folding in the Northern zone of the Rocroi Massif, at Fépin (Meilliez & Mansy, 1990) and at Laonry (Stasse, ms.1988; Malengreau *& al.*, in preparation) probably belong to the same D6 deformation stage. Further North, in the Southern zone of the Dinant Synclinorium, the same superposition of upright longitudinal folds on overturned subhorizontal folds was observed in the Middle Devonian series (Delvaux de Fenffe, 1985).

It is also worth to remember the existence of a series of normal faults which develop along the Southern zone of the Dinant Synclinorium, with maximum downthrow of more than 1000m (Delvaux de Fenffe, 1985 & 1990). They affects all the Variscan folds in that area, and they could represent the effects of a relaxation step in the Variscan deformation (D7).

VI. OTHER CALEDONIAN MASSIFS

In the Ardenne, Cambro-Ordovician rocks also outcrop in three other massifs: the Stavelot-Venn, in the North-East (see Bless *& al.*, 1990) and the small massifs of Serpont and Givonne.

In these regions, several workers have distinguished superimposed deformations. All Cambro-Ordovician rocks are affected by Caledonian regional folding, of kilometric scale, with an axial-plane cleavage and locally, a later crenulation cleavage. It is generally accepted that, during the Variscan event, cleavage development post-dates the peak of metamorphism. In some areas of the Stavelot-Venn Massif, Geukens (1969), Theunissen (1970), Mukhopadhyay (1974) and Fielitz (1985) distinguished two superimposed cleavages. The first one, a slaty cleavage, is older than the metamorphic porphyroblasts and the second one, a crenulation cleavage, post-dates porphyroblast growth. However, only Geukens (1969) considered that the first cleavage was Caledonian. More recently, Spaeth *& al.* (1985) demonstrated that slaty cleavage and a system of microfolds are both affected by the Helle tonalite intrusion, dated at 381±16 Ma (Lower Devonian) by Kramm & Buhl (1985). In addition, Geukens (1969 & 1976) reported slate clasts, with one or two cleavages, in an intrusion, and clasts of slate and magmatic rock, in the basal Lower Devonian conglomerate. In the Konsen borehole, in the SE part of the Stavelot-Venn Massif, Fielitz (1985) recognized two distinct tectonic events, each with cleavage development and folding.

In the Stavelot-Venn Massif, Schreyer (1975) and Kramm *& al.* (1985) consider that metamorphism in the Cambro-Ordovician rocks was mainly of Variscan origin. Illite crystallinity profiles across the NW unconformity (Spaeth *& al.*, 1985) suggest that metamorphism was not initiated during the Caledonian event. The maximal thermal effect recorded, reached only the anchimetamorphism level of Kubler (1967).

In the small Serpont and Givonne Caledonian massifs, the structure is poorly exposed and little information is available. The only good outcrop in the Serpont Massif shows the existence of Caledonian folds and axial-plane cleavage, reactivated during the Variscan deformation (Geukens & Richter, 1964; Beugnies, 1976).

VII. CONCLUSIONS

Although the detailed structural evolution of the Caledonian massifs of High-Ardenne is incompletely known, it appears that the general deformation history is relatively similar for all of the Ardenne's basement. This work reasserts the importance of the Caledonian event in the Ardenne and highlights the overprint of the Variscan deformation in the basement, as a reactivation of the Lower Palaeozoic Caledonian structures.

The regional effects of the Caledonian event in the Ardenne are (1) intense large to minor scale folding, (2) slaty cleavage development, (3) extension fractures, and (4) a limited thermal effect (probably less intense than the Variscan thermal metamorphism). In the Rocroi Massif, a more detailed survey allows the subdivision of the Caledonian event into three successive deformations:

D1- a regional deformation, with slaty cleavage and the formation of N-S sub-isoclinal folds and N-S tension fractures (subhorizontal E-W extension);

D2- regional large to medium scale ENE-WSW folding, with axial-planes moderately dipping to the South, and development of new cleavage, or reactivation of the pre-existing one;

D3- later N-S upright folding, which has only been recognised in the Rocroi Massif (subhorizontal E-W compression).

The tectonically quiescent period between Caledonian and Variscan major deformation phases is marked by Late Ordovician to Early Devonian magmatism and Devonian-Carboniferous sedimentation in an extensional context:

D4- intermediate period of extension with a mainly Devonian magmatic activity.

