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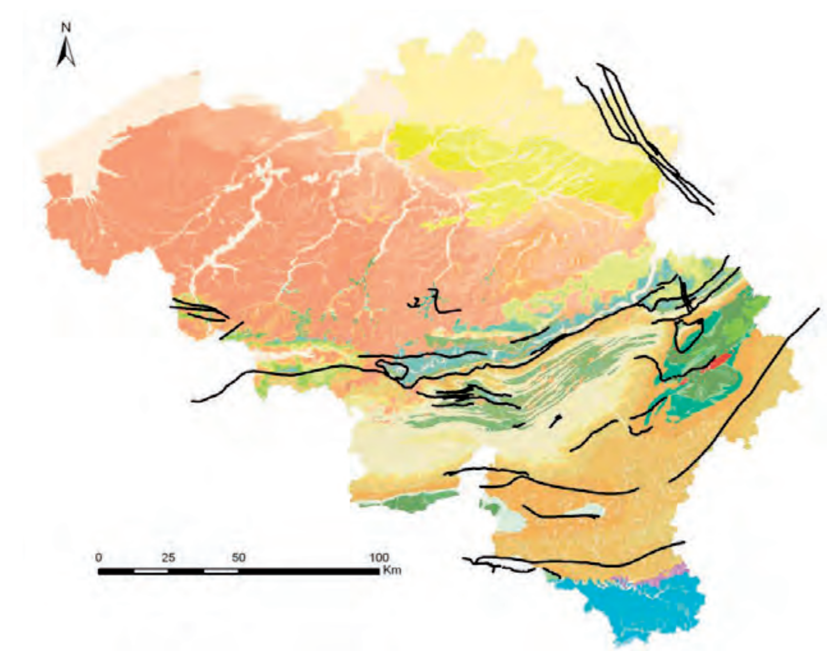
SYSTEMATIC INVENTORY AND ORDERING OF FAULTS IN BELGIUM
PART 2

Geoffrey CAMBIER & Léon DEJONGHE



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(215 pages, 154 figures, 3 table)

Cover illustration: map of the faults studied in Cambier & Dejonghe, 2010 (Faults project – part 1) and in this work (Faults project – part 2). The data used on this map are the most recent or the more coherent. The lithostratigraphic background is modified from http://www.onegeology.org and the legend corresponds to the International Stratigraphic Chart (http://www.stratigraphy.org).

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SYSTEMATIC INVENTORY AND ORDERING OF FAULTS IN BELGIUM – PART 2

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ABSTRACT. The inventory of major Belgian faults began with the so-called "Faults project" which aims to produce a catalogue compiling structural information and tectonic interpretations of each fault studied. Some results have already been published in a first volume released by Cambier & Dejonghe (2010); this paper constitutes the second part of this work. For each fault, the state of knowledge is established. The evolution of ideas and the divergent points of view presented in the literature over the years are described without taking a position for one of them unless a generally accepted consensus has emerged. Bibliographic research constitutes the basis of the work, the completion of which will clarify a large and scattered literature. Results from this work will also be published as a national-scale structural map of the Belgian fault network and as an electronic open access database.

Keywords: inventory, faults, Belgium

8. Foreword

This paper constitutes the second part of the "Faults project" written in the continuation of the part 1 (Cambier, G. & Dejonghe, L., 2010 – Systematic inventory and ordering of faults in Belgium – Part 1, Professional Paper 2010/1, N°307, Geological Survey of Belgium). We refer, therefore, to this paper for the introduction note of the Faults project (objective, terminology, method, geological setting).

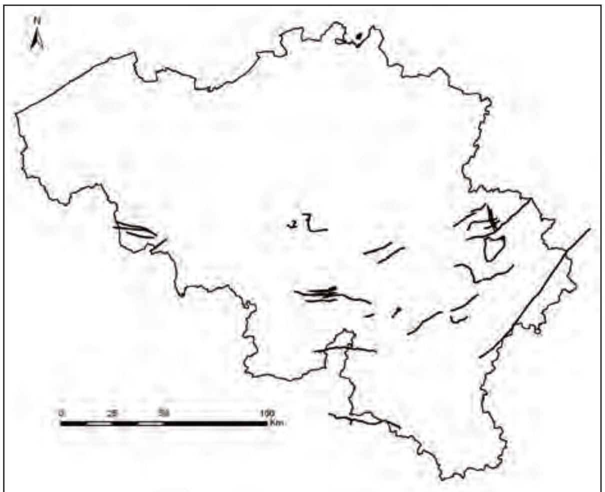


Fig. 120. Map of the faults studied in the first fault-dedicated Professional Paper (Cambier & Dejonghe, 2010).

Fig. 120 presents the faults studied in the first part of the work (Cambier & Dejonghe, 2010). Based on this second part, the structural map and the summary table previously released are updated and provided in Fig. 273 and Table 4 respectively.

The detailed index presented at the end of the book refers to faults tackled both in Cambier & Dejonghe, 2010 and in this work.

9. Descriptive data sheets of the faults (part 2)

9.1. Centre Fault

Location

The Centre Fault is introduced by Smeysters in 1887 under the name of the *"Grande faille du Centre"*. The fault was of economic significance as it separates two coal sub-basins known as the *"Centre-Nord"* and *"Centre-Sud"* basins (Hainaut, Namur Synclinorium) where workable strata were designated as *"Maîtresses-allures du Nord"* and *"Maîtresses-allures du midi"* respectively. The Centre Fault is related to many other faults and is therefore grouped in a faulted zone together with the secondardy fractures that connect to it.

Lithology and stratigraphy of the country rocks

The Centre Fault disrupts the *"Houiller"* Group of Namuro-Westphalian age. The rocks are mainly shales, siltstones and sandstones interlayered with coal seams.

Geometry

According to colliery studies, Smeysters (1887) believes in the existence of a large tectonic discontinuity separating two workable and superimposed coal seam groups, the *"Maîtresses-allures"* of the *"Centre-Nord"* and of the *"Centre-Sud"* being markedly separated by the *"Grande faille du Centre"*. Already in 1887, the fracture is considered as a complex of many faults rather than a single fault.

Briart (1894b) traces the fault through various coal concessions that are from the west to the east, the Sart le Moulin, the la Rochelle, the Amercoeur, the Jumet and the Saint-Marc concessions. The south-dipping (40-45°) and reverse Centre Fault is not a simple discontinuity but presents other numerous secondary accidents connected and located to the north or to the south of it. Those local south-dipping faults have a reverse offset.

On the basis of physical similarities, Briart establishes the continuity between the coal seams of the *"Centre-Nord"* and the *"Centre-Sud"* basins. The displaced coal seams on either side of the Centre Fault allow Briart to estimate a reverse offset, from the south to the north, of at least 1000 m.

Briart also envisages a western continuation of the Centre Fault. In this case, the fracture would be recognized in the *"comble Nord du Couchant de Mons"* to the north of Namur (Vedrin, St Marc Fault) and could therefore be the most significant and extended tectonic discontinuity of the Hainaut coal basin and probably of the entire Belgian territory. The Saint-Quentin Fault is not tackled and is probably not yet recognized at this time (1894b).

In 1900 and 1905, Smeysters suggests that the identification of the Centre Fault confirms and justifies the distinction between the *"Centre-Nord"* and the *"Centre-Sud"* basins (the second being uplifted over the first). Smeysters says that the Centre Fault consists of a thrust in which the offset may exceed 1200 m. Moreover, the fault has a second but quite significant satellite fracture called the Saint-Quentin Fault. This fault constitutes the northern branch of the Centre Fault. The structural map of the Hainaut coal-basin of Smeysters (1900) shows the Centre Fault to be at least 39 km long, but it is probably longer as mapping does not cover either lateral continuation. The Saint-Quentin Fault is traced over 27 km but is also probably longer.

Fig. 121 shows the Centre Fault as dipping to the south with an inclination of 45° at ground surface but reducing to 30° at a depth of about 500 m. The displacement of the "Léopold" coal seam, measurable on the cross-section below, indicates a reverse offset of at least 800 m.

The "Appaumée-Ransart" cross-section in Fig. 122 presents the Saint-Quentin Fault as a reverse south-dipping fracture in which the inclination evolves from 45° at surface to 10-15° at a depth of 500-600 m where the discontinuity connects with the Centre Fault. The reverse offset estimated by the displaced coal seams is about 140-150 m.

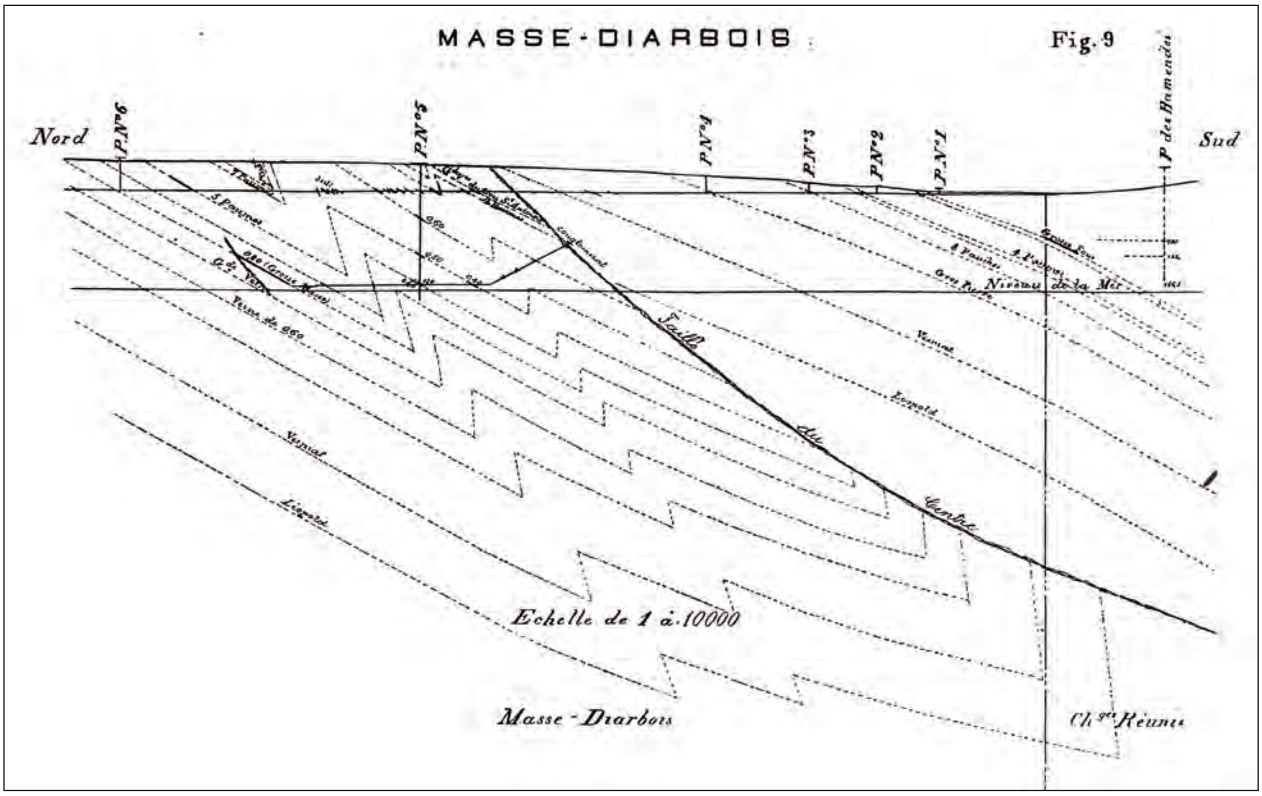


Fig. 121. The "Masse-Diarbois" cross-section of Smeysters (1900).

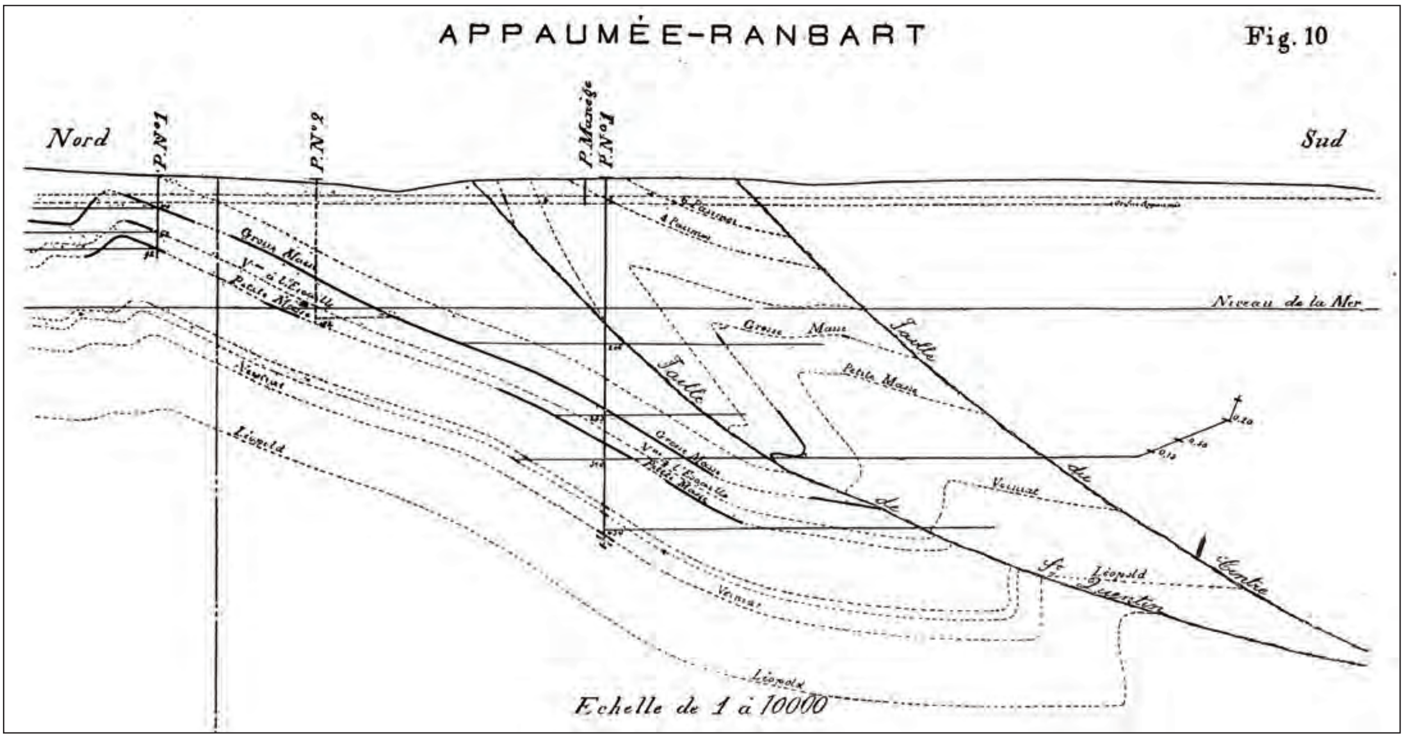


Fig. 122. The "Appaumée-Ransart" cross-section showing the relationship between the Centre and the Saint-Quentin faults (Smeysters, 1900).

Fig. 122. The "Appaumée-Ransart" cross-section showing the relationship between the Centre and the Saint-Quentin faults (Smeysters, 1900).

The geological maps of Briart & Bayet (1904, Fontaine-l'Évêque – Charleroi, n° 153) and of Stainier et al. (1904, Tamines – Fosse, n° 154) illustrate the Centre Fault over a length of 31 km from the south of Courcelles (in the west) to the north of Soye (in the east). Western and eastern continuations are supposed to exist but are not established. The Quaternary cover is not disrupted by the fault.

Cambier (1912) envisages the Centre Fault as consisting of a complex of faults, of a crushed and faulted zone. He emphasizes, therefore, the difficulty of drawing a precise trace of the "fault" (the trace of the Centre Fault corresponds to the upper limit of the faulted zone). He also suggests that the Canal Fault could be the western continuation of the Centre Fault.

In 1919, Fourmarier (1919a) proposes an offset of more than 1000 m and a dip to the south that is less than of the dip of the strata on either side of the fault. The Centre Fault is drawn over 28.5 km long but extensions are not known (Fig. 123). Fourmarier considers the Centre Fault as "one of the most important" fault in Belgium because of its displacement and especially because of its trace length. Indeed, the Centre Fault would continue eastward beyond the meridian line of Namur and would be connected to a discontinuity to the north of Marche-les-Dames that brings Silurian and Carboniferous rocks into contact. This fracture is known as the Landenne Fault (see Cambier & Dejonghe, 2010) on the geological maps of Stainier et al. (1901a, 1901b). In this case, the Centre Fault would have an extension of more than 70 km even without considering the probable continuation to the west.

Cambier (1920) draws the Centre Fault over a length of at least 40 km (Fig. 124). He provides detailed geometrical data collected in the numerous coal concessions where the fracture is observed. The decreasing inclination relative to the depth is clearly established. Within the Amercoeur concession, for example, the inclination decreases from 36° above 500 m depth to 22° at 685 m depth. The author believes that the Centre Fault constitutes a "first order" tectonic feature of the Hainaut coal-basin. The displacements are variable from place to place and are probably very significant but not measurable. Cambier also specifies the folding of the Centre Fault, which, in the vicinities of Gilly and Tamines, is displaced by over 900 m to the north (see Fig. 124 below, which illustrates the bending of the trace).

Marlière (1950) considers the "Centre Massif" as bounded by the south-dipping Centre Fault to the north (limiting the base of the "massif") and by the south-dipping Carabinier Fault to the south (limiting the "Carabinier Massif", thrust over the "Centre Massif"). The Centre Fault has a steep dip while the Carabinier Fault is a low-angle (about 12°) fracture.

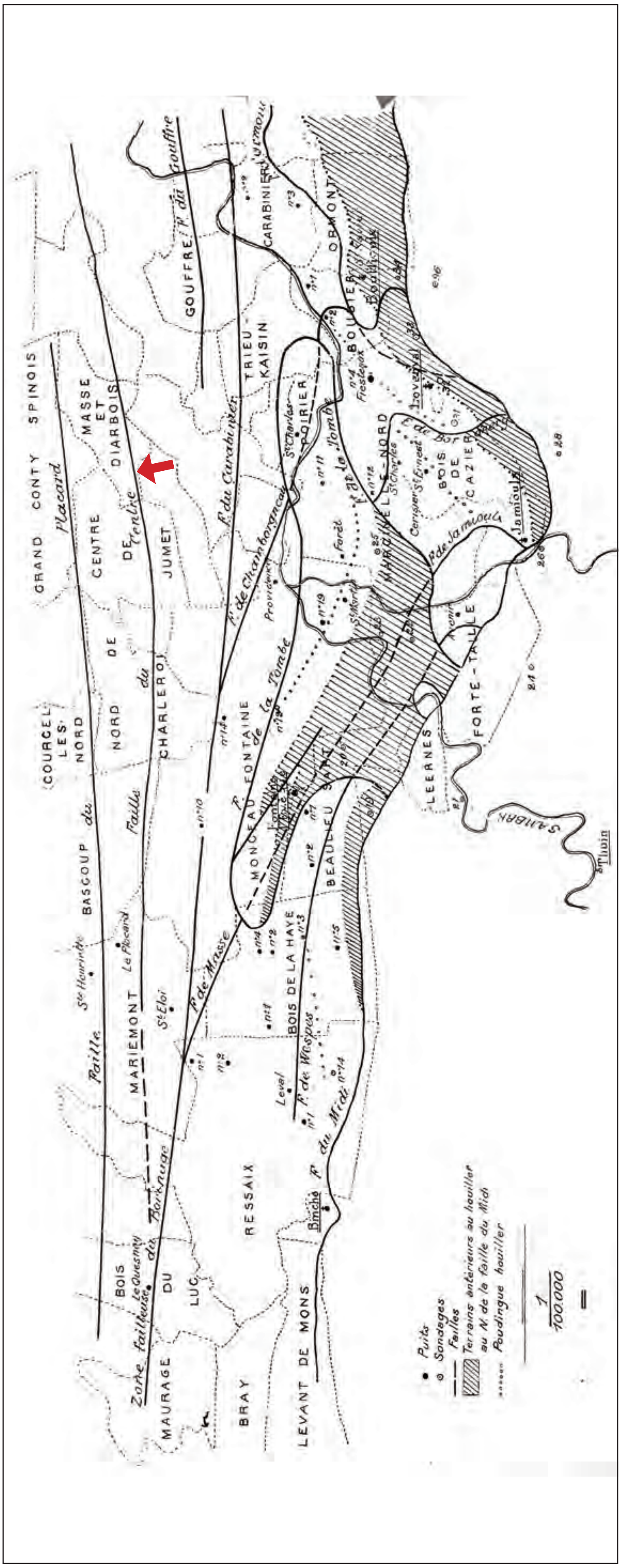


Fig. 123. Structural map of the Charleroi and Centre coal districts (Fourmarier, 1919a).

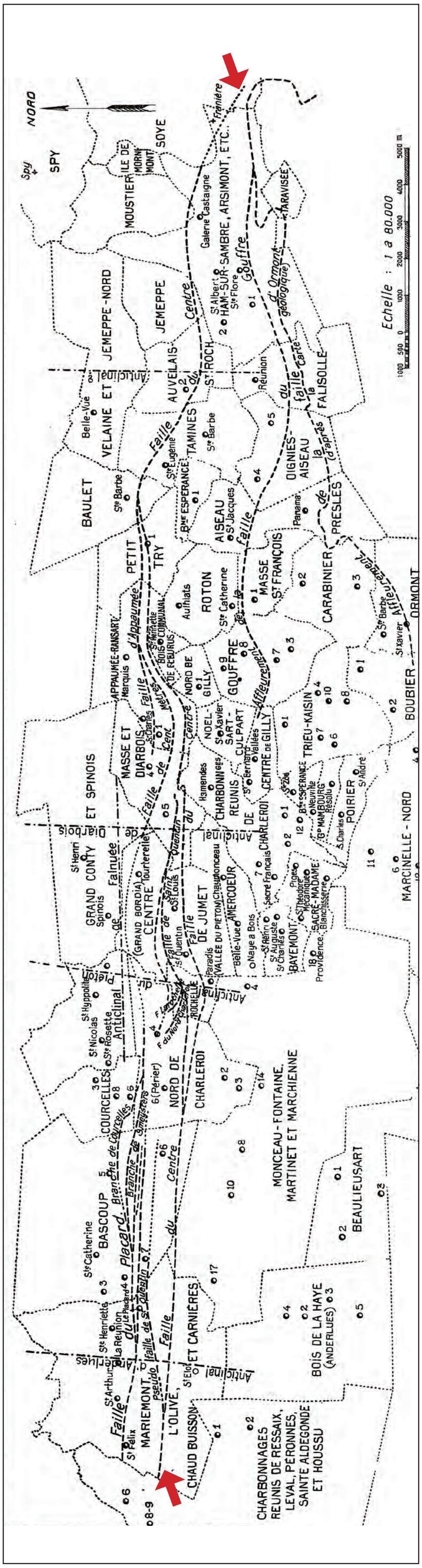


Fig. 124. Map of the coal concessions superimposed to the trace of the Centre Fault (Cambier, 1920).

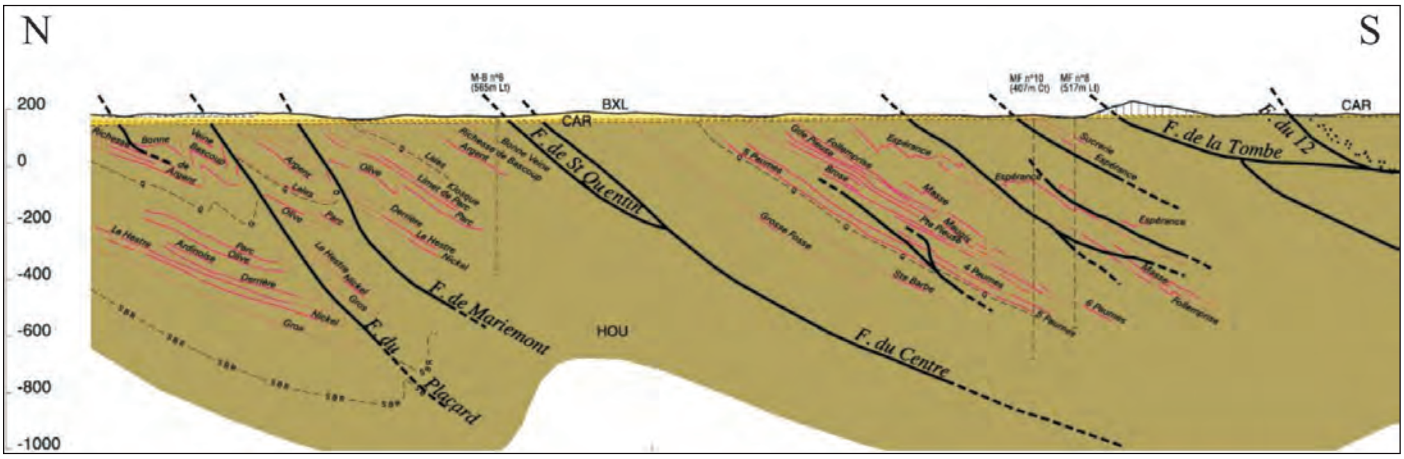


Fig. 125. North-south cross-section along the line of longitude through Fontaine-l'Évêque (Delcambre & Pingot, 2000). HOU = "*Houiller*" Group (Namurian-Westphalian). CAR & BXL = Carrières and Bruxelles formations (Eocene cover).

Delmer indicates in 1997 and 2004 the tectonic structure of the Hainaut coal-basin. From the south to the north, it comprises the allochthonous "Midi Massif" overlying a large allochthonous superficial unit, resting itself on "imbricated subautochthonous units". The Centre Fault belongs to the imbricate "massifs" that are separated from each other by reverse faults of moderate (i.e. measurable) offset (Fig. 128, see interpretations).

Delcambre & Pingot (2000) revise the geological mapping of the Fontaine-l'Évêque – Charleroi sheet. The tectonic structure in the vicinity of Charleroi is marked by three major units. From the north to the south these are the parautochthon "massifs" of Namuro-Westphalian rocks (of which the "Centre Massif" belongs), the thrust "massifs", subdivided into few tectonic stacks and the "Midi Massif" of Caledonian and Lower Devonian rocks. Over the entire length of the map (16 km), the Centre Fault marks the boundary between the "Placard Massif" to the north and the "Centre Massif" to the south.

The cross-section in Fig. 125, (from Delcambre & Pingot, 2000) shows the Centre Fault with a southern dip of about 40°. The depth attained by the discontinuity may reach 1000 m (1400 m on another section). The connection with the Saint-Quentin Fault is also visible.

Delcambre & Pingot (2000) also propose a particular evolution of the tectonic structure within the "Centre Massif" (Fig. 126). To the west, along the meridian line of Fontaine-l'Évêque, the "massif" is disrupted and stretched by a subhorizontal fault network (Fig. 126, 1), while to the east, along the meridian line of Gilly and Châtelet, an imbricate structure of superimposed "massifs" is observed (Fig. 126, 3&4). A progressive evolution of these structures is observed between the two meridian lines.

Interpretations

Smeysters (1900) considers the Saint-Quentin and other "secondary" faults to be related to the major thrust in the region, the Centre Fault. All of these discontinuities would appear progressively in response to a same compressive event acting from the south to the north.

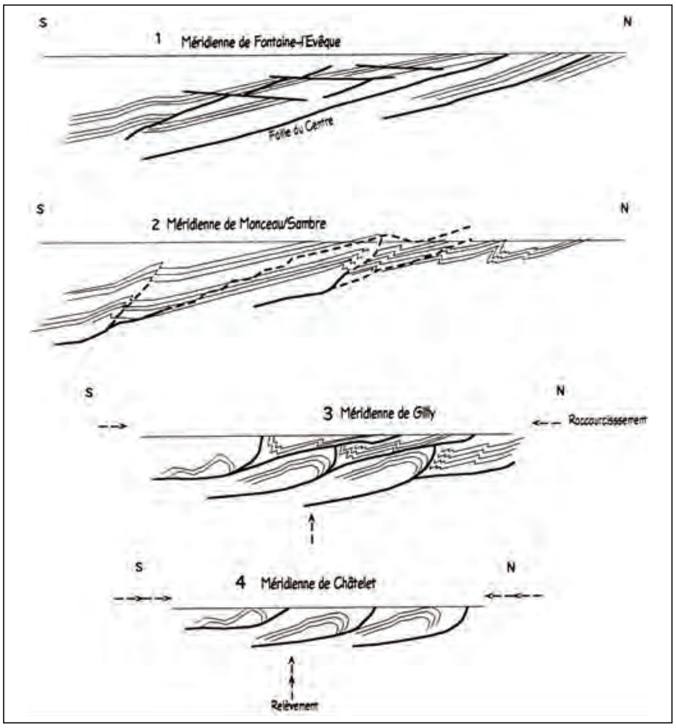


Fig. 126. Evolution of the deformation within the "Centre Massif", between the meridian lines of Fontaine-l'Évêque and Châtelet (Delcambre & Pingot, 2000).

As stated previously, the Centre Fault is folded (Cambier, 1920). The deviation of the strike is clearly visible on Fig. 127, which shows a northward displacement of the Centre Fault of 900 m in the vicinities of Gilly and Tamines. Cambier interprets the curving of the fault as resulting from enhanced contractional stress acting from the south to the north. Those constraints were active immediately before the onset of the Midi Fault (see section 9.4).

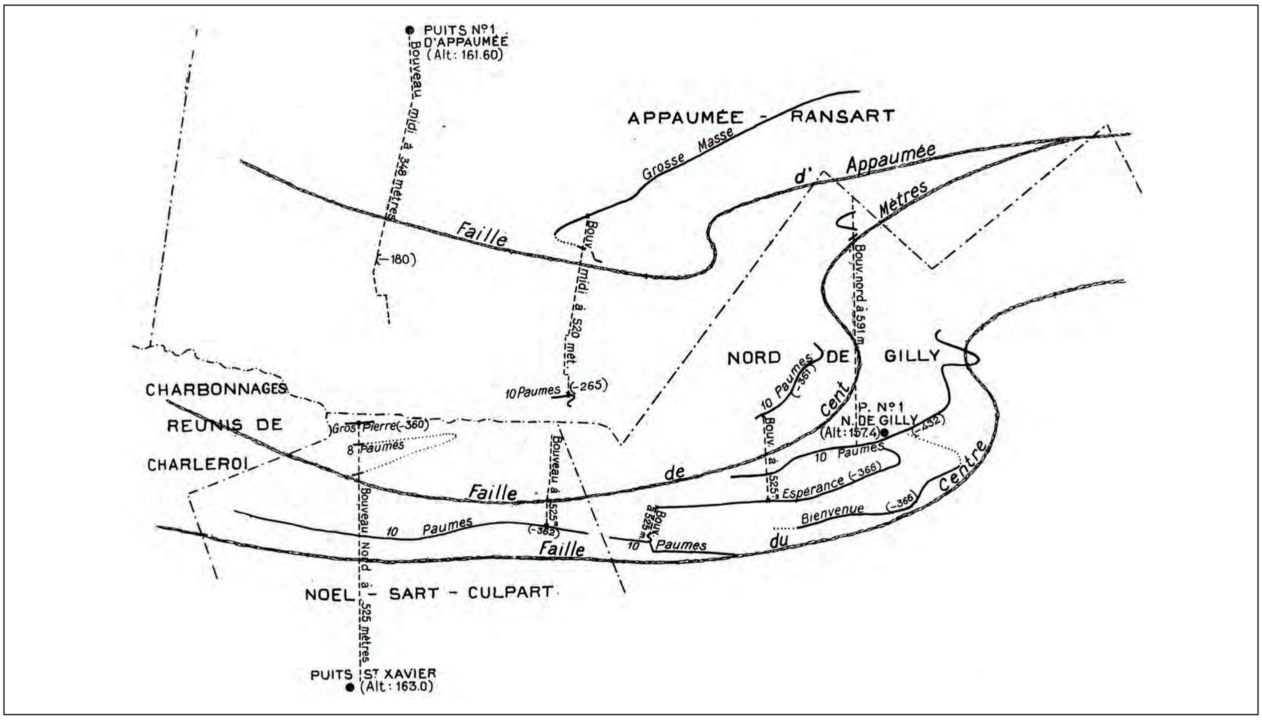


Fig. 127. Traces of the Centre Fault and its satellites (the 100 Mètres and the Appaumée faults) at a depth of 360 m (Cambier, 1920).

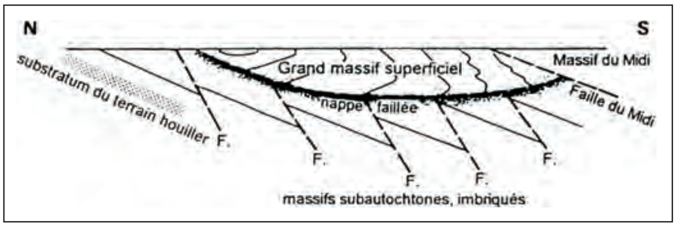


Fig. 128. Schematic cross-section through the Hainaut coal-basin (Delmer, 1997).

Marlière justifies in 1950 the geometrical aspects of the Centre Fault according to the theoretical point of view of Pruvost (1934, 1939). Studying the Saint-Etienne Coal Basin, Pruvost suggests that within homogeneous terrains a progressive transition can be observed between brittle deformation at depth and ductile deformation in the subsurface. In other words, the faults present in the deeper parts of the orogens are progressively "weaker" towards the surface and converted into anticlines. Moreover, Pruvost (1934) suggests that the longitudinal reverse faults associated with inclined or overturned anticlines have a maximum offset in the hinge areas of the folds. The displacement reduces and disappears in the plunging areas of the anticlines. Marlière indicates that the Centre Fault has a particular variable offset that is at a maximum in the vicinity of Charleroi but progressively decreases laterally and finally disappears.

In 2004, Delmer indicates that the imbricate subautochthonous units, the northernmost part of the Hainaut coal-basin (Fig. 128), are subdivided by south-dipping faults of low or moderate offset (the Centre, the Placard and the Carabinier faults being examples of those). The reverse displacement that characterizes these faults increases from the faults located to the north to those located to the south. Delmer interprets the structure of the subautochthonous units as resulting from a "frontal deadening" of the Variscan compression. The sequences within the units show conformity between the *"Houiller"* and the underlying Carboniferous substratum. All of these features indicate a piggy-back type sequence.

References

Briart, 1894b.

Briart & Bayet, 1904.

Cambier, 1912.

Cambier, 1920.

Cambier & Dejonghe, 2010.

Delcambre & Pingot, 2000.

Delmer, 1997.

Delmer, 2004.

Fourmarier, 1919a.

Marlière, 1950.

Pruvost, 1934.

Pruvost, 1939.

Smeysters, 1887.

Smeysters, 1900.

Smeysters, 1905.

Stainier et al., 1901a.

Stainier et al., 1901b.

Stainier et al., 1904.

9.2. Feldbiss Fault Zone

with references to the Rotem (or Rothem), Heerlerheide, Elen (or Eelen), Neeroeteren, Feldbiss (sensu stricto), Grote Brogel, Geleen, Reppel, Peel (Boundary) etc. faults

Location

The Feldbiss Fault Zone constitutes the southwest margin of the Roer Valley Graben (Fig. 129). The 20 km wide and 130 km long Roer (Dutch and French) or Rur (German) Valley Graben, (or Roermond Graben, or formerly Central Graben) is part of the Lower Rhine Rift System. The Roer Valley Rift System comprises three tectonic units, from southwest to northeast respectively: the Campine and South Limburg Blocks, the Roer Valley Graben and the Peel and Venlo Blocks. The Feldbiss Fault Zone can also be considered as the inner border of the Roer Valley Graben, with the Campine Basin as a transition zone to the Brabant Massif. The eastern part of the Campine Basin is intersected by a series of NW-SE faults with NE downthrow and smaller vertical Cenozoic displacements than the Feldbiss Fault Zone. The Rijen – Rauw - Beringen amalgamated fault system constitutes the western boundary of the neotectonic Campine ‘Graben Shoulder’, also bordering the Quaternary landscape recognised as the Campine Plateau (Fig. 129). The Roer Valley Rift System is located at the western margin of the Lower Rhine Embayment or Lower Rhine Graben and constitutes one of the most seismically active areas of the European Cenozoic Rift System.

The seismically active intra-plate area of the Roer Valley Graben is located in the southern Netherlands, in northeastern Belgium and continues further southeast in Germany (southwestern part of North Rhine – Westphalia) where it crosscuts the Variscan Midi-Aachen Thrust. The mapped fault structure (Fig. 129) is of Neogene age but is derived from a much older Cimmerian tectonic phase and a late Paleozoic depocenter; moreover the Roer Valley Graben has been subject to tectonic inversion during the late Cretaceous and some minor inversion events during the Cenozoic (Rossa, 1986; Tys, 1980).

The first geologists (e.g. Forir, 1904; Briquet, 1907; Stainier, 1911) to have recognised the faulted nature of this zone already highlighted the structural complexity of the southwest margin of the Roer Valley Graben. This margin is built up from multiple faults that have received different names and have different geometric attributes (e.g. strike) depending on the area and depth of investigation and the time slice studied. The faults display changing direction and displacement along the strike, which explains the divergent geometrical opinions of geologists over time. Dusar et al. (2001) indicate that no fixed framework of fault planes can be proposed for the southwest border of the Roermond Graben and therefore prefer to describe this structure as the “Feldbiss Fault Zone”.

Therefore, the name Feldbiss may denote a particular fault plane (the Feldbiss fault, which is only well expressed in South Limburg) or the entire fault zone forming the southern margin of the Roer Valley Graben.

Several references used in this descriptive data sheet are from the special issue of Geologie en Mijnbouw (van Eck, T. & Davenport, C.A. (eds), vol. 72(2-4), 1994) entitled “Seismotectonics and seismic hazard in the Roer Valley Graben; with emphasis on the Roermond earthquake of April 13, 1992”. We refer to this volume for extensive interpretations and references. The Feldbiss Fault Zone figures on the geological maps of, for example, Kimpe et al., 1978; Sels et al., 1999; Buffel et al., 1999; Langenaeker, 2000; on two important Dutch geological maps covering the Roer Valley Graben: the 1971 1/100 000 map of Kuyl and the 1975 1/600 000 map of Van Montfrans; and also on the 1/250 000 scale geological maps of Sittard-Maastricht (NITG, 1999) and Breda-Valkenswaard and Oss-Roermond (NITG, 2001).

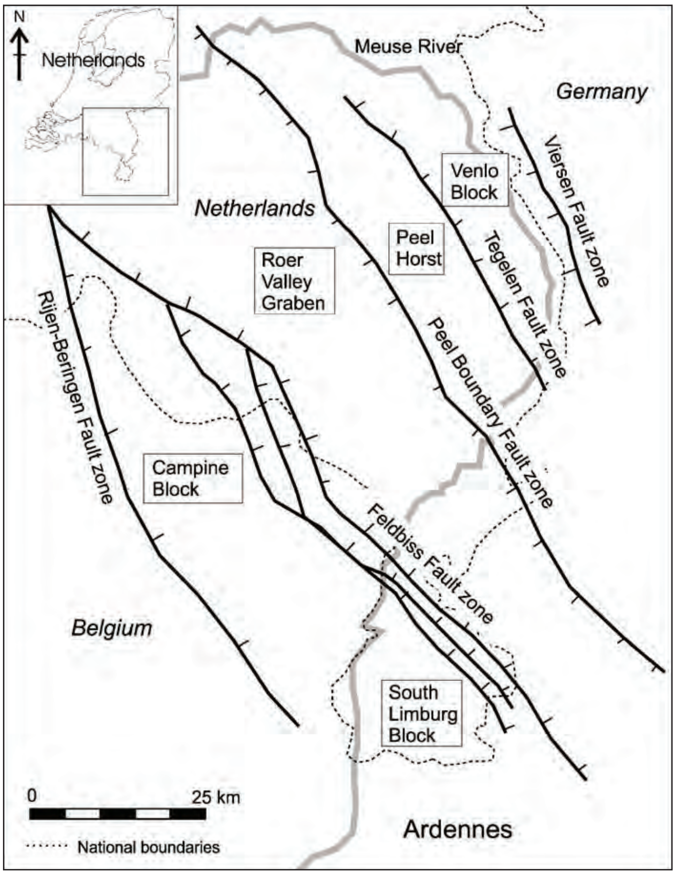


Fig. 129. Simplified structural map of the Roer Valley Rift System (Houtgast et al., 2002).

Stratigraphy and lithology of the country rocks

In the Sittard area, near the Belgian-Dutch border and close to the Feldbiss Fault Zone, the rupturing of the superficial part of the substratum, composed of successive Meuse terrace deposits, is evidence for the active subsidence of the Roer Valley Graben. The observation of altitude variations of the tabular terraces (mainly in drillholes) are interpreted in term of tectonic movement and offset along the faults.

On the left bank of the Meuse river north of Maastricht (i.e. in Belgian Limburg), Paulissen (1973) identifies 5 terrace levels between the alluvial flood plain and the "main terrace": (1) the Geistingen Terrace near the alluvial plain, (2) the Mechelen-aan-de-Maas Terrace covered by Weichselian aeolian sands (i.e. sandy loess), (3) the Caberg-Pietersem and (4) Eisden-Lanklaar Terraces covered by aeolian sands and separated by an erosion level and (5) the Lanaken Terrace. These terraces display a similar lithological composition, which generally includes gravels and coarse sands in the lower part and coarse to fine sands and loams in the upper part. The typical depositional paleoenvironment is a braided river system. Excepted for the Lanaken Terrace that was formed during the Holsteinian Interglacial (Mindel-Riss), the fluvial deposits constitute "climatic terraces" formed during cold periglacial periods (Paulissen, 1973).

The Meuse terraces of the Dutch South Limburg are generally subdivided into 4 groups: the East Meuse Terraces, the Main (or Higher) Terraces, the Middle Terraces and the Lower Terraces. Houtgast et al. (2002) provide a summary table of the fluvial terraces (Table 2).

Fig. 130 shows the subcrop map below the Upper Cretaceous of NE Belgium (Demyttenaere, 1989). According to Demyttenaere, the Rotem(-Heerlerheide) and Grote Brogel (Paulissen, 1973) faults (parts of the Feldbiss Fault Zone) delimit the Roer Valley Graben in the northeast from the Campine Block in the southwest. This figure shows the conservation of Jurassic rocks exclusively within the graben area. The footwall block (i.e. the Campine Basin, southwest of the northeast-dipping Feldbiss Fault Zone) is made up of Upper Paleozoic rocks, mainly of Westphalian age, overlain by Permian-Triassic rocks towards the north.

Table 2. Stratigraphic position of the Meuse fluvial terraces (In: Houtgast et al., 2002; data from Zonneveld, 1974; Van den Berg, 1996; Felder et al., 1989; Van Balen et al., 2000 & Houtgast et al., 2002).

Age of the terraces

	Terrace name		Period		Estimated Age (ka BP)			
	Van den Berg, 1996	Felder & Bosch, 1989	Van den Berg, 1996	Felder & Bosch, 1989	Van den Berg, 1996	Felder & Bosch, 1989	Van Balen et al., 2000	This paper
	Holocene Floodplain	Oost Maarland 3	Holocene	Holocene	3			
Lower terraces	Geistingen	Oost Maarland 2	Late Glacial	Late Glacial	11	13		11
	Mechelen a/d Maas	Oost Maarland 1	Weichselian	Weichselian	14		14	14
	Eisden Lanklaar	Gronsveld	Saalian	Eemian/ Late Saalian	130		130	130
Middle terraces	Caberg 3	Caberg		Saalian	245	250	250	250
	Caberg 2		Elsterian		330			330
	Caberg 1		Cromerian		420			420
	Rothem 2	Rothem 2		Elsterian/ Holsteinian	510	520	430	530
	Rothem 1	Rothem 1		Cromerian	620			600
	's Gravenvoeren	's Gravenvoeren			715		650	650
Higher terraces	Pietersberg 3	Pietersberg 2			780			710
	Pietersberg 2		Bavelian		870			750
	Pietersberg 1	Pietersberg 1			955	700	720	780
	Geertruid 3	Geertruid 3			1030		850	850
	Geertruid 2	Geertruid 2	Menapian	Bavelian	1090			980
	Geertruid 1	Geertruid 1	Waalian	Menapian	1280	1050	1100	1100

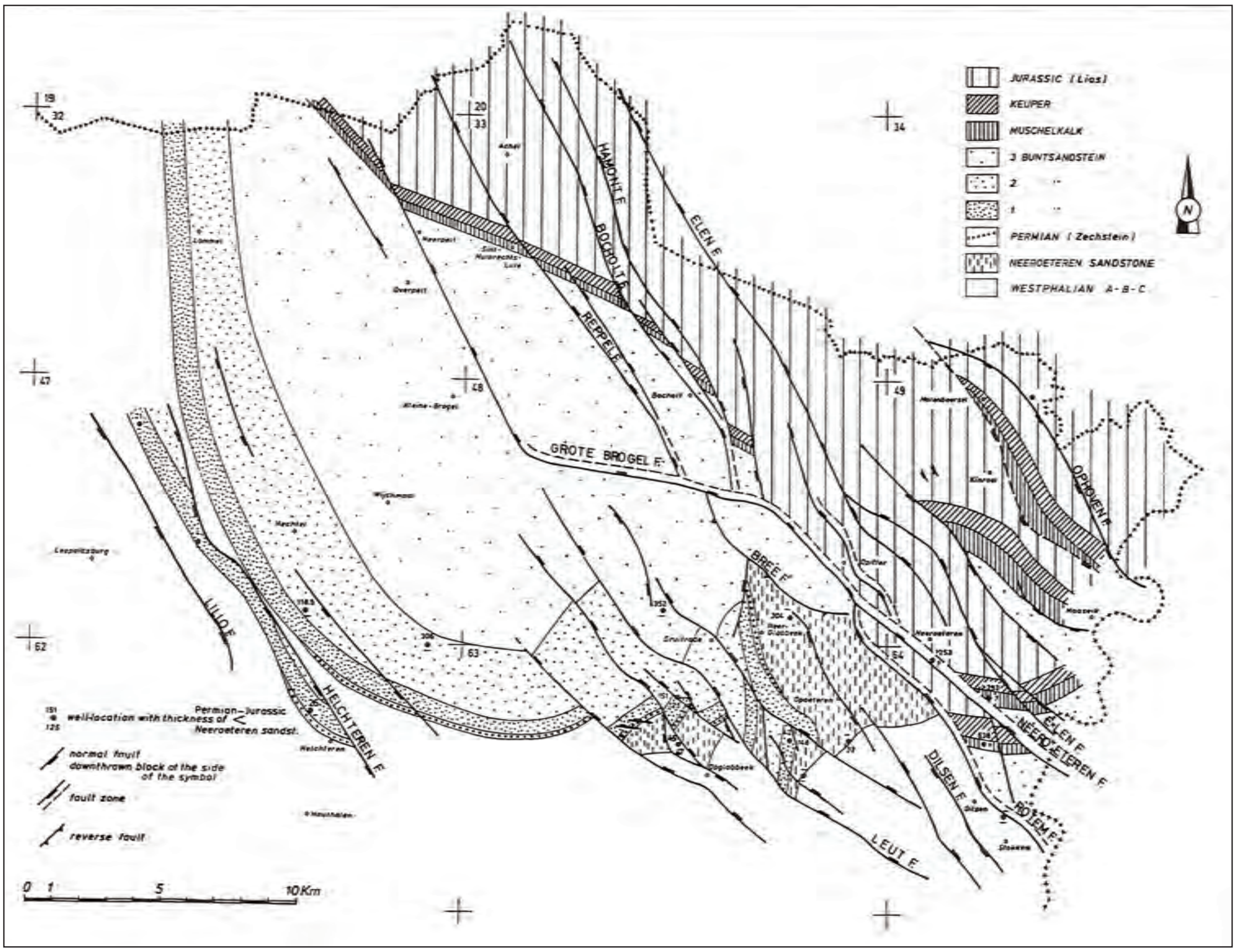


Fig. 130. Subcrop map below the Upper Cretaceous in northeastern Belgium (in Demyttenaere, 1989).

We refer the reader to the geological maps of Sels et al. (1999), Buffel et al. (1999) and Langenaeker (2000) and their explicative notes for details on the lithostratigraphy.

Geometry

According to borehole data, Forir (1904) identifies the Rothem Fault in Belgian Limburg. The major altitude difference of the Cretaceous, within boreholes located on either side of a supposed tectonic discontinuity, enables him to identify a fracture and moreover to estimate an offset of about 170 m. Indeed, the base Cretaceous is detected at -551 metres below the surface in the Eelen borehole (K on Fig. 131) and at -382.30 metres below the surface in the Dilsen borehole (A on Fig. 131). The northeastern block is downthrown.

Forir (1904) also highlights the structural complexity of the southern border of the Roermond Graben. He does not believe in a rectilinear character of the normal faults and supposes complex relations (i.e. splitting and joining) between them. The Feldbiss Fault is recognized (in 1904) by Dutch and German geologists but only in the Dutch-German border zone (Fig. 131). The Belgian continuation of the Feldbiss Fault is not envisaged.

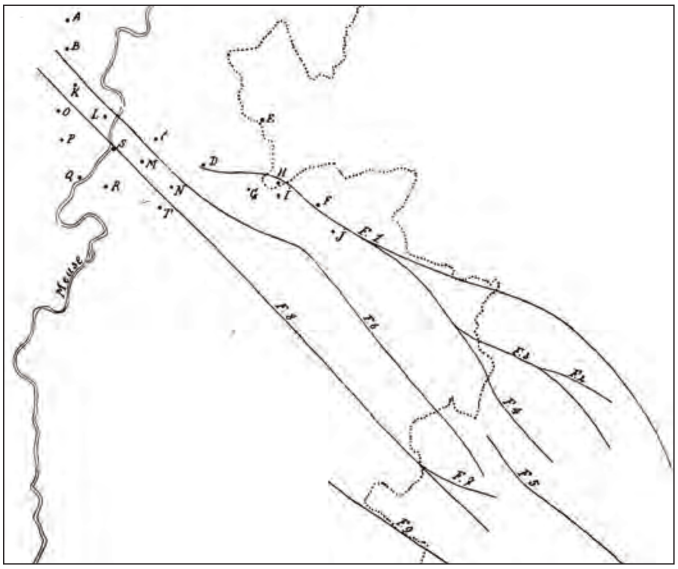


Fig. 131. Structural map of the main faults in the Aachen, South Limburg and East Campine coal basins (Forir, 1904). The Rothem Fault (F6) has a strike length of 36 km. F1 = "Sandgewand". F2 = "occidentale" Fault. F3 = Often Fault. F4 = "Feldbiss". F5 = "Münstergewand" or "Grosser Biss". F6 = Rothem Fault (or Uersfeld Fault). F7 = Richterich Fault. F8 = Dilsen Fault. F9 = Vaals Fault. A = Eelen borehole. K = Dilsen borehole.

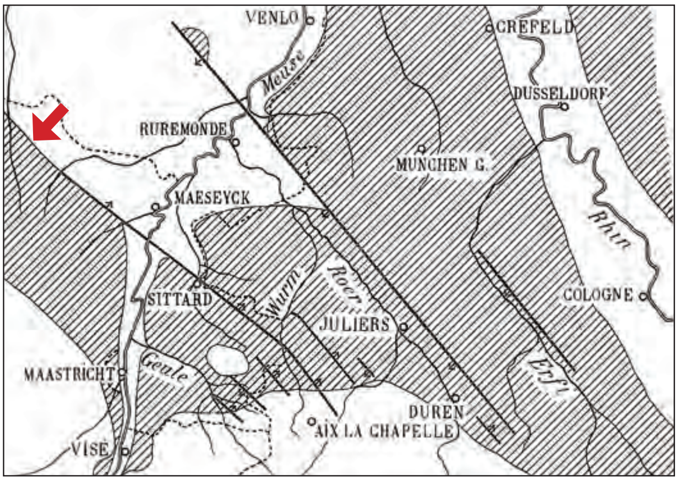


Fig. 132. Recent faults between the Meuse and Rhine rivers (Briquet, 1907). Hatched zones correspond to the outcrops of old alluvium. The arrow indicates the 62 km-long Feldbiss fault trace.

The work of Briquet, published in 1907, establishes the Belgian continuation of the Feldbiss Fault initially recognized by Dutch and German geologists in South Limburg. In the vicinity of Sittard, the NW-SE-striking Feldbiss Fault interrupts the continuity of the Meuse terraces (Fig. 132), which are downthrown (for about 30 to 40 m) to the north of the fault (Briquet, 1907). The discontinuity is easily discernable as it coincides topographically with a major scarp, the Neeroeteren-Bree Scarp (nowadays known as the Opitter Scarp). The Feldbiss Fault strikes over a distance of 62 km through Eygelshoven, Nieuwenhagen, Brunssum, Hillensberg and Sittard. Both the lower and upper terraces, including the uppermost or most recent terrace of the Meuse river are disrupted by the Feldbiss Fault.

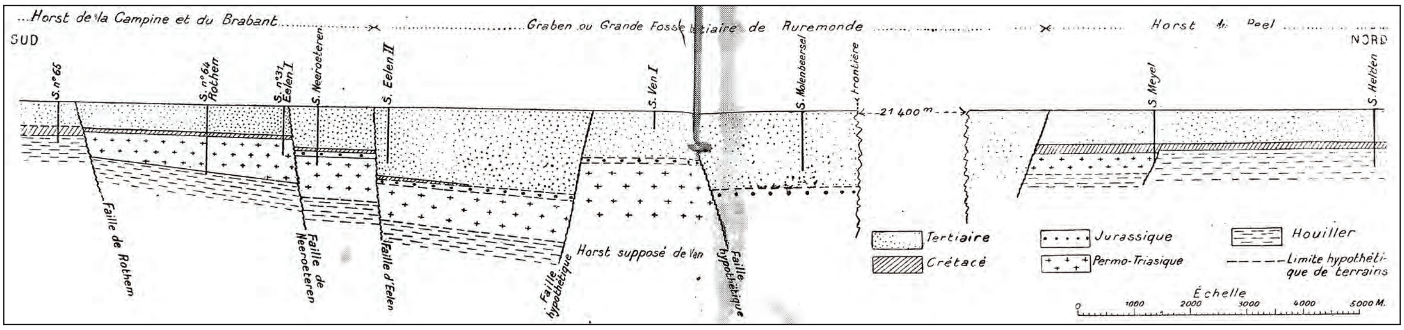


Fig. 133. S-N cross-section through the Roermond (or Ruremonde) Graben (Stainier, 1911). The location of the southwest part of this section is given in Fig. 134 ("Plan de coupe A-B").

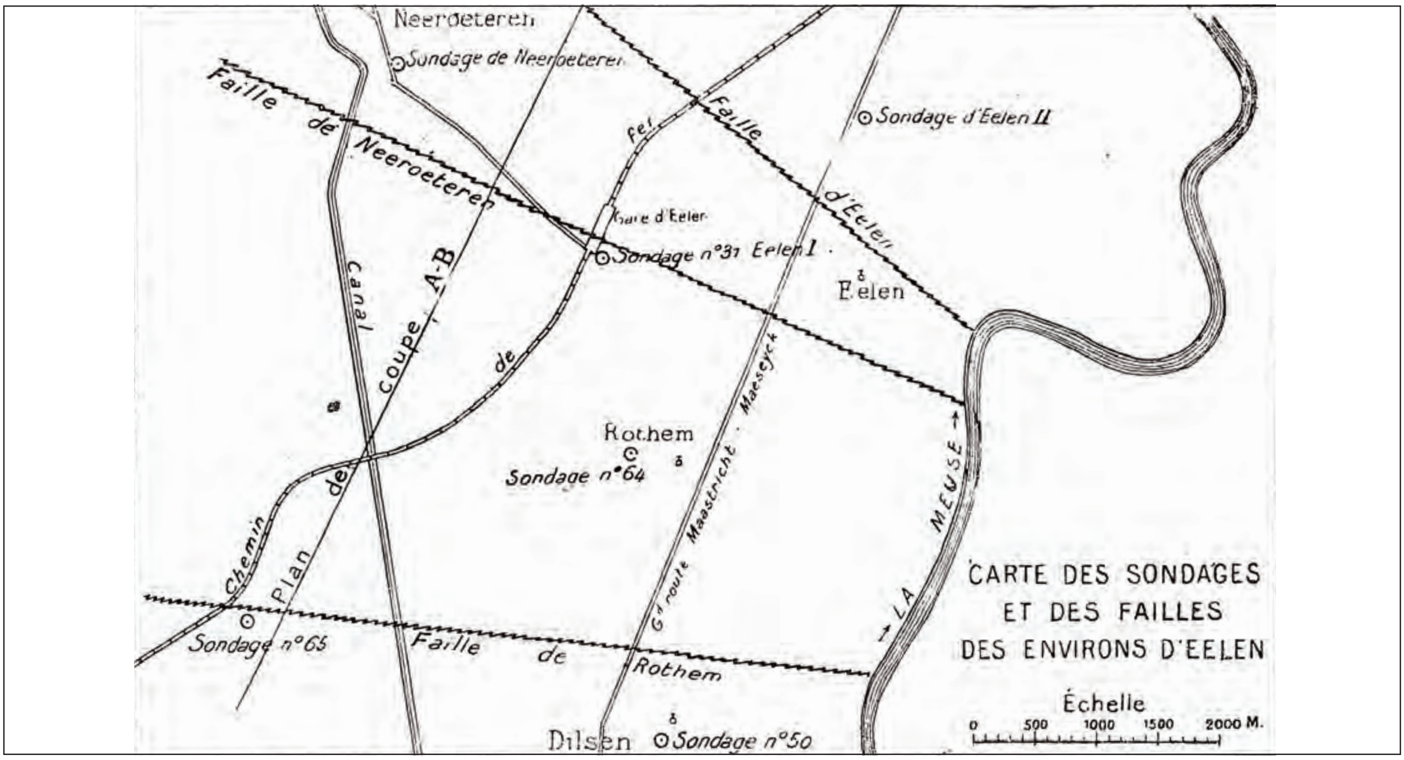


Fig. 134. Structural map of the vicinity of Eelen (Stainier, 1911). Boreholes are shown. Cross-section A-B is presented in Fig. 133.

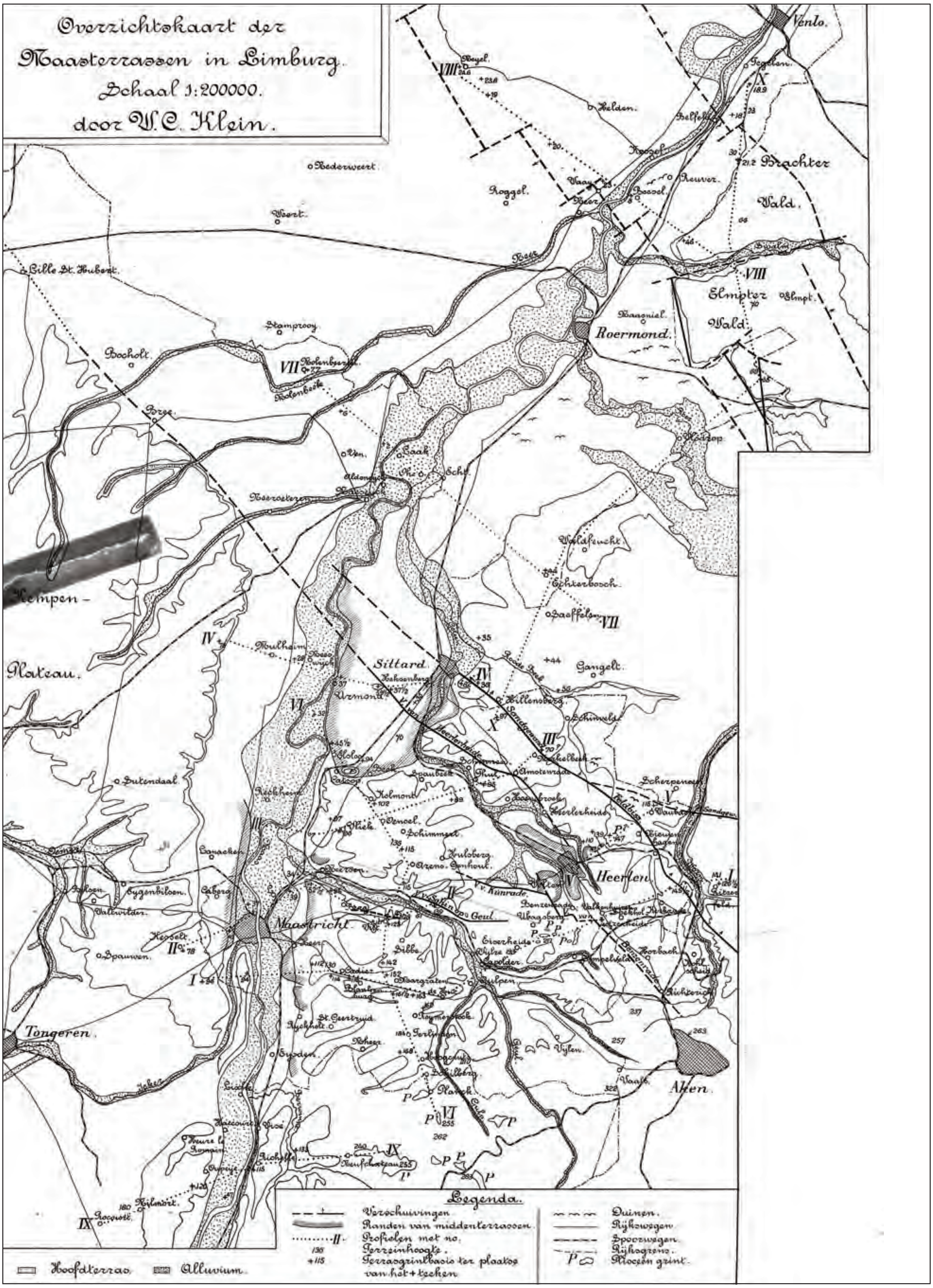


Fig. 135. Simplified geological map around the Meuse river in the vicinity of Maastricht (Klein, 1914). The main faults and terrace deposits are presented. A new structural feature is the connection between the Belgian Rothem and Dutch Heerlerheide faults. The Belgian continuation of the Feldbiss Fault until Sittard is also represented.

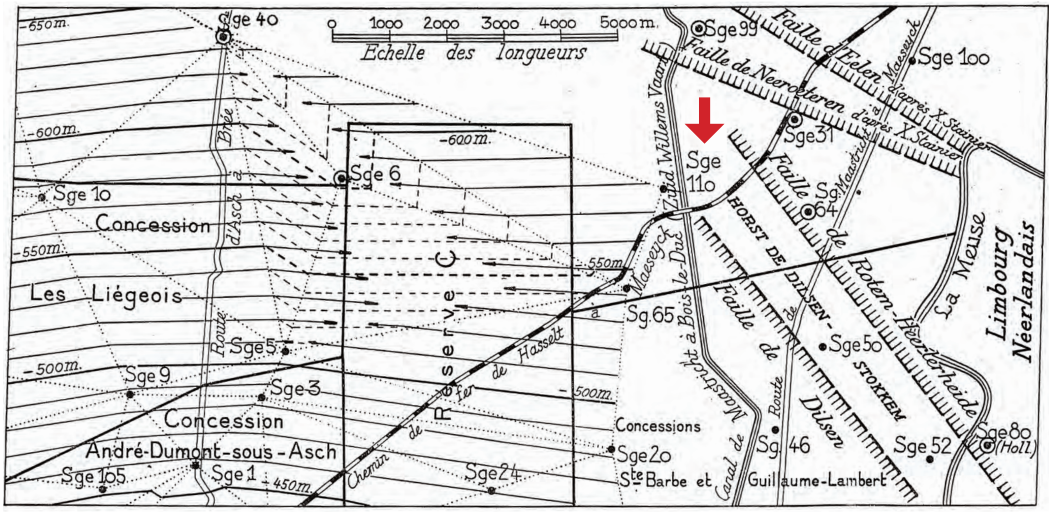


Fig. 136. Structural map of the vicinity of the Rotem borehole ("Sge 110", see the red arrow; Grosjean, 1939). The base of the Cretaceous is presented (countours at 10 m intervals). Boreholes where "red rocks" were identified are represented on the map by a double circle. Subsiding blocks are indicated by hatching.

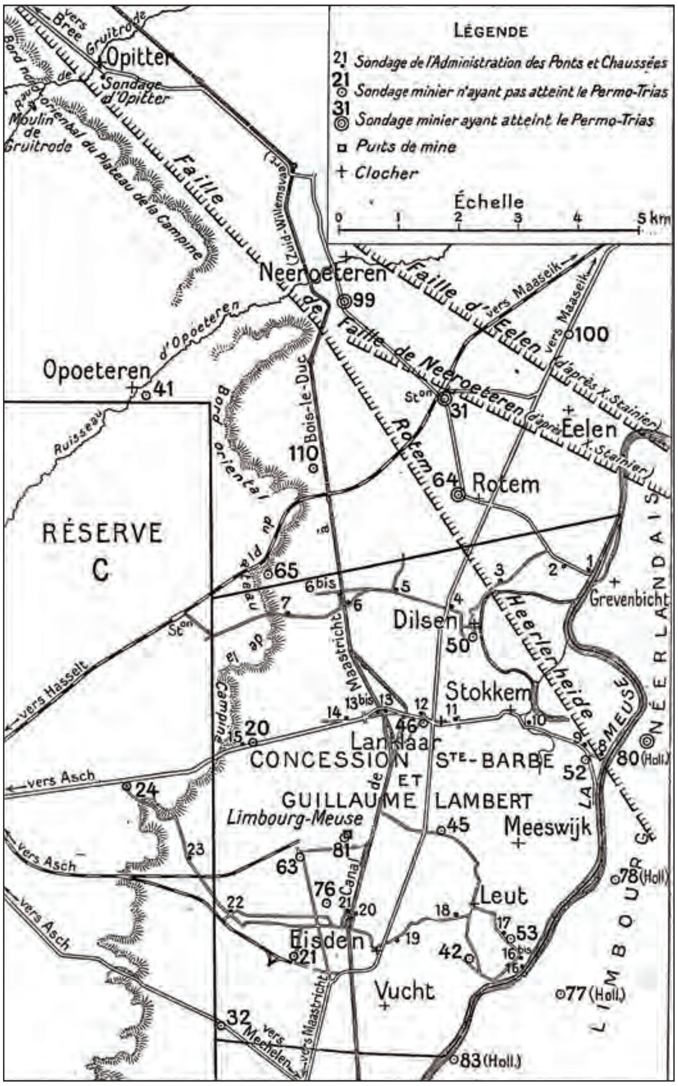


Fig. 137. Fault cartography in the vicinities of Rotem and Opitter (Grosjean, 1942). The Eelen and Neeroeteren fault traces are from Stainier (1911). Important boreholes are shown.

Based on borehole data, Stainier (1911) derives a detailed cross-section of the Tertiary Roermond Graben (Fig. 133). The differences in strata altitude established between the "n° 31 Eelen I" and Neeroeteren boreholes (Fig. 133) allows Stainier to introduce the Neeroeteren Fault, while the data from the Neeroeteren and "Eelen II" boreholes enables him to identify another discontinuity called the Eelen Fault. The Neeroeteren-Bree Scarp is attributed to the topographical expression of the Eelen Fault.

The different depths at which the Cretaceous is observed within the boreholes "n ° 31 Eelen I" and S. 99 Neeroeteren, indicate a subsiding movement of the northern block and a normal offset of about 150 metres along the Neeroeteren Fault. The normal displacement that affects the Rothem Fault is estimated at between 100 and 200 m and that of the Eelen Fault to about 450 m (the latter being measured on the cross-section in Fig. 133).

Stainier (1911) also proposes a structural map covering an area located to the west of the Meuse river, i.e. in Belgian territory (Fig. 134). No attempt at connecting the Belgian and Dutch segments of the Roermond Graben faults is proposed.

In 1936, Grosjean establishes the continuity between the Belgian Rothem Fault to the west of the Meuse valley and the Dutch Heerlerheide Fault to the east of the Meuse valley. The link between the Rothem and Heerlerheide faults was already envisaged by Klein in 1914 (Fig. 135). The western segment of the discontinuity (sensu Grosjean, 1936) corresponds to the Rothem fault trace (sensu Stainier, 1911). The Rothem-Heerlerheide Fault separates a southern block where Cretaceous rocks lie over Carboniferous deposits from a northern block where "red rocks", of Permian-Triassic age, are wedged between Cretaceous and Carboniferous deposits. A total vertical offset of approximately 800 metres is measured just west of the Meuse river.

In 1937, Grosjean admits mistakes regarding the cartography of the Rothem Fault (western segment) that he published the year before. Grosjean (1937) confirms that the Rothem Fault is the Belgian equivalent or westward continuation of the Heerlerheide Fault already identified by the Dutch geologists in South Limburg. He also reiterates that the trace of the Dutch Heerlerheide segment is well constrained by the observation of the "red rocks" while the trace of the Rothem segment remains hypothetical because of the lack of knowledge regarding the "red rocks" in that part of the fault.

In 1939, studying new borehole data, Grosjean constrains the trace of the Rotem-Heerlerheide Fault (Fig. 136). The absence of Permian-Triassic "red rocks" markers within the Rotem borehole (referenced "Sge 110" in the Fig. 136) enables Grosjean to attribute a linear, NW-SE striking trace to the fault. The Neeroeteren and Eelen faults are also represented in Fig. 136, their traces being taken from the work of Stainier (1911) as no new information had been brought forward.

Grosjean (1942) proposes an extension to the Rotem-Heerlerheide Fault further to the northwest beyond Opitter (Fig. 137). He also attributes the Neeroeteren-Bree Scarp to the Heerlerheide Fault.

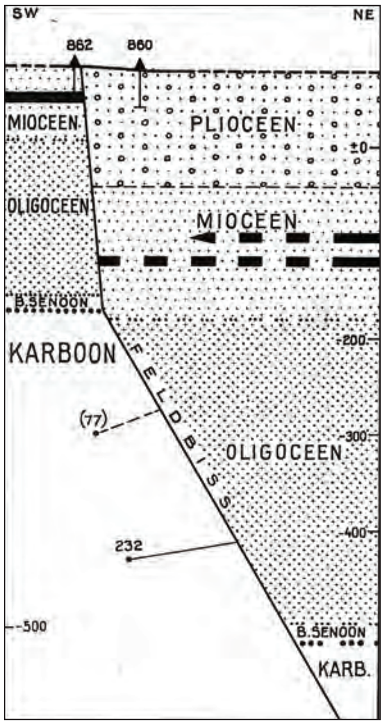


Fig. 138. SW-NE cross-section through the Feldbiss Fault in the region SE of Sittard (Rutten, 1943).

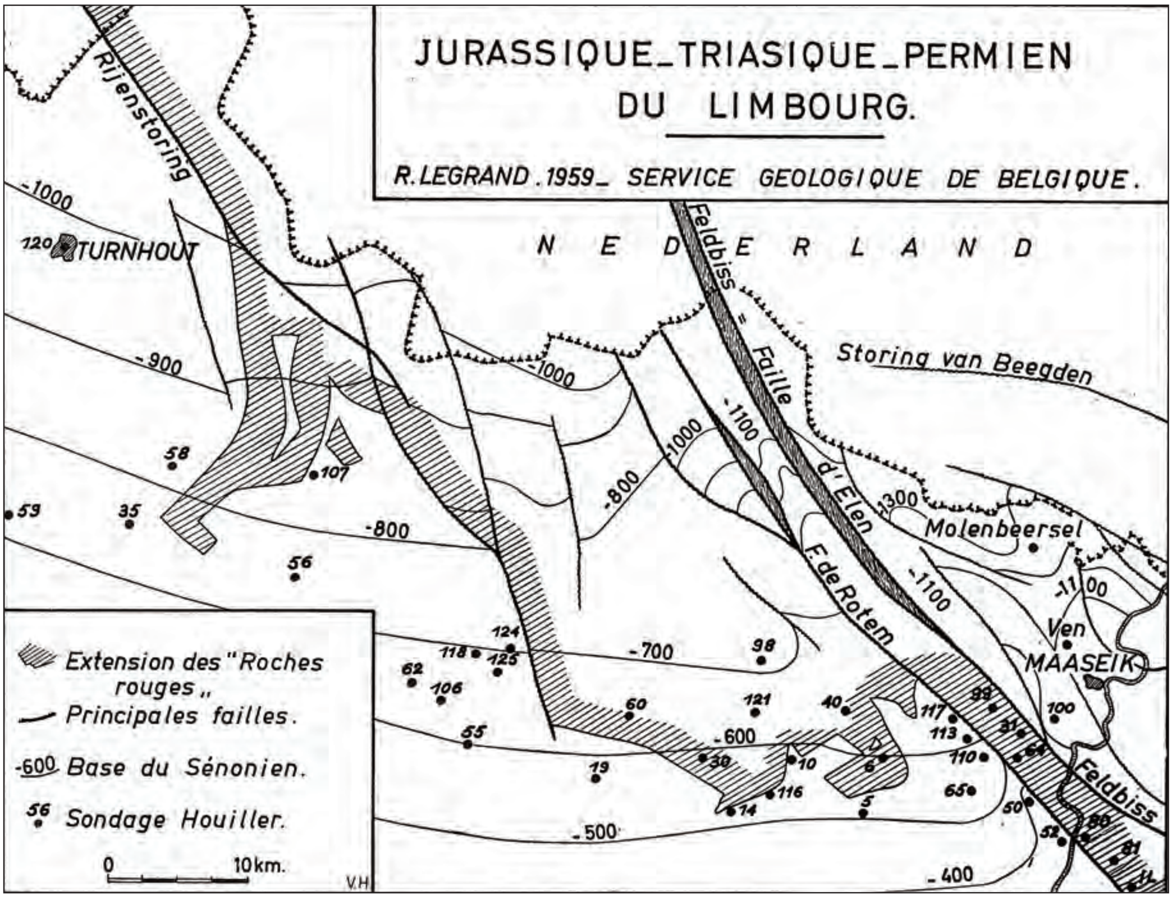


Fig. 139. Simplified geological map of the Jurassic, Triassic and Permian of the Belgian Limburg (Legrand, 1961). Two main faults, the Elen (or Feldbiss) and the Rotem (or Heerlerheide) faults, delimit the Campine area in the southwest from the Roermond Graben in the northeast.

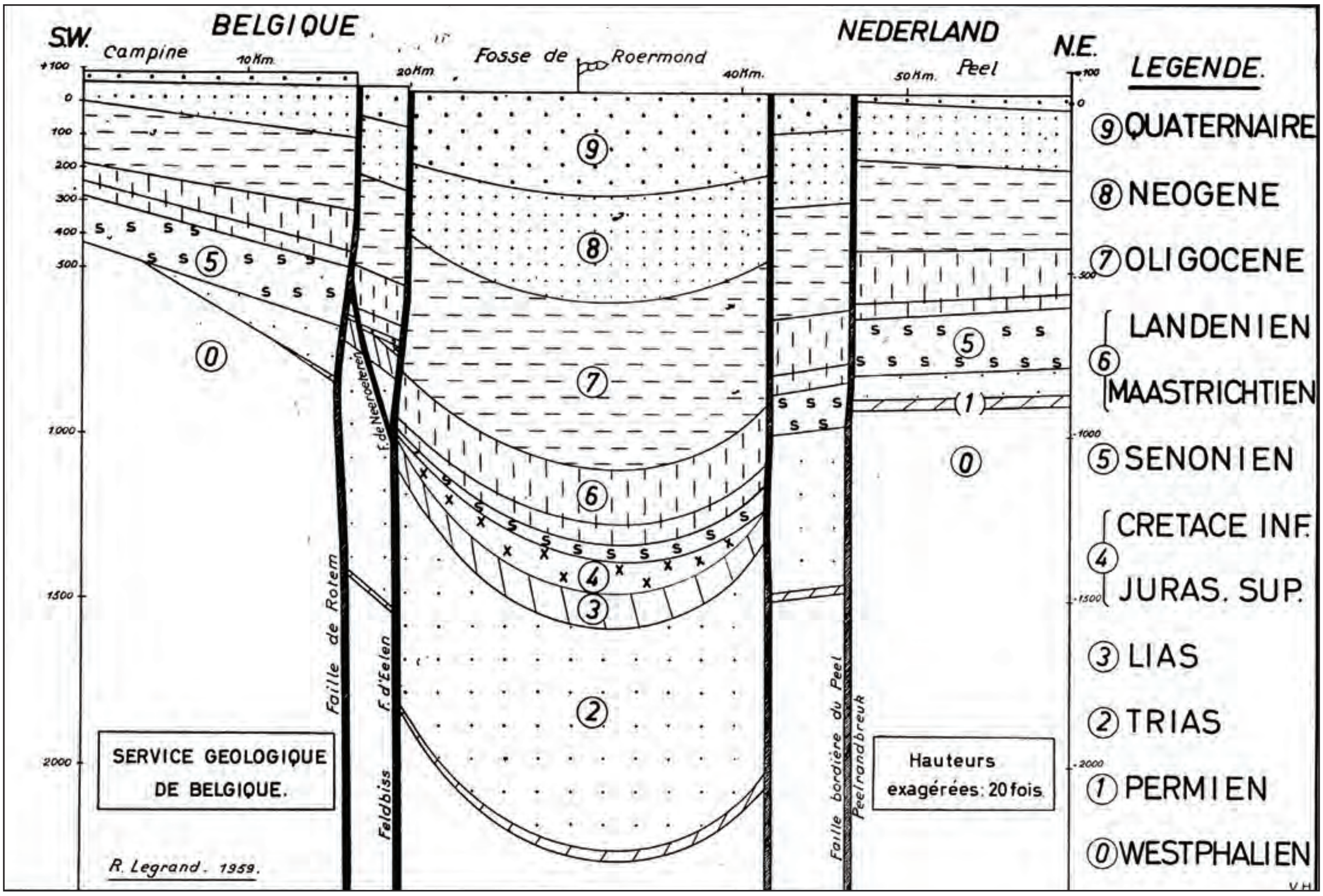


Fig. 140. SW-NE cross-section through the Roer Valley Graben (Legrand, 1961).

In the area southeast of Sittard, in South Limburg, Rutten (1943) proposes a steep northern dip for the Feldbiss Fault. The fault inclination evolves from 85° in the upper part of the substratum to 60° at depth where an anomalous contact between Carboniferous and Oligocene rocks is presumed (Fig. 138). The cross-section in Fig. 138 shows a normal displacement of about 400 metres.

In 1947, Heybroek considers that there has been no horizontal movement along the Feldbiss Fault, which has only a major dip-slip (normal) component.

Legrand (1961) considers the Elen Fault as the continuation of the Dutch Feldbiss Fault in Belgium (Fig. 139) and the Rotem Fault as the Belgian equivalent of the Dutch Heerlerheide Fault. Legrand also provides a general cross-section of the Roermond Graben reproduced in Fig. 140. Triassic-Permian red rocks extend within the graben but are absent southwest of the Rotem Fault close to the Belgian-Dutch border.

According to a seismic reflection survey, Bouckaert et al. (1981) identify the Bree Uplift, a folded horst structure parallel to the NW-SE-striking faults of the Roer Valley Graben located in the Bree area. The Bree Uplift exposes strata dragged towards the fault plane and therefore exhibiting a strong local dip of up to 65° striking parallel to the fault plane. The Bree Uplift represents a Carboniferous updoming structure rising under a reduced thickness of Cenozoic to Triassic cover. The structure of the Bree Uplift is confirmed by data from the borehole Opitter 48E0294 (Geological Survey of Belgium) drilled in the middle of this structure.

Bouckaert et al. (1981) also suggest that the strike of the Heerlerheide Fault is influenced by the Bree Uplift: the fault (which is actually subdivided into a northern and southern segment in the area between the Meuse river crossing and the city of Bree, Fig. 141) displays a bend to the north circumscribing the Bree Uplift and then links with the Feldbiss Fault (Fig. 141). The border faults of the southwestern part of the graben are responsible for the downthrown movement of the top of the Carboniferous for about 950 metres (Delmer, 1963). The top of the Carboniferous is detected at a depth of 657 m below the surface (-581 m MSL) in borehole 117, at only 556 m (-505 m MSL) at borehole 48E0224 on the Bree Uplift further north, and below the constructed (not drilled) level -1560 m in borehole 99 (Fig. 141). A normal offset of about 400 metres (disrupting the Carboniferous rocks) is attributed to the Heerlerheide and to the Feldbiss faults.

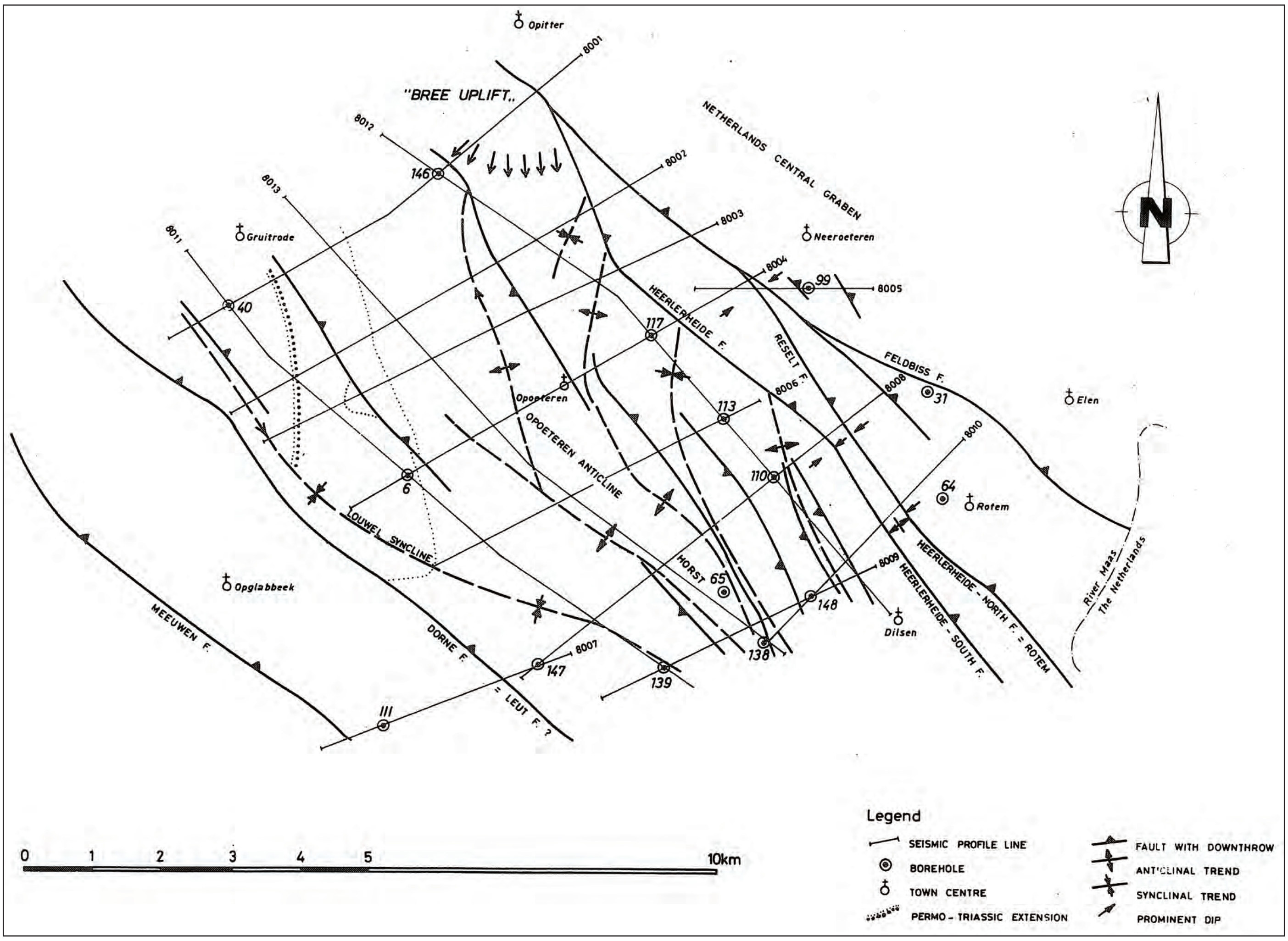


Fig. 141. Structural map of the Neeroeteren-Rotem coal exploration area (Bouckaert et al., 1981).

In 1985, Paulissen et al. consider the Roermond Graben as being bounded in the southwest by a fault zone that is composed of three main NW-SE subparallel faults: from north to south respectively the Feldbiss, the Geleen and the Heerlerheide faults (Fig. 142). According to a geophysical (electrical tomography) survey in the Rotem area, Paulissen et al. identify the buried Bichterweert Scarp, a major discontinuity at the base of the bottom gravels of the Maas valley (Fig. 143). The buried scarp is responsible for a sudden increase in gravel thickness that delimits gravel extraction towards the south. It has a curved surface strike ranging from an E-W direction in the Maas valley to a NW-SE direction between Rotem and Neeroeteren (Fig. 144). To the north of the Bichterweert Scarp, the base of the Maas valley bottom gravel is downthrown for 7-11 metres with respect to the southern block (Fig. 143). The Bichterweert Scarp represents the only major fault activity in the (subrecent) Maas valley gravels (other tectonic disturbances exist, see for example Brabers & Dusar, 1999) between Neeroeteren and Born and is attributed to the northwestward continuation of the Feldbiss Fault (Paulissen et al., 1985; Fig. 144).

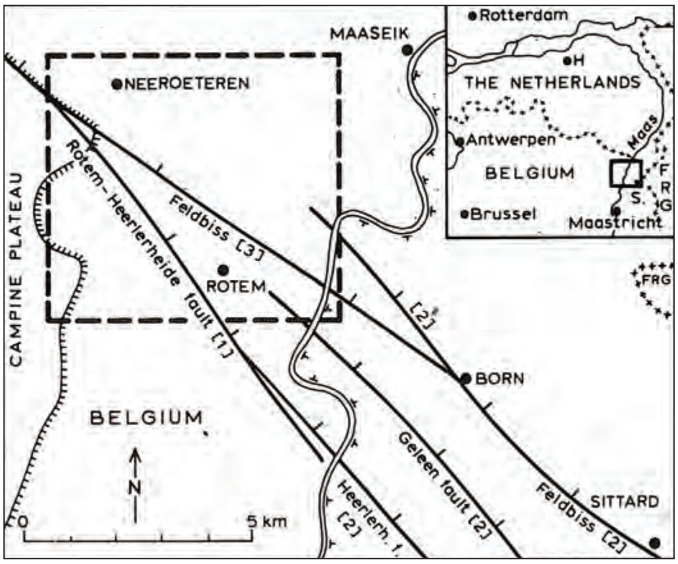


Fig. 142. The fault system of the southwest margin of the Roer Valley Graben in the Meuse valley (in Paulissen et al., 1985; according to Grosjean (1942) [1]; Kuyl (1971) and Van Montfrans (1975) [2] and Paulissen (1973) [3]).

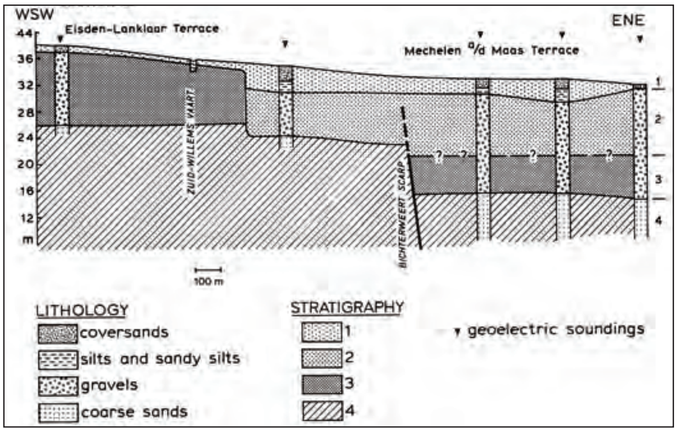


Fig. 143. WSW-ENE cross-section through the Bichterweert Scarp (Paulissen et al., 1985). 1 = Aeolian deposits (Weichselian Middle/Upper Pleniglacial). 2 = Mechelen-aan-de-Maas terrace (Weichselian). 3 = Eisdan-Lanklaar terrace (Saalian). 4 = Tertiary subsoil.

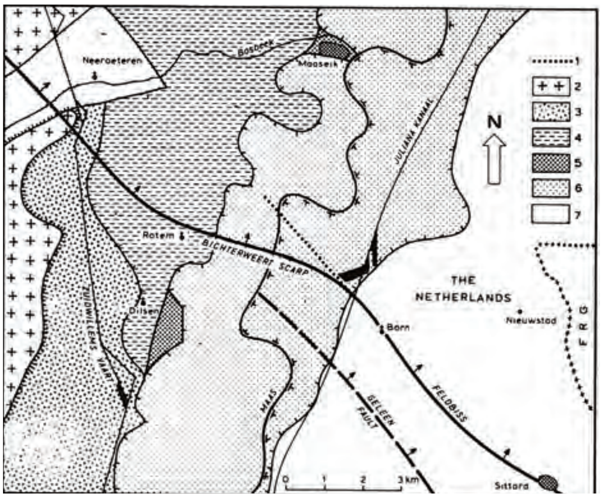


Fig. 144. Fault and terrace locations in the Maas valley (in Paulissen et al., 1985), according to Kuyl (1971), Van Montfrans (1975) and Paulissen (1973). 1 = Feldbiss pattern not recorded in this work (Paulissen et al., 1985). 2 = Main terrace deposits. 3 = Eisden-Lanklaar terrace. 4 = Mechelen-aan-de-Maas terrace. 5 = Geistingen terrace. 6 = Alluvial plain of the Maas. 7 = Deposits of the Bosbeek.

In his work of 1989, Demyttenaere points out the complexity of the fault mapping in the southwest margin of the Roer Valley Graben. He provides two schematic structural maps of the southwest border faults of the graben (Fig. 145 & 146), giving a summary of the structural ideas from the beginning of the 20th century.

Basing on the work of Paulissen (1973), Demyttenaere (1989) considers the Belgian southwest border of the Roer Valley Graben as formed by three main faults: the Rotem, the Neeroeteren and the Elen faults (Fig. 130 & 147). This threefold subdivision has been adopted in most later publications dealing with the subsurface geology. According to seismic data, these Belgian fault segments may be traced both northwestward and southeastwards to the Dutch border. Connections with their Dutch counterparts were established as follows: the Rotem Fault corresponds to the Heerlerheide Fault, the Neeroeteren Fault to the Geleen Fault and the Elen Fault to the Feldbiss sensu stricto.