The Variscan deformation in the basement was greatly influenced by the pre-existing anisotropy. The principal direction of compression was at high angle to the pre-existing anisotropy (cleavage and bedding planes). The basement reacted in a relatively passive manner, by renewed shortening in a direction normal to the existing anisotropy, and by N-S extension with shear movements along the anisotropy. The deformation of the Lower Devonian cover took place mainly by folding and cleavage development, but was also greatly influenced by the Lower Palaeozoic structures. The Variscan progressive deformation is divided into three recognizable steps:

D5- major Variscan deformation with reactivation of the Caledonian structure in epizonal metamorphic conditions;

D6- minor kink band formation in the Neufchâteau Synclinorium and hectometric-scale E-W upright folding in the Northern zone of the Rocroi massif and in the Southern part of the Dinant Synclinorium;

D7- important normal faulting in the Southern part of the Dinant Synclinorium, as a relaxation step.

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REFERENCES

ANDRE L. & S. DEUTSCH (1984) - Les porphyres de Quenast et de Lessines: Géochronologie, Géochimie isotopique et contribution au problème de l'âge du socle précambrien du Massif du Brabant (Belgique). Bull. Soc. belge Géol., Bruxelles, 93, 375-384.

ANDRE L., S. DEUTSCH & J. HERTOGEN (1986) - Trace-element and Nd isotopes in shales as indexes of provenance and crustal growth: the Early Palaeozoic from the Brabant Massif (Belgium). Chemical. Geol., 57,, 101-115.

ANDRE L., J. HERTOGEN & S. DEUTSCH (1986) - Ordovician-Silurian magmatic provinces in Belgium and the Caledonian orogeny in Middle Europe. Geology, 14, 879-882.

ANTHOINE R. (1940) - Contribution à l'étude du Massif cambrien de Rocroi. Mem. Acad. roy. Sc. Belg., Bruxelles., 12(4).

ASSELBERGHS E. (1946) - L'Eodévonien de l'Ardenne et des régions voisinnes. Mém. Inst. géol. Univ. Louvain, 14, 598p.

BAUDET D. (1981) - Elements structuraux du Cambrien des carrières du Bois de Fumay (Massif de Rocroi, Ardenne française). Ann. Soc. géol. Nord., 104, 99-108.

BEUGNIES A. (1963) - Le Massif cambrien de Rocroi. Bull. Serv. carte géol. France, 270, 155p.

BEUGNIES A. (1972) - Le site géologique de la station de transfert d'énergie de Revin. Ann. Soc. géol. Belg., Liège, 95, 335-343.

BEUGNIES A. (1976) - Structure et métamorphisme du paléozoïque de la région de Muno, un secteur-clef du domaine hercynien de l'Ardenne. Ann. Mines Belg., 6, 481-509.

BEUGNIES, A. (1969) - Les roches à quatrz dihexaédriques du Franc-Bois de Willerzie. Bull. Soc. belge Géol., Bruxelles, 72, 311-329.

BEUGNIES A., P. DUMONT, F. GEUKENS, G. MORTELMANS & M. VANGUESTAINE (1976) - Essai de synthèse du Cambrien de l'Ardenne. Ann. Soc. géol. Nord. Lille, 96, 263-273.

BLESS M., J. BOUCKAERT, T. CALEMBEECK, L. DEJONGHE, A. DEMOULIN, C. DUPUIS, P.J. FELDER, F. GEUKENS, F. GULLENTOPS, L. HANCE, J. JAGT, E. JUVIGNE, U. KRAMM, A. OZER, A. PISSART, F. ROBASZYNSKI , R. SCHUMACKER, A. SMOLDEREN, G.SPAETH, Ph. STEEMANS, M. STREEL, G. VANDENVEN, M. VANGUESTAINE, R. WALTER & M. WOLF. 1990. The Stavelot Massif from Cambrian to Recent. A survey of the present state of knowledge. Ann. Soc. géol. Belg., Liège, 113, 1-21.

BLESS M., J. BOUCKAERT, M.A. CALVER, L. DEJONGHE, J.M. GRAULICH, M. HORN, W. KIMPE, J. KULLMANN, J.-P. MEESSEN, D. NAYLOR, J.T. OLIVERA, E. PAPROTH, F. PARIS, J.C. PERDIGAO, A. RIBEIRO, M. ROBARDET, L. SANCHEZ de POSADA & J. TRUYOLS (1977) - Y a-t-il des hydrocarbures dans le Pré-Permien de l'Europe Occidentale? Serv. géol. Belg., Prof. Paper, 148, 11, 24-39.