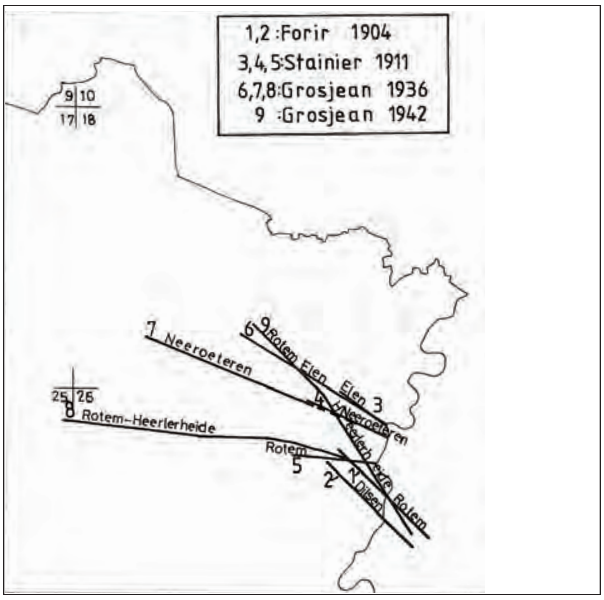


Fig. 145. Evolution of views on the structure of the southwest margin of the Roer Valley Graben from 1904 to 1942 (Demyttenaere, 1989). Structure number 8 on the figure attributed to Grosjean (1936) was actually considered not to exist by him. It is the continuation of structure number 5 (of Stainier, 1911).

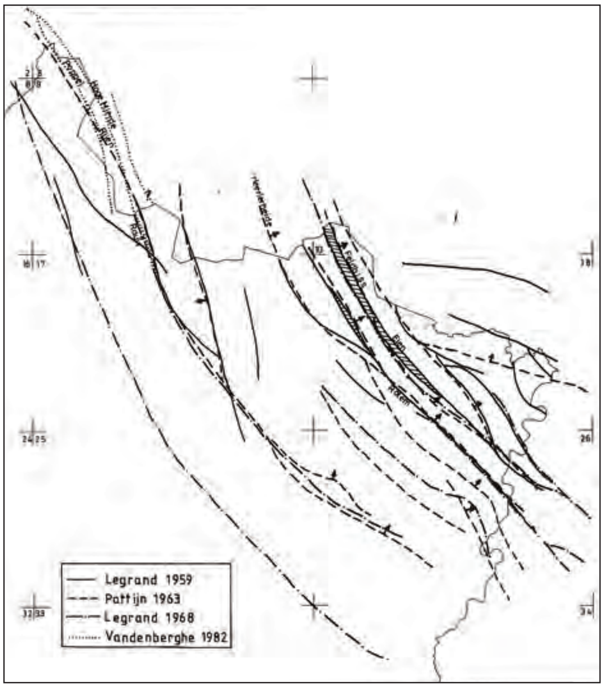


Fig. 146. Evolution of views on the structure of the southwest margin of the Roer Valley Graben from 1959 to 1982 (Demyttenaere, 1989).

The northwest segment of the Rotem Fault links up with the Neeroeteren Fault, which itself, further to the northwest, splits into multiple branches (where the Roer Valley Graben widens) of which the most westerly is the Grote Brogel Fault, followed by the Reppel and Bocholt-Hamont faults (Fig. 147). This fault framework was established by Paulissen in 1973 based on the topographic expression of the faults (steps in the landscape) and to a lesser extent on the influence of the faults on the distribution of Pliocene-Pleistocene deposits (Rhine or Meuse sediments). The geomorphological boundary between the Campine Plateau (graben shoulder) in the southwest and the Bocholt Plain (Roer Valley Graben) in the northeast follows the Neeroeteren fault, passing into a stepwise passage from plateau to plain between the Grote Brogel, Reppel and Bocholt-Hamont faults (Fig. 150).

Geluk et al. (1994) consider the Roer Valley Graben as asymmetric. The graben is bounded by several parallel and antithetic faults, which are the Feldbiss, the Neeroeteren and the Heerlerheide faults at the southwest margin (partly in Belgium, with offsets of about 100-400 metres at the base of the Tertiary) and the single Peel Boundary Fault at the northeast margin (Fig. 148).

According to geomorphologic analyses of the Belgian part of the Roer graben, Paulissen (1997) identifies 5 scarps of tectonic origin, 1 problematic scarp and 2 buried or covered scarps (Fig. 150). The steep Bree Fault Scarp (1 in Fig. 150) strikes NW-SE and is already recognized topographically as a tectonic fault scarp (the topographic expression of "the Felbiss") by Briquet in 1907. The Zutendaal gravel, i.e. the main Meuse river terrace deposit on the Campine Plateau, is displaced (downthown) between 20 and 25 metres to the north. The Berg Fault Scarp (2 in Fig. 150) is a 5 metres scarp delimiting the Tertiary sands to the south from the Pliocene sands to the north. The Grote Brogel Fault Scarp (3 in Fig. 150) displays a maximum offset of 15 metres. To the north of Bree the Reppel Fault Scarp (4 in Fig. 150) displaces the top of the Main Terrace by about 5 m. The gentle Bocholt Fault Scarp (5 in Fig. 150) has an offset of only 2 m. Fig. 150 also shows the Bichterweerd scarp (7 in Fig. 150, refer to the view of Paulissen et al. (1985) above), the poorly constrained Hamont Scarp (8 in Fig. 150) and the Waterloos Scarp (6 in Fig. 150) which remains problematic from a landscape genesis point of view.

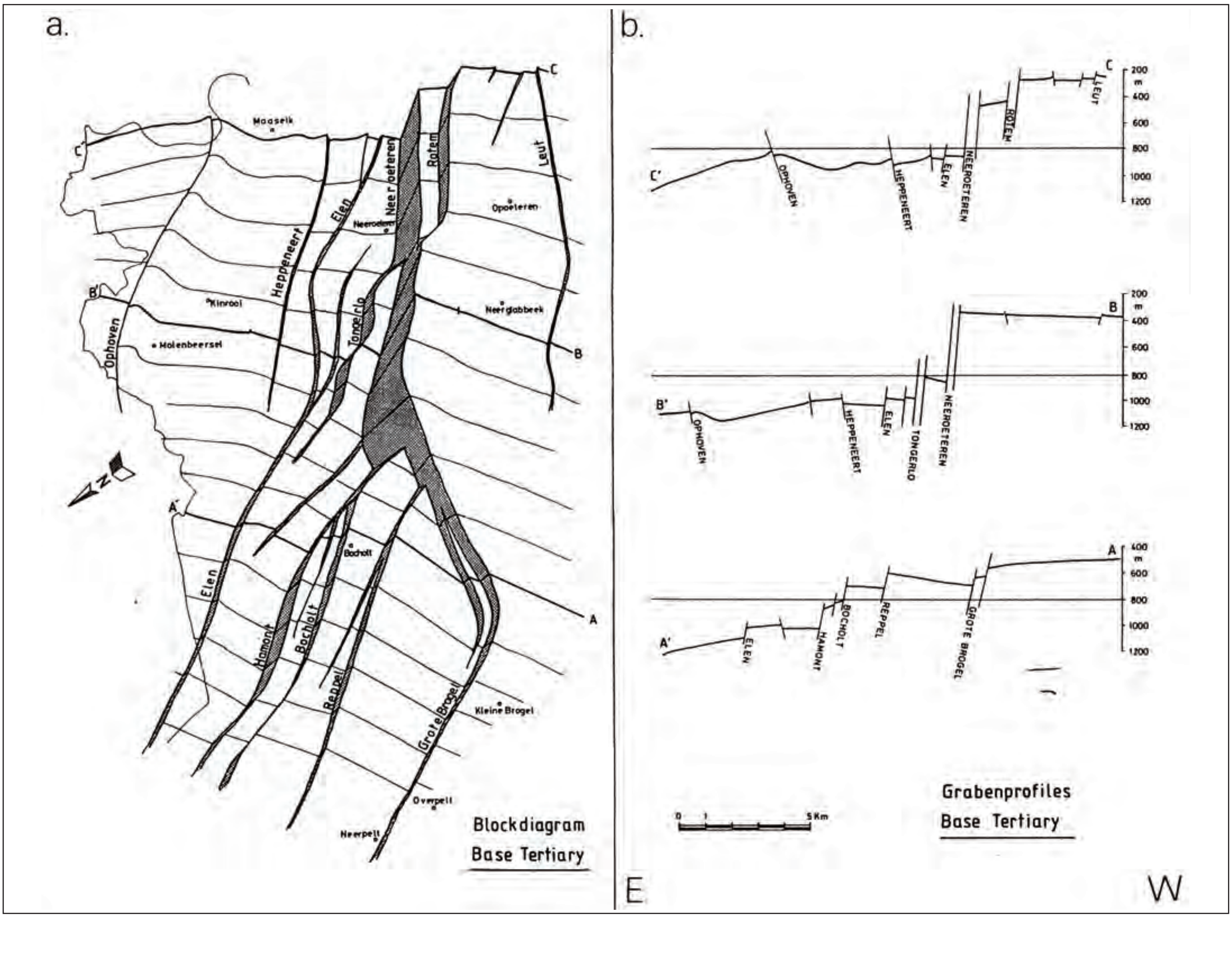


Fig. 147. A. Block-diagram for the base of the Tertiary in the vicinity of Neeroeteren showing the NE-dipping fault scarps of the Belgian Rotem, Neeroeteren, Elen (etc.) faults. B. E-W cross-section through the Roer Valley Graben (Belgian part), the surface presented constitutes the base of the Tertiary. Locations of the profiles are shown on the left (A) (Demyttenaere, 1989).

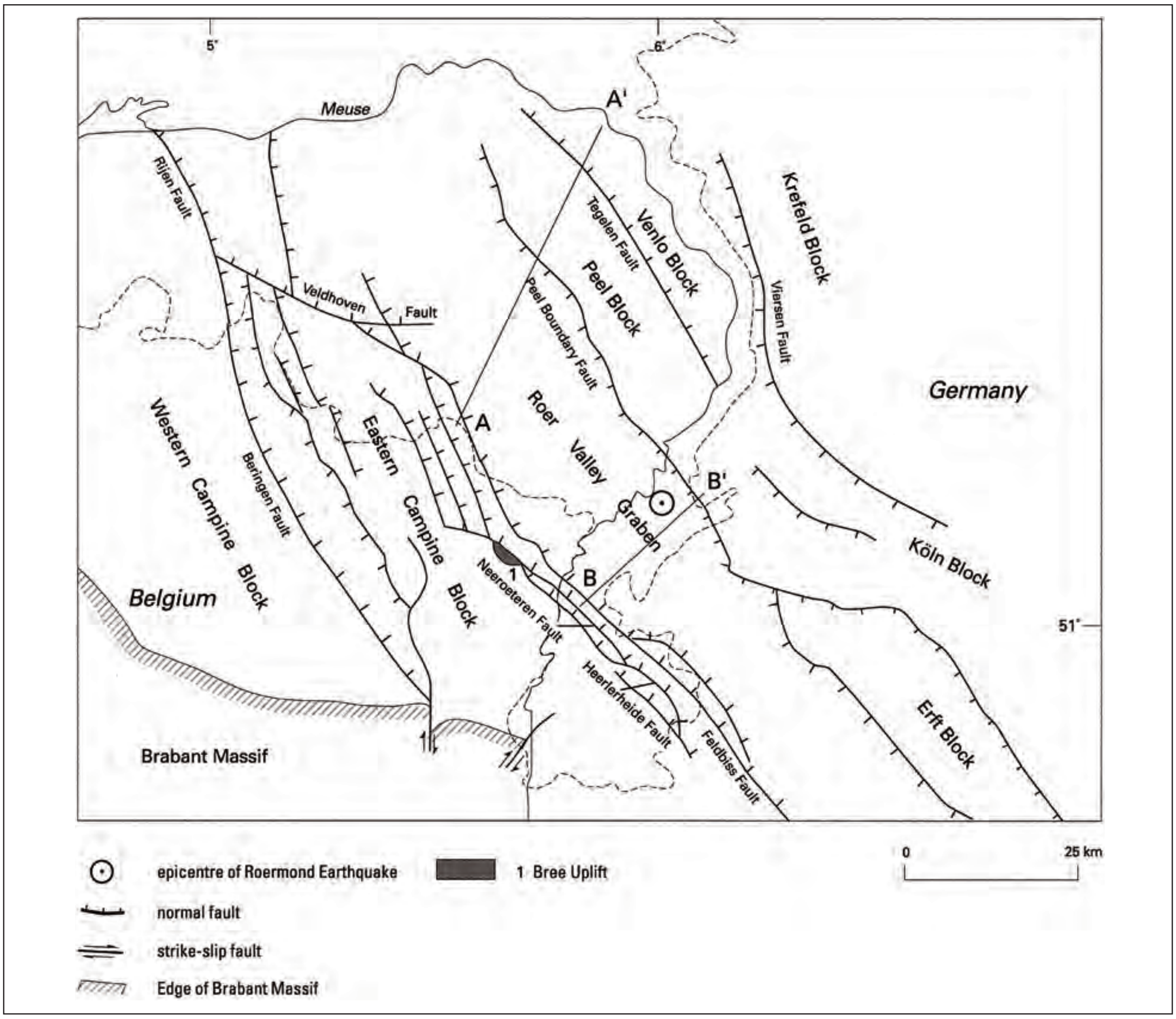


Fig. 148. Structural map of the Roer Valley Graben (in Geluk et al., 1994; modified from maps of Demyttenaere, 1989; Langenaeker & Dusar, 1992; Geological Survey of Belgium; Geologisches Landesamt Nordrhein-Westfalen, 1988 and Van Doorn & Leyzers Vis, 1985). B-B' cross-section is given below in Fig. 149.

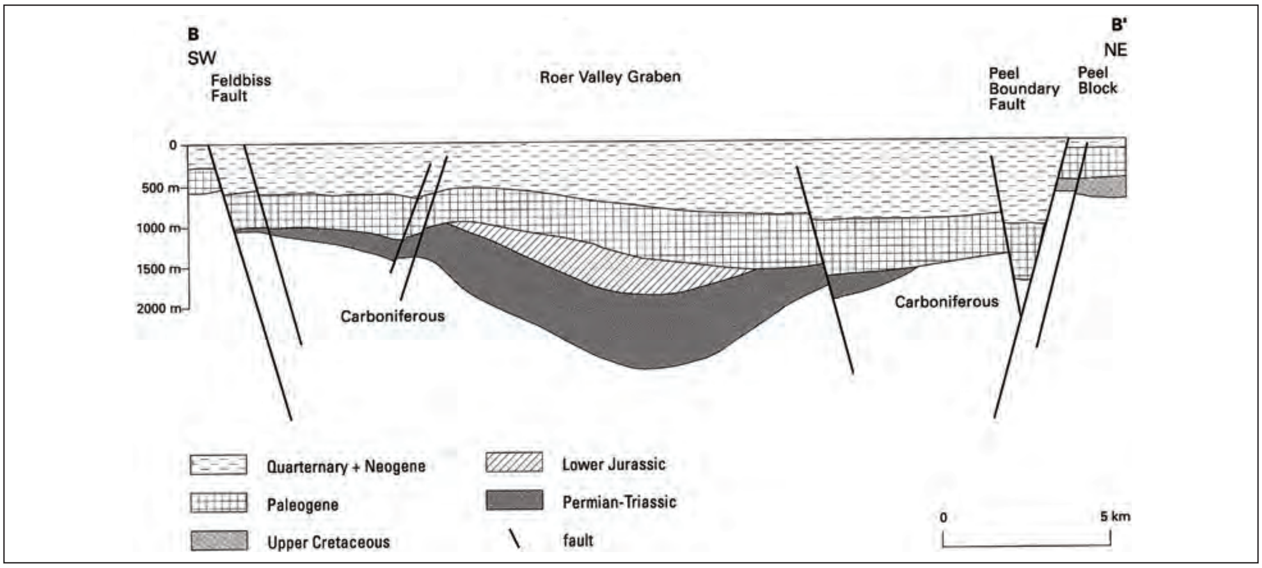


Fig. 149. B-B' SW-NE section through the Roer Valley Graben (Geluk et al., 1994). The location of the section is shown in Fig. 148.

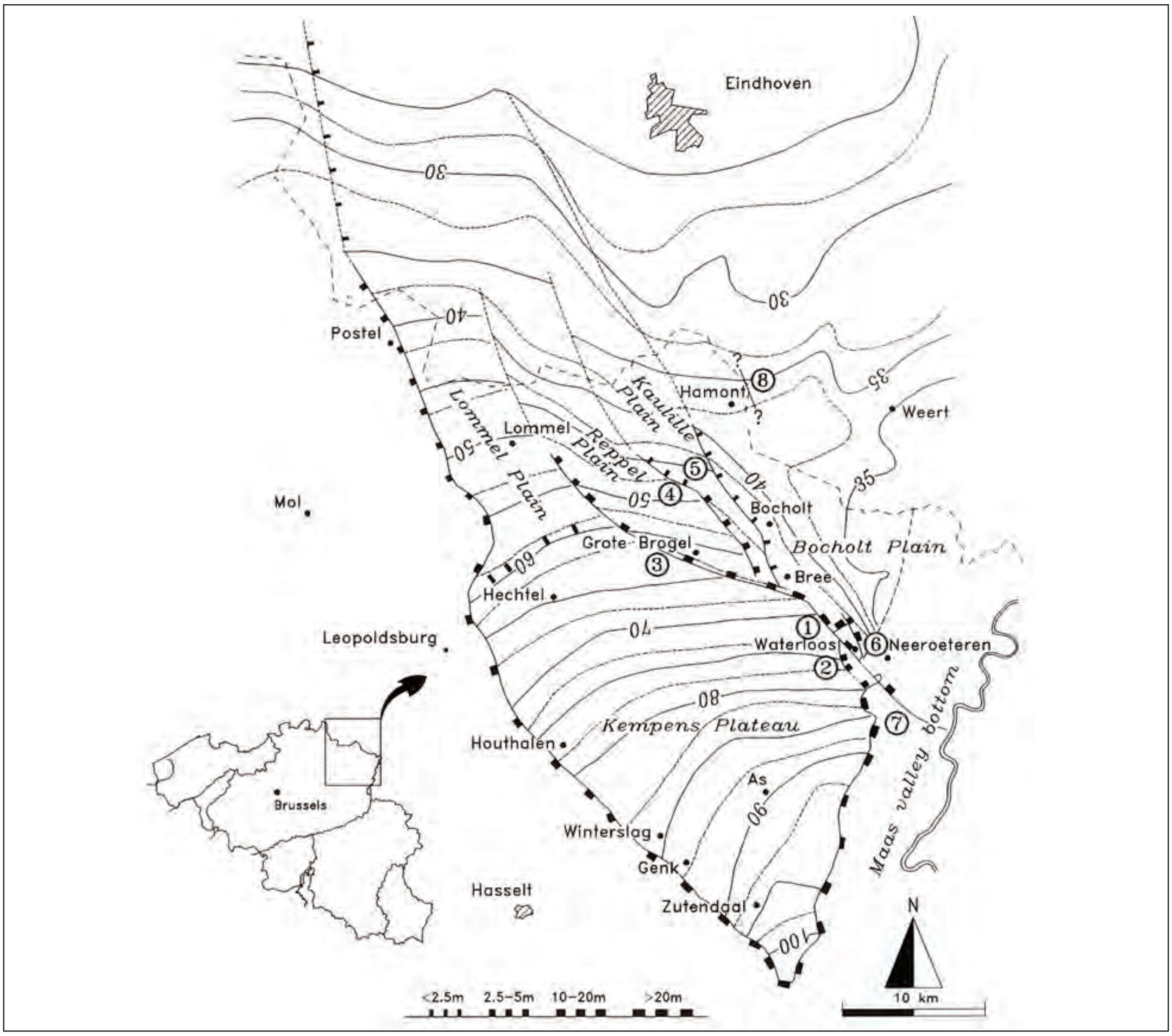
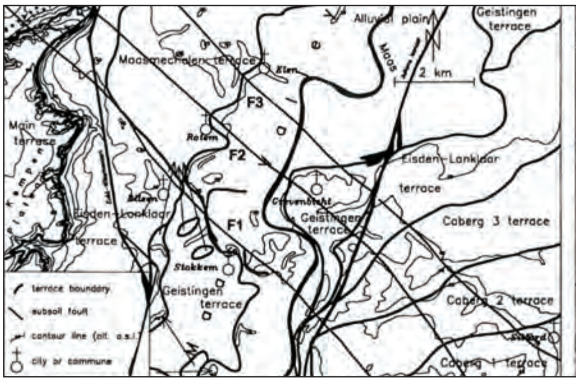


Fig. 150. Map of the morphological features of the Belgian part of the Roer Valley Graben (Paulissen, 1997). 1 = Bree Fault Scarp. 2 = Berg Fault Scarp. 3 = Grote Brogel Fault Scarp. 4 = Reppel Fault Scarp. 5 = Bocholt Fault Scarp. 6 = Waterloos Scarp. 7 = (buried) Bichterweert Scarp. 8 = (buried) Hamont Scarp.

In 1999, Beerten et al. summarize the divergent points of view regarding the structural evolution of the Feldbiss fault system or “Feldbiss Bundle”. The authors consider the fault zone as composed of three main fractures, from south to north, F1, F2 and F3 (Fig. 151).

A summary of the evolution of structural concepts relating to the Feldbiss Fault Zone as compiled in 1999 by Beerten et al. follows:

- from coal prospecting in the Limburg coal field, Forir (1904) and Stainier (1907, 1911) recognize normal NW-SE-striking faults;
- based on seismicity, Patijn (1961, 1963) identifies three faults disrupting the top of Carboniferous deposits, the Heerlerheide Fault (F1), the Geleen Fault (F2) (considered of minor significance in the 1963 paper) and the Feldbiss Fault (F3);
- Kuyl (1971 – Dutch geological map of South Limburg) and Van Montfrans (1975 – Dutch geological map of the Netherlands) reproduce the fault traces of Patijn;
- Demyttenaere (1988) observes three faults at the base of the Miocene and attempts to connect them with their Dutch equivalent. F1 is called the Rotem-Heerlerheide Fault, F2 the Neeroeteren-Geleen Fault and F3, the Elen-Feldbiss Fault;
- Geluk et al. (1994) consider the three main fractures as follows: the Heerlerheide Fault (F1) in the south, the Neeroeteren Fault (F2) and the Feldbiss Fault (F3) in the north;
- Langenaeker (1998) identifies three faults in Permian to Jurassic rocks in Belgium. From south to north respectively these are the Heerlerheide Fault (F1), the Feldbiss Fault (F2) and the Elen Fault (F3).



	Belgian Limburg	Belgian and Dutch Limburg	Dutch Limburg
Fault 3	Demyttenaere (1988) Elen fault	Geluk et al. (1994) Feldbiss fault	Pattijn (1961 and 1963) Feldbiss fault
	Langenaeker (1998) Elen fault		Paulissen et al. (1985) Feldbiss fault
Fault 2	Paulissen et al. (1985) Feldbiss fault	Geluk et al. (1994) Neeroeteren fault	Pattijn (1961 and 1963) Geleen fault
	Demyttenaere (1988) Neeroeteren fault		
	Langenaeker (1998) Feldbiss fault		
Fault 1	Demyttenaere (1988) Rotem fault	Geluk et al. (1994) Heerlerheide fault	Pattijn (1961 and 1963) Heerlerheide fault
	Langenaeker (1998) Heerlerheide fault		

Fig. 151. Left. Simplified map of the Rotem area showing the topography, the Meuse terraces and the three main faults (F1, F2, F3) of the Feldbiss fault system (Beerten et al., 1999). **Right.** Table of the names attributed to the three faults (1, 2 and 3) according to different authors.

Table 3. Correlation of the Feldbiss Bundle faults between the Meuse valley (four first studies) and the Dutch South Limburg (Geluk et al., 1994). Dusar et al. (2001) consider the Bichterweerd Scarp as a transpressional structure that originated as a transfer fault between the Elen and the Feldbiss faults located west and east of the Meuse respectively.

Langenaeker (1998)	Demyttenaere & Laga (1988)	this work	Beerten et al. (1998)	Geluk et al. (1994)
Elen		Elen		
	Elen	Bichterweerd	F3	Feldbiss
Feldbiss	Neeroeteren	Geleen	F2	Geleen
Heerlerheide	Rotem	Heerlerheide	F1	Heerlerheide

In 2001, Dusar et al. highlight the structural complexity of the southwest margin of the Roer Valley Graben, which is composed of many fractures. These faults generally present changing directions and offsets along strike during successive tectonic activations. The many geologists that in the past focused on particular stratigraphic-structural levels in the graben therefore have divergent opinions regarding the splitting and/or joining of the faults (Table 3). As a consequence, Dusar et al. (2001) indicate that beyond the term "Feldbiss Fault Zone", no simple structural scheme can be proposed.

Interpretations

In Belgium, the Roer Valley Graben does not appear abruptly at the margin of the Campine Palaeozoic plateau (Stainier, 1911). The central and deepest part of the graben is the result of successive subsiding blocks where movement is conducted along steeply north-dipping normal faults.

Legrand (1961) and Patijn (1963) suggest continuous subsidence within the Roermond Graben from the Upper Oligocene (Chattian) onwards. They also identify inversion tectonics resulting in a reduction in thickness of the Cretaceous (see for example the Senonian thickness reduction in the Roer Valley Graben in Fig. 140).

Subsidence of the graben and tilting of its flanking blocks would result from 4 main stages (Bouckaert et al., 1981): (1) a first Asturian stage of block faulting, (2) a second Kimmeric (Jurassic) stage of normal faulting and graben initiation, (3) a third Laramide (Cretaceous) stage of inversion of the subsided blocks and (4) a fourth Tertiary stage of normal faulting in the same sense as stages 1 and 2. Evidence from the Cretaceous rocks that are apparently not involved in the Carboniferous uplift in the Bree area (i.e. the Bree Uplift) suggests that the Roer Valley Graben formation probably postdates the Bree Uplift. Actually, Bouckaert et al. (1981) agree with the view of Legrand (1961) for whom the Upper Cretaceous of the Jurassic basin (i.e. stage 2) was structurally reverse and therefore characterized by either a non-deposition or the denudation of the most of the Cretaceous cover.

Paulissen (1973) studies the morphology, and the top and base of the Meuse Valley fluviatile deposits and brings new insights to the Quaternary fault activity along the southern border of the Roermond Graben. The Feldbiss Fault, which coincides with the tectonic escarpment limiting the Campine plateau to the northeast, has been active during different stages. The most recent activity occurs during the deposition of the Mechelen a/d Maas and Geistingen terraces. The influence of the Geleen Fault remains unknown and no activity is suspected along the Rotem-Heerlerheide Fault during or after the deposition of the Eisden-Lanklaar terrace.

A geoelectric survey done by Vandenberghe (1982) in the Late and Middle Quaternary close to the Belgian-Dutch border (near Eindhoven) provides evidence for younger fault activity (i.e. of Middle and Upper

Pleistocene age) at the southwest margin of the Roer Valley Graben. It appears that Quaternary activity is weak (fault movements of less than 10 metres) and continued to recent times (i.e. until probably a few hundred thousand years ago).

In 1985, Paulissen et al. suggest that the Maas terraces constitute good reference levels for dating the Quaternary fault activity on the southwest border of the Roermond Graben. Main results are:

(1) the most important younger Pleistocene fault activity is synchronous with the aggradation of the Eisden-Lanklaar terrace (i.e. the youngest of the Saalian terraces, referenced number 3 in Figs. 143 & 144); and

(2) minor posterior faulting is dated to between the Eemian and Weichselian Upper Pleniglacial periods. No post-Upper Pleniglacial activity is detected.

Rossa (1986) presents a detailed study of the Upper Cretaceous and Tertiary tectonic inversion for both the Campine area and the western Rhenish-Westphalian coal district (Germany). A chronology of the events is as follows: *"Inversion tectonics possibly started in the Cenomanian, certainly in the Turonian, culminated in the Coniacian to Campanian, gradually decreased in the Maastrichtian and Lower Tertiary"*. Another conclusion of Rossa's 1986 work concerns the Cenozoic normal faulting that, subsequent to the inversion, has compensated and obliterated the reverse displacements at the level of the Cretaceous, except for the southwestern boundary of the Bree Uplift.

On the basis of seismic campaigns and geophysically logged wells, Demyttenaere (1989) proposes a post-Paleozoic tectonic history of north-eastern Belgium; special attention being made to the Roer Valley Graben: (1) Carboniferous, Triassic and Lower Jurassic times are marked by significant subsidence of the entire Campine Basin; (2) as a consequence of the Early Kimmerian tectonic phase, differential subsidence affects the north-eastern part of Belgium during the Lower Jurassic; (3) the Late Kimmerian tectonic phase, Upper Jurassic (Lower Malm) in age, is responsible for the formation of the Roer Valley Graben; (4) a period of relative tectonic quiescence; (5) the Upper Cretaceous is marked by a tectonic inversion resulting from a contractional stress regime and during which the graben is converted into a structural high; (6) Paleocene, Eocene and Lower Oligocene times constitute a relatively quiet tectonic period with local weak inversion; and (7) the return of a period of subsidence of the graben from the Upper Oligocene until today.

Demyttenaere (1989) considers the Rotem Fault as the most important fault of the southwest margin of the graben during the Mesozoic Kimmerian tectonics (as was already stated in Tys, 1980). Mesozoic deposits are well preserved in the graben as a result of the strong subsidence along the fault (Fig. 147A gives the block-diagram of the top of the Mesozoic that is progressively downthrown towards the northeast). Afterwards, the Neeroeteren Fault becomes the most significant fault (normal offset more than 400 m) during the Tertiary Alpine tectonics.

Paulssen et al. (1992) provide the first interpretations of a strong earthquake of intensity VII on the MSK scale occurring on April 13, 1992 in the region of Roermond in the Netherlands. The Roermond earthquake is the strongest earthquake recorded since the onset in 1904 of seismic measurements in the Netherlands and was felt in large areas of the Netherlands, Belgium, Germany, France and England. Paulssen et al. (1992) measure an epicenter of 51°10.2'N – 5°58.3'E (i.e. within the Roer Valley Graben; Fig. 152) and a focal depth of 21 km.

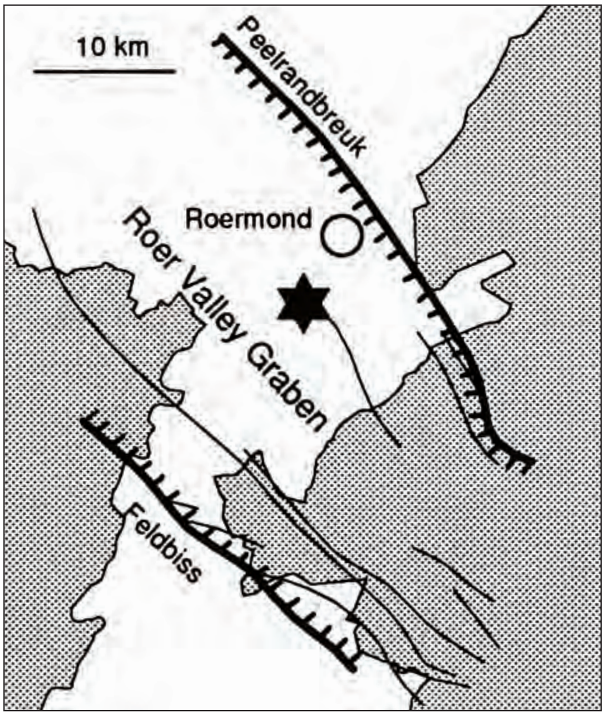


Fig. 152. Location (see the black star) of the Roermond earthquake epicenter (Paulssen et al., 1992). The Feldbiss and Peel Boundary faults are given for Dutch territory.

The earthquake most probably occurred during a pure dip-slip movement along a 124° striking and 70° southwest dipping fault plane. The Roermond earthquake is interpreted as a normal faulting event along the Peel Boundary Fault revealing the ongoing subsidence of the Roer Valley Graben in an extensional tectonic regime (Paulssen et al., 1992).

In their 1994 paper, Camelbeeck et al. present a compilation of the source parameters of the Roermond earthquake (main shock) derived from multiple studies.

Camelbeeck et al. (1994) envisage a moment magnitude (Mw) of 5.4 and a local Richter magnitude of 5.8. The normal dip-slip movement causing the earthquake is compatible with the tectonic model of the Roer Valley Graben which consists mainly of a NW-SE-trending maximal horizontal principal stress that is responsible for the active rifting of the graben (Camelbeeck & van Eck, 1994).

Ahorner (1994) counts 5 major earthquakes with magnitude ML > 5.0 affecting the Roermond graben since the 1755 (i.e. the “historical earthquake” in Fig. 153); a sixth major seismic event, the strongest earthquake of the 20th century, being the MW 5.4, 1992 Roermond earthquake.

Geluk et al. (1994) consider the development of the Roer Valley Graben to be related to Late Jurassic basin development and the subsequent Late Cretaceous tectonic inversion. The Late Oligocene marks the onset of differential subsidence, which increases during the Miocene, Pliocene and Quaternary and which is accommodated in the northeast solely by the Peel Boundary Fault. No significant movement along the Feldbiss Fault Zone has occurred before the Late Oligocene. Finally, intermittent strike-slip and dip-slip movements along the Peel Boundary Fault would be responsible for spasmodic subsidence during the Late Neogene and Quaternary.

On the basis of the work of Rossa (1986), Geluk et al. (1994) deduce the rates of fault movement on the SW border of the graben. The Late Cretaceous tectonic inversion rate (reverse movement) is lower than 0.01 mm/yr. The Tertiary deepening rate of the graben (normal movement) is estimated to 0.015 mm/yr. Geluk et al. (1994) indicate that the southwestern border of the Roer Valley Graben also has a dextral strike-slip component. The authors interpret the Bree Uplift, along the Neeroeteren Fault, as a positive flower structure.

Paulissen (1997, after 1973) identifies multiple tectonic fault scarps (see above) that are all later than the Main Terrace gravels of the Meuse river (Middle Pleistocene in age). The tectonic activity of these scarps is subdivided into 4 main phases:

- **phase 1:** tectonic activity along the Feldbiss fault system producing an offset of 5 to 8 m along the Berg (which afterwards becomes inactive) and Bree Fault Scarps;
- **phase 2:** tectonic activity of Saalian age (i.e. during the deposition of the valley bottom gravels of the Meuse river) along the Feldbiss fault system with displacements of about 3 to 5 m. The Bichterweert Scarp and the Bree Fault Scarp develops;
- **phase 3:** tectonic activity of Weichselian age (i.e. during the deposition of the Mechelen a/d Terrace) along the Feldbiss fault system with displacements of about 5 m. The Bichterweert and Bree Fault Scarps develop further;
- **phase 4:** “*historical tectonic displacement along the Bree Fault Scarp*” as already suggested by Camelbeeck & Meghraoui (1996).

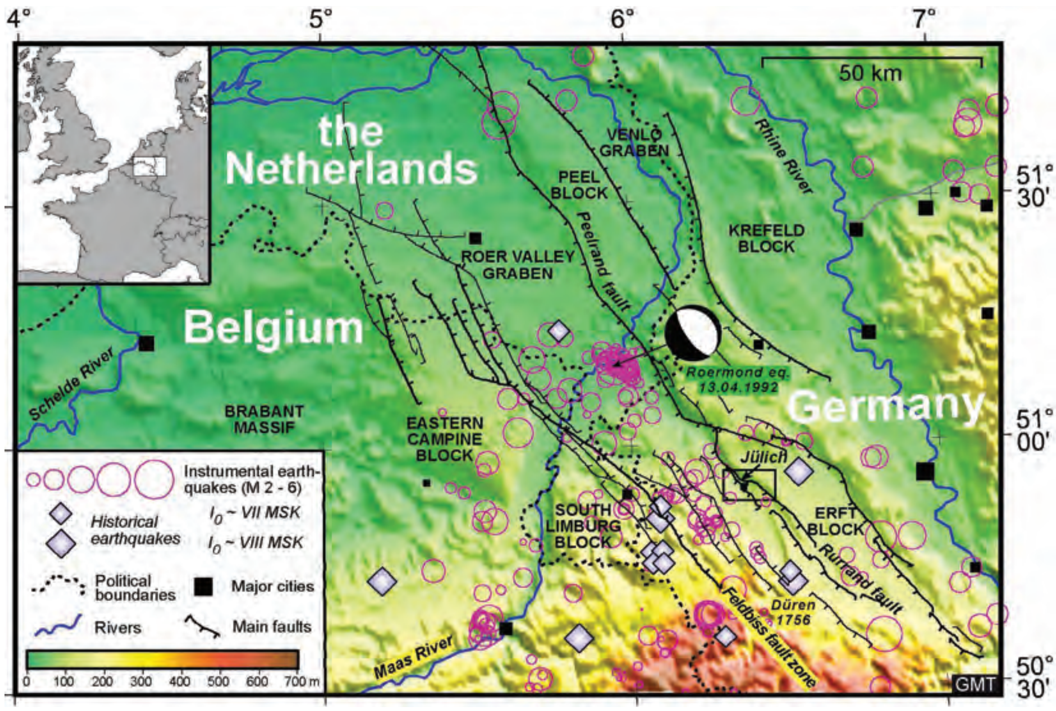


Fig. 153. Seismotectonic map of the Lower Rhine graben area (http://www.seismologie.be/SAFE/Rotem/). Main Quaternary faults and instrumental (from 1911 to 1999) and historical earthquakes are presented. See Vanneste et al. (2001) and Vanneste & Verbeeck (2001) for details.

In 1998, Camelbeeck & Meghraoui show evidence for seismic surface faulting along the Feldbiss Fault (actually along the Bree Fault Scarp) occurring during the Late Pleistocene and Holocene. According to ^{14}C dating evidence, the last earthquake event along the Bree Fault Scarp would have occurred between 610 and 890 AD. A vertical coseismic offset of between 0.5 and 1 metre and a moment magnitude (Mw) of at least 6.3 are proposed. Focusing on the three newly defined surface-faulting earthquakes during the last 28,000-35,000 years BP along the Bree Scarp, Camelbeeck & Meghraoui (1998) propose an earthquake return period of about 12 ± 5 ky and a vertical deformation rate of about 0.06 ± 0.04 mm/yr. Later, in 2000, according to paleoseismic analysis along the NW-striking and 70° dipping Bree Fault Scarp, Meghraoui et al. identify three large earthquakes during the last 45 ky along the scarp. A relative vertical (normal) deformation rate of 0.07 mm/yr is inferred.

Langenaeker (1999, 2000) does not agree with the view of Rossa (1986) who considers the Bree Uplift as the result of the inversion of a significant Cimmerian normal fault located at the SW border of the uplift. Langenaeker (1999) does not observe this Cimmerian fault or even any overthrust at the SW border of the Bree Uplift. He therefore proposes a more complicated tectonic model that includes transpressional movements (dextral strike-slip component) along the graben border faults. The SW boundary of the Roer Valley Graben is not a straight line but displays two major bends: one located south of Opitter and another located west of Bree. Based on the observation of multiple strike-slip faults (Harding, 1985), Langenaeker (1999) suggests that a right-lateral wrench component on the Heerlerheide-Feldbiss-Grote Brogel Fault system (during the "Sub-Hercynian" tectonic stage) would result in both a "restraining bend" (observed to the west of Bree) and a "releasing bend" (observed to the south of Opitter, Fig. 154). Agreeing with the paper of Geluk et al. (1994), Langenaeker (1999) considers the Bree Uplift as a half-flower structure resulting from a pop-up effect at the "restraining bend" of the Feldbiss Fault and from a reverse movement on both the southwestern border fault of the Bree Uplift and the Feldbiss Fault. Dusar et al. (2001) suggest that the Bichterweerd Scarp could be a buried example for such transpressional event in the relay zone between two faults.

In 1999, Vanneste et al. identify multiple soft-sediment deformation along the Bree Fault Scarp (i.e. the Feldbiss Fault). The genesis of small-scale normal faults, asymmetric folds and load structures for example, as well as exceptional liquefaction of the sediments constitute evidence for several earthquakes. These seismogenic deformations in the superficial deposits are probably related to at least three distinct events in the last 30,000 years and one event before. These events would be analogous in size or even exceeding the Ms 5.3 Roermond earthquake in 1992. Later, in 2001, Vanneste et al. make a new investigation of a trench along the Bree fault escarpment and highlight 6 paleo-earthquake events since the late Pleistocene. A seismotectonic map (Fig. 153) of the Roer Valley Graben and of its neighbouring tectonic blocks is proposed in 2001 by Vanneste & Verbeeck.

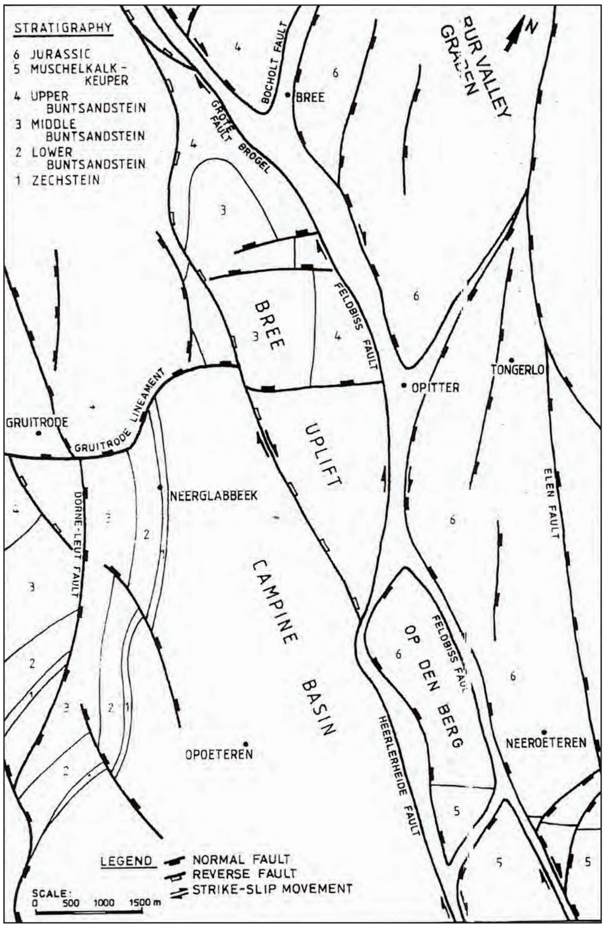


Fig. 154. Map of the Bree Uplift showing the Pre-Cretaceous subcrop (Permian to Jurassic, numbers 1 to 6; Langenaeker, 1999).

Houtgast et al. (2002) carry out a geomorphological survey in the Netherlands (in the vicinity of Sittard) in order to estimate the displacement history of terrace deposits that are disrupted by the faults of the Feldbiss Fault Zone and to locate these faults (Fig. 155). During the Middle and Late Pleistocene, the whole fault system (composed of the Heerlerheide, Geleen and Feldbiss faults) has an average displacement rate between 0.041 and 0.047 mm/yr. Individual faults show an average rate ranging between 0.010 and 0.035 mm/yr (Fig. 156). The authors consider the Heerlerheide Fault to contribute less than 10% to the total offset along the Feldbiss Fault zone.

The Feldbiss Fault Zone is considered as a system of overstepping faults, actually Paleozoic strike-slip faults reactivated in a normal way. From Sittard to the Meuse river (i.e. in a direction towards Belgium), the contribution of the Feldbiss Fault to the total displacement of the Feldbiss Fault Zone decreases while the contribution of the Geleen Fault increases (Fig. 156). The extensional strain is therefore transferred from the Feldbiss Fault to the Geleen Fault in a northwestward direction.

Based on correlations established between the Feldbiss and Geleen faults and on previous paleo-seismological studies in the Roermond Graben, Houtgast et al. (2005) suggest that a single or multiple, moderate to large earthquake event(s) occur(s) about 15,000 years ago. This seismic event was followed by an increase in fault activity (i.e. increased displacement rates) between 15,000 and 10,000 years BP.

Considering both the timing and the extent of the enhanced fault activity, Houtgast et al. (2005) propose a deglaciation origin. Van den Berg et al. (2002) already suggest a relationship between faulting and crustal unloading following the melting of the Weichselian ice-sheet (Fig. 157). Ice-sheet (un)loading can have effects on stresses in the crust as far as 300-500 km from the ice-sheet margins, the largest effects being related to the fore-bulge about 150 km from an ice-sheet margin (Fig. 158;

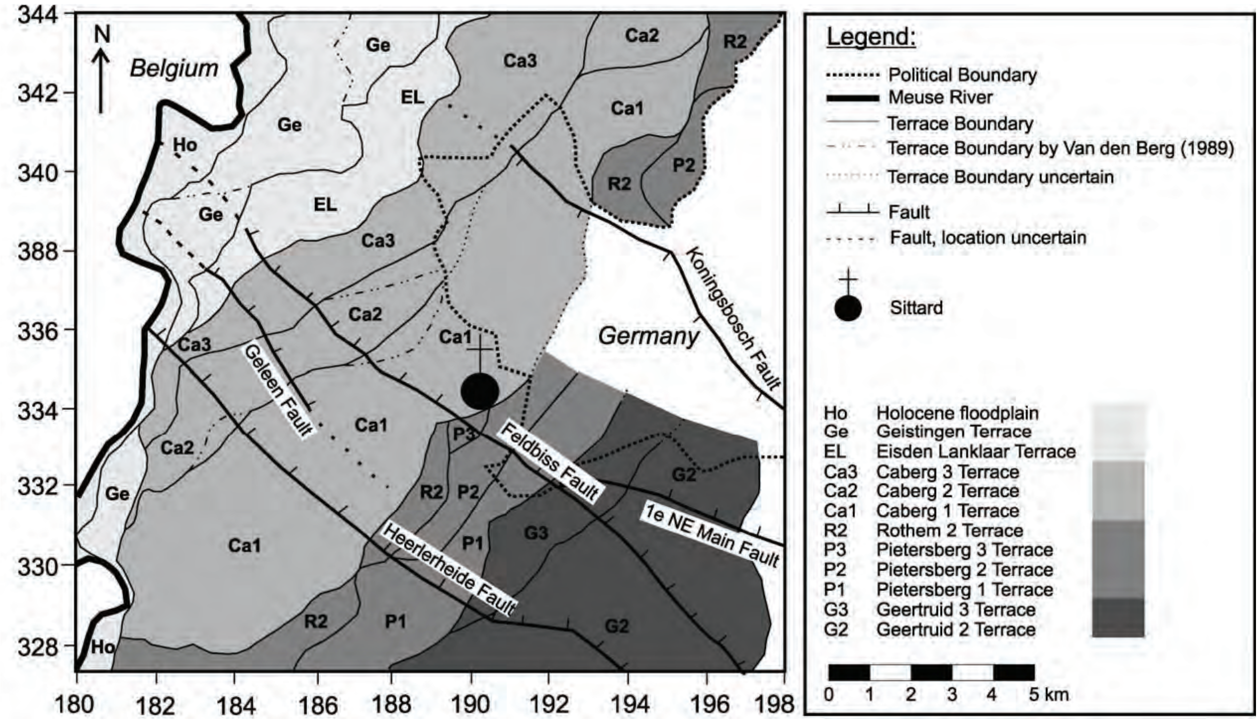


Fig. 155. Dutch map of the terrace deposits in the Sittard area showing the three main faults (Feldbiss, Geleen and Heerlerheide faults) constituting the Feldbiss Fault Zone (Houtgast et al., 2002).

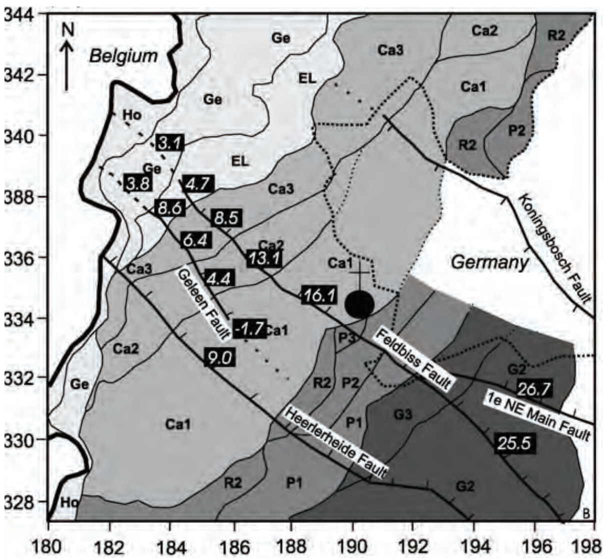


Fig. 156. Displacement (in mm/ky) of the base of the terrace deposits along the faults of the Feldbiss Fault Zone (Houtgast et al., 2002). See Fig. 155 for legend.

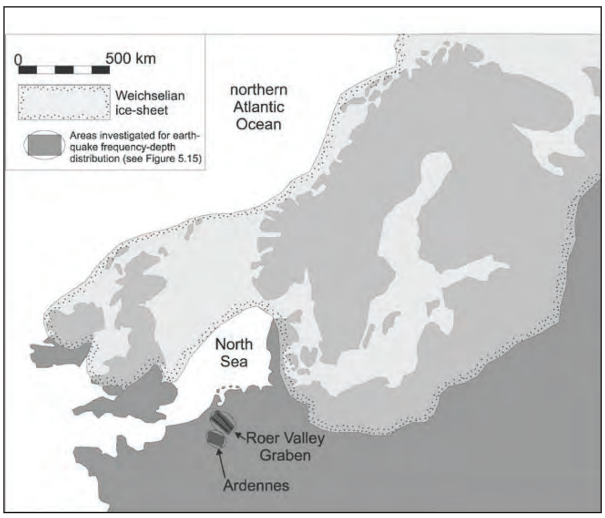


Fig. 157. Simplified map of the extent of the Weichselian ice-sheet during the Last Glacial Maximum (after Boulton et al., 2001 and Sejrup et al., 2000; In Houtgast et al., 2005). The Roer Valley Graben location is shown.

e.g. Muir-Wood, 2000). Houtgast et al. (2005) consider that the fore-bulge of the Late-Weichselian continental ice-sheet affected the Roer Valley Graben. During the deglaciation (i.e. glacial unload), the fore-bulge on the Roer Valley Graben began to collapse between 20,000 and 15,000 years BP. The collapse enabled the release of the extensional constraints that had built up in the crust during the glacial period (Fig. 158). Hypothetically, such unloading-induced stress could explain the increase fault activity along the main faults of the Roer Valley Graben during the initial stage of glacial unloading.

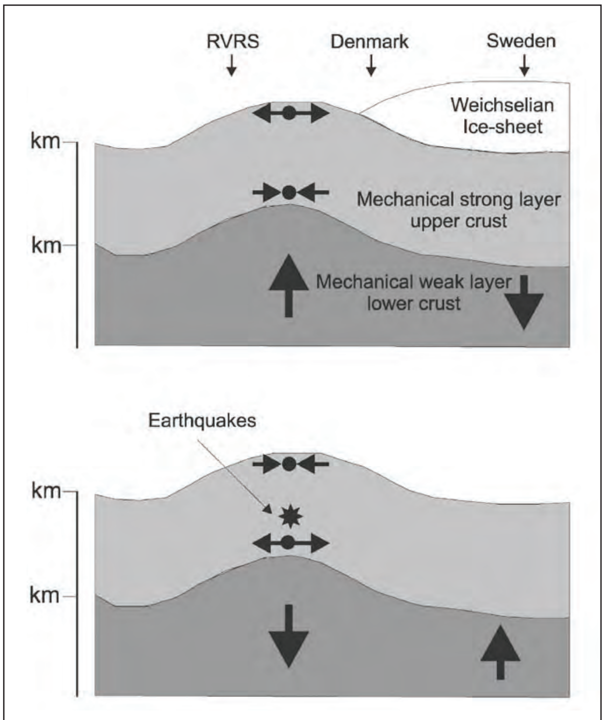


Fig. 158. Simplified model of the glacial loading (**top**) and unloading (**bottom**) at the margins of the Weichselian ice-sheet and the various stresses induced (Houtgast et al., 2005). Not to scale. RVRS = Roer Valley Rift System.

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9.3. Herbeumont Fault

Location

The Herbeumont Fault was identified by Asselberghs in 1921 over a strike length of 25 km between Bouillon and Straimont and was considered to be the continuation of the Aiglemont Fault (see Cambier & Dejonghe, 2010) that was introduced by Gosselet in 1883. The latest mapping of the Herbeumont Fault (Fig. 159) was released in 1954 by Asselberghs (In Fourmarier, 1954). The lineament is recognized for at least 75 km from Aiglemont in the west (France) to Martelange in the east (Belgium-Luxembourg border). Both western and eastern extensions remain possible. The fault is an important thrust in southern Belgium that displaced the Givonne Anticline (i.e. the southernmost Cambrian Inlier of the Ardenne Allochthon) over the Eifel Synclinorium.

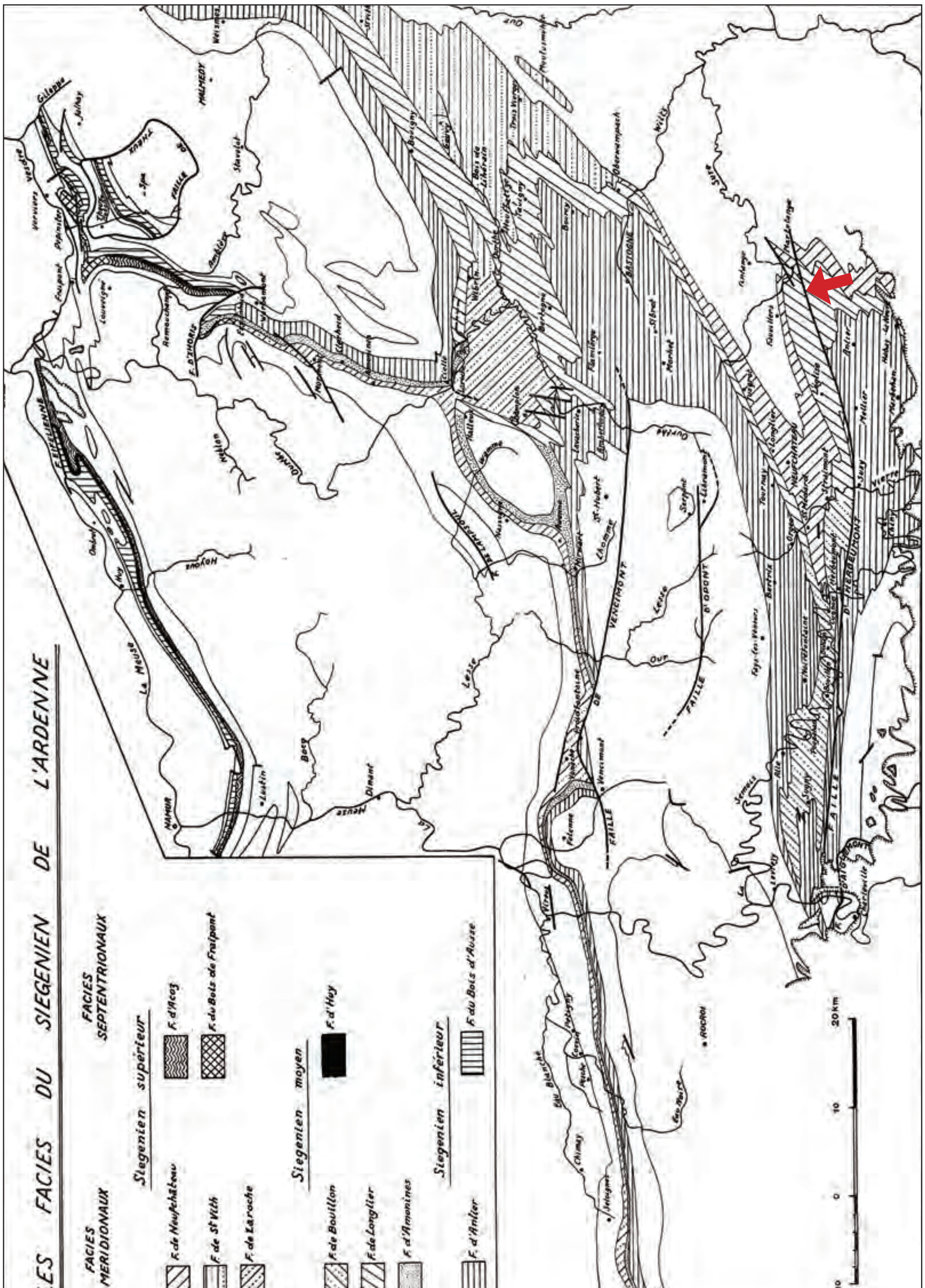

Fig. 159. Regional geological map of the Ardenne Allochthon to the east of the Meuse river showing the main thrusts and the Pragian (i.e. "*Siegenien*", Lower Devonian) terrain (Asselberghs, 1946, 1954).

Lithology and stratigraphy of the country rocks

Asselberghs's geological map of 1946 and 1954 shows that the hanging wall block at surface is mainly composed of Lower Praguian rocks (the former Lower Siegenian). These terrains correspond to the Saint-Hubert and the Mirwart formations that are made up of green shales, quartzites, siltstones and sandstones. The Mirwart and the Villé formations (Praguian in age) constitute the northern autochthonous block. The Villé Fm comprises dark blue shales and slates with characteristic fossiliferous carbonate sandstones.

Geometry

In 1921, Asselberghs identifies a fault between Bouillon and Straimont which subsequently will be recognised as the Herbeumont Fault. Even though Asselberghs does not directly observe the presence of this discontinuity to the west of Bouillon, he proposes to extend and connect it with the Aiglemont Fault located about 15 km westward. This extension is therefore badly constrained. One year later, in 1922, Asselberghs discredits the existence of the Aiglemont Fault in the area concerned and applies the name of Herbeumont Fault to a fault zone recognized between Corbion (western vicinity of Bouillon) and Straimont.

Asselberghs indicates later (1924) the continuity between the Aiglemont and the Herbeumont faults. The author insists on the structural complexity in the Vrigne valley where an "apparent transverse" fracture is identified. This N-S-striking fault segment cannot be prolonged farther northward and displays a probable curve to the east (Fig. 160). The fault has therefore an E-W-trending trace that probably extends eastward and connects with the Herbeumont Fault in the Bouillon region. Asselberghs also gives an estimate of the horizontal (thrust) offset of at least 10 km in the vicinity of Vrigne.

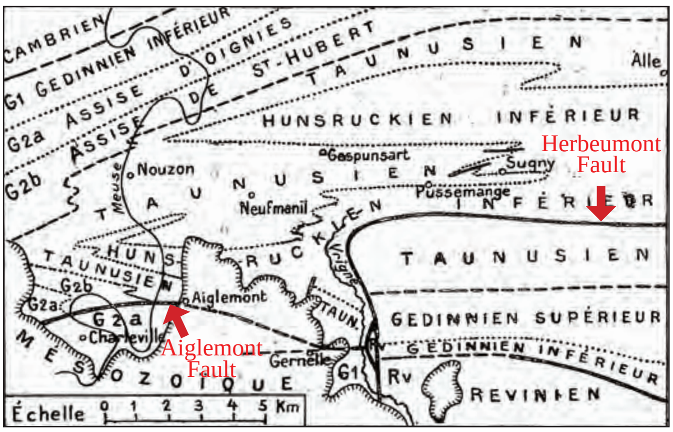

Fig. 160. Structural and stratigraphic map in the Vrigne valley vicinity (Asselberghs, 1924). Note the N-S-trending trace of the western termination of the Herbeumont Fault that disappears under the Mesozoic rocks.

In 1927, Asselberghs envisages a N-S-striking fault in the Vrigne valley that he temporarily calls the Vrigne Fault. This fracture has a strong bend to the east and therefore acquires an E-W strike (Fig. 161). A quite significant reverse offset, estimated at several km in the Vrigne valley, has led the author to propose a connection between the Vrigne and the Herbeumont faults (despite the 12 km separating their respective terminations). He also indicates a low-angle dip of 15° to the south in the Bouillon vicinity and a reverse offset of at least 10 km at the meridian line of the Vrigne river.

Quiring proposes in 1933 to extend the Herbeumont Fault a long away eastward to the Luxembourg Oesling region and to Germany (100 km east of Martelange). The relay would be the "Sauer-Uberschiebung". No arguments were given.

Macar envisages in 1933 a continuity between the Aiglemont and Herbeumont faults. From this assumption, the N-S fault observed in the Vrigne valley would be explained as due to a tectonic wedge trapped between the main thrust and the substratum. In addition, the strong facies differences on hanging wall blocks of both the Aiglemont and Herbeumont faults would be simply explained by facies lateral variations. The author emphasizes the hypothetical character of the connection.

In 1936, Macar introduces the "Ruisseau des Gravis" Fault, a low-angle thrust-type fracture with a dip of less than 10° to the south running from SW of Neufmanil in France to SE of Sugny in Belgium. This fault was already studied by Asselberghs under the name of the Vrigne Fault but this name does not appear in the work of Macar. The fault constitutes one probable extension of the Herbeumont Fault. Two other faults located to the south, i.e. the "Moulin du Gigue" and the "Ruisseau de Borne" faults, (Fig. 162) were also detected. These constitute other possible extensions of the Herbeumont Fault. The western termination of the Herbeumont Fault is therefore divided into 3 branches and the cumulate reverse offset indicates a northward thrust of about 6.5 km. The "Ruisseau des Gravis" Fault shows a N-S strike in the Vrigne valley where it disappears under Mesozoic rocks and, farther southward, probably under the Aiglemont Fault itself. No connection between the Aiglemont and the Herbeumont faults is therefore assumed, which is a completely different interpretation of Macar's publication of 1933.

Asselberghs reiterates in 1940(a) that the western extremity of the Herbeumont Fault has thrust Lower Praguian rocks of the Givonne Syncline over the Praguian terrains of the Eifel Syncline. This segment has an offset of at least 2500 m and displays a N-S-striking trace that disappears under the Mesozoic rocks of the Paris Basin and, further southwards, under the Aiglemont Fault.

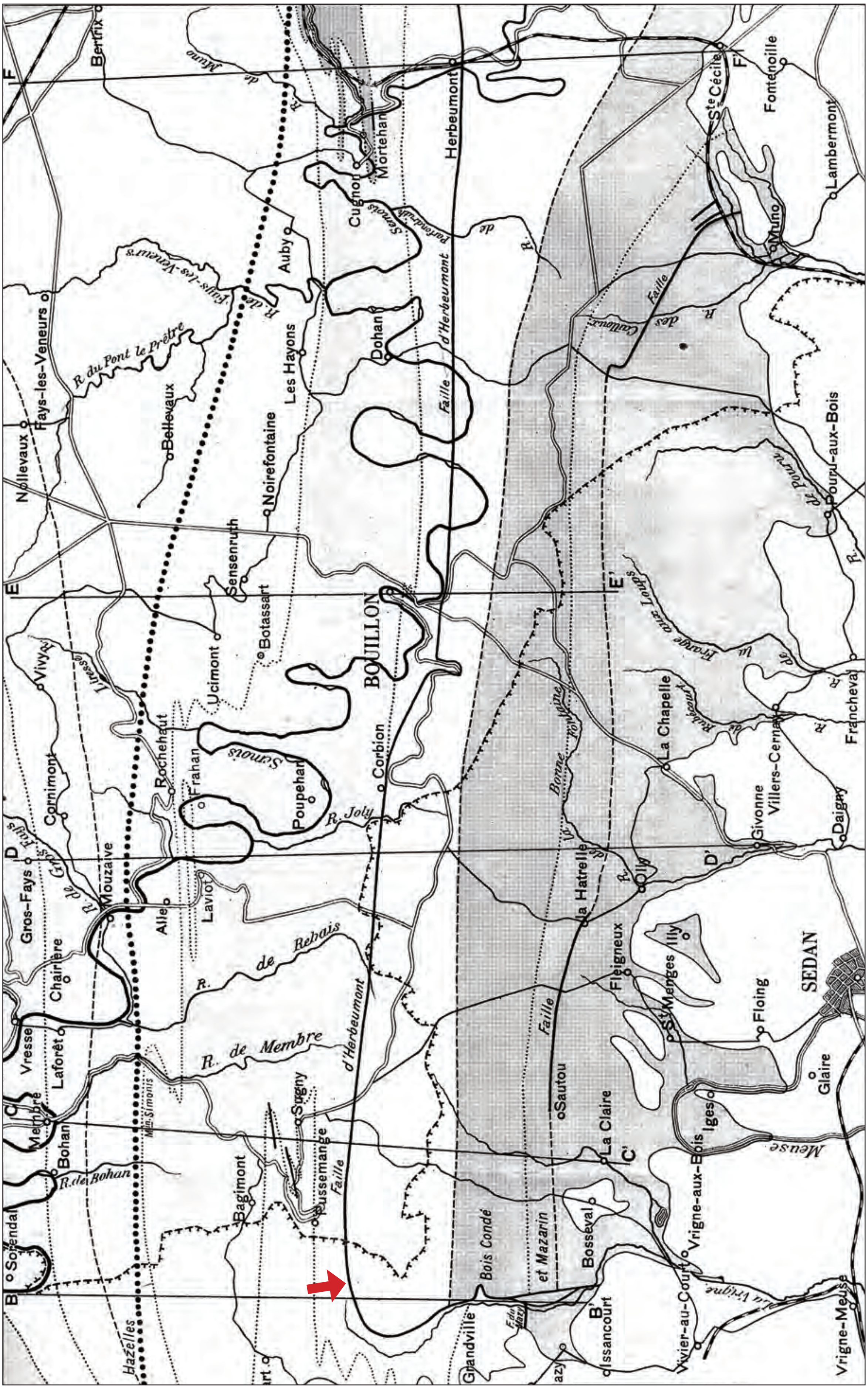

Fig. 161. Extract of the geological map of Asselberghs (1927) showing the particular trace of the Herbeumont Fault. The low-angle fault plane explains the transverse aspect of the trace in the Vrigne valley.

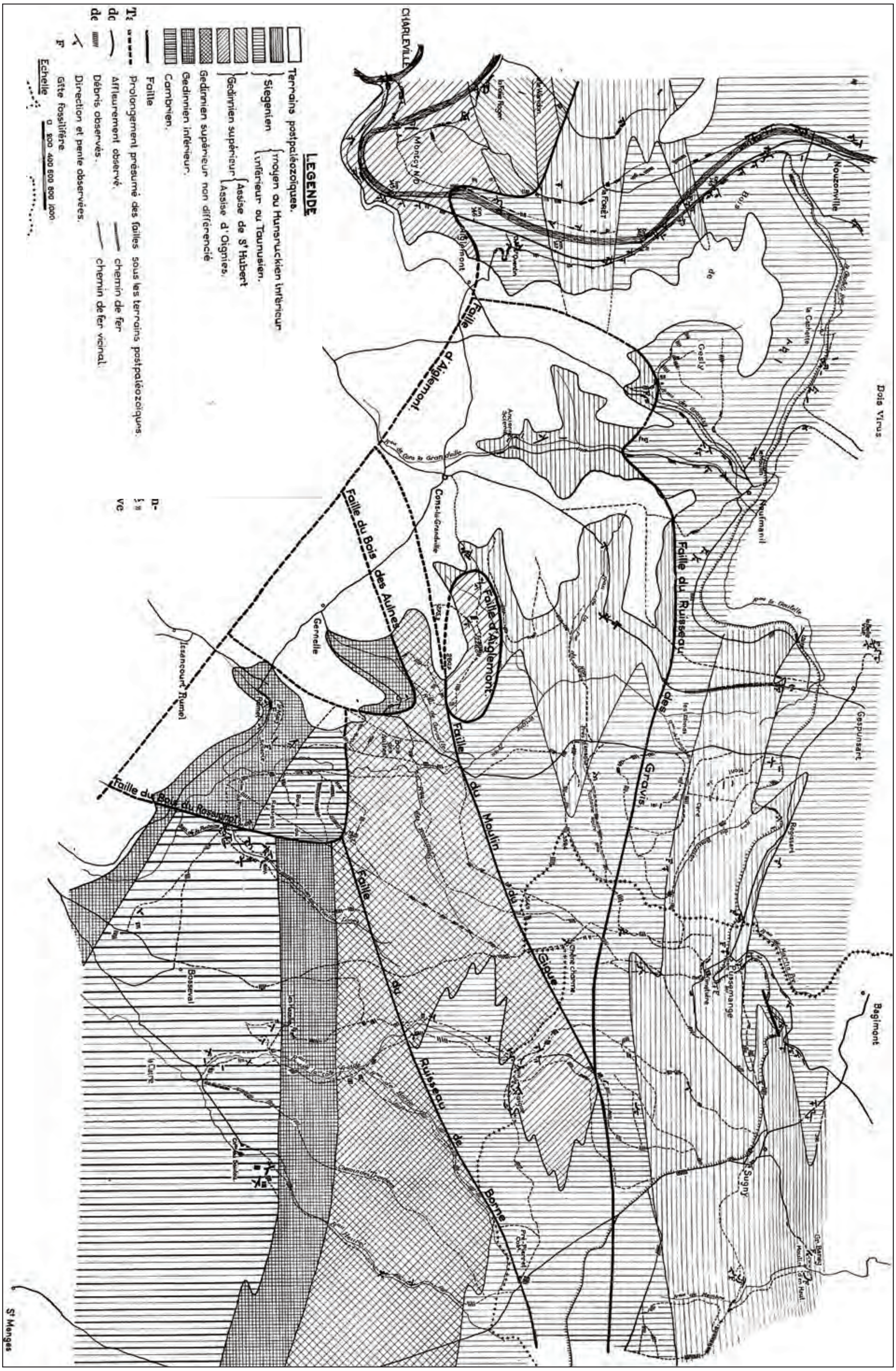

Fig. 162. Structural and geological map in the vicinity east of Charleville (Macar, 1936). The map shows the relation between the Aiglemont and the Herbeumont (i.e. "Ruisseau des Gravis") Faults.

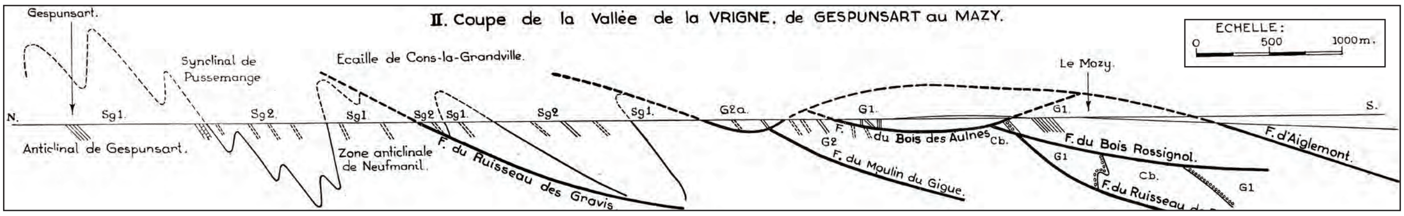

Fig. 163. Cross-section in the Vrigne valley (Macar, 1936). The eastern termination of the Herbeumont Fault (i.e. the Ruisseau des Gravis segment) passes under the Aiglemont Thrust. G1 = Lower Lochkovian. G2(a) = Upper Lochkovian. Sg1 = Lower Pragian. Sg2 = Middle Pragian. Cb = "*Coblencien*" (Lower Devonian).

In 1946, Asselberghs proposes an extension of the fracture eastward from Straimont to beyond Martelange. The Martelange Fault, initially considered as a separate fracture located east of the locality of the same name, has a significant reverse (thrust) offset of about 4 km. The fault is recognized along 6 km. The author has no doubt about the connection between the Herbeumont and the Martelange faults. The latter becomes, therefore, the eastern extremity of the Herbeumont Thrust. The author also indicates that as no observations of the fault were collected in Luxembourg territory he ignores the continuation of the Herbeumont Fault further to the East. Anyway, the Herbeumont fracture is a major longitudinal thrust rightly recognized over a distance of 75 km from Aiglemont to Martelange. The author observed a low-angle dip of 15° in the vicinity of Bouillon.

Interpretations

The Herbeumont Fault is an important structural feature of the south-Belgian regional geology as already recognized in 1921 by Asselberghs. From the Meuse river in France to the Grand Duchy of Luxembourg, the Eodevonian rocks of the Givonne Anticline are thrust over the Lower Devonian formations of the Eifel Synclinorium. The Herbeumont Fault is likely to be related to the Variscan Orogeny and to the setting-up of the Ardenne Allochthon. Macar suggests in 1933 that the northward thrust of the Givonne Anticline may have taken place during the Lower Carboniferous (during the early stage of Variscan shortening).

In 1954, Fourmarier envisages the Herbeumont and the Aiglemont faults as two strictly different and separate thrust fractures. A continuity between those faults is discounted because of facies differences observed between the respective hanging wall blocks. According to Fourmarier, the Aiglemont Fault defines an inner thrust sheet located within another sheet bounded by the Herbeumont Fault (Fig. 163). In other words, he proposes that the thrust region that affects the southern part of the Eifel Synclinorium is composed of two stacked thrust sheets separated by the Aiglemont Fault.

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Macar, 1936.
Quiring, 1933.

9.4. Midi (– Condroz – Eifelian – Aachen) Fault (or Thrust)

With references to the Boussu, Anzin, "Cran de retour", Carabinier, Tombe, Masse, Barrois, Maulenne, Boussale, Ormont, Bois de Presle, Ry du Chapelain, Tihange, Huy, Aguesses-Asse, Tunnel, Theux, Juslenville, Ormont, Kinkempois, Vaux, Ourthe, Streupas … faults.

Location

The Midi Fault constitutes the most significant structural feature of the entire Belgian fault network as it coincides with the Variscan frontal thrust, delimiting the Ardenne Allochthon to the south which is overthrust on the Brabant Parautochthon to the north (Fig. 164). In Belgium and surroundings areas, from the Channel and the French Boulonnais region in the west to Aachen and the Roer Valley Graben in the east, the Midi-Aachen Fault strikes over a distance of about 370 km. It actually constitutes a segment of the northern frontal system of the Western and Central European Variscides recognized for more than 2000 km from SW Ireland to Poland.

Actually, the front zone of the Rhenohercynian fold-and-thrust belt, the northernmost external part of the Variscides in Western Europe, is not simply constituted by a major thrust plane but by a set of many thrusts. Indeed, the separation between the Variscan allochthonous Ardenne domain and the Brabant parautochthonous area is not a sharp contact but a transition frontal zone where many thrusts and superimposed thrust sheets are present. Despite many faults being involved in the complex and "sliced" Variscan Front Zone, two main (historical) faults in Belgium are generally considered to represent the main thrust components of the displacement of the Ardenne Allochthon fold-and-thrust belt over the relatively undeformed Brabant Parautochthon: the Midi Fault and the Eifelian Fault.

Mining works in the Walloon collieries during the 19th century has significantly helped to improve the geological and structural knowledge concerning the elongated Upper Carboniferous Walloon coal basin (*"sillon houiller Haine-Sambre-Meuse"*). The well-developed mining industry was also responsible for the discovery of many fault contacts that were problematic in term of prospecting for coal and various ores, leading to a better understanding of the geometry and origin of important fractures. The Midi Fault in the Hainaut coal-basin and the Eifelian Fault in the Liège coal-basin were particularly important for the miners as these discontinuities limit to the south the industrially significant and workable coal-bearing Namur basin. Stainier (1920a) has published a history of knowledge relating to the thrust faults in the *"sillon houiller"* emphasizing the influence of industrial activity and the observations made in the numerous mines and collieries.

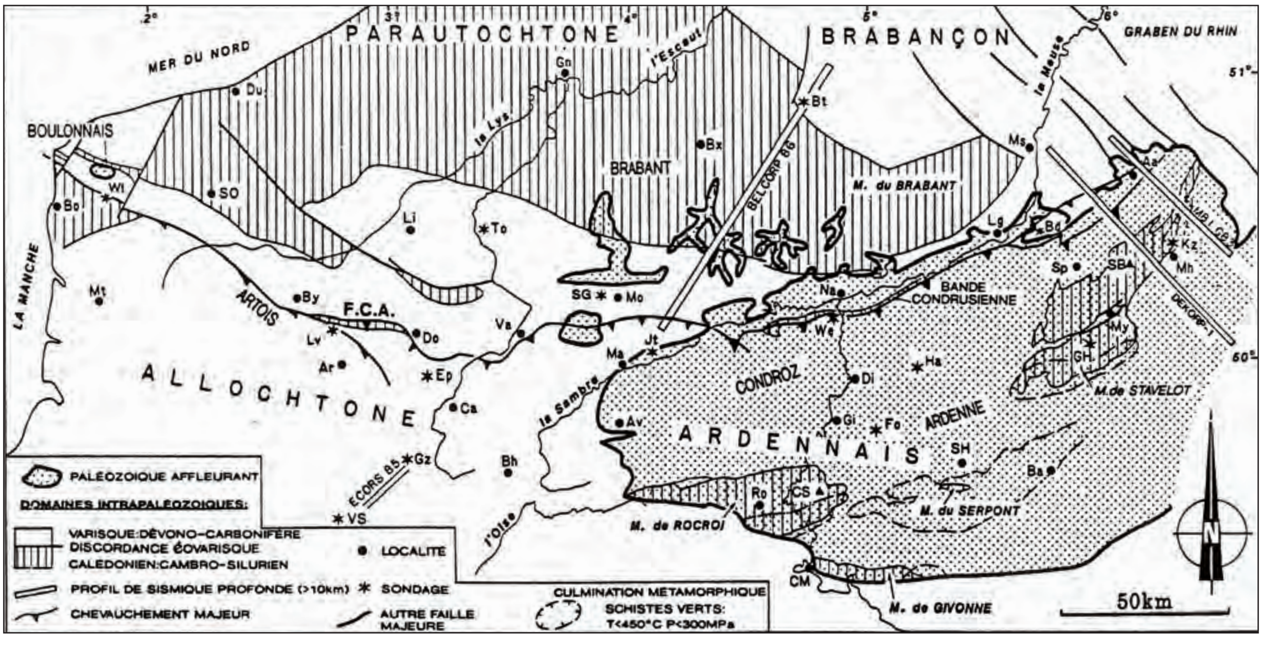


Fig. 164. Locations of the Paleozoic outcrops between the Channel and the Rhine river (Meilliez et al., 1991). The Midi Fault (here referenced as "F.C.A." for "front de chevauchement ardennais") separates the allochthonous Ardenne domain from the Brabant Parautochthon. Boreholes: Bt = Boischoot. Bd = Bolland. Ep = Epinoy. Fo = Focant. GH = Grand-Halleux. Gz = Gouzeaucourt. Ha = Havelange. Jt = Jeumont. Kz = Konzen. Lv = Liévin. SG = Saint-Ghislain. To = Tournai. Vs = Vermandovilliers. We = Wépion. Wt = Le Waast. Deep seismic profiles: M81 (e.g. Meissner et al., 1981; see below). ECORS (e.g. Cazes et al., 1985; see below). BELCORP86 (e.g. Bouckaert et al., 1988; see below). DEKORP-1 (e.g. DEKORP Research Group, 1991). Localities: Aa = Aachen. Ar = Arras. Av = Avesnes/Helpe. Ba = Bastogne. Bh = Bohain. Bo = Boulogne. By = Bruay-en-Artois. Bx = Bruxelles. Ca = Cambrai. CM = Charleville-Mézières. Di = Dinant. Do = Douai. Du = Dunkerque. Gn = Gent. Gi = Givet. Lg = Liège. Li = Lille. Ms = Maastricht. My = Malmédy. Ma = Maubeuge. Mh = Monschau. Mo = Mons. Mt = Montreuil. Na = Namur. Ro = Rocroi. Sp = Spa. SO = Saint-Omer. Va = Valenciennes.

Briefly, from an historical point of view, Dumont (1832) identifies a fault to the south of the Liège coal basin, which is named Eifelian Fault by Malherbe in 1873. Dufrénoy & Elie de Beaumont (1841) detect another fault in the Hainaut coal basin (Valenciennes, Mons and Charleroi regions) that will later be considered by Gosselet (1860a,b) as the continuation of Dumont's fault and called *"Grande Faille"*. Cornet & Briart (1876) designate the fracture as the Midi Fault or Eifelian Fault but suggest in 1877 a preference for the Midi Fault for the entire strike length. The term *"Grande Faille du Midi"* may also be observed in the literature.

According to Stainier (1920a), the first occurrence of the Midi Fault in the literature is dated back to 1777. Actually, based on a work of Sir Laurent, Sir Castille establishes a map of the coal seams from Charleroi to Monchecourt (between Douai and Valenciennes in France). The existence of the Midi Fault is well recognized in 1777 and was simply called the *"crant"*. The fault has a particular significance in the collieries as it displaces the coal measures to the north and has therefore an impact on their exploration.

The northern front of the Rhenohercynian fold-and-thrust belt in Belgium is composed of several relaying segments and the connections between them is the subject of many debates. For example:

- the link between the well-established Midi Fault in the Hainaut area and the Eifelian Fault in the Liège area is made through a problematic narrow strip of Ordovician and Silurian terrains where no major overthrust fault contact is clearly identified. This particular Lower Paleozoic inlier is variously called in the Belgian literature "Bande du Condroz", "Bande de Sambre-et-Meuse", "Bande de Dave", "Bande condruzienne" (Fig. 164), "Bande Silurienne du Condroz", "Anticlinal du Condroz" or "Ride du Condroz" and is generally referred to in English as the "Condroz Inlier" or "Sambre-et-Meuse Strip".

- the connection between the Midi-Eifelian Fault in the vicinity of Liège and the Aachen Fault to the east in Germany remains highly controversial. For Graulich (e.g. 1955, 1984), Graulich et al. (e.g. 1984, 1986) and Hollman & Walter (1995), the Aguesses-Asse Fault is the eastern continuation of the Eifelian Fault and therefore a frontal segment of the Variscan Rhenohercynian fold-and-thrust belt. For other geologists, the connecting feature between the Midi and Aachen thrusts is elsewhere (e.g. Michot, 1986, 1988, 1989). For example, the continuation is assigned to the Tunnel Fault further to the south (Hance et al., 1999). Therefore, we refer readers to the Aguesses-Asse and Tunnel faults (described in detailed in Cambier & Dejonghe, 2010) for extensive explanations of the Variscan Front Zone between Liège and Aachen.

Bibliographic references relating to the Midi-Aachen Fault and to the Variscan Front Zone in Belgium are plentiful and we have based our work on a representative part of them. The list of references below is therefore non-exhaustive. Be aware that other historical reviews have been made such as those of Bouroz et al. (1961) and Sintubin (1992).

Stratigraphy and lithology of the country rocks

Simply, the Midi and Eifelian faults characterize a major anomalous stratigraphic contact between the siliciclastic Lower Devonian rocks of the northern border of the Dinant Synclinorium to the south with Coal Seams bearing-Upper Carboniferous or *"Houiller"* rocks of the southern border of the Namur basin to the north.

The detailed stratigraphy and lithology along the Midi-Eifelian Fault as shown on the old Belgian geological maps at 1:40 000 scale are described below in the Geometry section.

Geometry

In 1832, Dumont considers the Carboniferous Liège coal-basin as being limited to the south by a fault at the contact with the Devonian terrain. This unnamed discontinuity is considered of local significance and is drawn over 11 km from west of the Seraing coal-basin until Angleur in the east. Further to the southwest, the Carboniferous rocks are no longer separated from the Devonian by this fault but instead by an anticlinal wrinkle of old terrains that corresponds to the Lower Paleozoic Condroz Inlier (Dumont, 1835). From a stratigraphic point of view, the Lower Devonian hanging wall of the unnamed fault is assimilated into the *"Système quartzo-schisteux eifélien"* of the *"Terrain anthraxifère"*. Later, in 1873, this particular stratigraphy of the country rocks will be used to justify the name of Eifelian Fault chosen for this fault by Malherbe (see below).

According to Dufrénoy & Elie de Beaumont (1841), the Upper Carboniferous (*"Houiller"*) Valenciennes coal-basin (in French territory) is limited to the south by the *"Poudingue de Burnot"*, a red conglomerate of Lower Devonian (Emsian) age. Despite the contact between the Carboniferous coal-basin and the red "Poudingue de Burnot" not being observable due to an overlying tabular Mesozoic-Cenzoic cover (Fig. 165), Dufrénoy & Elie de Beaumont (1841) present an hypothesis of a fault discontinuity. They also envisage an eastward continuation of the fault into Belgian territory; this fault would therefore also constitute the southern limit of the Mons and Charleroi Carboniferous coal-basin. The fault is drawn subvertically (Fig. 165).

Following the views of Dufrénoy & Elie de Beaumont (1841), Delanoüe (1852) confirms that in the Valenciennes area, the red Devonian rocks are "pulled up" over the Carboniferous ("*Houiller*") terrain through an unnamed fault. This fault would continue in an easterly direction until at least the city of Binche in Belgium. In 1856, Godwin-Austen also agrees with this hypothesis for the fault.

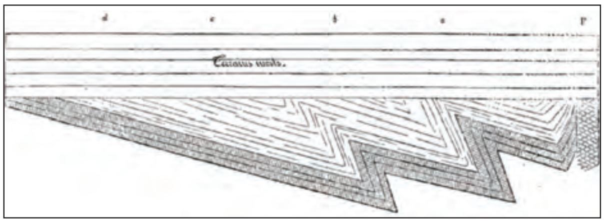


Fig. 165. N-S cross-section of the Carboniferous Valenciennes basin in the Anzin area (Dufrénoy & Elie de Beaumont, 1841). The (unknown) contact between the Carboniferous rocks to the north and the Devonian rocks to the south (F on the figure) is hidden by the flat-lying Mesozoic-Cenozoic cover. The type of contact therefore remains unknown so that the hypothesis of a fault discontinuity is envisaged but not certain.

In 1860(a,b), Gosselet envisages continuity between the fault of Dumont (1832) in the Liège coal-basin and the fault of Dufrénoy & Elie de Beaumont (1841) in the Valenciennes coal-basin. The so-called *"Grande Faille"* would have a principal E-W strike extending from Liège in the east to Mons and Valenciennes in the west and probably extending further westwards under the Cretaceous cover. Gosselet (1860) considers the *"Grande Faille"* to be a key feature of the structural framework of Belgium.

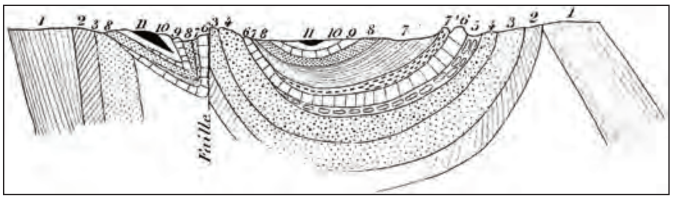


Fig. 166. N-S cross-section of the "*anthraxifère* basin" (Gosselet, 1860a). The subvertical "*Grande Faille*" marks a major regional discontinuity between the northern (Namur) and southern (Dinant) secondary basins. 1 = Silurian. 2 = Gedinnian (i.e. Lochkovian). 3 = "Grauwacke à *Leptoena Murchisoni*". 4 = "*poudingue de Burnot*". 5 = "*schistes à calcéoles*". 6 = "*calcaire de Givet*". 7 = "*schistes de Famenne*". 8 = "*psammites du Condroz*". 9 = "*calcaire de Tournai*". 10 = "*calcaire de Visé*". 11 = "*Houiller*" (i.e. Upper Carboniferous).

The *"Grande Faille"* subdivides the "*anthraxifère* basin" into two secondary basins (Gosselet, 1860a): the southern basin (wide and regular) and the northern basin (narrow and irregular but industrially significant, Fig. 166). These basins correspond to the Dinant and Namur synclinoria respectively. The northern block is principally composed of Upper Carboniferous (*"Houiller"*) rocks but also of Lower Carboniferous, Upper and Middle Devonian terrains while the southern block is mainly made up of the red Burnot conglomerate. Fig. 166 constitutes one of the first sections in which the (subvertical) *"Grande Faille"* appears. In 1860(b), Gosselet specifies: *"le plissement a été suivi d'une faille qui s'étend de Liège à Mons, et peut-être même plus loin, et d'un renversement presque général du bord sud du bassin septentrional"*. In other words, the southern border of the Namur basin is almost entirely overturned to the north.

In 1862, Dormoy suggests that from the Channel to the Prussian area, only the northern half of the Upper Carboniferous coal-basin outcrops. Indeed, he remarks that the strata constituting the coal-basin (i.e. a syncline) always dip to the south and therefore belong entirely to the northern limb of the basin. The southern limb which must exhibit a general northern inclination is not observed. In the Pas-de-Calais, Valenciennes, Charleroi and Liège areas, the coal-basin is actually a "half coal-basin" of which the 20 km wide southern limb has been "removed". Dormoy (1862) explains this particular structure by invoking the upheaval of the southern half of the basin (Fig. 167, *"Ligne de soulèvement et limite sud du Bassin actuel"*) followed by the erosion of the uplifted massif (see interpretations below). In other words, the southern limit of the coal-basin coincides with an unnamed 280 km long plane of upheaval (see Fig. 167 for its direction) that coincides with the *"Grande Faille"*.

In 1863, important observations were made in the vicinity of Liège where mining (colliery) works have shown the continuity of the Upper Carboniferous (*"Houiller"*) terrain under the Lower Devonian rocks that stratigraphically belong to the "système quartzo-schisteux eifélien" of Dumont (1832). This led Briart & Cornet (or Cornet & Briart) in 1863 to suggest the northward thrust of the Devonian over the Carboniferous rocks. In French: *"à une époque géologique postérieure à la formation houillère, le grès rouge a été soulevé et poussé vers le Nord, en glissant sur le terrain houiller dont il a ainsi recouvert une large bande"*.

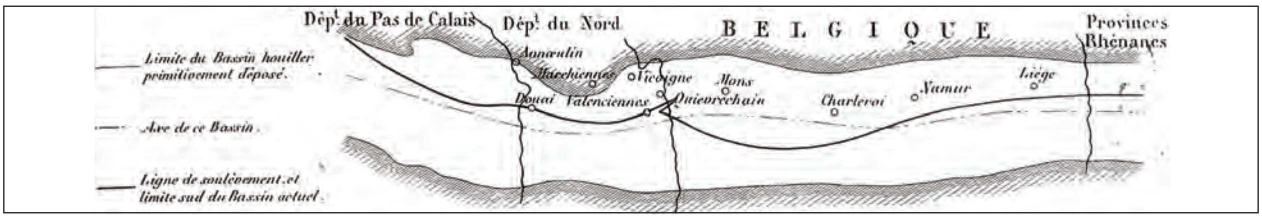


Fig. 167. Schematic map of the Upper Carboniferous coal-basin extension (Dormoy, 1862). The southern half of the basin has been uplifted and removed by erosion (see interpretations below). The 280 km long upheaval trace coincides with the *"Grande Faille"*.

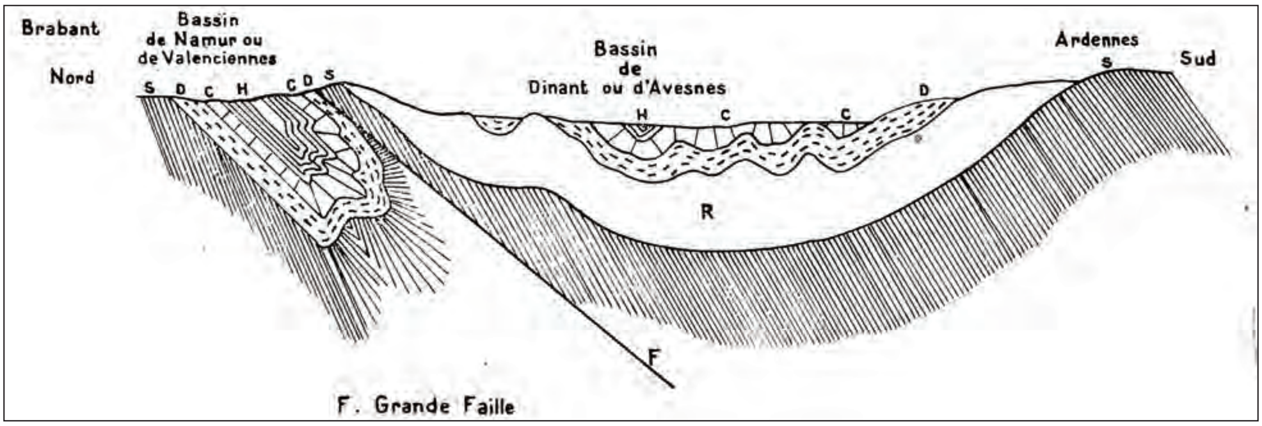


Fig. 168. N-S cross-section through the Namur and Dinant basins (Gosselet, 1874). This section is inspired by the work of Cornet & Briart in 1863.