BLESS M.J.M., J. BOUCKAERT & E. PAPROTH (1989) - The Dinant nappes: A model of tensional listric faulting inverted into compressional folding and thrusting. Bull. Soc. belge Géol., Bruxelles, 98, 221-230.

DANDOIS P. (1981) - Diagenèse et métamorphisme des domaines calédonien et hercynien de la vallée de la Meuse entre Charleville-Mézières et Namur (Ardenne franco-belges). Bull. Soc. belge Géol., Bruxelles, 90, 299-316.

DANDOIS P. (ms.1984) - Le métamorphisme des terrains paléozoïques de la partie médio-occidentale de l'Ardenne. Une approche de l'extension de la zone métamorphique par l'étude des minéraux phyllosilicates de la fraction argileuse des roches silico-alumineuses. Thèse, Univ. Louvain-la-Neuve, 200p.

De BETHUNE, P. (1986) - Esquisse historique des théories sur le métamorphisme en Ardenne. Ann. Soc. géol. Nord., 105, 115-119.

DELVAUX de FENFFE D. (1985) - Géologie et tectonique du Parc de Lesse et Lomme au bord sud du bassin de Dinant (Rochefort, Belgique). Bull. Soc. belge Géol., Bruxelles, 94, 81-95.

DELVAUX de FENFFE D. (1990) - Structures tardi-et post-hercyniennes dans le bord sud du Synclinorium de Dinant, entre Han-sur-Lesse et Beauraing (Belgique). Bull. Soc. belge Géol., Bruxelles, 112, 317-325.

DELVAUX de FENFFE, D. (1991). Kink bands from the Neufchâteau Synclinorium, french Ardenne. Meuse-Rhine Geologist Meeting at Mont-Rigi (Hautes-Fagnes, Stavelot Massif), 18-19 may 1990. Ann. Soc. géol. Belg., 113 (2) : 126-127.

DELVAUX de FENFFE D. & D. LADURON (1984) - Analyse structurale au bord sud du Massif de Rocroi (Ardenne françaises). Bull. Soc. belge Géol., Bruxelles, 93, 11-26.

DELVAUX de FENFFE D. & D. LADURON (1986) - Orogenèses calédoniennes et hercyniennes dans le bord sud du Massif de Rocroi. In: Groupes de Contacts F.N.R.S.. "Structure et évolution de la croûte terrestre", Bruxelles, 19 avril, 354-387.

DELVAUX de FENFFE D. & D. LADURON (1987) - Réponse à Messieurs Cl. LE CORRE et H. HUGON concernant leur article "Au sujet de l'analyse structurale au bord sud du Massif de Rocroi (Ardenne françaises)". Bull. Soc. belge Géol., Bruxelles, 96, 53-54.

DUCHATEAU J. (ms.1983) - Etude structurale au bord sud du Massif de Rocroi dans la région de Naux. Mém. Licence, Univ. Louvain-la-Neuve, 45p.

FIELITZ W. (1985) - Structural inventory or the core profile ot the research borehole Konzen, Hohes Venn (West Germany). N. Jb. Geol. Paläont. Abh., 171, 63-73.

FOURMARIER P. (1909) - Le contact du Dévonien et du Cambrien dans la vallée du ruisseau de Pernelle, au sud de Couvin. Ann. Soc. géol. Belg., Bruxelles, 36, 211-214.

FOURMARIER P. (1931) - Les plissements calédoniens et les plissements hercyniens en Belgique. Ann. Soc. géol. Belg., Liège, 54, B364-384.

FOURMARIER P. (1944) - Une anomalie de la schistosité dans le Dévonian de la Semois. Ann. Soc. géol. Belg., Liège, 67, B29-36.

FOURMARIER P. (1951) - L'âge de la schistosité du Cambrien du Massif de Stavelot. Bull. Acad. Roy. Belg. Cl. Sc., sér.5, 37, 341-347.

FOURMARIER P. (1954) - Tectonique. In Prodrôme d'une description géologique de la Belgique. Soc. géol. Belg. (ed.). Liège: Vaillant-Carmanne, 826p., 1 carte.

FOURMARIER P., J.M. GRAULICH & L. LAMBRECHT (1962) - Les effets d'une phase tardive du plissement hercynien sur le versant nord du synclinorium de Neufchâteau. Ann. Soc. géol. Belg., Liège, 85, 357-370.