Again in 1863, Cornet & Briart report important field observations made in a quarry near Binche. They observe northerly overturned *"Houiller"* rocks, steeply dipping (50-60°) to the south, surmounted by non-overturned Devonian rocks gently dipping (10°) to the south. The fault discontinuity is clearly visible in the quarry and consists of irregular strata composed of fragments of both Upper Carboniferous and Devonian age. Like previous geologists, Cornet & Briart develop a unitary (or continuous) concept of the fault. They write: *"Après la formation du terrain houiller dans notre pays, il y a eu, depuis la frontière française jusqu'à la frontière prussienne et même au-delà de ces limites, un mouvement horizontal de translation de l'Ardenne vers le Nord"*. In other words, Cornet & Briart (1863) suggest a major northward "horizontal translation" of the Ardenne, occurring in the region between the French-Belgian border in the west (Valenciennes) and Germany in the east (the "Prussian" border, Aachen), in which extended Devonian cover overlies Upper Carboniferous rocks.

In 1873, Malherbe calls the discontinuity recognized between Clermont-sous-Huy and Angleur that separates the Carboniferous Liège coal-basin to the north from the Eodevonian rocks to the south, the Eifelian Fault. The Eodevonian hanging wall was stratigraphically assimilated into the *"Système quartzo-schisteux eifélien"* of the *"Terrain anthraxifère"* of Dumont (1832), which justifies the choice of the name Eifelian Fault. The fault has a moderate dip of about 45° to the south. The reverse offset along the northeastern segment of the Eifelian Fault is estimated to be 200-250 metres. In his work of 1875 (which is complementary to his work of 1873), Malherbe writes: *"La Faille Eifélienne forme la limite méridionale du Bassin houiller dit de Seraing. Prenant son origine à la pointe de calcaire Eifélien près de Ramet, elle se poursuit en prolongement du contact du système quartzo-schisteux Eifélien avec une pente de 45°.* [...] *Il est plus que probable que la faille Eifélienne se poursuit vers l'est jusqu'au contact du calcaire au nord de Montzen"*.

In 1874, Gosselet proposes that the Silurian Condroz crest, a former paleogeographic high between the northern Namur and the southern Dinant basins probably had a "weak point" corresponding to an old fault of Silurian age. When the contractional stresses occurred, the weak point broke and a major fracture, here called *"grande faille"*, appeared in the area between Liège and Marquise (near Boulogne in the French Boulonnais). The fault actually separates the Silurian Condroz crest in the south from the northern Namur basin (Fig. 168). Gosselet (1974) adds that the northern part of the Namur basin, resting firmly on the Brabant unit, was relatively undeformed compared to the southern part which was markedly disrupted and displaced to the north (Fig. 168).

Taking inspiration from the views of F.-L. Cornet, Gosselet (1874) indicates that the Condroz crest moved up along the inclined plane of the "*grande faille*" and had therefore pushed away in his front some Carboniferous carbonated masses that had been exhumed from deeper level. These tectonic thrust sheets are lying between the red Devonian sandstones to the south and the "*Houiller*" rocks to the north.

Already in 1874, Gosselet makes an analogy between the French-Belgian and the English coal-basins. The Upper Carboniferous French-Belgian coal-basin must exist under the Channel and connect further to the west with the *"Houiller"* terrains of Bristol and of South Wales. Indeed, both the chemical composition and the structural framework are similar. The same analogy is established between the southern Dinant basin and the Paleozoic Devon and Cornwall basin as paleontological and petrographical similarities are identified.

In 1876 and 1877, Cornet & Briart indicate that the *"Houiller"* terrain of the southern part of the Hainaut coal-basin is overturned and plunges under Lower Devonian rocks. This anomalous "weird" superposition is regional in extent, as envisaged over 200 km from Liège in the east to the Pas-de-Calais in the west. It actually corresponds to a south-dipping fault producing a major reverse offset of the Devonian rocks to the north that is called Midi Fault in the Hainaut or Eifelian Fault in the province of Liège.

However, as a result of a revision of the stratigraphic subdivisions, the Eifelian Fault would no longer disrupt the *"système quartzo-schisteux eifelien – Terrain anthraxifère"* but the *"Terrain rhénan"* instead, making the term *"Faille Eifelienne"* inconsistent. Cornet & Briart (1877) therefore propose to name the discontinuity *"Faille du Midi"* for the entire strike. Geologists and engineers working in the Hainaut (Mons) collieries actually were used to call the discontinuity *"Faille du Midi"* as the industrially interesting *"Houiller"* coal strata of the Namur basin are delimited to the south (i.e. *"au Midi"*) by this fracture. Cornet & Briart (1877) also specify that both the offset and the parallelism of the fault and strata orientations cannot be constant everywhere along the trace.

In 1876, Cornet & Briart suggest that the northern (or Hainaut or Namur) basin is separated from the southern (or Dinant) basin by a ridge of Silurian rocks that outcrops along a narrow strip between Châtelet and Huy. The outcrop of the Ordovician-Silurian Condroz Inlier is attributed to the particular combination of the Midi Fault, the Anzin *"cran de retour"* and the Boussu Fault.

In 1877 and 1879, Macar suggests a continuation of the Eifelian Fault further eastward beyond Chênée within the Herve basin (Fig. 169). The reverse offset decreases from the southwest to the northeast, evolving from at least 1000 metres in the Yvoz-Angleur area to at most 100-200 metres within the Herve basin. The southern dip of the fault plane is also highly variable: from a gentle dip of 19° in the vicinity of Angleur to locally 60-70° elsewhere. The inclination is generally more than 45°.

Using field observations and borehole data, Faly (1878) follows precisely the trace of the Midi Fault for 15 km from the vicinity of Binche to the Sambre river. The Midi Fault is observed in an abandoned quarry near Binche where Devonian sandstones are anomalously present above Carboniferous limestones. The contact dips gently to the south.

In 1878, Gosselet publishes a map of the eastern termination of the Midi Fault in the area southwest of Liège, between Hermalle and Angleur (Fig. 170). He indicates that the 200 km long Midi Fault, separating the Namur and the Dinant basins, is disrupted between Hermalle and Angleur by three main transverse, NNW-SSE striking faults (see for example on Fig. 170 the dextral Kinkempois transverse fault that displaces the Midi Fault for 2 km). Gosselet indicates that a transverse fracture, the Vaux Fault (C on Fig. 170), located in the Vesdre valley, constitutes the eastern termination of the *"crête du Condroz"* as well as that of the Midi Fault. He believes, therefore, that research into the continuation of the Midi Fault to the east of the transverse Vaux Fault within the Herve Massif is useless.

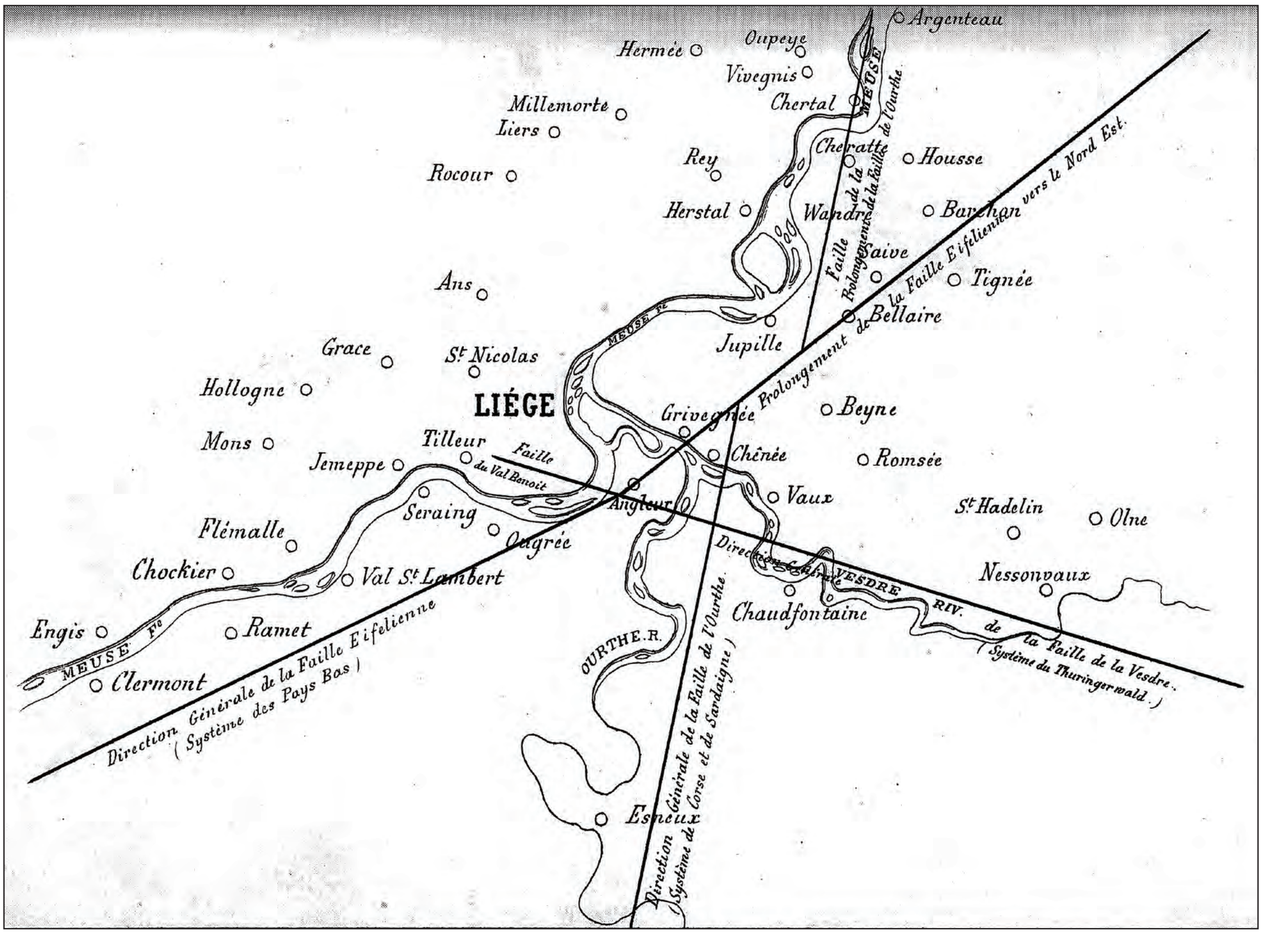


Fig. 169. Structural map of the main faults in the vicinity of Liège (Macar, 1877).

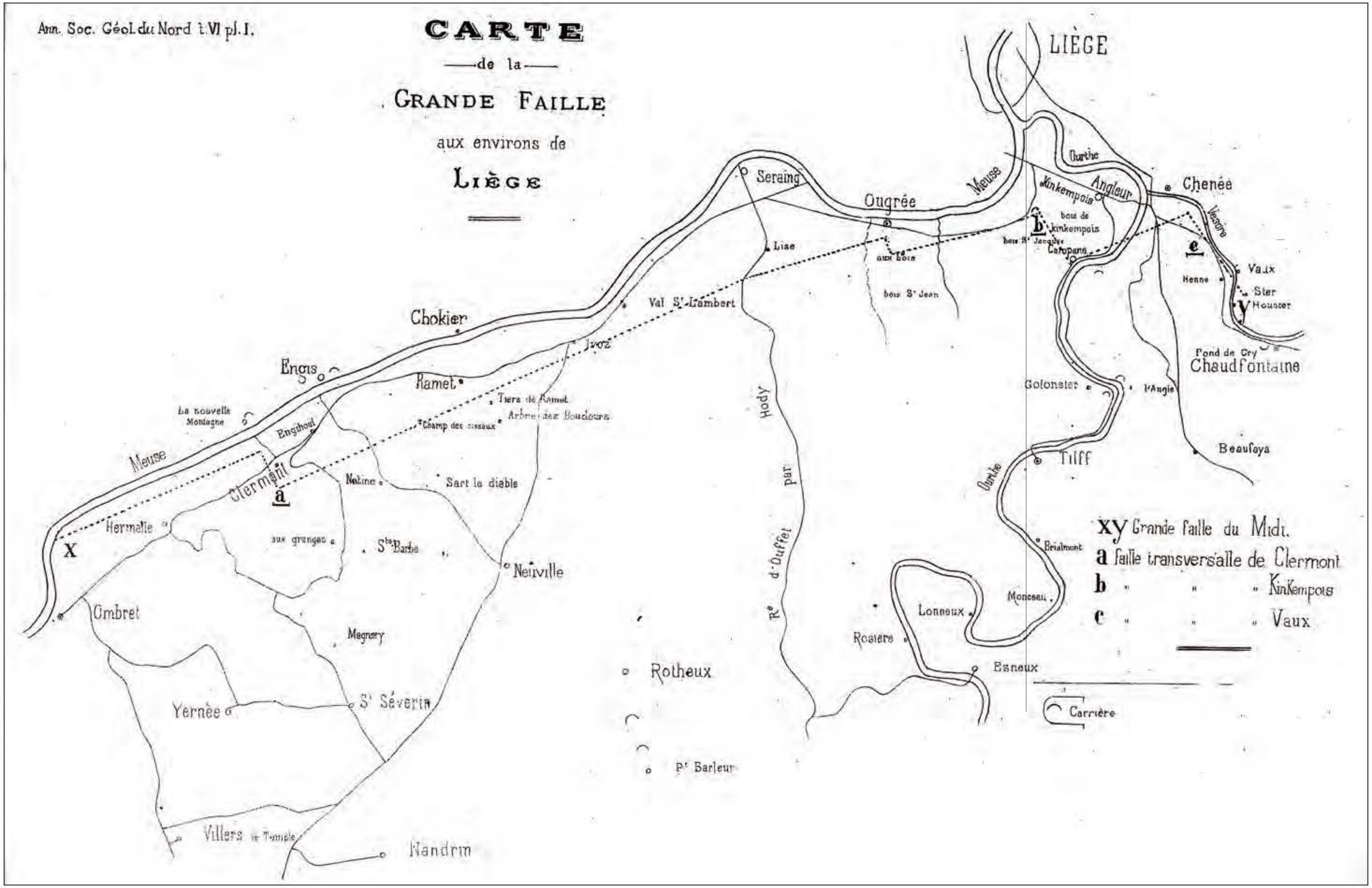


Fig. 170. Cartography of the "*Grande Faille*" in the vicinity of Liège (Gosselet, 1878).

Gosselet (1879, 1880) indicates that the *"Grande Faille"* has a changing southern dip. Strata in the northern footwall block are systematically overturned to the north and plunge under the Lower Devonian terrain (i.e. of "Gedinnian" or Lochkovian age) of the southern hanging wall (Fig. 171). Gosselet also identifies a thrust tectonic stack or *"lambeau de poussée"* isolated by the *"Grande Faille"* and the *"Faille limite"* (Fig. 171). Actually, when the Lower Devonian glided upward on the *"Grande Faille"*, some rock masses were removed from the Namur basin and pushed up along the thrust plane to the north. He also indicates on Fig. 171 the *"Faille de retour"* or *"Cran de retour"*, which is located within the *"Houiller"* basin and separates the southern folded and overturned terrain from the northern non-deformed part of the Namur basin.

In 1879, Dewalque proposes to extend the Eifelian Fault to the east where the fracture would mark the discontinuity between the Liège and Herve "massifs". In that area, the fault would be less significant as the two blocks on either side of the fault display similar Upper Carboniferous (*"Houiller"*) rocks. Multiple arguments for this continuation are provided: (1) the difficulty of connecting the Coal Measures between the Liège and Herve basins, (2) the field observation (at surface) of faults, and especially (3) the analogy of the geological structure between the Condroz and the Herve "Massifs". The Eifelian Fault would even continue further to the east near Aachen in Germany, separating the Upper Carboniferous Rolduc-Worm (or Wurm) and Eschweiler-Inde basins in the NW and SE respectively. Dewalque (1879) therefore considers that the Condroz, Herve and Eischweiler basins belong to the same major regional unit (to the south), characterized by regular, symmetrical folds, separated by the Eifelian Fault from the Liège and Rolduc basins (to the north), of which the southern border shows irregular inclined to overturned folds.

In his famous work of 1888, Gosselet retains his ideas developed in 1879 and in 1880. He suggests that the western termination of the Midi Fault, in the Boulonnais area, has not been identified and that the eastern termination is formed by the transverse Vaux Fault located in the Vesdre valley. The *"Grande Faille"* would not continue further eastwards. Gosselet (1888) adds that it is wrong to say that the *"Grande Faille"* strikes from Boulonnais to the city of Liège. Indeed, along a 65 km long segment of the trace between Sart-Saint-Eustache (or Sart-Eustache, near Châtelet) and Engis, the Midi Fault is replaced by a significant fold of which the axis is formed by the Silurian Condroz ridge (Fig. 172). No real fault discontinuity is found along this segment while laterally, i.e. to the west of Sart-Saint-Eustache and to the east of Engis, the anticline "breaks off" and "associates" with the reverse south-dipping Midi Fault.

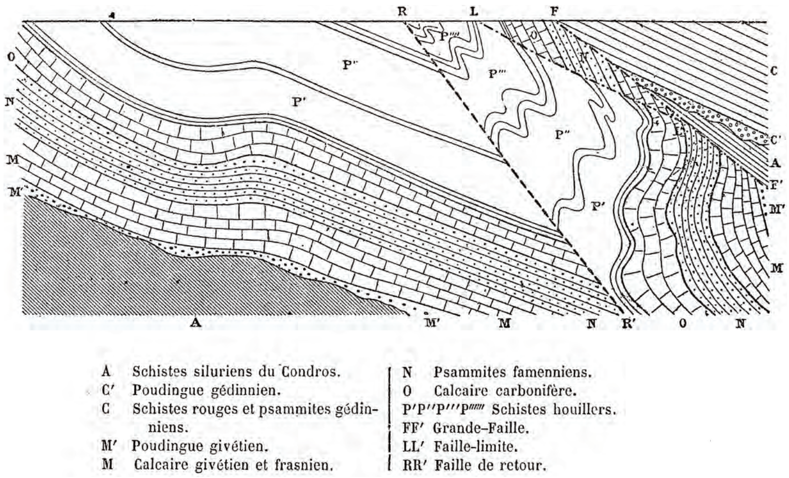


Fig. 171. Theoretical N-S cross-section through the French-Belgian *"Houiller"* coal-basin (Gosselet, 1879).

Fig. 172. Mapping of the Midi and Eifelian faults after Gosselet in 1888. The western Midi segment is separated from the eastern Eifelian segment by the Silurian Condroz ridge where no fault contact has been observed.

In 1891 and 1892, De Dorlodot indicates that a small western part of the narrow Silurian Sambre-et-Meuse strip is delimited to the north by the Midi Fault. The hanging wall of the Midi Fault is therefore composed of Silurian rocks from the western termination of the Condroz ridge (near Couillet) in the west to the eastern termination of the Midi Fault (near Fosse) in the east. Further to the west, beyond the western termination of the Sambre-et-Meuse Strip, the Silurian of the hanging wall of the Midi Fault is replaced by Lochkovian rocks of the northern border of the Dinant basin.

The views of De Dorlodot include, therefore, a distinction between the Midi and Eifelian faults and the continuation of the Midi Fault on a partial segment of the northern border of the Condroz Inlier. In other words, de Dorlodot suggests the non-continuity of the *"grande faille"* and is therefore opposed to the structural concepts of a continuous Midi-Eifelian Fault as supported by Gosselet, Dewalque and later (see below) Fourmarier.

From 1897 to 1904, geological cartography of the Midi Fault, the Silurian Condroz narrow strip and the Eifelian Fault at the 1:40 000 scale was undertaken by several Belgian geologists. The Midi Fault is identified over distance of 69 km, the Silurian of the Condroz Inlier for 63 km and the Eifelian Fault for 15 km, thus totalling a length of 147 km from the French-Belgian border (Quiévrechain in France) to Angleur in the vicinity of Liège (Fig. 173). Twelve geological maps cover the Belgian part of the Midi-Condroz-Eifelian segments. From the west to the east respectively, the maps of: Quiévrain – Saint-Ghislain (N°150; Rutot & Cornet, 1902a), Mons – Givry (N°151; Rutot & Cornet, 1902b), Binche – Morlanwelz (N°152; Briart, 1899), Fontaine-l'Évêque – Charleroi (N°153; Briart & Bayet, 1904), Gozée – Nalinnes (N°164; Bayet, 1900), Tamines – Fosse (N°154; Stainier et al., 1904), Malonne – Naninne (N°155; Stainier et al., 1901a), Gesves – Ohey (N°156; Stainier et al., 1901b), Andenne – Couthuin (N°145; Stainier et al., 1901c), Huy – Nandrin (N°146; Dewalque et al., 1898), Jehay-Bodegnée – Saint-Georges (N°133; Stainier et al., 1899) and Seraing – Chênée (N°134; Forir & Mourlon, 1897).

The state of knowledge relative to the Midi, Condroz and Eifelian structural segments at the beginning of the 20th century as it appears on the old geological maps cited above is as follows (Fig. 173):

- To the west, close to Quiévrechain (in France), the fracture, called here *"Grande faille du Midi"*, has a main E-W strike. It is justified by an anomalous stratigraphic contact between Lower Devonian (Pragian) siliciclastic rocks to the south (i.e. "Cb2 – *Coblencien*") and Upper Carboniferous rocks to the north ("H1b & H2 – *Houiller*"). The fault strikes through Elouges, Dour, Eugies, Genly and Harmignies and disappears there (to the east of Harmignies) under the tabular Cretaceous cover which is not disturbed by the fracture.
- The fault re-appears in Waudrez, in the area west of Binche, and runs towards Anderlues where its strike bends to the southeast. The fracture, here called *"Faille du Midi"*, brings into contact mainly Lower Devonian sandstones and shales (either Lochkovian, Gdb – *Gedinnien* or Pragian, Cb1 – *Coblencien*) to the south with Carboniferous rocks (either Visean, V2c or Namurian, H1b – *Houiller inférieur*) to the north.
- The fault continues towards the south-east and then acquires a northeast strike directed towards Châtelet. The southward bend of the Midi Fault limits the so-called *"Anse de Jamioulx"* where again Visean, Namurian and Westphalian rocks are in anomalous contact with Lochkovian and Pragian rocks to the south.
- Further eastwards, the Midi Fault returns to an E-W direction in the area south of Châtelet. There, the lithostratigraphy of the Midi Fault country rocks shows a major new characteristic as the hanging wall is no longer made up of Lower Devonian rocks but of Silurian instead, indicating the beginning of the Lower Paleozoic Condroz Inlier. The Midi Fault goes along the northern limit of the Silurian ridge until the locality of Presles where it disappears. The footwall block also shows changes in lithology and stratigraphy. Upper Carboniferous rocks are no longer present immediately north of the fault but are replaced by Givetian, Frasnian, Famennian and Visean terrains.

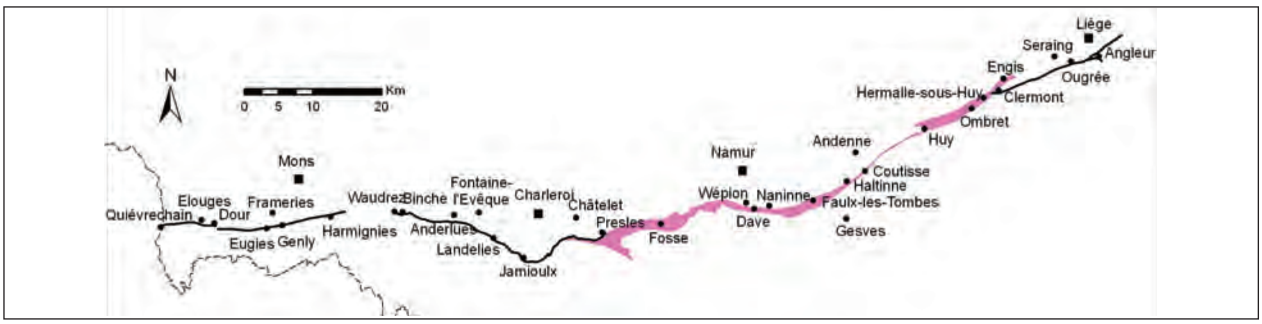


Fig. 173. Simplified map of the Midi-Eifelian Fault and the Silurian of the Condroz Inlier according to Belgian geological maps at 1:40 000 scale published between 1897 and 1904 (see the text above for a description and references).

- After running over 69 km from Quiévrechain to Presles, the Midi Fault can not be identified further eastwards. The narrow Silurian Condroz (or Sambre-et-Meuse) Inlier seems to represent the continuation of the Midi Fault. The inlier strikes through many villages such as Fosse, Wépion, Dave, Faulx-les-Tombes, Coutisse, Huy, Clermont and Engis, totalling therefore a length of 63 km. The Condroz strip is quite narrow, being about 700 m wide in the Wépion region and at most 120 m wide locally in the vicinity of Coutisse.

- Near Clermont, the fault, here called *“Faille Eifelienne”*, appears. It has a mainly ENE strike direction running from south of Ramet, south of Ougrée and continuing until a point west of Angleur. The western segment of the fault shows an anomalous contact between the Lower Devonian to the south (Cb1 & Cb2 – *Coblencien* or Pragian) and Frasnian – Famennian – Visean rocks to the north (F2c, Fa2b & V2a respectively). The eastern segment shows Pragian sandstones (Cb3) in the hanging wall and Namurian-Westphalian micaceous sandstones and shales (H2) in the northern footwall. The Eifelian Fault is identified over a distance of 15 km.

- In the vicinity of Angleur, the Eifelian Fault subdivides into 2 fault segments: a first NE-striking fault in the Upper Carboniferous Herve “Massif” and a second ENE- then S-striking fault within the Ourthe valley. The eastern termination and/or continuation of the Eifelian Fault has been the subject of much debate. We refer readers to the Aguesses-Asse Fault and the Tunnel Fault, both described in Cambier & Dejonghe (2010) for further details. Hance et al. (1999) suggest considering the Tunnel Fault, to the southeast of Liège, as the true continuation of the Midi-Eifelian Fault.

According to the divergent opinions regarding the eastern termination of the Eifelian Fault, Forir (1899) summarizes the views of his colleagues in 3 groups (Fig. 174):

(1) According to both G. Dewalque and the *“Carte générale des mines”*, the Eifelian Fault continues beyond Angleur (Kinkempois) within the *“Houiller”* terrain of the Herve basin with a general ENE strike (A on Fig. 174);

(2) According to J. Gosselet, the Eifelian Fault (B on Fig. 174) is disrupted by a transverse, NNW-SSE-striking secondary fault (C on Fig. 174) and is therefore translated to the south. There, the Eifelian Fault continues within the Ourthe valley and butts out into another NNW-SSE-trending secondary (Vesdre) fault (D on Fig. 174);

(3) According to H. Forir, the eastern termination of the Eifelian Fault bends to the southeast (E on Fig. 174) and is interrupted by a secondary (Ourthe) fault (H on Fig. 174). The Eifelian Fault is displaced to the east in an apparent sinistral strike-slip movement and continues with a SE strike in the Vesdre valley. Then, near Chaudfontaine, the Eifelian Fault returns to an ENE direction where it separates the *“Houiller”* Herve basin in the north from the Devonian-Lower Carboniferous terrain of the Condroz.

Also in 1899, Forir tackles the problematic Aguesses-Asse Fault (see the complete description in Cambier & Dejonghe, 2010). He identifies an ENE-striking and 50° south-dipping fault (G on Fig. 174) in the Aguesses colliery that he considers to be too different to the Eifelian Fault to be its continuation. Forir (1899) already understands the significance of the Aguesses Fault depending on how it is conceived, either as a major thrust (e.g. Dewalque) forming the eastern continuation of the Eifelian Fault and separating the Liège (or Namur) and the Dinant (including the Herve basin) units, or as a secondary minor fault (e.g. Forir) separating the Liège and Herve basins both located north of the Eifelian Fault within the Namur basin.

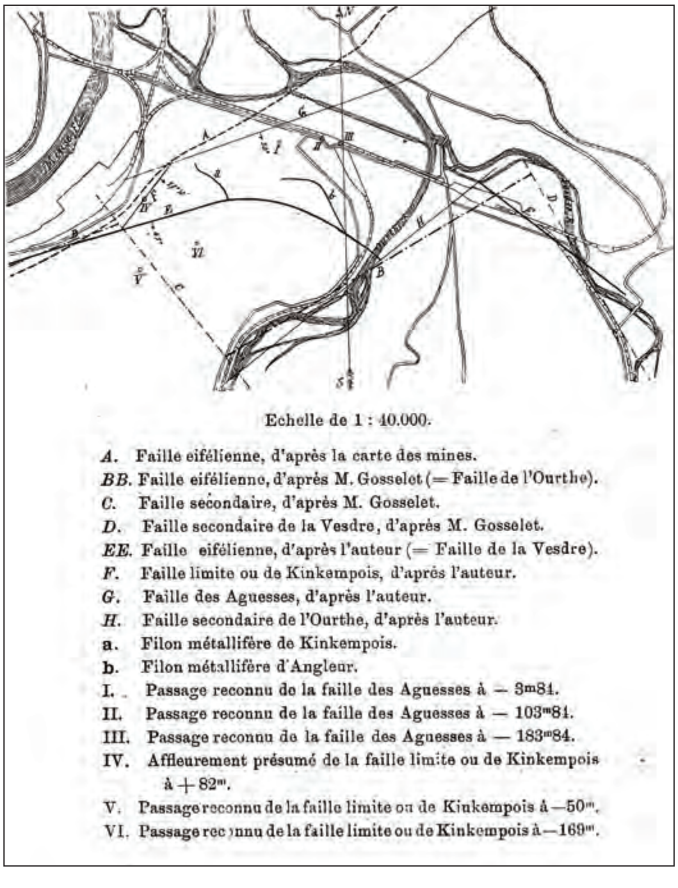


Fig. 174. Map showing the hypotheses regarding the eastern continuation of the Eifelian Fault according to various geologists (Forir, 1899).

In 1901, de Dorlodot envisages the "Theux basin" as an "eyelet" of a large autochthonous zone probably resulting from the erosion of a previously overlying thrust nappe. He therefore interprets the "Theux basin" as a tectonic window within the main northward translated thrust nappe. This structural description constitutes one of the first contributions to the development of the theories of nappe transport and thin-skinned tectonics applied to the Rhenohercynian fold-and-thrust belt.

De Dorlodot (1901) also suggests that the Theux Fault (see Cambier & Dejonghe, 2010) coincides and joins at depth with the Eifelian Fault. De Dorlodot does not specifically use the term "tectonic window" that will be employed later by Fourmarier in 1904 and 1905 (i.e. "*Fenêtre de Theux*").

Fourmarier (1904, 1905, 1906a) proposes a particular relationship between the south-dipping Eifelian Fault and the north-dipping Theux Fault. Actually, both fractures would join at depth therefore delimiting an upper allochthonous unit defined as the Condroz Nappe (Fig. 175).

Indeed, Fourmarier (1904) believes that all the tectonic stacks (the *"lambeaux de poussée"*) belong to a same large thrust nappe (*"nappe de charriage"*). Consequently, at the regional scale, he proposes that the entire Devonian-Carboniferous terrain of the Vesdre valley constitutes a huge thrust nappe that moved and glided over the *"Houiller"* terrain at depth.

Fourmarier (1904) therefore envisages a connection at depth between the Upper Carboniferous of the Herve basin and that of the Theux area under the Devonian-Carboniferous thrust nappe. The Theux Massif, called *"fenêtre"* in his paper of 1904, is rightly considered as a tectonic window. In other words, the Eifelian-Theux Fault is directly implied in the major regional northward thrust of the Dinant Synclinorium over the Namur basin of which the Theux Window (or "*Massif de Theux*" on Fig. 175) is intimately related.

In 1904 and 1908a, Fourmarier studies the eastern continuation of the Eifelian Fault. From Kinkempois, the Eifelian Fault bends to the east then to the south. The discontinuity is called the Streupas Fault here, and continues with a SW-NE strike along the Ourthe Fault. Fig. 176 presents the complex structural framework in the area of the eastern continuation of the Eifelian Fault beyond Liège.

Actually, to the west of Angleur, the direction of the simple, easily recognizable Eifelian Fault, is sub-parallel to the Meuse valley while to the east of Angleur, the trace becomes more complex and seems to subdivide into several branches that bound tectonic stacks thrust over each other (Fourmarier, 1908a; Fig. 177; *"ces failles limitent une série de lambeaux ou mieux d'écailles de poussée, refoulées les unes sur les autres"*). These fractures are low-angle south-dipping reverse faults related to the overturning and breaking of folds.

Just like Forir in 1899, Fourmarier (1904, 1908a) did not think of the eastern continuation of the Eifelian Fault as a simple and single discontinuity but as multiple fractures delimiting multiple tectonic stacks.

In 1906(a,b), Fourmarier presents an important consideration, the connection between the Midi and the Eifelian faults. The junction would occur through the narrow Silurian Condroz Inlier that limits the Namur and the Dinant basins. Fourmarier actually lost the trace of the Eifelian Fault to the west of Engihoul near Clermont within the Silurian terrain of the Condroz anticline. However, Fourmarier (1906a) believes that the Eifelian discontinuity continues within the Sambre-et-Meuse Inlier, retaining its "major significance" and bringing into contact Silurian rocks on each side of the fault plane. The whole Condroz anticline is longitudinally cross-cut by the low-angle south-dipping Eifelian Fault which connects with the Midi Fault to the south of the Hainaut coal-basin.

The hypothesis of a link between the Midi and the Eifelian faults has the advantage of gathering together all the structural observations and thrust phenomena (faults, thrust sheets …) between the Namur and the Dinant basins together in one model that is the overthrust of the Dinant syncline over the Namur syncline. In this way, Fourmarier (1906a,b) establishes an analogy between the folded Ardennes domain and the younger high mountain ranges like the Alps where huge thrust nappes are also observed. The Ardennes area is distinguished from other younger folded belts by the erosion that has removed parts of the thrusts and obliterated their observations.

The passage of a major discontinuity within the Condroz anticline is justified by the major facies differences between the Paleozoic rocks of the Namur and the Dinant basins (e.g. the absence of Lower Devonian rocks within the Namur basin and the thick sequence of such to the south) separated from each other by the Condroz Inlier.

In 1907, de Dorlodot publishes a structural map of the western Condroz Inlier between Fosse and Wépion where a major fracture, called the Maulenne Fault is considered to have a connective function between the Midi and Eifelian faults and with which they share similarities. The Maulenne Fault is therefore a segment of the overthrust of the Condroz anticline over the Namur basin. De Dorlodot (1907) actually specifies that the link between the Midi and the Eifelian faults within the Ordovician-Silurian Condroz anticline is composed of three faults that are the Maulenne, the Ormont and the Boussale faults (see Cambier & Dejonghe, 2010). The "break thrust" type Maulenne Fault (or *"faille de rupture"*) has a low-angle dip to the south and a reverse displacement probably exceeding 2 km to the south of Malonne that strongly decreases laterally.

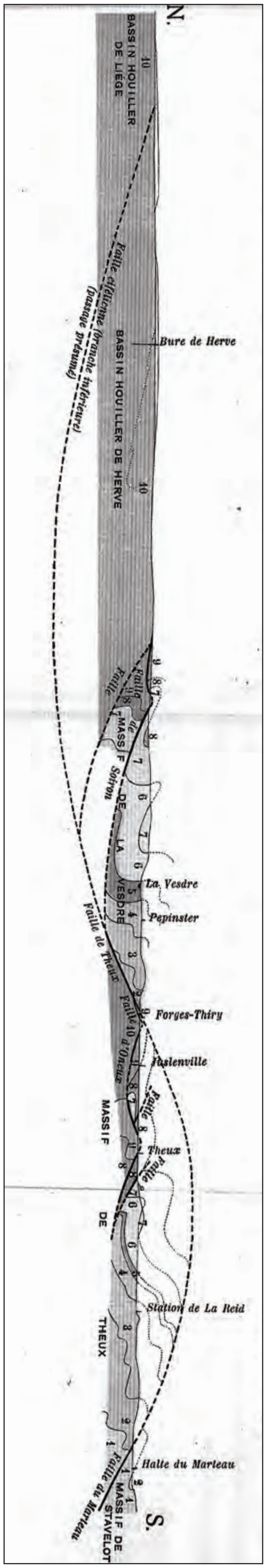


Fig. 175. N-S cross-section through the undulating Eifelian-Theux Thrust and the Theux Window (Fourmarier, 1906a).

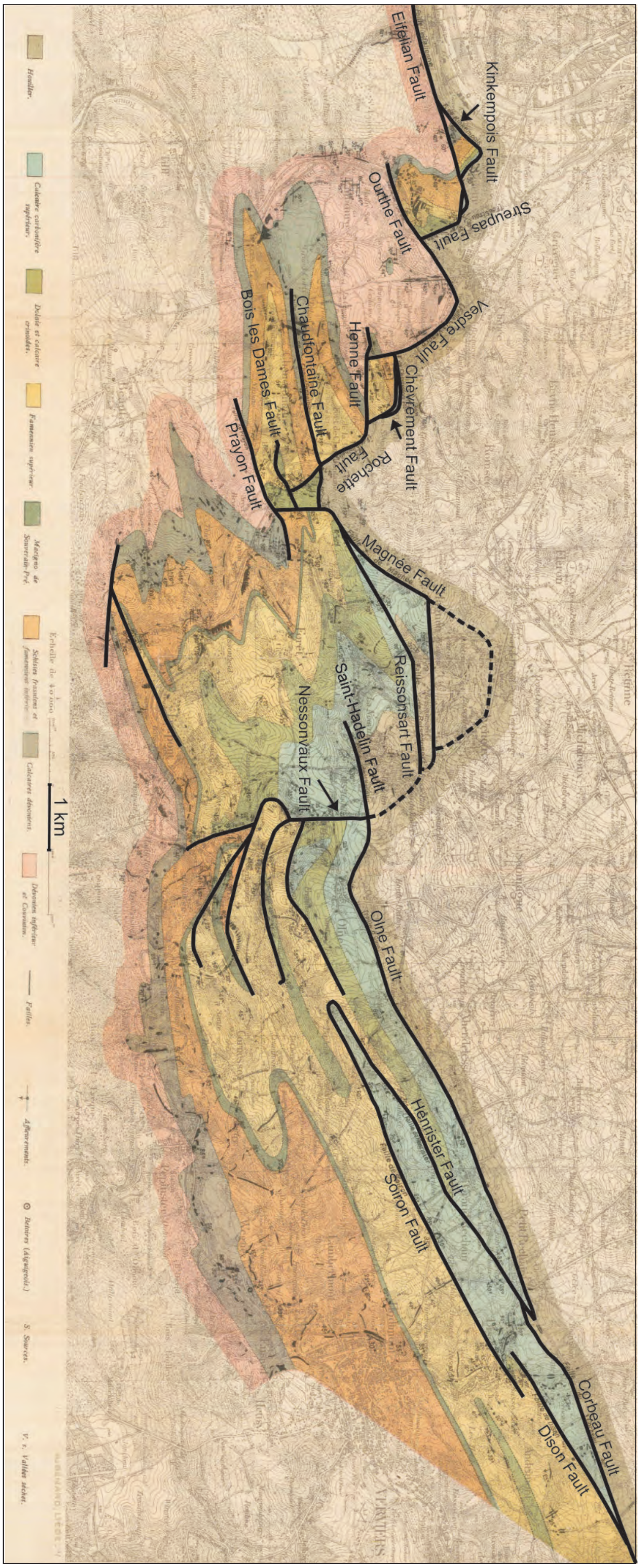


Fig. 176. Geological map of the eastern continuation of the Eifelian Fault (Fourmarier, 1904).

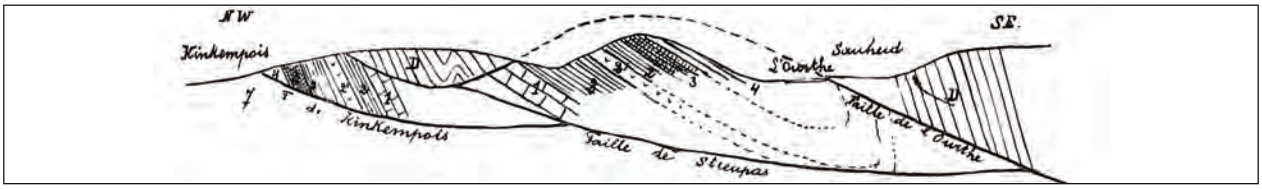


Fig. 177. NW-SE cross-section through the eastern continuation of the Eifelian Fault in the vicinity of Angleur (Fourmarier, 1908a). The Eifelian Fault splits into several gently south-dipping thrust faults enabling the superposition of multiple thrust sheets. 1 = Frasnian limestone. 2 = Famenne shales (Fa1ab). 2' = sandstones interbedded within the shales (2). 3, 4 & 5 = Micaceous sandstones of Famennian age (Fa1c, Fa2b & Fa2c respectively). 6 = Lower Carboniferous dolostones (T). 7 = "*Houiller*" (H1).

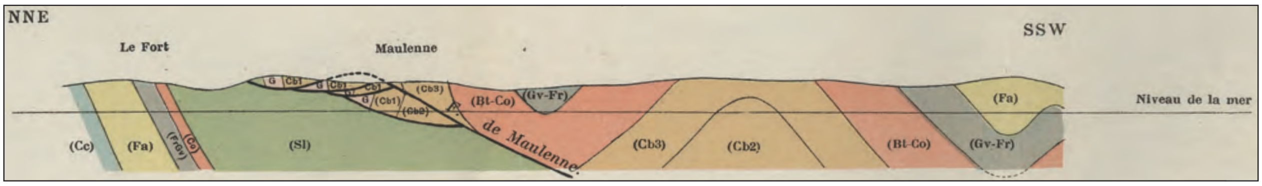


Fig. 178. N-S section of the Silurian Condroz Inlier and the Maulenne Fault (i.e. the Midi Fault) at the meridian line of Floreffe (Foumarier, 1908b). Sl = Silurian. G = "*Gedinnien*" (i.e. Lochkovian). Cb1, Cb2 & Cb3 = Lower, Middle and Upper "*Coblencien*" (i.e. Lower Devonian). Bt-Co = "*Burnotien*" and "*Couvinien*" (i.e. Eifelian). Gv-Fr = Givetian & Frasnian. Fa = Famennian. Cc = Carboniferous limestones. H = "*Houiller*" (i.e. Upper Carboniferous)

In 1908b, Fourmarier believes that the region between Fosse and Wépion is characterized by many fractures of which the Maulenne and the Ormont faults are parts. He concludes that this complex network of faults is related to the overthrust of the Dinant basin over the Namur basin. The Floreffe region is therefore marked by the passage of the *"grande faille"* that strikes longitudinally along the whole Silurian Condroz ridge and of which the Maulenne Fault is a segment. The section in Fig. 178 across the Silurian Condroz anticline shows, in the vicinity of Floreffe, the folded, 2000 m offset-Maulenne Fault that thrusts the Lower Devonian border of the Dinant basin over the Silurian terrain. Several tectonic stacks compose the footwall.

In 1910, Stainier publishes some geometrical aspects of the Eifelian Fault that miners have observed both in the Bois d'Avroy and in the Ougrée collieries (in the south-western area of Liège). In the Bois d'Avroy colliery, the Eifelian Fault has a southern dip of about 35-45° and a strike direction of about N35°E. In the Ougrée colliery, dips of between 27° and 39° were reported. Stainier (1910) actually observes a reduction in the dip with depth.

According to Lodin (1911), the northward push (i.e. the Variscan compression) has resulted in the formation of several low-angle faults showing a systematic upward movement of the hanging wall over the footwall. The sum of the reverse displacement of each fault is estimated at about 5 km in the Chaleroi area and about 4 km in the vicinity of Mons. Just like Marcel Bertrand, Lodin considers the *"cran de retour"* as the continuation of the Midi Fault of the French Pas-de-Calais department. A large northerly inclined anticline, which is the origin and the first stage in the establishment of the thrust nappe, broke and formed the thrust front that moved the "*Houiller*" terrain to the north over Upper Silurian schists.

In 1913, Fourmarier summarizes various geometrical observations. Near Eugies (SW of Mons), the Midi Fault has a southerly dip of about 25°; near Harmignies (SE of Mons), a dip of about 23-24° and near Waudrez (W of Binche), a dip more than 30°. In the vicinity immediately east of Binche, the Midi Fault has been recognized at 335 m below the surface in the Mahy-Faux borehole but at 611 m depth near Montifaux a short distance to the south. The inclination of the fault plane is estimated at between 7 and 8° between those two points meaning that the dip, of about 25° at the surface, is closer to horizontal at depth.

Fourmarier (1913) also considers the Boussu, the Belle-Victoire and the Fontaine-l'Évêque – Landelies thrust sheets (Fig. 179) as initially belonging to the same single large tectonic stack. The south-dipping fault that limits this major tectonic stack at its base is not planar but displays undulations. Erosion has allowed an apparent parcelling of the large thrust mass into three smaller thrust sheets. Also as a result of the erosion, Fourmarier (1913) believes that the outcropping *"Houiller"* terrain of the Hainaut area was initially covered by the Dinant nappe. He estimates the thrust displacement of the nappe to be about 15-20 km although it was probably greater than this when the thrust happened.

Fourmarier (1913) proposes a comparison between the structures of the Upper Carboniferous Hainaut and Liège coal-basins. The Hainaut basin is formed by numerous superimposed thrust sheets. Concerning the Eifelian Fault, which is of similar significance to the Midi Fault in the Hainaut area, Fourmarier believes that the Liège basin also has a structure produced by the superposition of tectonic wedges. Fig. 179 summarizes the views of Fourmarier in 1913. It also gives the possible extension of the Upper Carboniferous below the Midi Fault and the minimal thrust of the Dinant nappe of the Namur basin.

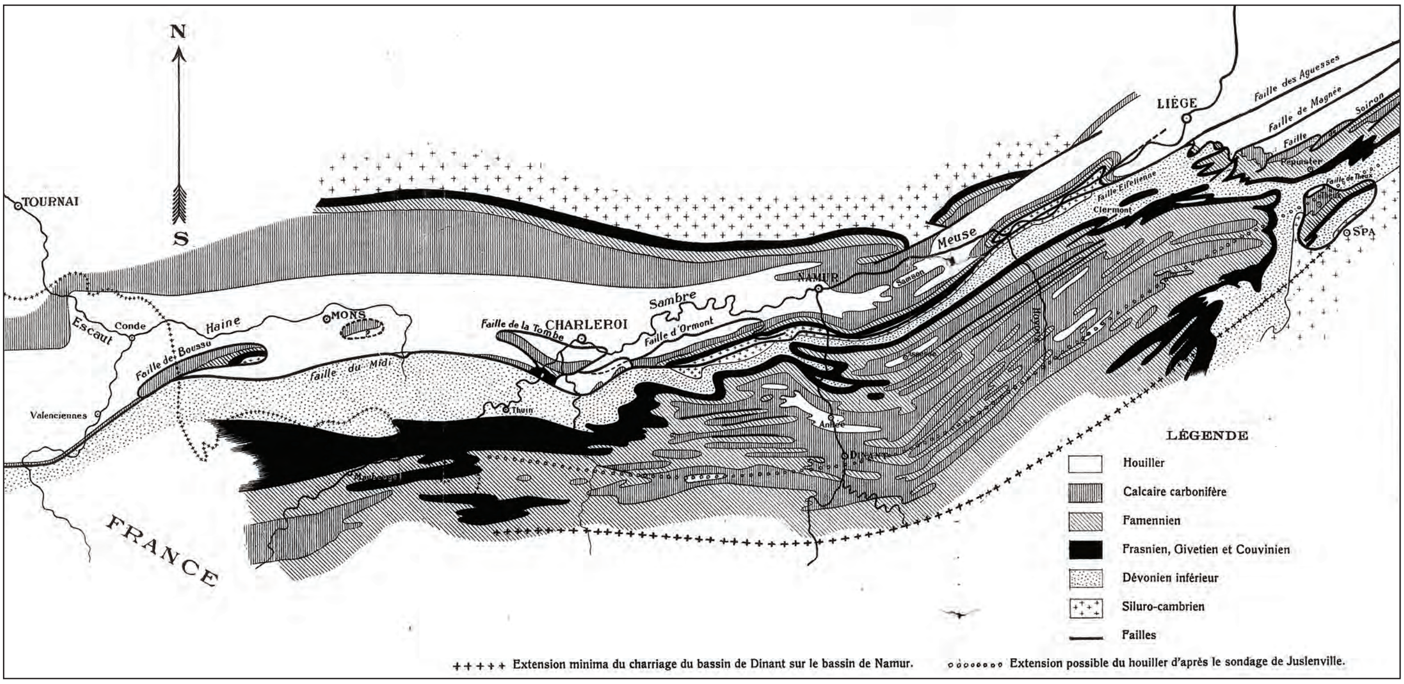


Fig. 179. Geological map of the Namur basin and of the northern part of the Dinant basin (Fourmarier, 1913).

In 1914, Fourmarier studies the complex structure of the Silurian Sambre-et-Meuse Strip between Presles and Vitrival (SE of Charleroi). The western part of the Silurian inlier shows many fractures delimiting tectonic stacks thrust over each other but for which the offsets are difficult to estimate. The Lochkovian Bois-de-Presles thrust sheet, located in the middle part of the Ordovician-Silurian Condroz Inlier, proves that the Midi Fault continues within the inlier and subdivides there into many branches that separate several wedges.

In 1919b, Fourmarier proposes a revision of the Kinkempois thrust stack (*“lambeau de poussée de Kinkempois*”) structure. In the vicinity of Angleur, Fourmarier suggests that the Eifelian Fault has a particular S-shaped trace punctuated by two tectonic stacks: the stack of Kinkempois to the west and of Streupas to the east. Fig. 180 shows that the Kinkempois “massif” is made up of Middle and Upper Devonian rocks separated from the Lower Devonian of the Dinant basin to the south by the Eifelian Fault and separated from the Upper Carb oniferous of the Liège basin to the north by the Kinkempois Fault.

In 1920b, Stainier makes a structural analogy between the Silurian Condroz Inlier and the Mendips Hills in southern England (to the south of Bristol). Stainier indicates the similarity between the southern border of the Namur basin and the southern border of the Bristol basin on one hand and between the Condroz strip and the Mendip Hills on the other. Stainier believes that the Carboniferous Mendip Hills are separated from the *“Houiller”* Bristol basin to the north by a fault (Fig. 181) having the same thrust role as the Ormont Fault in Belgium. To the south, the Devon syncline (to the south of the “Quantock Hills” on Fig. 181) is separated from the Mendips to the north by a major thrust-type fault comparable to the Midi Fault. Just like in Belgium, no Lower Devonian rocks are observed to the north of the major thrust and a Silurian terrain constitutes the core of the ridge.

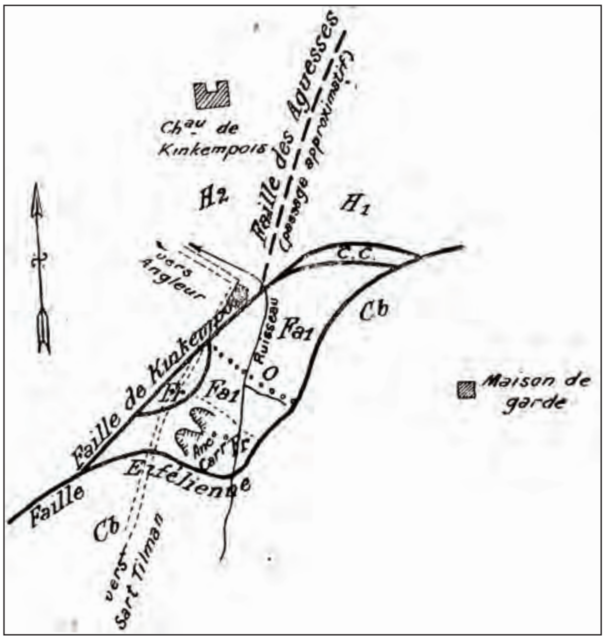


Fig. 180. Simplified geological map of the Kinkempois thrust sheet (Fourmarier, 1919b). H2 = Westphalian. H1 = Namurian. C.C. = Carboniferous limestones. Fa1 = Lower Famennian. O = oolitic ironstones. Fr = Frasnian. Cb = “*Coblencien*” (i.e. Pragian).

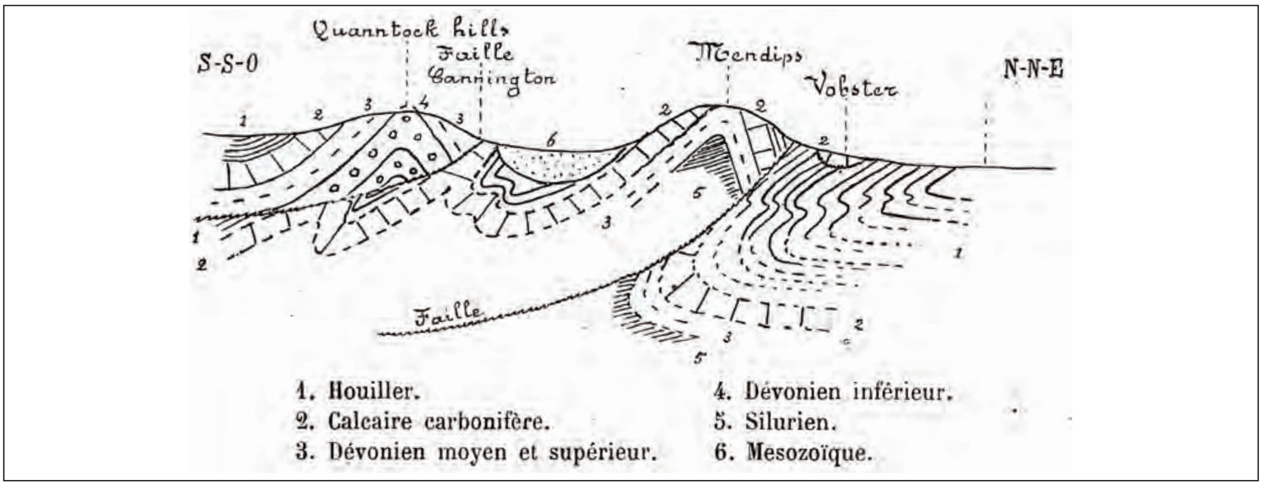


Fig. 181. SSW-NNE section through the Mendips region (Bristol) in southern England (Stainier, 1920b).

According to these observations in southern England and to this structural analogy between England and Belgium, Stainier (1920b) proposes new ideas on the Silurian Condroz crest. Firstly, he confirms the intense faulting within the Silurian inlier ("*la crête du Condroz se montre comme une longue zone profondément déchiquetée par des failles de refoulement qui établissent un trait d'union entre la faille du Midi et la faille eifelienne*"). Secondly, he specifies that the southern border of the Namur (or Sambre-et-Meuse) syncline is not known as this part of the syncline plunges under the plane of the Midi Fault.

In order to avoid confusion between the terms of Eifelian and Midi faults, Fourmarier (1923) proposes calling the major Belgian discontinuity "*charriage du Condroz*" or Condroz Thrust as it runs alongside one of the most striking structural feature of the Belgian geology that is the Condroz Inlier.

On the basis of facies comparisons between the Theux Window and the associated surrounding thrust nappe (i.e. the eastern border of the Dinant Synclinorium), Fourmarier (1923) indicates an offset of about 10 to 12 kilometres along the Condroz Thrust. This displacement applies only to the thrust of the Vesdre Massif over the Theux Massif, i.e. to the Theux Fault (see Cambier & Dejonghe, 2010). After taking into consideration the eroded masses and the secondary thrusts that are also involved in the transport of the Dinant basin, the total displacement can be estimated at 15-20 km. The comparison of Lower Devonian facies of the Theux and the Dinant units enables Fourmarier (1923) to propose an offset of "much more" than 30 km. This length is the distance that initially separated the place where the Lower Devonian of Theux formed and the northernmost limit attained by these rocks during the thrust of the Dinant basin.

Fourmarier (1923) also suggests that the Paleozoic domain of Belgium is formed by the superposition of two large thrust nappes. The upper nappe is made up of the Dinant syncline and is limited to the north by the Condroz Thrust. The lower nappe only crops out in the Theux Window and is the result of the "Juslenville Thrust" (Fourmarier, 1923; Fig. 182).

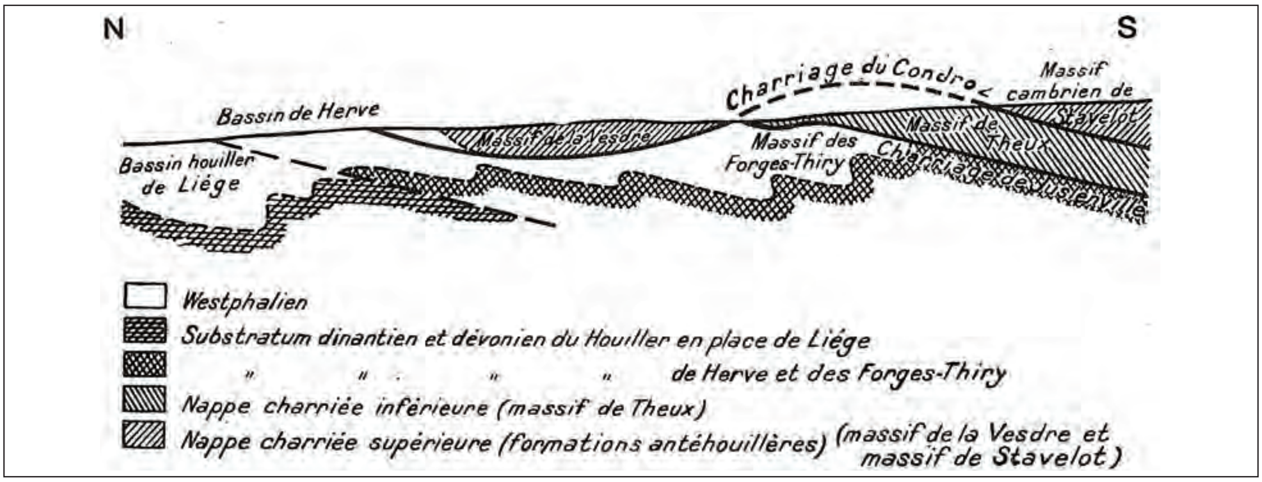


Fig. 182. N-S cross-section through the Theux Window (Fourmarier, 1923).

The total displacement produced by the thrusts in Belgium (Condroz, Juslenville, etc.) is more than 50 kilometres (Fourmarier, 1923). Indeed, the offset of the upper nappe along the Midi Fault – Condroz Thrust over the lower nappe is about 15 km and the offset of the lower nappe along the Juslenville Thrust over the underlying terrains is about 30 to 40 km. Moreover, these underlying terrains are quite unknown and may not necessarily constitute a true autochthonous domain. It is probable, therefore, that other deeper, unknown thrusts may exist. Foumarier (1923) adds that the Variscan thrusts of Belgium are exactly comparable to these of the "true" and young mountain ranges where thrust phenomena are more obvious because erosion does not yet levelled them.

Near Engis (between Huy and Liège), Fourmarier (1925) identifies the Eifelian Fault within a 30 metre thick crushed zone. The fracture has a notable dip as it is almost subvertical. Fourmarier explains this kind of inclination for a thrust fault by the presence of a small tectonic stack. Fourmarier (1925) also suggests that the Condroz Thrust has a similar structural architecture in both the Hainaut and Liège areas. In the Hainaut, the Midi Fault is punctuated and marked by numerous well-developed thrust sheets (*"lambeaux de poussée et lames de charriages nombreux et développés"*), whilst in the Liège basin, the aspect of the Eifelian discontinuity seems simpler. Actually, the Eifelian Fault is equally punctuated by a series of thrust stacks, but the erosion, more intense in the vicinity of Liège, has left only few traces of these tectonic stacks.

In 1924, Stainier reports observations of the Midi Fault near Elouges (between Mons and Valenciennes). The fault is identified in a borehole at between 75 and 79 metres depth. This thick faulted zone is composed of dark shales (a mix of Devonian and Upper Carboniferous rocks). The dip of the fault is about 10° to the south.

Michot (1932) presents a structural analyse of the Silurian Sambre-et-Meuse Strip between Huy and Ombret. Three tectonic units comprise the Silurian Inlier (Fig. 183): the "northern massif" (to the north of the Tihange Fault), the "median massif" (between the Tihange and the Huy faults) and the "southern massif" (to the south of the Huy Fault). The Tihange Fault is a discontinuity between the Lower and the Upper Wenlockian (Late Early Silurian) while the Huy Fault is justified by an anomalous contact between the Ordovician to the south and the Upper Wenlockian to the north. The stratigraphic offset along the latter fracture is therefore significant. Michot also reports several secondary fractures considered as satellites of the main Huy Fault and displaying a very low-angle plane of about 5° to the south. The Huy Fault is viewed as a horizontal fault of Late Variscan age allowing, like a few other fractures within the Silurian Inlier, a northward thrust of the hanging wall.

Michot (1932) assimilates the northern and the median massifs (Fig. 183) into the "Namur Synclinorium" and the southern massif into the Dinant Synclinorium. He did not find any fractures between the Huy Fault and the Lochkovian northern border of the Dinant basin so that he considers the Huy Fault to be part of the Condroz Thrust – Midi Fault enabling the movement of the Dinant Synclinorium over the Namur "Synclinorium".

In 1933 and 1934, Fourmarier proposes a revision of the estimate of the offset along the Condroz Thrust. Based on sedimentological and structural (fold) arguments, Fourmarier estimates the displacement at 30 km. However, in the Hainaut area, the Namur basin is formed by the superposition of many tectonic wedges separated by as many thrusts. These secondary thrusts belong to the large faulted zone of the Condroz Thrust for which the total displacement is therefore probably more than 30 km in the Hainaut region.

In 1933 and 1936, Kaisin proposes a detailed study of the tectonic structures of the Namur Basin in the Namur area (Fig. 184). The section across the Namur Basin in Fig. 184 shows (1) a large southern unit composed of imbricated thrust sheets (= *"nappes méridionales"*), (2) a 2 km wide zone of intense faulting (= *"zone failleuse du Condroz"*) which thrusts the southern domain to the north, (3) an Upper Carboniferous (*"Houiller"*) south-dipping domain, also intensely faulted (= *"bande namurienne de Namur"*), (4) a less-deformed Devonian-Lower Carboniferous terrain covering the Brabant basement (= *"couverture dinantienne et dévonienne du massif Siluro-Cambrien du Brabant"*) and (5) the Brabant foreland (= *"avant-pays brabançon"*). The direction of the tectonic push is supposed to come from the south.

In 1933 and 1936a, Kaisin indicates that the Silurian Sambre-et-Meuse narrow strip is composed of superimposed thrust sheets separated by as many thrust faults. He also believes that the exact direction and strike of the Midi Fault within the Silurian Condroz Inlier is not fundamental. However, the aspect of the fault surface itself is far more important, namely a boundary between the northern Dinant basin and its basement. Kaisin (1936a) adds that if the south-to-north directed tectonic push is envisaged, then the Midi Fault must have an increasing offset towards the south at depth. Also the Midi fault plane must continue under the Ardenne domain and further to the south under the Mesozoic-Cenozoic Paris Basin before finally joining the lower face of the rigid crustal block. This block would have played a hinterland role pushing the imbricated thrust sheet domain (i.e. the *"nappes méridionales"*) to the north.

Again in 1936a, Kaisin provides a simplified geological map of the Brabant Massif (Fig. 185). The Brabant Massif extends from Wales to Belgium where it plunges and disappears in the Maastricht area. The Midi Fault is very long and extends from Wales to eastern Belgium. Fig. 185 represents the *"Houiller"* coal-basins surrounding the Lower Paleozoic Brabant basement (in black) and the trace of the long Midi Fault which is partially hypothetical.

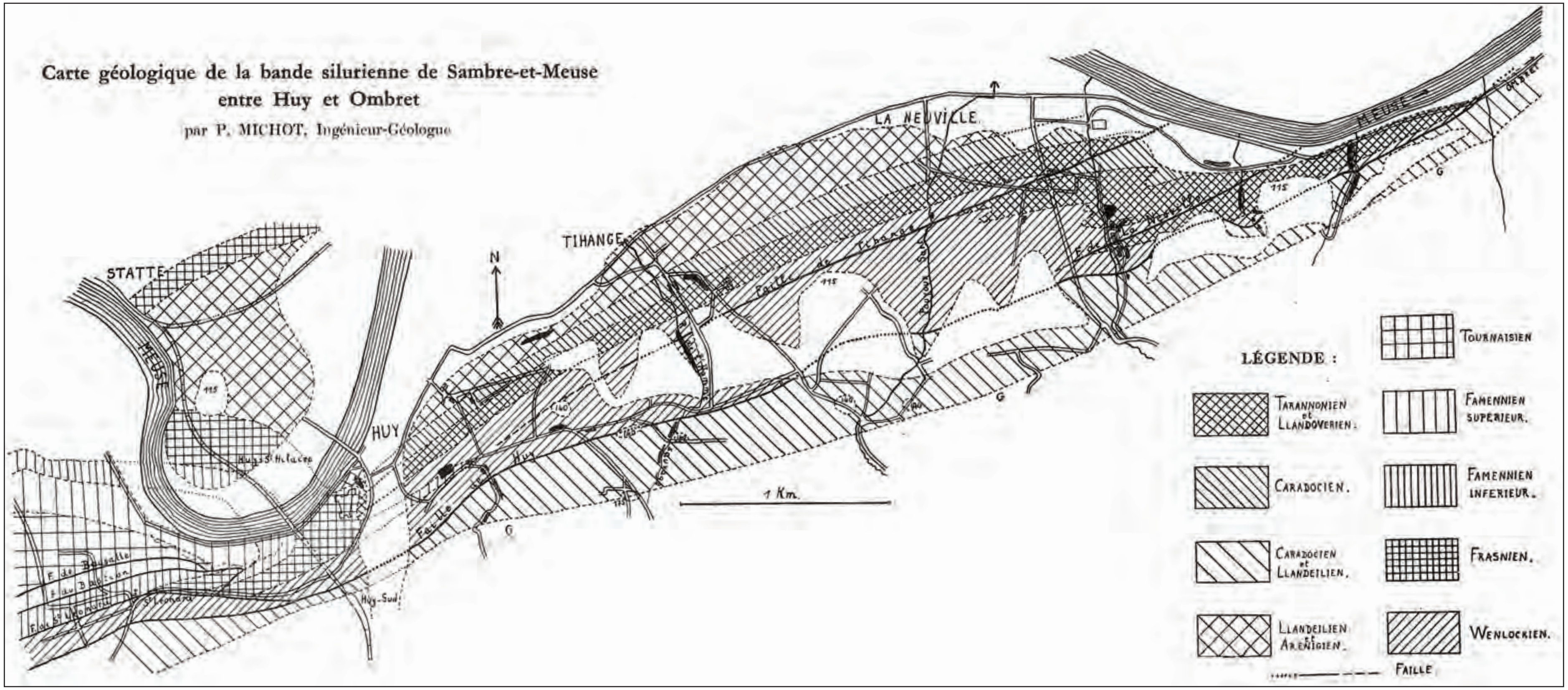


Fig. 183. Geological map of the Silurian Sambre-et-Meuse Inlier between Huy and Ombret (Michot, 1932). Two main fractures are shown: the Tihange Fault to the north and the Huy (= Midi-Eifelian) Fault to the south.

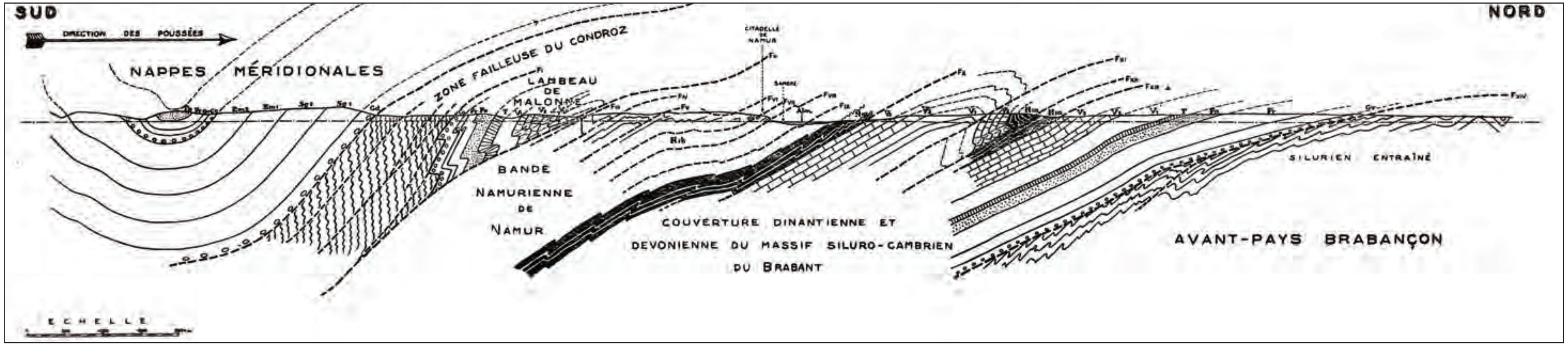


Fig. 184. S-N cross-section through the Namur Basin at the meridian line of Namur (Kaisin, 1933).

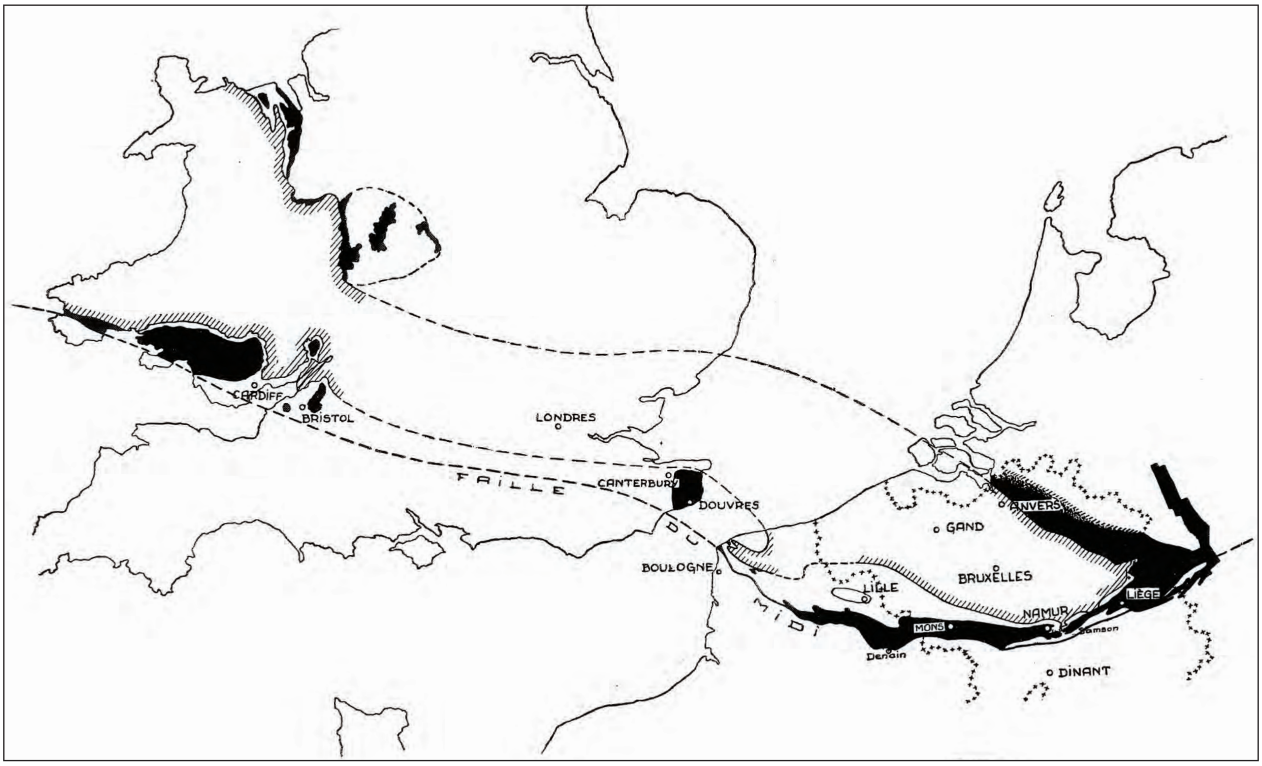


Fig. 185. Schematic map of the Brabant Massif and the surrounding Upper Carboniferous coal-basins (Kaisin, 1936a). The French-Belgian Midi Fault is extended westwards to Wales.

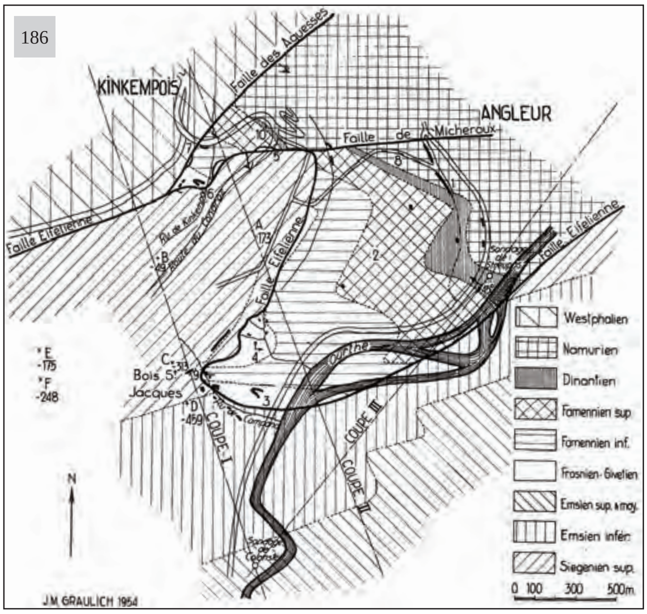
In 1941, Humblet suggests that the southern limit of the Liège coal-basin is marked by the 13 km long Eifelian Fault, striking from Neuvilles-en-Condroz (SE of Engis) to Kinkempois (near Angleur). The fault dips to the south and ranges between 30° near Angleur, 35° near Ougrée and 40° near Seraing.

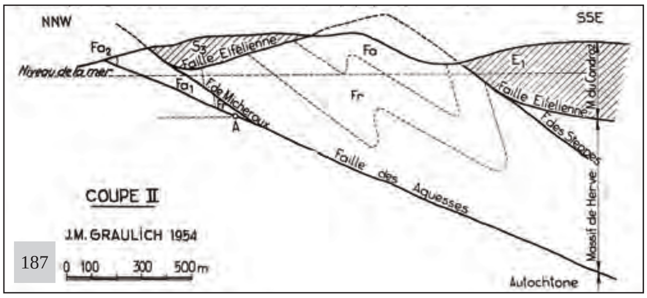
Fourmarier (1954) presents the following observations regarding the Silurian Condroz (or Sambre-et-Meuse) Inlier. The inlier appears as an anticline structure bordered on both sides by rocks that have very different ages. For example, near Huy, two contrasting Devonian facies, Lochkovian to the south and Frasnian to the north, are separated by only a narrow strip of Silurian rocks which is locally only 120 metres wide. On the southern side, more than 2500 m thick of Lower and Middle Devonian rocks are present while these are absent a few dozen metres to the north of the Silurian ridge within the Namur basin. This particular lithostratigraphic disposition around the Condroz strip enables Fourmarier (1954) to consider as "infinitely probable" the presence of a thrust surface within the Silurian terrains therefore making the connection between the Midi and the Eifelian faults.

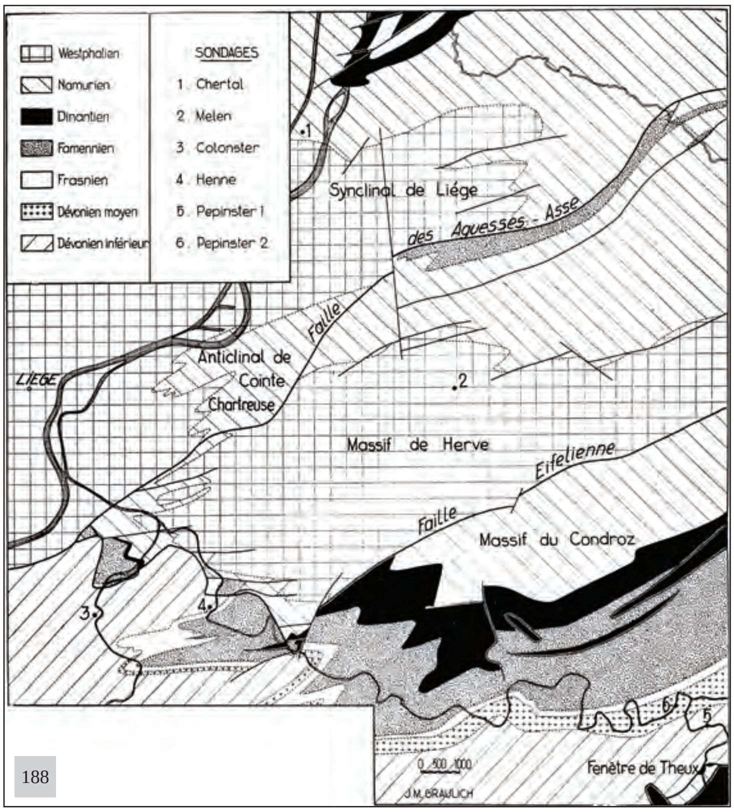
Fourmarier (1954) adds that compared to the Midi and the Eifelian faults, the structural understanding of the Ordovician-Silurian Condroz Inlier has been quite delayed. The lack of continuous cross-sections within the Silurian ridge and the difficulty in determining a clear and detailed Silurian stratigraphy contribute to the difficulty in elucidating the structure of the Condroz narrow strip.

Graulich (1955) no longer considers the Streupas and the Kinkempois "massifs" as tectonic stacks but as perisynclinal reappearances of the pre-Upper Carboniferous substratum of the Herve basin or, in other words the western termination of the Herve basin plunging to the west. This concept for the tectonic structure around the Eifelian Fault in the vicinity of Liège is well illustrated by the geological map in Fig. 186 and by the section in Fig. 187 showing that the pre-Upper Carboniferous rocks below the Eifelian Fault are not arranged in thrust sheets but belong to the Herve basin itself. This view necessitates folding of the Eifelian Fault. Further to the east, Graulich (1955) doubts the existence of an Upper Devonian Chèvremont thrust sheet and he again considers these rocks as a part of the Herve basin. The fault has a low-angle dip in the Ougrée area of about 30-35°.

Graulich (1955) also suggests a link between the Aguesses-Asse Fault at depth with the Eifelian Fault. The Eifelian discontinuity would limit the Condroz Massif to the north and the Aguesses-Asse discontinuity would underline the Herve Massif, therefore separating it from the Liège Syncline further to the north (Fig. 188). In other words, the true front of the allochthon thrust on to the parautochthon would be located at the regionally-significant Aguesses-Asse Fault. We refer the readers to the Aguesses-Asse Fault (Cambier & Dejonghe, 2010) for the development of the structural ideas of Graulich (1984) and Graulich et al. (1984, 1986).







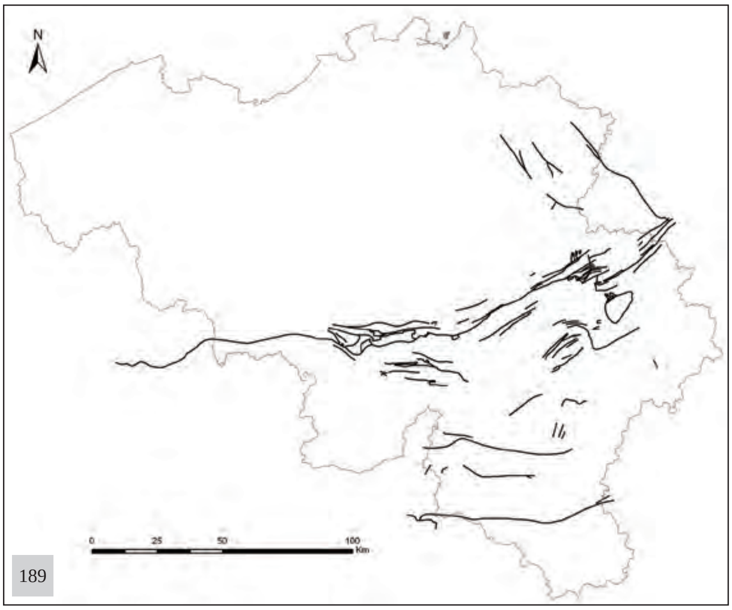


Fig. 186. Geological map of the Angleur region (Graulich, 1955)

Fig. 187. NNW-SSE cross-section (number II) through point A on Fig. 186 (Graulich, 1955). The Eifelian Fault is folded. The pre-Upper Carboniferous terrain in the footwall of the Eifelian Fault belongs to Herve basin and is therefore not composed of independent tectonic stacks (i.e. the Streupas and Kinkempois thrust sheets). S3 = Upper Pragian. E1 = Lower Emsian. Fr = Frasnian. Fa1 = Lower Famennian. Fa2 = Upper Famennian.

Fig. 188. Geological map of the eastern continuation of the Eifelian Fault (Graulich, 1955) separating the Condroz Massif to the south from the Herve Massif to the north.

Fig. 189. State of knowledge of the Belgian fault network compiled and mapped by de Béthune (1961).

In 1961, Bouroz et al. establish that the frontal zone of the Variscan orogeny is composed in the French-Belgian border region of six thrust sheets transported over each other from south to north. The southernmost sheet is the Midi massif. The Midi massif belongs to the northern border of the Dinant Synclinorium and is thrust along the *"Grande Faille du Midi"*. Drillhole data suggest a steepening of the dip at depth therefore providing evidence for a vertical uplift of the Midi massif.

In 1961, de Béthune publishes a geological map of the Belgian territory that shows the state of knowledge of the fault network (Fig. 189).

According to de Béthune (1961), the Variscan (or Hercynian) belt is subdivided into (1) the foreland, (2) the parautochthon frontal zone, (3) the Condroz nappe and (4) the Herbeumont nappe. Respectively from the north to the south:

(1) The foreland area of the Belgian Variscan belt constitutes an immobile basement relative to the displaced tectonic units of the Variscan orogeny. The Campine basin, the Caledonian Cambrian-Silurian Brabant Massif and the northern border of the Namur basin comprise the foreland.

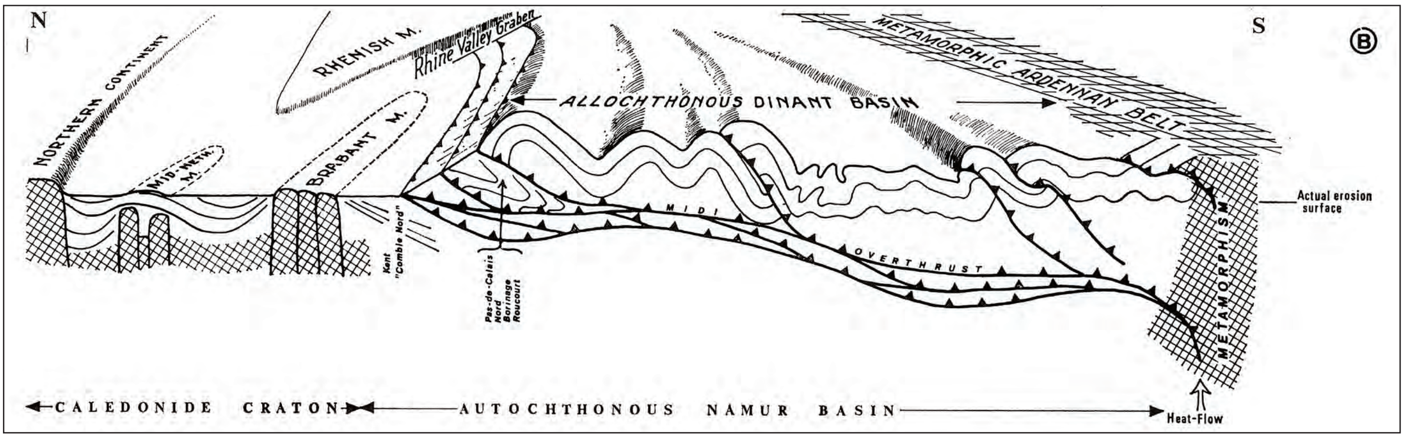


Fig. 190. N-S cross-section through the Variscan front (Bless et al., 1977). The Midi overthrust has transported the allochthonous Dinant Basin to the north over the autochthonous deposits of the Namur Basin.

(2) The frontal zone of the orogen is made up of superimposed thrust sheets. The tectonic wedges are thrust over each other from south to north and are northerly overturned. The frontal zone comprises the central and southern parts of the Namur basin. Actually, the southern border of the Namur basin has been removed at depth and transported to the north by the nappe movement. De Béthune (1961) therefore considers the Namur basin to be a northerly-overturned syncline compressed under a thrust nappe.

(3) The Midi Fault is a thrust surface along which the Dinant basin is transported over the foreland and under which the stacked thrust sheets of the frontal zone have been carried away. In the French Artois and Valenciennes regions and also in the Hainaut area, the Midi Fault consists of an anomalous stratigraphic contact between the Lower Devonian of the Condroz nappe thrust over the Upper Carboniferous frontal zone. Between Châtelet and Clermont, the Midi Fault is "lost" within the narrow Silurian strip of the Condroz Inlier. De Béthune (1961) explains the dissymmetry between the two "limbs" of the Condroz "anticline" by a major thrust within the inlier. The southern "limb" belongs to a northerly thrust nappe. Beyond Clermont, the Midi (or Eifelian) Fault again explains the contact between Lower Devonian and Upper Carboniferous rocks. Between Liège and Aachen, the fault subdivides into many branches.

(4) The Condroz nappe, thrust to the north along the Midi Fault consists of the Dinant basin, the Vesdre massif, the Ardenne anticline area and the Neufchâteau basin.

(5) The Herbeumont nappe is overthrust to the north along the Herbeumont Fault (see section 9.3). To the south, the Paleozoic nappe is covered by the tabular Mesozoic strata of the Paris basin which here marks the end of the Variscan belt outcrop in Belgium.

In 1977, Bless et al. consider that the Namur Basin continues for at least 35 km to the south under the Dinant Synclinorium. The Namur Basin would therefore be present under the Dinant-Givet region and may again continue under the Ardenne metamorphic belt. Bless et al. (1977) assimilates the Dinant Synclinorium into a large allochthonous nappe overthrust to the north onto the autochthonous Namur Basin (Fig. 190). The translation would have operated along the Midi overthrust. The transported Dinant nappe would be limited to the east by the Rhine Valley Graben where strike-slip faults were probably active during the Variscan tectonism. Fig. 190 shows the paleogeography subsequent to the Variscan shortening (Upper Carboniferous times) and the resulting structural relationships between the autochthonous deposits of the Namur Basin and the allochthonous Dinant Basin.

In 1981, Geukens proposes two cross-sections of the Belgian Variscan belt, a first in the central part (between Namur and Wellin) and a second in the eastern part (across the Theux Window and the Stavelot-Venn Massif). No new ideas regarding the Midi Fault are presented except that Geukens suggests a greater net slip along the Midi Overthrust than along the Condroz and "Eifel-Asse" (= Eifelian) Overthrusts. In the central part near Namur, the exact position of the 45° south-dipping Condroz Overthrust is not known and is drawn within the Ordovician-Silurian terrain of the Sambre-et-Meuse Strip.

Raoult & Meilliez (1985, 1987) provide a simplified geological map covering a large area between the Channel to the west and the Meuse valley to the east (Fig. 191). They indicate that the northern front of the Variscan Orogen is marked by a major overthrust called the Midi Fault in the west that continues to the east through the Silurian Condroz Inlier then through the Eifelian Fault. North of the front, the "Namur Synclinorium" unconformably overlies the Brabant Massif and may not be considered as autochthonous but as parautochthonous due to the imbricated tectonic wedges. South of the

Variscan front, the Dinant Synclinorium unconformably overlies the Cambrian-Ordovician basement (i.e. Rocroi, Givonne and Stavelot inliers) and can be considered as an allochthonous unit overthrust to the north. Based on borehole data, Raoult & Meilliez (1985) suggest an average low-angle dip of 15° for the Midi overthrust. Indeed, the Epinoy borehole (Fig. 191), located 7 km south of the surface emergence of the Midi Fault, intersects this fault at a depth of 2100 m while the Jeumont borehole (Fig. 191), 10 km south of the Midi fault trace, intersects the fault at 2400 m depth. Raoult & Meilliez (1985, 1987) indicate that the offset of the Midi Fault, or the net translation vector of the Dinant Nappe, is at least 50 km and possibly as much as 150 km. Both geologists (1985, 1987) also propose a detailed cross-section of the Ardenne (Fig. 214) given and described below in the Interpretations section.

As in his previous papers of 1980, 1986 and 1988, Michot (1989) reiterates that the Eifelian Thrust is independent from the Aguesses-Asse Fault of which the Aachen Fault constitutes the eastern continuation. The Aachen Fault is therefore of low significance and has no link with the Eifelian Fault. Michot (1989) estimates the total transport of the Condroz Nappe along the Midi-Eifelian Fault from the French to German borders to be at most 15 km and perhaps only 10 km.

In 1990, Meilliez & Mansy apply the model of thin-skinned deformation to the Ardenne domain. Using these concepts, the Paleozoic basement between the English Channel and the Rhine river is constructed from two structural domains separated by a major thrust (Fig. 192). Both the Midi and the Aguesses-Asse faults belong to this separation tectonic complex. The northern domain is called the Brabant Parautochthon (= *"Parautochtone brabançon"*) and the southern domain, thrust onto the northern one, the Ardenne Allochthon (= *"Allochtone ardennais"*). These new terms replace the old designations of synclinorium and anticlinorium which apply only to the Paleozoic cover while the new terms apply to both the Paleozoic cover and to the basement of a larger area from France to Germany. Simply, the Brabant Parautochthon corresponds to the former "Namur Synclinorium" and Liège Syncline while the Ardenne Allochthon corresponds to the Dinant, Vesdre and Neufchâteau synclinoria and to the Ardenne and Givonne anticlinoria.

In 1994, Geluk et al. indicate that the eastern extremity of the Variscan Midi-Aachen Thrust is crosscut and nearly obliterated by the Roer Valley Graben, the northwestern branch of the Rhine Graben rift system (see the Feldbiss Fault Zone in section 9.2).