GEUKENS F. (1969) - De ouderdom der druksplijting in het caledonisch massief van Stavelot. Méd. Kon. Vl. Acad. Kl. Wet. Brussel, XXXI(4), 3-13.

GEUKENS F. (1976) - L'âge des roches éruptives dans le Massif de Stavelot. Ann. Soc. géol. Belg., Liège, 100, 615-618.

GEUKENS F. (1978) - Geologishe profil durch das Stavelot-Venn Massiv. Deutsche Geol. Gesell. Excursionführer. Aachen, 37-44.

GEUKENS F. & D. RICHTER (1962) - Problèmes géologiques dans le Massif de Serpont (Ardenne). Bull. Soc. belge Géol., Bruxelles, 70, 196-212.

GOFFETTE O. (ms.1986). Filons magmatiques du Massif de Rocroi (Ardenne française): Relations structurales avec l'encaissant, pétrologie structurale, pétrologie, géochimie. D.E.A. Univ. Lille, 77p.

GOFFETTE O., J.-P. LIEGEOIS & L. ANDRE (1991). Age U-Pb dévonien moyen à supérieur des zircons du magmatisme bimodal du Massif de Rocroi (Ardenne, France): Implications géodynamiques. C. R. Acad. Sci., Paris. 312, II, 1155-1161.

GOLDSTEIN A.G. (1988). Factors affecting the kinematic interpretation of asymmetric boudinage in shear zones. - J.Struct.Geol., 10, 707-715.

GOSSELET J. (1879) - La roche à Fépin. Ann. Soc. géol. Nord., Lille, 6, 66-73.

GOSSELET J. (1888) - L'Ardenne. Mém. expl. Carte géol. France. Paris, 848p., 9pl., 1 carte.

HANON M. (ms.1972) - Contribution à l'étude des porphyroïdes du Franc-Bois de Willerzie. Mém. Licence, Univ. Bruxelles, 55p.

HANMER S. (1986) - Asymmetrical pull-aparts and foliation fish as kinematic indicators. J. Struct. Geol., 8, 111-122.

HUGON H. & C. LE CORRE (1979) - Mise en évidence d'une déformation hercynienne en régime cisaillant progressif dans le Massif cambrien de Rocroi (Ardenne). C. R. Acad. Sc. Paris, 289, D: 615-618.

HUGON H. (1983) - Structures et déformation du Massif de Rocroi (Ardenne). Bull. Soc. géol. minéral. Bretagne, (C), 15, 109-143.

KLEIN, C. (1980) - L'intérêt tectonogénétique de la discordance post-calédonienne en Ardenne. Les notions d'hérédité mécanique et d'induction tectonique. Bull. Soc. belge Géol., Bruxelles, 89, 1-54.

KRAMM U. & D. BUHL (1985) - U-Pb Zircon dating of the Hill tonalite, Venn-Stavelot Massif, Ardenne. N. Jb. Geol. Paläont. Abh., 171, 329-337.

KRAMM U., D. BUHL & I.V. CHERNYSHEV (1985) - Caledonian or Variscan metamorphism in the Venn-Stavelot Massif, Ardenne? Arguments from a K-Ar and Rb-Sr study. N. Jb. Geol. Paläont. Abh., 171, 339-349.

KUBLER B. (1967) - La cristallinité de l'Illite et les zones tout à fait supérieures du métamorphisme. Etages tectoniques. Colloq. Neuchâtel, 105-122

LACASSIN, R. (1988) Large-scale foliation boudinage in gneisses. - J.Struct.Geol., 10, 643-647.

LAMBERT A. & J. BELLIERE (1976) - Caractères structuraux de l'Eodévonien aux environs de Bastogne. Ann. Soc. géol. Belg., Liège, 99, 283-297.

Le CORRE C. & H. HUGON (1987) - Au sujet de l'analyse structurale au bord sud du Massif de Rocroi (Ardenne françaises). Bull. Soc. belge Geol., Bruxelles, 96, 49-51.

LEGRAND R. (1967) - Ronquières, Documents géologiques. Mém. expl. Carte géol. & Min. Belgique., 6, 60p.

MALENGREAU B., V. STASSE, F. BRODKON & D. LADURON (in preparation) - Analyse structurale de la deformation polyphasée dans le bord nord du massif de Rocroi (région de Laonry).

MEILLIEZ F. (1981) - Filons magmatiques et structures plissées près de Revin (Ardenne). C.R. somm. Soc. géol. France, 3, 101-104.