Based on palynomorph reflectance data, Steemans (1994) provides arguments for the structural framework of the western part of the Sambre-et-Meuse Strip (Fig. 193). Within the Lower Paleozoic basement of the Condroz Inlier south of the Bois-de-Presles Fault, reflectance (Ro) values range between 4.32 and 5.50% decreasing to between 3.35 and 3.48% north of the fault (Fig. 193). The Bois-de-Presles Fault therefore limits two blocks of similar age that were buried at very different depths. The Bois-de-Presles Fault is interpreted as the eastern continuation and a segment of the Midi Fault, a hypothesis already formulated by Fourmarier in 1914 (see above).

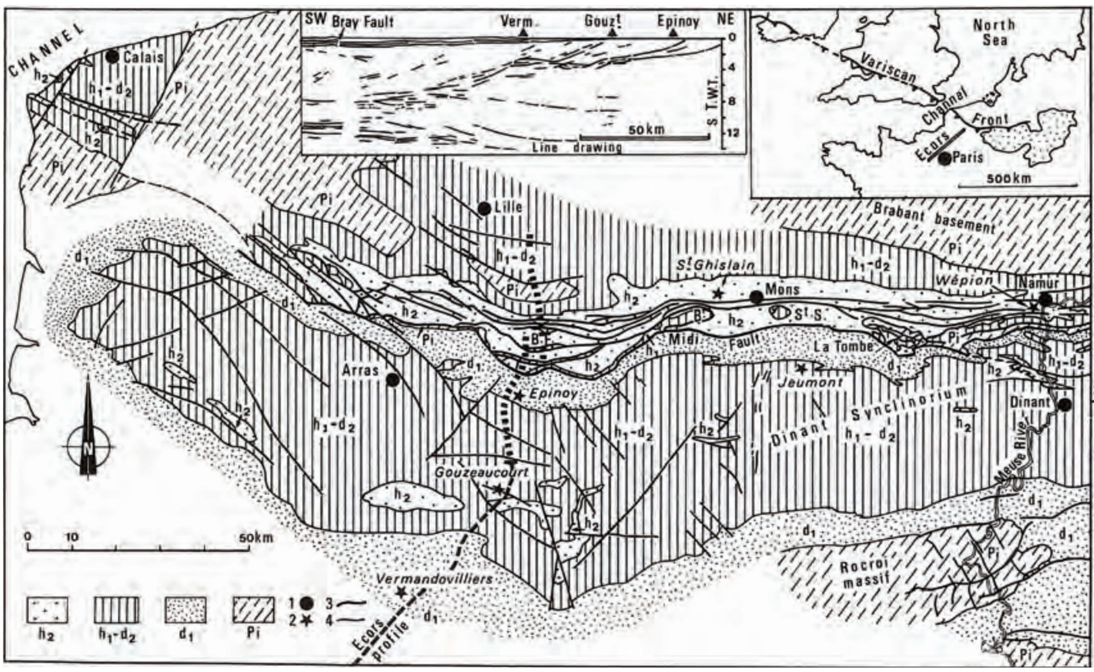


Fig. 191. Geological map of the Paleozoic subcrop between the Channel and the Meuse valley at the longitude of Dinant (Raoult & Meilliez, 1987). Insets show the location of the ECORS "Nord de la France" seismic profile and the northern half of the line drawing. H2 = Silesian. H1-d2 = Dinantian, Middle and Late Devonian. D1 = Early Devonian. Pi = Lower Paleozoic. 1 = city. 2 = borehole. 3 = fault. 4 = stratigraphic contact.

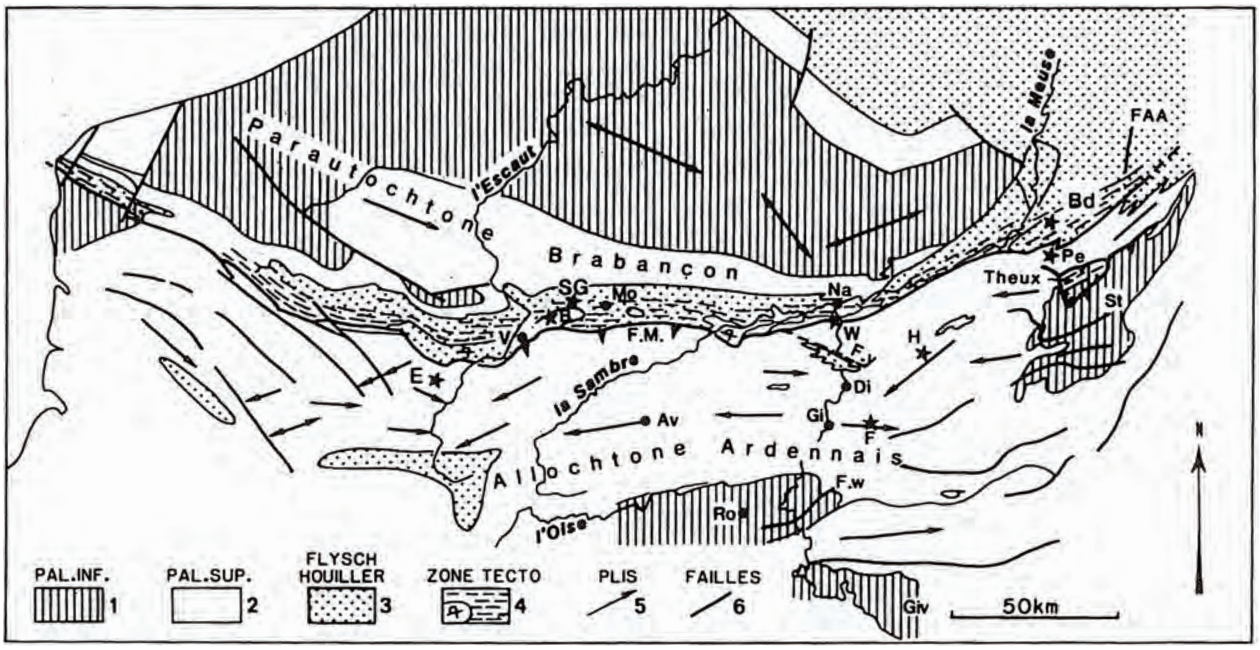


Fig. 192. Simplified geological map of the Paleozoic basement from the English Channel to the Rhine river (Meilliez & Mansy, 1990). Both the Brabant Parautochthon and the Ardenne Allochthon have a Lower Paleozoic basement (1) unconformably overlain by a Devonian-Lower Carboniferous cover and (2) an Upper Carboniferous (i.e. Silesian or *"Houiller"*) flysch. The allochthon is separated from the parautochthon by a frontal tectonised zone (4) which is the footwall of the major Midi Fault. Main folds (5), faults (6) and boreholes (star) are given. FM = Midi Fault. FAA = Aguesses-Asse Fault. Fy = Yvoir Fault. Localities: Av = Avesnes. Di = Dinant. Gi = Givet. Mo = Mons. Na = Namur. Ro = Rocroi. V = Valenciennes. Boreholes: Bd = Bolland. E = Epinoy. F = Focant. H = Havelange. Pe = Pépinster. SG = St-Ghislain. W = Wépion.

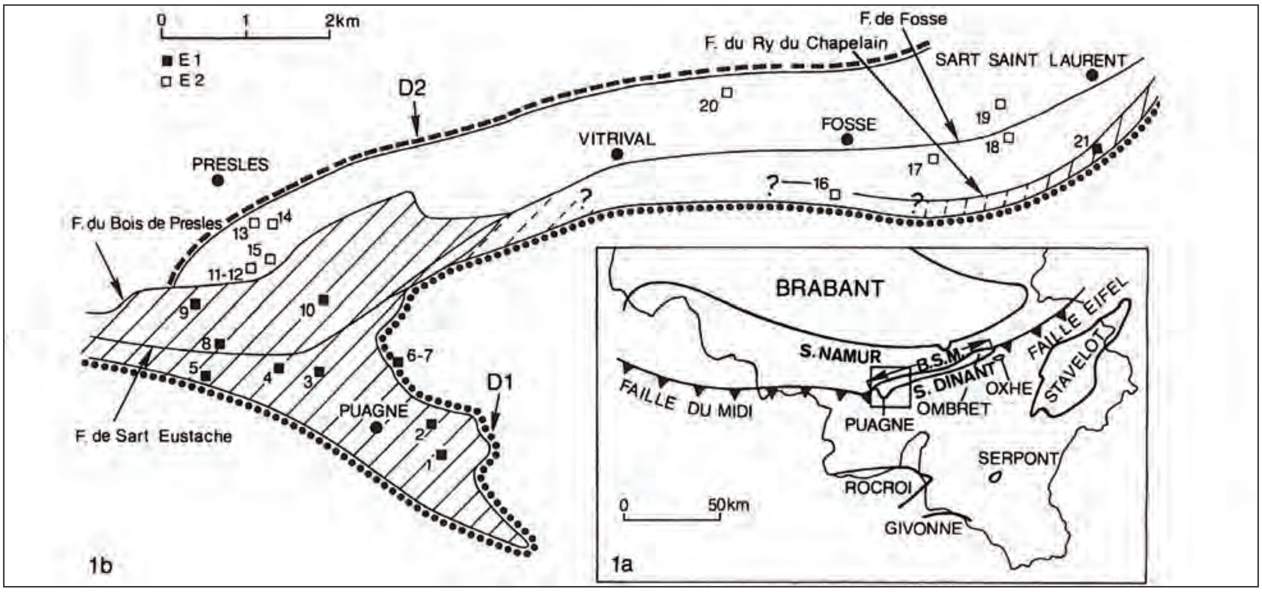


Fig. 193. 1a. Locations of the Caledonian massifs and the Namur and Dinant synclinoria. 1b. Geological structure of the western part of the Sambre-et-Meuse Strip (after Michot, 1934, 1944). Hatched area = presence of schistosity. D = Devonian conglomerates. D1 = Lower Devonian. D2 = Middle Devonian. E = sample number and their level of maturation of the organic matter; E1 = high Ro, E2 = mean Ro. F = faults. From Steemans (1994)

Further to the east near Fosse, reflectance data enables Steemans (1994) to consider the Ry du Chapelain Fault as the continuation of the Midi Fault. This connection was already proposed by Michot in 1944.

According to Hollmann & Walter (1995), the northern frontal system of the Rhenohercynian fold-and-thrust belt in the vicinity of Liège and Aachen is composed of 4 main tectonic units that are, from south to north: the Theux Window, the Vesdre Nappe, the Herve Imbricate Zone and the Liège Syncline (Fig. 194).

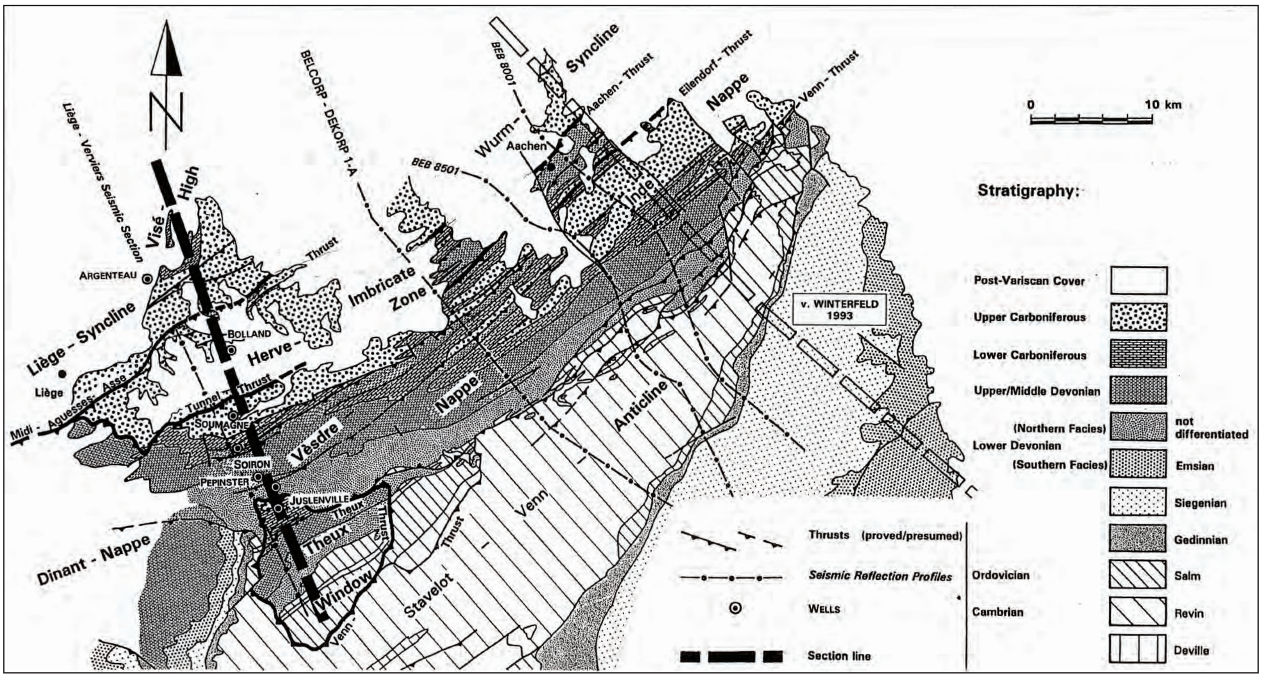


Fig. 194. Geological map of the Variscan Front Zone between Liège and Aachen (Hollmann & Walter, 1995). The northern foreland of the Stavelot-Venn Inlier is made up of the Theux Window, the Vesdre Nappe, the Herve Imbricate Zone (between the Midi-Aguesses-Asse and Tunnel thrusts) and the Liège Syncline. Main seismic profiles and drillholes are given.

Like the views of Graulich (e.g. 1955, 1984), Hollmann & Walter (1995) suggest that the eastern continuation of the Midi-Eifelian Fault beyond Liège is made through the Aguesses-Asse Fault (see Cambier & Dejonghe, 2010). The Aguesses-Asse Thrust separates the Herve Imbricate Zone to the south (an allochthonous unit) from the Liège Syncline to the north which is therefore assimilated within a parautochthonous unit. The entire hanging wall of the Aguesses-Asse Thrust is an allochthonous complex made up of the Herve Imbricate Zone, the Vesdre Nappe and the Theux Window.

Sintubin & Matthijs (1998) suggest that the eastern continuation of the Midi-Eifelian Fault beyond Liège splits into 4 major branches (Fig. 195): the Theux, Eupen, Xhoris and Venn faults (The Theux and Xhoris faults are described in separated data sheets published in Cambier & Dejonghe, 2010). These 4 fractures constitute the equivalent of the Variscan front thrust in the northern part of the Stavelot Inlier.

The views of Hance et al. (1999) differ greatly from the opinions of Hollmann & Walter (1995) for whom the Aguesses-Asse Fault is the continuation of the Midi Thrust. After Hance et al. (1999), the connection between the Midi Fault to the west (actually the Eifelian Fault in the Liège vicinity) and the Aachen Fault to the east is made by the Tunnel Fault (located to the south of the Aguesses-Asse Fault, see Cambier & Dejonghe, 2010). The south-dipping Tunnel Fault actually connects at depth with the north-dipping Theux Fault forming the Theux-Tunnel Thrust (Figs. 196 and 197). The Theux-Tunnel Thrust is therefore considered as a major component of the Variscan frontal thrust displacing the Ardenne Allochthon to the north over the Brabant foreland during the Asturian stage of the Variscan Orogeny. The Theux-Tunnel Fault continues to the south of the Theux Window with a southern dip and directly joins with a deep flat-lying reflector, which is envisaged as the downward continuation of the Midi-Aachen Thrust (Fig. 196 & 197).

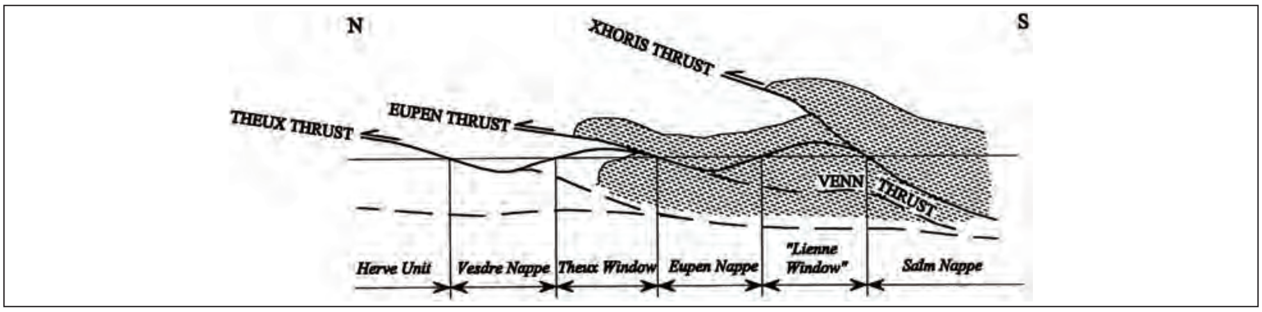


Fig. 195. Variscan tectonic units and subdivision of the Midi-Aachen Fault into four main thrusts (Theux, Eupen, Xhoris and Venn faults) in an overstep sequence to the north of the Stavelot-Venn Massif (Sintubin & Matthijs, 1998). Hatched area = Lower Paleozoic basement.

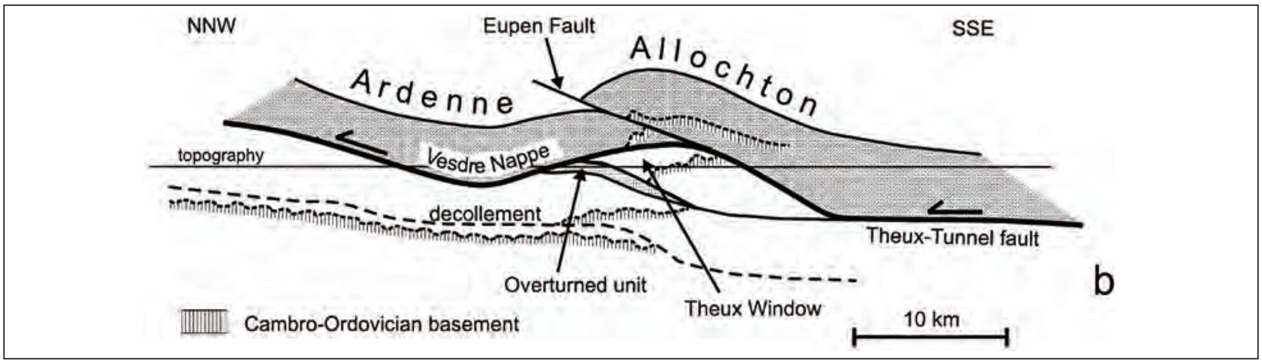


Fig. 196. Schematic cross-section through the Variscan frontal system in the east of Liège (Hance et al., 1999). The folded Theux-Tunnel Fault is a segment of the Midi-Aachen Thrust and enables the northward transport of the Ardenne Allochthon over the Brabant foreland.

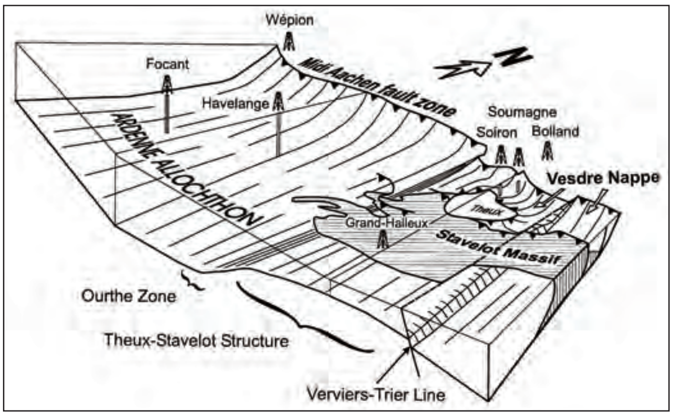


Fig. 197. 3D schematic model of the Ardenne Allochthon (Hance et al., 1999).

According to Delcambre & Pingot (2000), the south-dipping Midi Fault has a low-angle dip of 15° in the vicinity of Landelies (Fig. 198) that increases to about 40° near Bouffioulx. From a stratigraphic point of view, the hanging wall is made up of Lower Devonian sandstones and siltstones overthrust onto the carbonate Visean and siliciclastic, coal-bearing Upper Carboniferous (Namurian-Westphalian) footwall. Delcambre & Pingot (2000) define 3 structural units on the Fontaine-l'Évêque – Charleroi geological map: the parautochthonous massifs to the north, the thrust massif dislocated in superimposed thrust sheets in a middle position and the (allochthonous) Midi Massif to the south. With this concept, the Midi Fault separates the "sliced" thrust massif from the transported Midi Massif.

In 2001, Verniers et al. consider the Condroz Inlier as composed of at least four tectonic stacks transported along the Midi Thrust, a northeastern one (near Ombret), a large one in the central area and two smaller stacks in the southwest (in the Puagne and the Acoz areas). Siliciclastic rocks of Ordovician and Silurian age, forming height Ordovician and nine Silurian formations, compose the inlier (Verniers et al., 2001).

In 1997 and 2004, Delmer makes a revision of the structure of the Variscan front in the Hainaut and Namur area. The orogenic frontal zone is subdivided into three main regional units: a large allochthonous nappe or Midi Massif to the south, a second transported allochthonous unit called *"Grand Massif Superficiel"* in a median position and finally an Upper Carboniferous subautochthonous area (i.e. the *"massifs subautochtones imbriqués"*) to the north (Fig. 199). The tectonic relationships between those three units are represented on the cross-section in Fig. 250 (Tombe Fault in section 9.6). The Midi Massif, translated along the Midi Fault, is overthrust onto the *"Grand Massif Superficiel"* which itself is transported over the imbricated subautochthonous massifs.

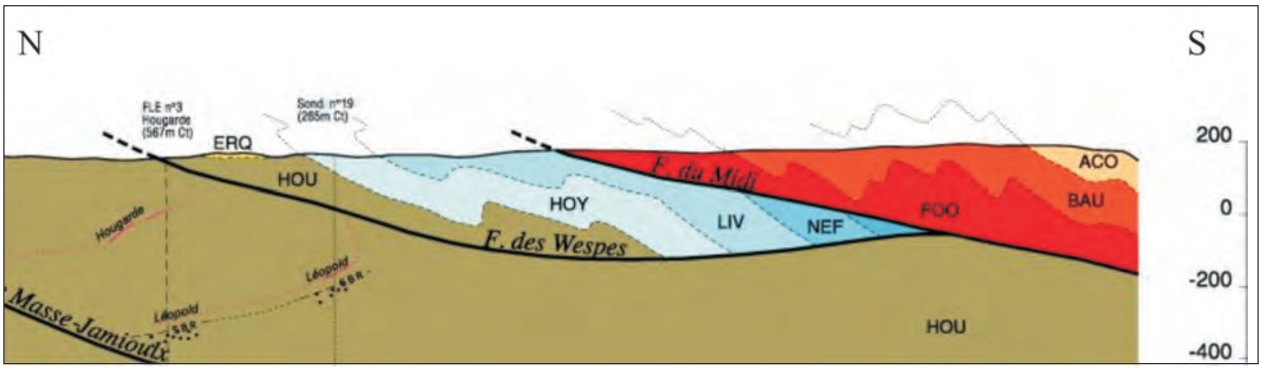


Fig. 198. N-S cross-section through the Midi Fault in the vicinity of Landelies (Delcambre & Pingot, 2000). FOO, BAU & ACO = Fooz, Bois d'Ausse and Acoz formations (Lower Devonian). NEF, LIV & HOY = Neffe and Lives formations and Houyoux Group (Visean). HOU = "*Houiller*" Group (Namurian-Westphalian).

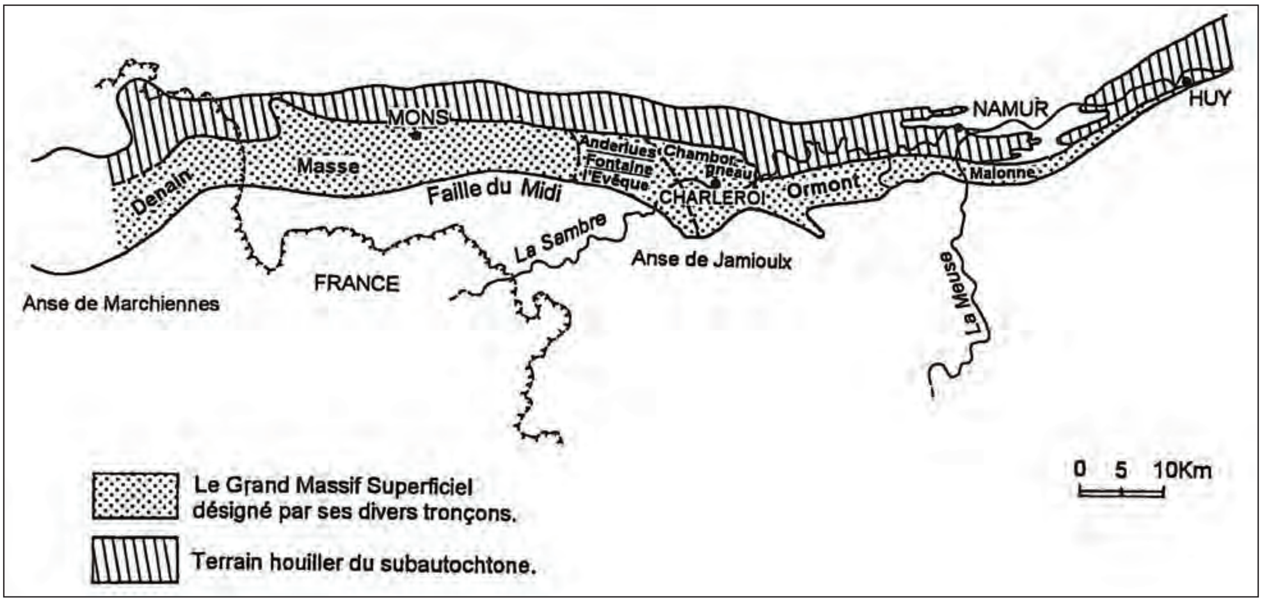


Fig. 199. Geological map of the Variscan Front Zone between the French-Belgian border and Huy (Delmer, 2004).

Interpretations

Dumont (1832) gives no explanations for the origins of the Belgian geological structures, which henceforth includes particular deformation structures, typically folds and faults. However, d'Omalius d'Halloy (the "father of Belgian geology" and reviewer of the Dumont's work; see Cauchy et al., 1832) provides explanations for the structures identified by Dumont. These constitute the first descriptions of thrust phenomena in Belgium (in the Province of Liège).

The existence of folded and faulted structures in the Belgian subsoil are considered to be the result of "dislocations of the earth's crust" and of "movements of separated blocks". D'Omalius d'Halloy also adds that strata, after their formation, have undergone violent movements and consequently acquired a specific structure that results from "gliding" on an inclined plane. The Belgian geologist does not talk explicitly about "thrusts" but properly alludes to their existence in the vicinity of Liège. Consequently, d'Omalius d'Halloy indicates that the work of Dumont contains observations in favour to the "Plutonian theory" of J. Hutton, which, in the beginning of the 19th century, trends to replace the "Neptunian theory" of A.G. Werner.

Dufrénoy & Elie de Beaumont (1841) recognize that the southern border of the Valenciennes coal-basin is intensely folded and shows successive fold limbs generally gently dipping in a same southerly direction (Fig. 165). Both authors also indicate that the plunge of the fold axes may be strongly variable including northerly overturned folds with subhorizontal fold axes. The folding stage occurring in a subhorizontal plane would be the result of a horizontal force acting in a NNW-SSE direction. No relation between the subvertical fault (i.e. the Midi Fault) and the horizontal force is envisaged.

According to the structural views of Dormoy (1862), the southern half part of the *"Houiller"* coal-basin does not exist that he explains by upheaval (in French, the *"soulèvement général du Midi"*) and erosion (Fig. 200). Uplifting has occurred longitudinally in the middle part of the *"Houiller"* coal-basin a short time after the deposition of the Upper Carboniferous rocks. Subsequently to the uprising of the southern half domain of the basin, a major cataclysm is invoked to erase and level the uplifted masses at the end of the *"Houiller"* period. Actually, the southern half would have been "swept out" and "carried away" to the south (Dormoy, 1862).

From the discovery of the Midi Fault (Dumont, 1832; Dufrénoy & Elie de Beaumont, 1841) until the works of Gosselet (1860a,b) and Dormoy (1862), geologists agree with an origin of the overturned folds in the southern border of the *"Houiller"* coal-basins as resulting from subhorizontal contractional forces. They also agree that there is a subvertical anomalous fault contact between the *"Houiller"* to the north and the *"anthraxifère"* terrain to the south and with the upward movement of the southern block. However, excepting the ideas of d'Omalius d'Halloy (in Cauchy et al., 1832), no attempt is made to explain the mechanism of the origin of the fault.

In 1863, Briart & Cornet propose an explanation of the structural discontinuity between Valenciennes and Aachen that accounts for the anomalous contact between Devonian and Upper Carboniferous rocks to the south and to the north respectively. The fault would result from the amplification and breaking of an anticline structure. They write: *"Le premier effet du mouvement de rapprochement de l'Ardenne a été la formation, au Sud du bassin, d'une voûte dont la partie septentrionale s'est renversée sur le terrain houiller qui, aussi probablement, s'est plié et renversé sur lui-même. La puissance de compression continuant à agir, il s'est produit une rupture vers la clef de voûte et la partie méridionale de celle-ci a été poussée vers le Nord en glissant sur le plan de rupture"*. The movement along the fault is clearly considered as resulting from contractional constraints.

The geological events that folded and faulted the Belgian Carboniferous rocks and that are responsible for their current structural disposition are divided, into 5 main stages (after Cornet & Briart (1876, 1877):

- the first stage (Fig. 201) comprises both the initial folding of the Lower Paleozoic Condroz Inlier and the subsequent deposition of Devonian and Carboniferous rocks within the basins. The Condroz crest separated the southern Dinant basin from the northern Namur basin where the Devonian thickness is less than in the Dinant basin;

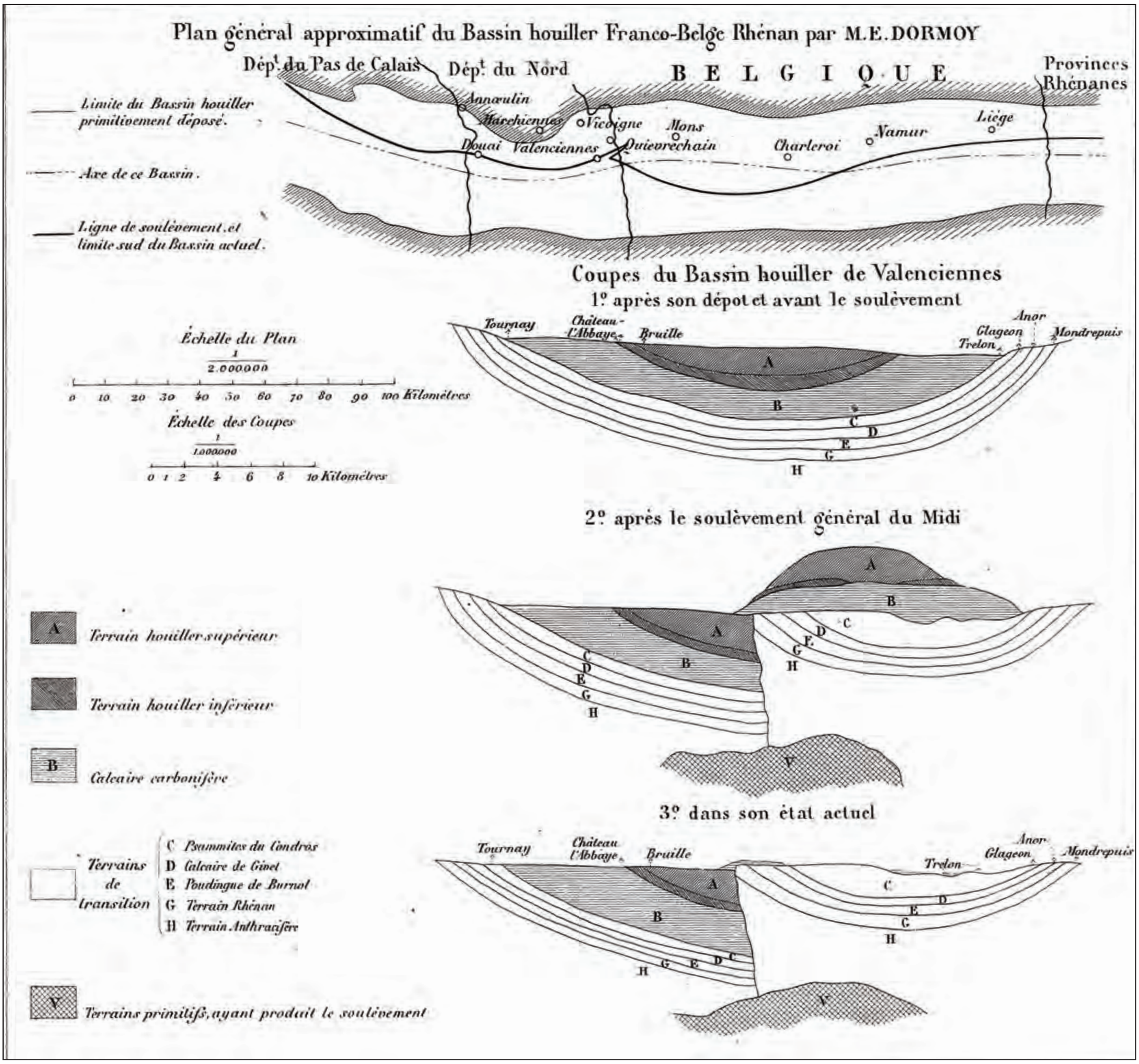


Fig. 200. N-S section across the Upper Carboniferous French-Belgian coal-basin at three different stages: (1) after the formation of the deposits and before their disruption, (2) the longitudinal uplifting of the southern half part of the basin, and (3) after the removal of the upraised masses (Dormoy, 1862).

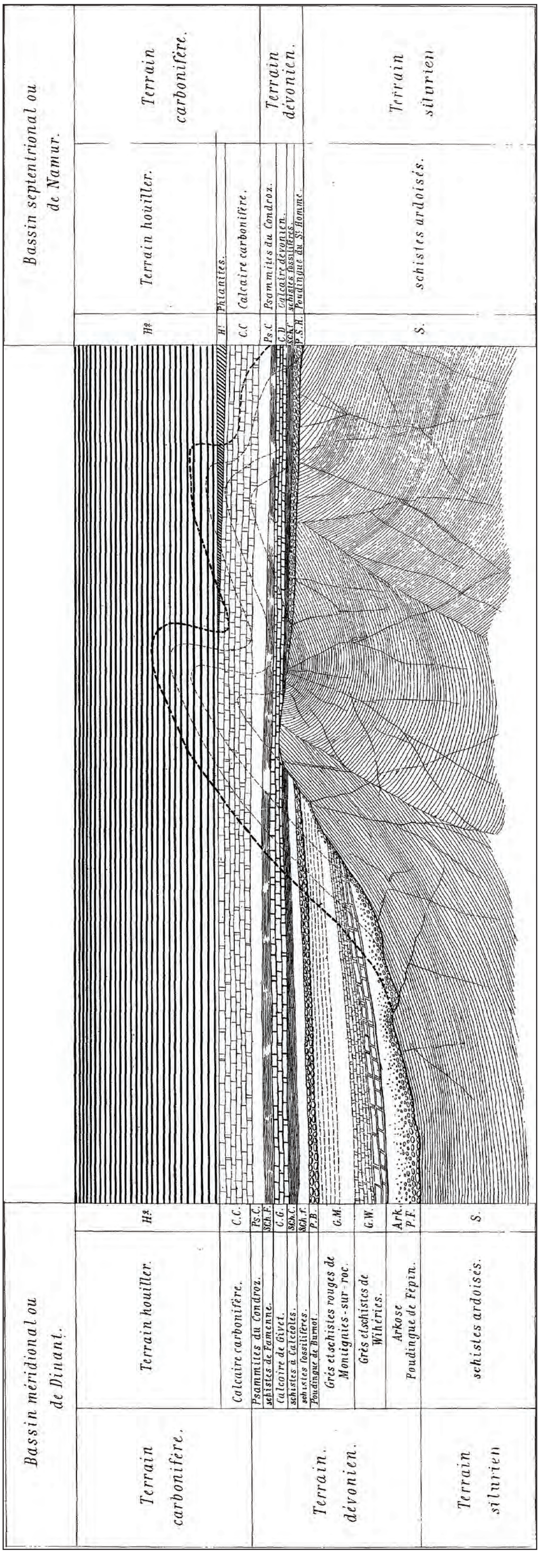


Fig. 201. First stage: first upheaval of the Silurian Condroz ridge and Devonian-Carboniferous deposition in the Dinant and Namur basins (Cornet & Briart, 1877).

- the second stage (Fig. 202) is the second folding of the Condroz crest at the end of Carboniferous times. The Devonian and Carboniferous rocks of the southern border of the Namur basin were "straightened up", overturned and folded but not faulted. For Cornet & Briart (1877), the rocks initially display a "plastic" behaviour (in response to the geological constraints) that evolves after lithification into a "breaking" behaviour (stages 3, 4 & 5);
- the third stage (Fig. 203) is the formation of the north-dipping Boussu Fault north of the Condroz Inlier and the northward recess of the northern Namur basin (i.e. normal displacement);
- the fourth stage (Fig. 204) is the generation of the south-dipping Anzin *"Cran de retour"* or Anzin Fault responsible for the recess of the southern part and upheaval of the northern part of the Namur basin (i.e. normal displacement). Cornet & Briart (1877) remark that the two folding stages (stages 1 & 2) of the Condroz crest are due to two different contractional events while the two faulting stages (stages 3 & 4) result from opposing extensional events acting during the southward movement of the Dinant basin;
- the fifth stage (Fig. 205) is a last faulting event responsible for the Midi (-Eifelian) Fault and imposing a strong relief to the Belgian terrain. The reverse Midi Fault enables the thrusting of "old" rocks of the Dinant basin over the Carboniferous Namur basin. Cornet & Briart (1877) suggest that Belgian territory had a mountainous relief before being entirely eroded, levelled and covered again by seas by the Cretaceous.

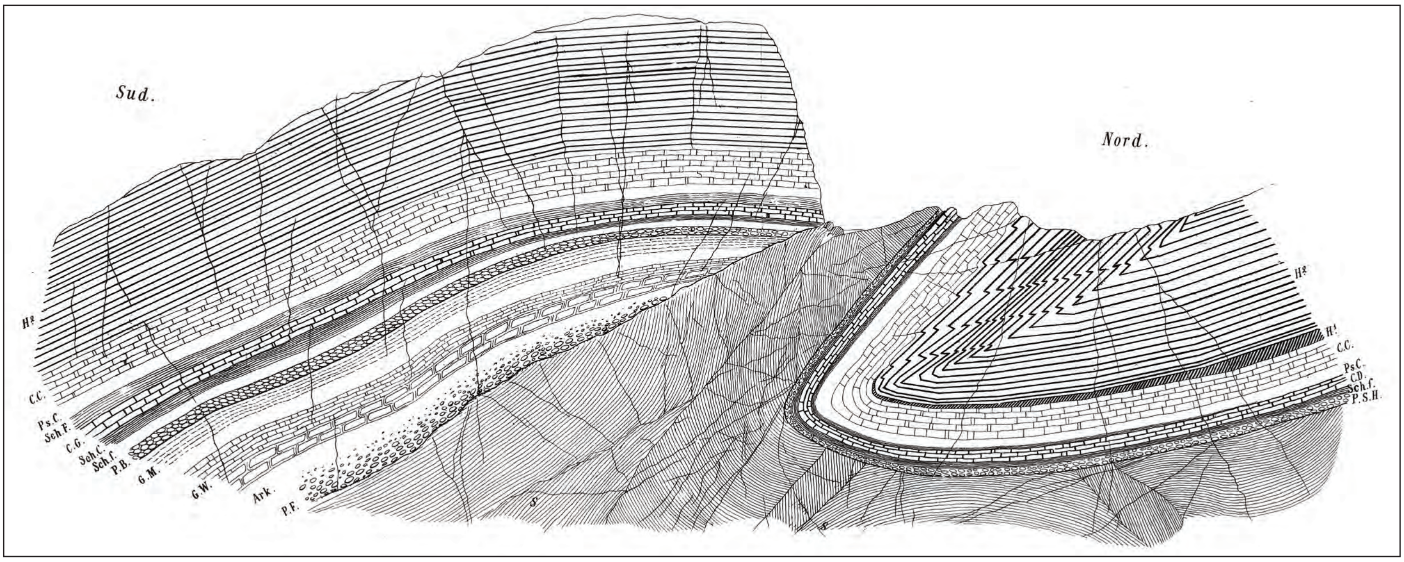


Fig. 202. Second stage: second upheaval of the Condroz crest and overturning of the southern part of the Namur basin (Cornet & Briart, 1877).

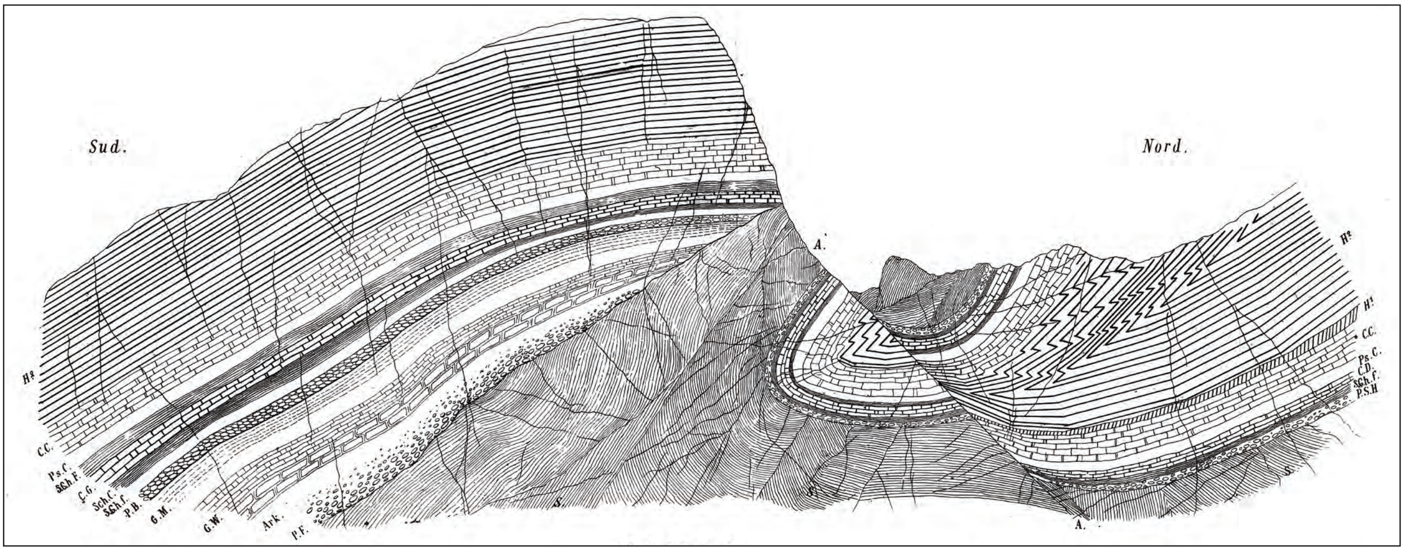


Fig. 203. Third stage: formation of the Boussu Fault (Cornet & Briart, 1877).

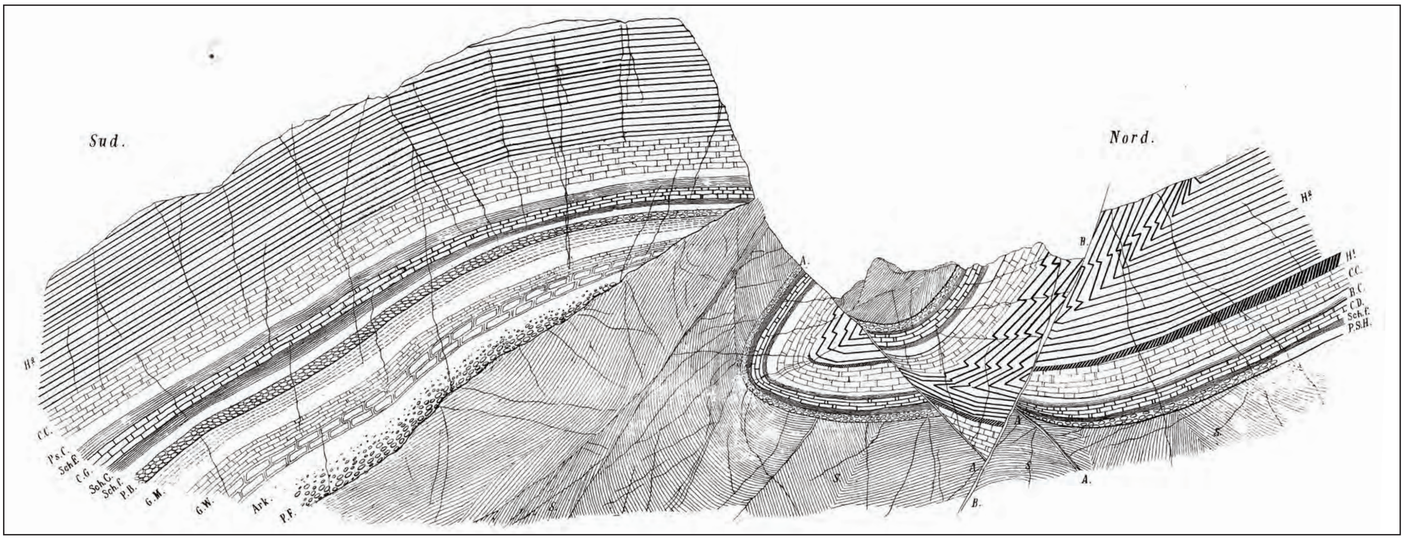


Fig. 204. Fourth stage: generation of the Anzin Fault (Cornet & Briart, 1877).