MEILLEZ F. (1989) - Tectonique distensive et sedimentation à la base du Dévonien, en bordure NE du Massif de Rocroi (Ardenne). Ann. Soc. géol. Nord, Lille, 107, 281-295.

MEILLIEZ F. & J.-L. MANSY (1990) - Déformation pelliculaire différenciée dans une série lithologique hétérogène: le Dévono-Carbonifère de l'Ardenne. Bull. Soc. géol. France, (8), VI, 177-188.

MICHOT P. (1980) - Belgique. In: Géologie des Pays européens, France-Belgique-Luxembourg., Paris: Dunod, 485-576.

MICHOT P. (1989) - Synclinorium de Herve versus (Synclinorium de Verviers". Faille des Aguesses-Asse, chevauchemennet ao grand charriage. Bull. Soc. belge Géol., Bruxelles, 98, 7-25.

MORTELMANS G. (1955) - Considérations sur la structure teconique et la stratigraphie du Massif du Brabant. Bull. Soc. belge Géol., Bruxelles, 64, 179-218.

MUKHOPADHYAY, D. (1974) - Textural relations of porphyroblasts in the salm phyllites from Vielsalm and Salm-Chateau, Ardenne, Belgium. Geol. Rdsch., 63, 609-618.

PAPROTH E. (1987) - The Variscan Front North of the Ardenne-Rhenish Massifs. Ann. Soc. géol. Belg., Liège, 110, 279-298.

PIQUE A., S. HUON & N. CLAUER (1984) - La schistosité hercynienne et le métamorphisme associé dans la vallée de la Meuse, entre Charleville-Mézières et Namur (Ardenne franco-belges). Bull. Soc. belge Géol., Bruxelles, 93, 55-70.

PLATT J.P. & R.L. VISSERS (1980) - Extensional structures in anisotropic rocks. J. Struct. Geol., 2, 397-410.

RAST N. (1956) - The origin and significance of boudinage. Geol. Mag., 93, 401-408.

RAOULT J.F. (1986) - Le Front Varisque du nord de la France d'après les profils sismiques, la géologie de surface et les sondages. Rev. Géol. Dyn. Géog. Phys., 27, 247-268.

RAMSAY J.G. (1967) - Folding and fracturing of rocks. - McGraw-Hill, New-York, 568p.

RAOULT J.-F. & F. MEILLIEZ (1987) - The Variscan Front and the Midi fault between the Channel and the Meuse River. J. Struct. Geol., 9, 473-479.

ROCHE M. (ms. 1985) - Contribution à l'étude du sondage et de la région de Willerzie. Mém. Licence, Univ. Bruxelles, 137p.

SCHREYER W. (1975) - New petrologic evidence for hercynian metamorphism in the Venn-Stavelot Massif, Belgium. Geol. Rundschau, 64, 819-830.

STASSE V. (ms.1988) - Contribution à l'étude structurale du bord nord du Massif de Rocroi (région de Laonry). Mém. Licence, Univ. Louvain-le-Neuve, 76p.

SPAETH G., W. FIELITZ & B. FRANK (1985) - Caledonian deformation and very low-grade metamorphism in the NorthEastern part of the Stavelot-Venn Massif. N. Jb. Geol. Paläont. Abh., 171, 297-310.

STEEMANS P. (1989) - Paleogéographie de l'Eodévonien Ardennais et des régions limitrophes. Ann. Soc. géol. Belg., Liège, 112, 103-119.

THEUNISSEN K. (1970) - L'andalousite et ses phases de transformation dans la région de Vielsalm. Ann. Soc. géol. Belg., Liège, 93, 363-382.

VANDER AUWERA J. & L. ANDRE (1985) - Sur le milieu de dépôt, l'origine des matériaux et le faciès métamorphique de l'Assise de Tubize (Massif de Brabant, Belgique). Bull. Soc. belge Géol., Bruxelles, 94, 171-184.

VANGUESTAINE M. (1986) - Progrès récents de la stratigraphie par acritarches du Cambro-Ordovicien d'Ardenne, d'Irlande d'Angleterre, du Pays de Galles et de Terre-Neuve orientale. Ann. Soc. Géol. Nord, Lille, 105, 65-76.

WATERLOT G. (1937) - Sur la stratigraphie et la tectonique du Massif cambrien de Rocroi. Bull. Serv. Carte géol. France, n°195, T.31, 77-131