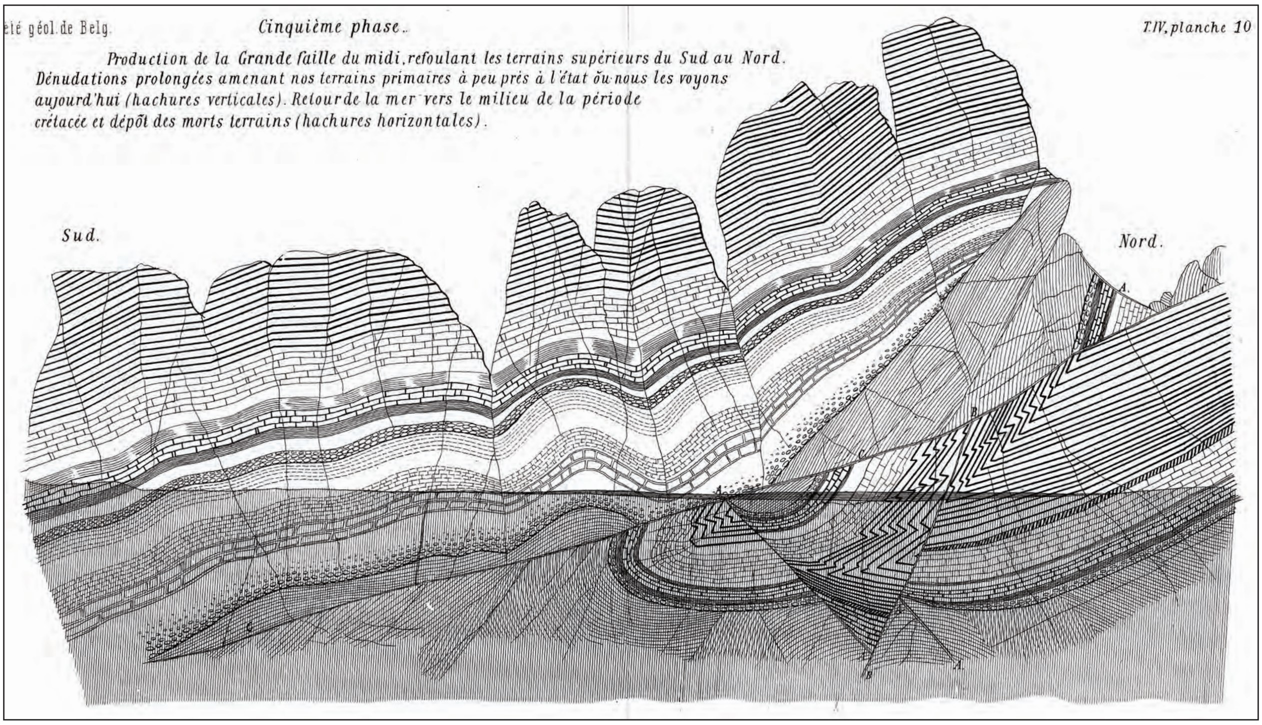


Fig. 205. Fifth stage: generation of the Midi Fault (Cornet & Briart, 1877).

In 1879, 1880 and 1888, Gosselet considers the geological structure of the French-Belgian *"Houiller"* coal-basin as a consequence of the so-called *"Ridement du Hainaut"* event. This stage of intense rock deformation results from a major horizontal "pushing" directed from south to north. The origin of the folding is explained as follows: *"La cause du ridement réside dans l'affaissement des parties centrales du bassin et dans le relèvement relatif des bords avec glissement des couches les unes sur les autres. L'affaissement lui-même est une conséquence du retrait constant de la croûte terrestre"*. Gosselet (1879, 1880) proposes the following succession of events:

(1) during Devonian-Carboniferous times, the Condroz Inlier constituted a topographic high (anticline) separating the northern Namur and southern Dinant basins where infilling with sediments was in progress (Fig. 206);

(2) the "Ridement du Hainaut" event caused the accentuation of the Condroz anticline and its overturning to the north (Fig. 207). Strata of the Namur basin were uplifted and overturned to the north in such a way as to seem to plunge under the Silurian Condroz anticline while the Lower Devonian strata of the Dinant basin moved to the north;

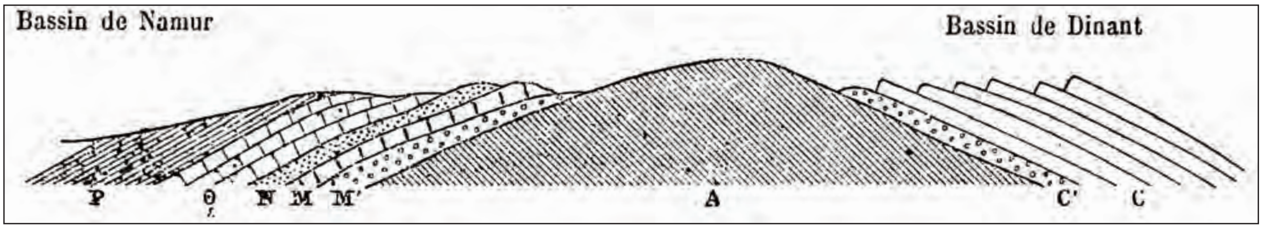


Fig. 206. N-S section across the Silurian Condroz Anticline (A) separating du Namur and Dinant basins during Devonian-Carboniferous times (Gosselet, 1879). A = Silurian shales of the Condroz. C' = Lochkovian conglomerates. C = Red shales and Lochkovian micaceous sandstones. M' = Givetian conglomerates. M = Givetian and Frasnian limestones. N = Famennian micaceous sandstones. O = Carboniferous limestones. P = Upper Carboniferous shales.

(3) to the west of the Condroz anticline, the effect of the north-directed pushing was more significant as the Lower Devonian terrain of the Dinant basin was unconformably placed over the Devonian-Carboniferous rocks of the Namur basin (Fig. 208). This unconformable discontinuity is called *"Grande Faille"* by Gosselet.

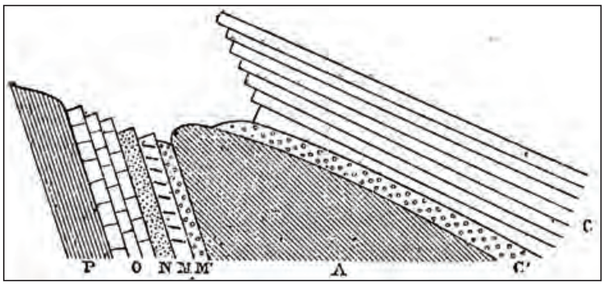


Fig. 207. Overturning of the Condroz Anticline with northward movement of the northern border of the Dinant basin and straightening up of the southern border of the Namur basin (Gosselet, 1879). See Fig. 206 for the legend.

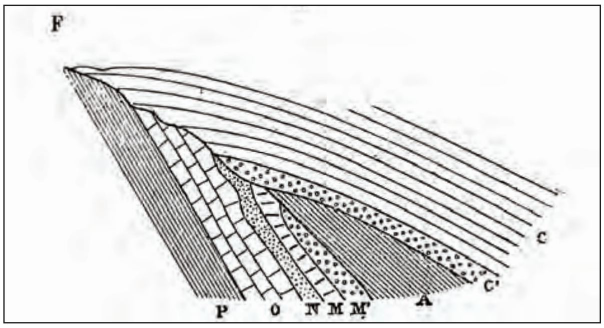


Fig. 208. Breaking of the Condroz Anticline and northward translation of the Dinant basin over the Namur basin (Gosselet, 1879). The fault discontinuity (i.e. the "*Grande Faille*") constitutes the northern limit of the Ordovician-Silurian Condroz Inlier. See Fig. 206 for the legend.

In 1894a, Briart proposes a detailed study of the thrust sheets in the vicinity of Fontaine-l'Évêque, Landelies and Marchiennes. He suggests a subdivision of the "Tombe Massif" into three stacked thrust sheets bounded to the south by the Midi Fault. In the chronology of the thrust-type faults (= *"failles de refoulement"*), the Midi Fault would be the last manifestation of the successive south-to-north thrust transports. We refer readers to Figs. 243 & 251 in section 9.6 showing the Tombe Fault.

In 1898, de Dorlodot presents a hypothesis regarding the formation of both the "Condroz crest" and the Midi Fault. Contrary to the views of Gosselet, the *"grande faille"* would not result from the amplification and breaking of the Condroz anticline but probably from the increase of the process that resulted in the genesis of the Condroz crest. The formation of the Midi Fault and the Condroz Inlier are interdependent as the uplifting of the Condroz crest was probably the cause of the formation of the *"grande faille"*. De Dorlodot (1898) adds: *"la grande faille et les failles analogues consistent formellement dans le refoulement de la crête du Condroz vers les dépressions de la plaine houillère qui s'étendait à ses pieds"* (Fig. 209).

In 1900 and 1905, Smeysters makes an inventory of the major faults disrupting the eastern part of the Upper Carboniferous Hainaut coal-basin (e.g. the Carabinier, Ormont and Tombe faults). Smeysters suggests that the Midi Fault constitutes the last manifestation of the contractional dynamics responsible for the current structure of the Hainaut coal-basin. Smeysters (1900) also indicates that recent seismic activity in Belgium results from movements along the active Midi Fault. The earthquakes of February 23rd, 1828; November 1881 and September 2nd, 1896 are described as "thrust seisms" (or *"séismes de chevauchement"*) related to the Midi Fault.

In 1906(a), Fourmarier publishes a note in which the connection is made between the Midi and Eifelian faults that were initially considered independent. The link would occur within the Silurian Condroz anticline. This hypothesis has the advantage that all the tectonic discontinuities between the Namur and the Dinant basins are packaged together in one very large but simple structural feature. Related to the tectonic compression and folding of the Ardenne area, the Condroz anticline is accuentuated until the formation of a large thrust sheet (*"grande nappe de charriage"*). The Dinant basin has therefore been transported and overthrust to the north over the Namur basin.

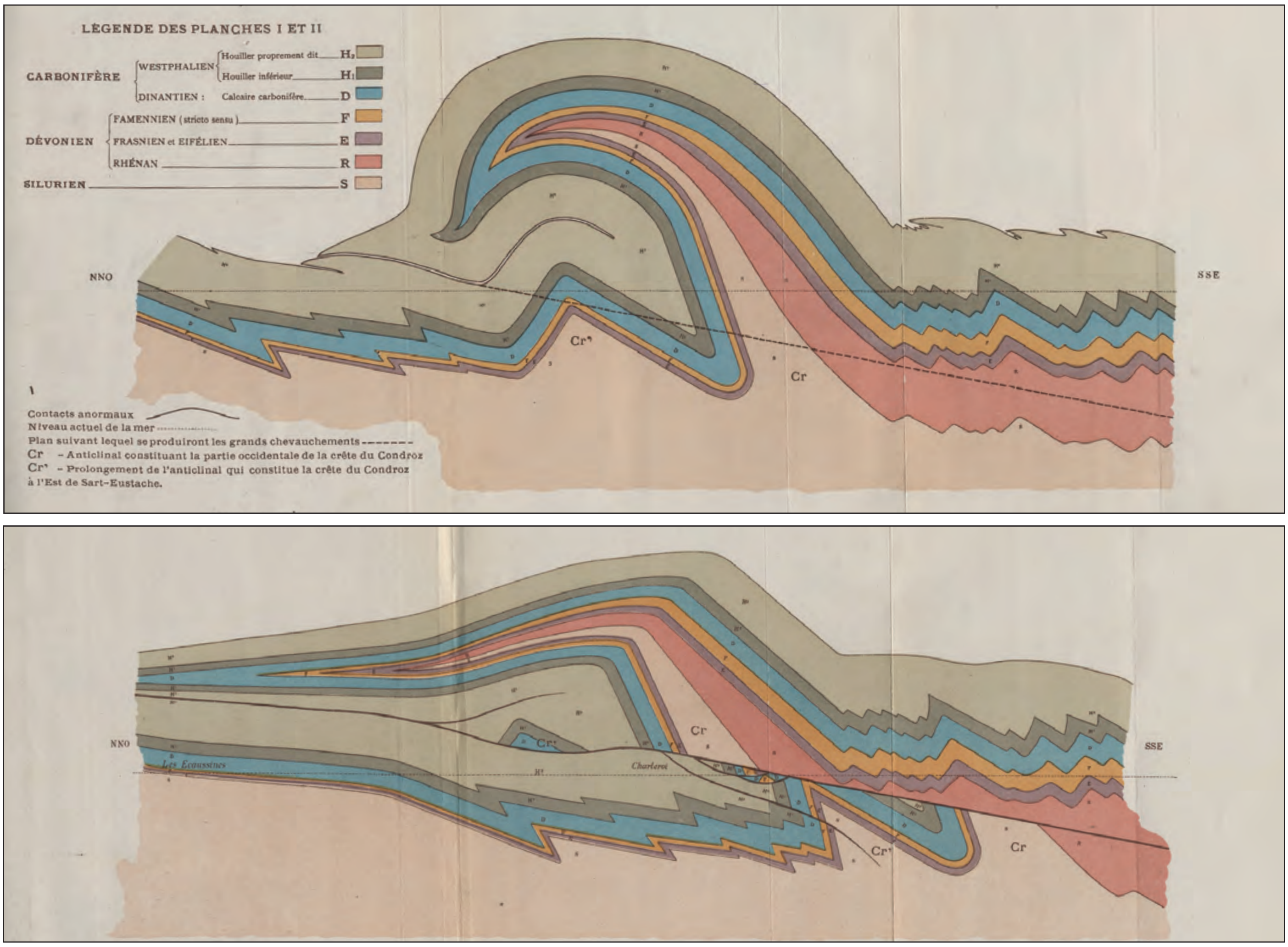


Fig. 209. A. N-S geological section in the vicinity of Charleroi representing the Condroz anticline and the folded structure prior to the major horizontal overthrusts. B. Same cross-section after formation of the horizontal overthrusts (de Dorlodot, 1898).

Based on the structural analogy between the Condroz Inlier and the Mendip Hills (southern England) established by Stainier in 1920(b) (see above), a new interpretation of the genesis of the Condroz ridge is proposed. To the north, the Sambre-et-Meuse or Namur syncline was a large basin formed by two longitudinal secondary basins (north and south) separated by the northern Silurian anticline ridge. The Namur syncline was separated from the Dinant syncline further to the south by the southern Ordovician-Silurian anticline ridge. During shortening, the Midi Fault appears and displaces the southern anticline ridge. The Dinant basin was therefore thrust in its entirety to the north over the southern secondary basins of the Namur syncline to butt against the northern Silurian anticline ridge. In other words, the Condroz Inlier would result from the superposition (fault contact) of two Silurian crests that were initially separated.

Fourmarier (1951) proposes the following succession of 4 events linked to the Variscan structures in the Liège area (Fig. 210):

(1) Production of first-order folds (I on Fig. 210). Folding of the Devonian-Carboniferous rocks marks the initiation of the Namur and the Dinant synclinoria as individual features.

(2) First stage of thrusting. Amplification and breaking of the Condroz anticline (II on Fig. 210) produces the first northward thrust (i.e. the Rocheux Fault, R on Fig. 210) and the first nappe. The Rocheux Fault is observed within the Theux Window and constitutes the southern limit of the Namur basin.

(3) Second stage of thrusting. Disruption and displacement of the first nappe (III on Fig. 210) by a second northward thrust (i.e. the Eifelian or Theux Fault) that enables the northward movement of a subsequent second nappe, which unconformably overlays the first nappe.

(4) Small folding event. Later folding of the terrains amplifies the undulations of the Theux-Eifelian Fault which acquires a characteristic listric trend.

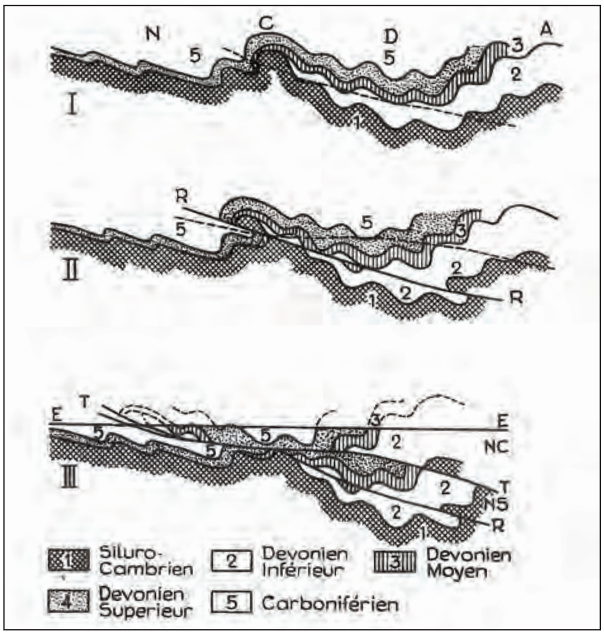


Fig. 210. Main stages of the Variscan Orogeny in the vicinity of Liège (Fourmarier, 1951). N = Namur basin. C. Condroz anticline. A = Ardenne anticline. D = Dinant basin R = Rocheux Fault. T = Theux (= Eifelian) Fault. EE = topographic surface. NC = Condroz Nappe. NS = Spa Nappe.

Fourmarier (1951) also tries to understand the "tectonic anomalies" to the east of Liège, probably related to the original heterogeneity of the Paleozoic basement. He expresses the following rule: *"une série sédimentaire parfaitement homogène soumise aux efforts orogéniques se déforme en plis réguliers ; si la déformation s'est réalisée de façon anormale c'est qu'il y avait un manque d'homogénéité de la matière soumise aux efforts"*. In other words, the Devonian-Carboniferous terrains east of Liège must have "homogeneity defaults" that are potential places for genesis of the effects of the shortening. The Paleozoic basement "anomalies" formed during the sedimentation stage have to be taken into consideration when studying the particular tectonic structure to the east of Liège.

The *"Grande Faille du Midi"* is identified by Clément in 1963 on seismic profiles where the emergence of a major reflector "F" coincides exactly with the Midi fault trace. The interpreted profile is located along the French-Belgian border between Solre-le-Château and Merbes-le-Château. The overthrust character of the Midi Fault as well its low-angle dip to the south are confirmed by the seismic profiles. Clément (1963) makes the hypothesis that the domain located under the seismic reflector "F" corresponds to the southern border of the Brabant Massif and to the deepest parts of the Namur Basin. The Midi Fault would run at a depth of more than 2000 metres while the Cambrian-Silurian basement would occur at 4500 metres depth.

The knowledge relative to the northern front of the Variscan Orogen is much improved in the eighties by the acquisition of geophysical seismic data. In 1981, Meissner et al. interpret a reflection-refraction profile carried out in 1978 through the Stavelot-Venn Inlier and demonstrate thin-skinned tectonics at the front of the Variscan Orogen. Thanks to this study, Meissner et al. are able to reveal thrust planes within parts of the northwestern Variscides and to analyze their nature from the reflection seismology.

Indeed, they identify a strong upper reflector at a depth of 3-4 km that is assimilated to the Midi Fault. The reflector is interpreted as *"a prominent and well lubricated thrust fault along which a huge horizontal nappe displacement took place during the last stages of the Variscan orogeny"*. They also propose a cross-section (Fig. 211) showing the deep flat-lying reflector of the Midi Fault. Meissner et al. make a comparison between the structures of the northwestern Variscides and those of the North American Appalachians where major thrusts have also been identified through seismic campaigns. The authors add that *"the formation of thin-skinned nappes riding over a rather undeformed subsurface along plane, well lubricated thrust faults may indicate a final stage of compressional tectonics"*.

In 1985, Durst identifies a strong and good quality seismic reflector at about 1.1 – 1.5 seconds in a seismic profile across the northeastern Stavelot-Venn Massif (Fig. 212). The reflector rises constantly to the north and emerges at the surface near Aachen where it coincides with a discontinuity of the Aachen overthrust (i.e. the eastern segment of the Midi-Aachen Fault, also known in Dutch as *"Aachener Überschiebung"*). Actually, the reflector lies at a depth of about 3000 to 4000 metres; its SW dip displays a variable inclination and is interrupted by several steeply south-dipping faults (Fig. 212).

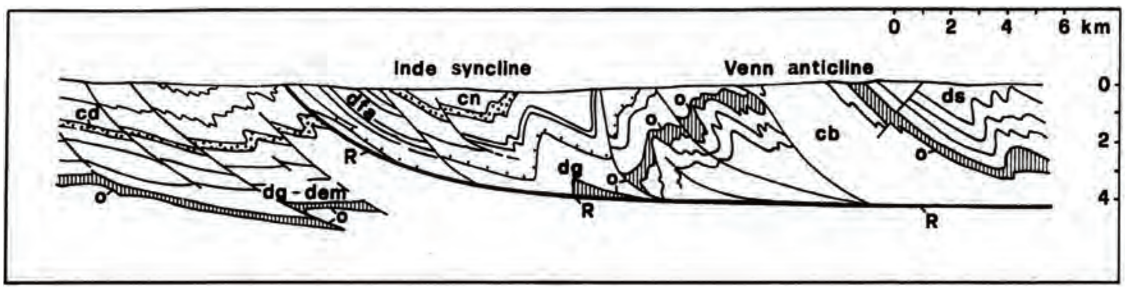


Fig. 211. Cross-section across the Stavelot-Venn Massif (Meissner et al., 1981). R = seismic reflector (Midi Fault), cb = Cambrian, o = Ordovician, dg = Gedinnian (i.e. Lochkovian), ds = Siegenian (i.e. Pragian), dem = Emsian, dfa = Famennian, cd = Dinantian, cn = Namurian.

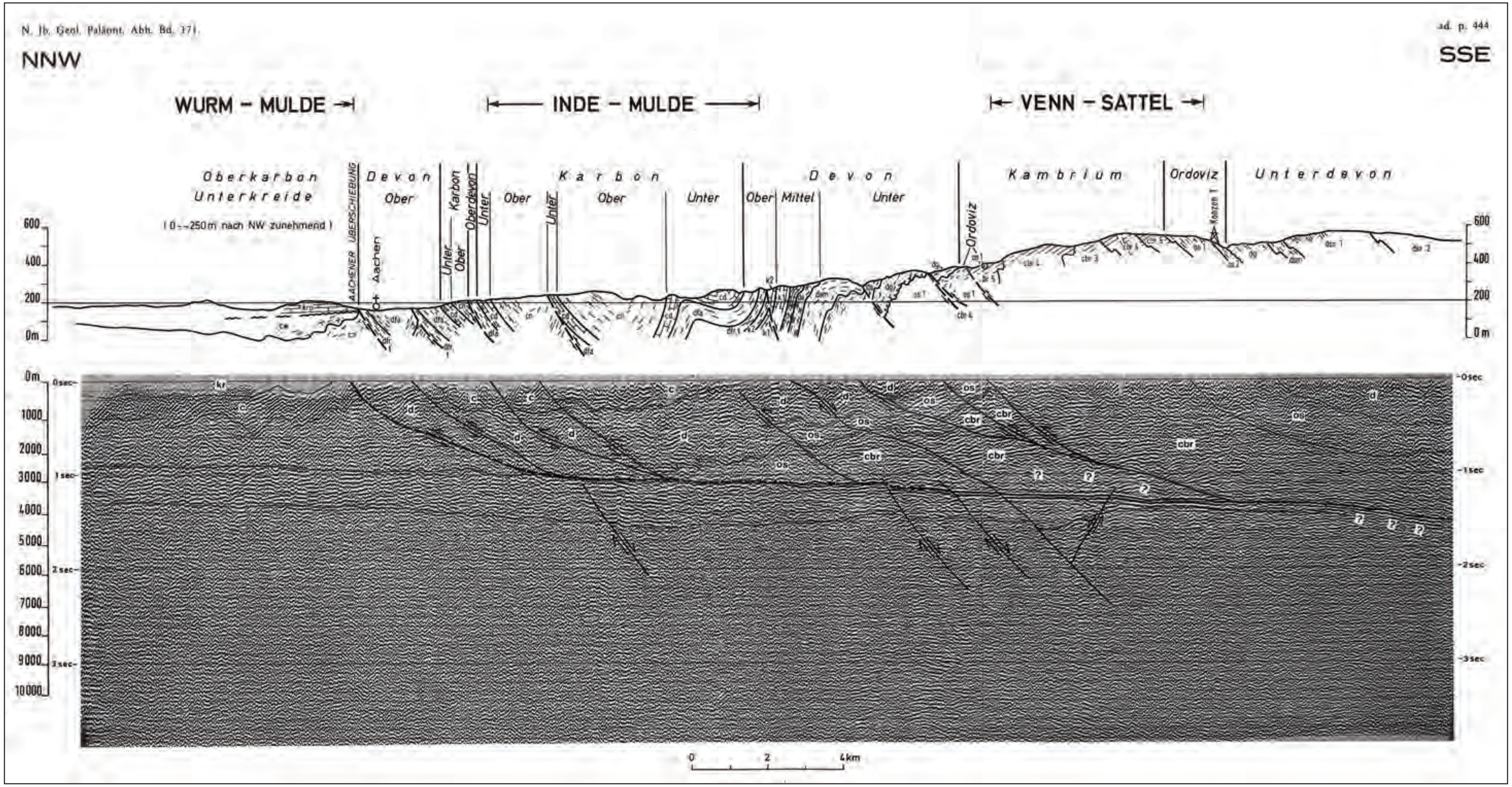


Fig. 212. Seismic profile and geological interpretations across the northeastern Stavelot-Venn Inlier (Durst, 1985). The major seismic reflector at about 1.1 – 1.5 seconds coincides with the Aachen thrust fault. cbr = Cambrian. os = Ordovician. d = Devonian. c = Carboniferous. kr = Cretaceous.

Interpretations of the seismic profile (Durst, 1985; Fig. 212) also show that: (1) the northern part of the Lower Paleozoic Stavelot-Venn Anticline (or *"Venn Sattel"*) is transported to the north and overthrust onto the Inde Syncline (or *"Inde Mulde"*), (2) steeply south-dipping extensional faults exist below the plane of the Aachen overthrust and (3) south-dipping contractional thrust-type faults delimiting imbricated thrust sheets constitute the allochthonous upper unit.

Within the framework of the French *"ECORS"* project (*"Etude Continentale et Océanique par Réflexion et réfraction Sismiques"*), a 228 km long and NE-SW oriented seismic profile was realized between Cambrai and Dreux in 1983 (Cazes et al., 1985). The main results and interpretations from this so-called *"Nord de la France"* deep seismic profile are published in Cazes et al. (1985), Raoult & Meilliez (1985, 1987), Mansy et al. (1997), Lacquement et al. (1999) and other papers.

The clear continuity of the subhorizontal seismic reflectors observed at about 3 seconds and their link with the Midi Fault located at the northernmost extremity of the ECORS profile (Fig. 213) enables Cazes et al. (1985) to define an upper allochthonous unit or *"Nappe de Dinant"* made up of the Dinant Synclinorium and the Ardenne Anticlinorium. Below 5 seconds, the whole part of the lower crust may correspond to the folded Caledonian Precambrian and Lower Paleozoic basement or to the southward continuation of the Brabant Massif which, furthermore, is unaffected by the Variscan shortening. The central part of the ECORS profile shows at between 3 and 8 seconds, multiple gently south-dipping (20-30°) reflectors that are interpreted as imbricated thrust sheets where the Dinant Nappe "roots" (Fig. 213).

Cazes et al. (1985) consider the Midi Fault to be a major thrust leading to the northward overthrusting of the "Ardenne-Dinant nappe" onto the autochthonous Caledonian Brabant massif and its relatively undeformed Devonian-Carboniferous cover. The overthrust of the Dinant Nappe is observed over a distance of 125 km from its deep roots to its front (which is the Midi Fault). The average depth of the Moho is between 35 and 40 km making this the deepest Moho discontinuity in Western Europe.

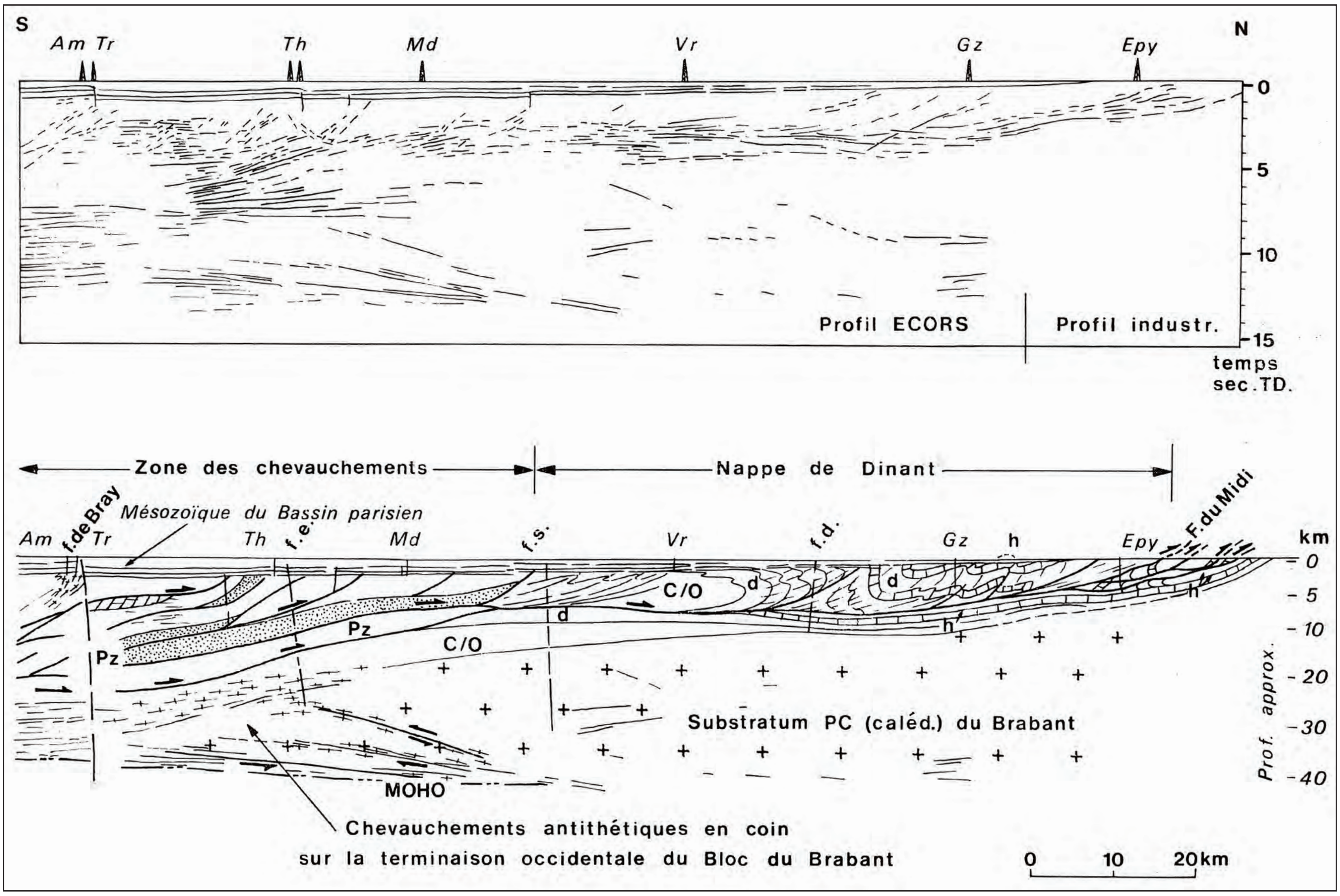


Fig. 213. Northern half of the ECORS *"Nord de la France"* deep seismic profile and geological interpretations (after Cazes et al., 1985; in: Raoult & Meilliez, 1985). Faults: f.d. = Doullens Fault. f.e. = Eu Fault. f.s. = Somme Fault. Boreholes: *Am* = Aux-Marais borehole. *Epy* = Epinoy borehole. *Gz* = Gouzeaucourt borehole. *Md* = Montdidier borehole. *Th* = Thieux borehole. *Tr* = Troussencourt borehole. *Vr* = Vermandovilliers borehole. Geological succession: PC = Precambrian. C/O = Cambrian-Ordovician. d = Devonian. h = Carboniferous. Pz = undifferentiated metamorphic Paleozoic rocks (dotted = magnetic rocks).

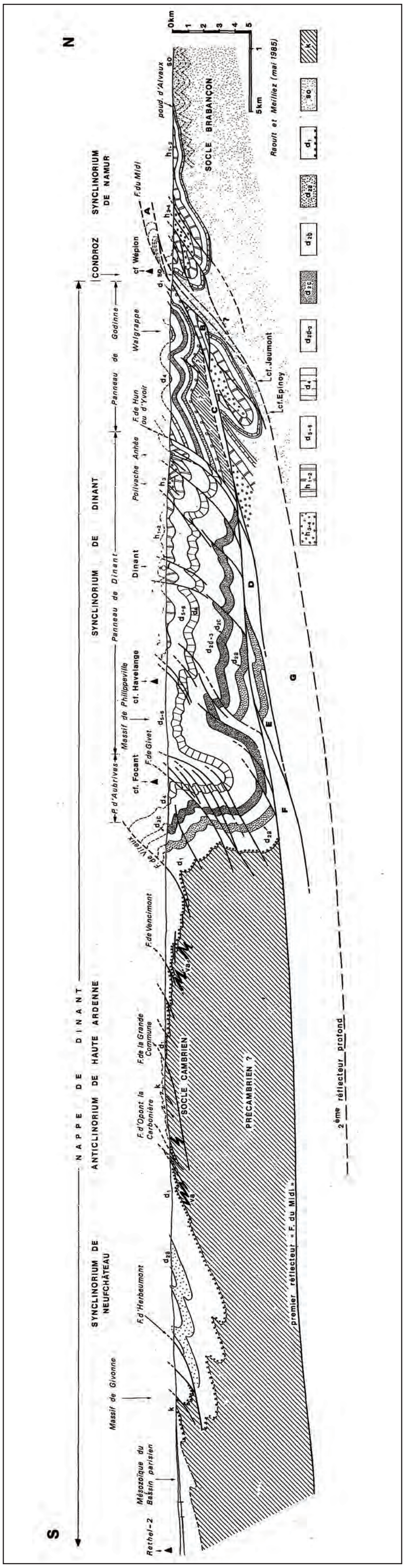


Fig. 214. Deep cross-section of the Variscan front system (Ardenne Massif and Brabant Massif) at the longitude of Dinant (Raoult & Meilliez, 1985). k1 = Cambrian. so = Silurian-Ordovician. d1 = Lochkovian. d2a = Lower Pragian. d2b = Upper Pragian. d2c = Lower Emsian. d2d-3 = Upper Emsian-Eifelian. d4 = Givetian. d5-6 = Frasnian-Famennian. h1-2 = Dinantian. h3-4 = Namurian-Westphalian.

Fig. 214 is the Ardenne cross-section proposed by Raoult & Meilliez (1985, 1987), based mainly on borehole and seismic (ECORS "*Nord de la France*" profile) data. The major Midi overthrust is drawn (from the north to the south) according to the following principles: (1) the fault would outcrop between the Ordovician-Silurian Condroz Inlier and the Lower Devonian northern border of the Dinant Synclinorium (see also Fig. 191 above), (2) thanks to borehole observations, the depth of the Midi Fault (known as the "first reflector") around 5-6 km south of its surface trace occurs at between 2000 and 3000 metres, (3) in the Havelange borehole, the Midi Fault reflector is at about 5-6 km depth (Graulich, 1982), (4) in the Focant borehole, a second deep reflector appears at about 1000-1500 m below the first Midi Fault reflector (Bless et al., 1977), and (5), below the Cambrian Rocroi Massif, in the Ardenne Anticlinorium, the Midi Fault deepens again to the south and occurs at a depth of 8 km while the second reflector is at about 11 km (Fig. 214). Raoult & Meilliez (1985) add that both the first Midi and the second reflector were formed at a greater depth before their subsequent uplifting as a result of the late isostatic readjustment.

Raoult & Meilliez (1985, 1987) also represent the structural organization as shown on the geological map in Fig. 191 and the cross-section in Fig. 214. The northern front of the Variscan Orogen is formed by two regional domains, the autochthonous and parautochthonous domain to the north of the Midi Fault and the allochthonous Dinant Nappe to the south. Each of these domains is composed of a number of major structural units.

The northern domain is formed by the schistose Brabant basement unconformably overlaid by the Silurian-Ordovician Condroz Inlier and by the stratigraphic succession and imbricated thrust sheets located below the Midi Fault between the first Midi and the second deepest reflectors (Fig. 214). The "Namur Synclinorium" is considered to be a large syncline made up of several tectonic stacks thrust on each other and overturned to the north. The Silurian Condroz strip is located below the "Namur Synclinorium" and cannot be considered as an ancient paleogeographical high limiting the Namur and Dinant areas (Michot, 1979, 1980). The Condroz Inlier is viewed as a "sliced" anticline formed by the imbrication of tectonic stacks where the removed cover must be located within tectonic wedges like wedge **A** on Fig. 214. This kind of structural disposition is coherent and is compared, for example, to the Tombe Massif further to the west.

The allochthonous Dinant Nappe is formed in the north by the Dinant Synclinorium, itself made up of three subunits (or "*panneau*" in French): the Godinne, the Dinant and the Aubrives subunits. The two first are separated by the Yvoir Fault described in detailed later in this volume. Briefly, the Dinant Synclinorium is formed by a folded succession that is inclined or overturned to the north and associated with many thrust discontinuities delimiting tectonic wedges thrust over each other. To the south, the Dinant Nappe is composed of the Haute-Ardenne Anticlinorium then the Neufchâteau Synclinorium and the Givonne Massif. Due to the great depth of the Midi Fault (8-9 km), the presence of Precambrian rocks within the Dinant Nappe remains possible. The Mesozoic series of the Paris Basin cover the Paleozoic rocks at the southern extremity and constitute the end point of the Ardenne cross-section.

According to Raoult & Meilliez (1985, 1987), the second deepest reflector bounds an upper series of imbricated thrust sheets (that lie between the two deep seismic reflectors) from a lower Caledonian basement. The Devonian-Carboniferous upper series is folded, faulted and more or less transported along its basement so that it is described as “parautochthonous”. The second deep seismic reflector probably constitutes this décollement level separating the parautochthonous Namur series from the more autochthonous Brabant basement.

Raoult & Meilliez (1985, 1987) make an estimate of the shortening which they consider had been generally underestimated. Shortening is estimated at 30-35% taking into account the major thrusts and folds but without taking into consideration any internal diastrophism and the offset of the Midi Fault (i.e. the net translation vector of the Dinant Nappe). The net translation vector of the Midi Fault is at least 40-50 km and at most 100-120 km (Cazes et al., 1985), possibly 150 km from the work of Raoult & Meilliez (1987). The cross-section in Fig. 214 covers about 90 km. The original (palinspastic) length of the section before the effects of the Variscan shortening is estimated at between 300 and 400 km. This value is obtained after “removing” the folds, the numerous thrusts including the offset of the Midi Fault and the internal deformation. Actually, the internal strain (cleavage, third-order folds) is difficult to account for despite its contribution in the total shortening and explains why the total shortening is thought to be an underestimate.

The pre-deformation length of 300-400 km representing the width of the former Devonian continental slope and margin is quite comparable to the dimensions of modern day continental margins.

Raoult & Meilliez (1987) add that the allochthonous Dinant Nappe is marked by a pervasive cleavage and well-crystallized illite while the absence of cleavage and weak-crystallized illite is characteristic of the parautochthonous “Namur Synclinorium”. The transport of the Dinant Nappe along the Midi Fault is therefore younger than generation of the cleavage in the Dinant Nappe. The authors also propose in 1987 a possible kinematic history (Fig. 215) linked to the formation of structures in the Upper Carboniferous coal-basin to the north of the Epinoy borehole (Fig. 191). Two stages are distinguished (Fig. 215): (1) formation of a large anticline overturned to the north. The dashed lines indicate the positions of future shear planes, ramps and décollement levels; (2) initiation of the major overthrusting (on the Midi Fault) and northward transport of the Dinant Nappe. Erosion may have removed a 2 or 3 km thickness of rocks meaning that the base of the overturned anticline was probably located deeper than today at a depth of probably 6 km.

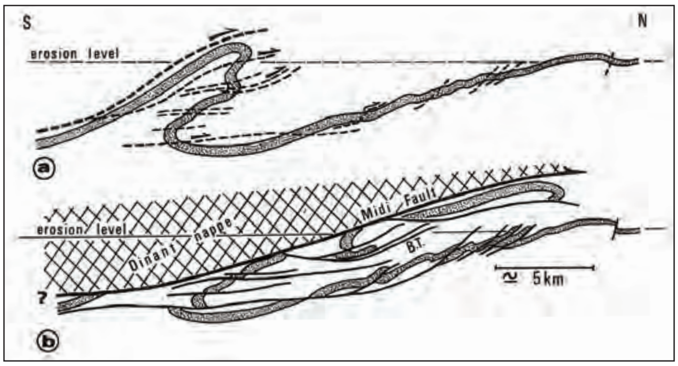


Fig. 215. Theoretical and hypothetical kinematic history for the structures within the Upper Carboniferous coal-basin north of the Epinoy borehole (Raoult & Meilliez, 1987). BT = Barrois Fault.

In 1988, Bouckaert et al. publish the first results of the BELCORP deep seismic campaign. Near Jeumont, they observe a first strong reflector dipping to the south that is interpreted as the Midi overthrust.

In 1989, Bouroz provides new insights regarding the geodynamic framework of the Variscan belt. Contrary to the general ideas of many other geologists, the Variscan structure would not result from a tectonic “push” from south to north but the mechanism and the origin of the deformation should be “sought at depth”. Indeed, the strong deformation of the southern border of the Namur “Synclinorium” located under a weakly deformed Dinant Synclinorium does not argue for a south-to-north directed stress but can only be explained by subduction processes.

Bouroz (1989) gives a new interpretation of the “*Nord de la France*” ECORS seismic profile (Fig. 216). The 40° south-dipping reflectors located between 55 and 80 km to the south of the Epinoy borehole (CFP on Fig. 216) are correlated with the bending downwards at depth of the Namur “Synclinorium” as it is subducted under the Dinant Synclinorium and under the Midi Fault. The “*Grande Faille du Midi*” was initiated during a period of maximal constriction and is considered as a subduction plane under which the Namur “Synclinorium” has plunged. A subsequent extensional stage has generated late-Variscan normal faults along which igneous rocks were emplaced. Indeed, many steeply south-dipping faults with a dip-slip component crosscut and displace the previously continuous surface of the Midi Fault.

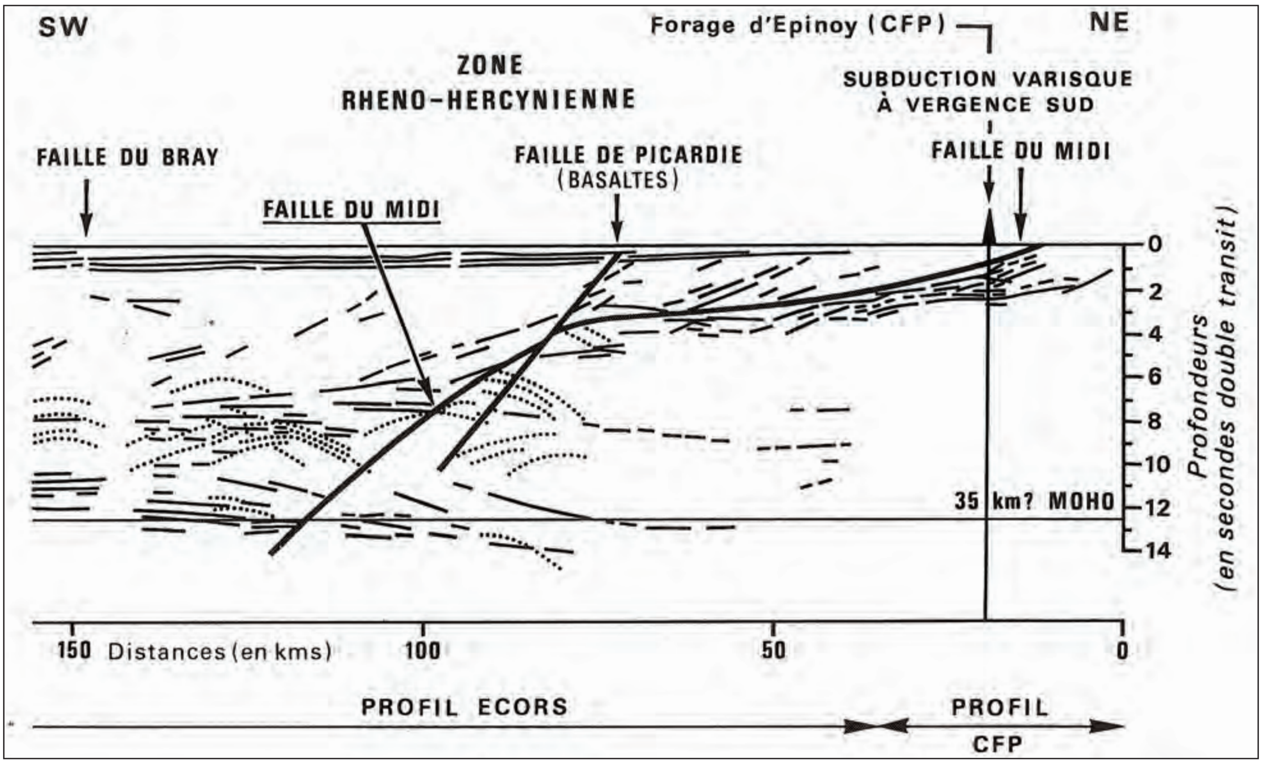


Fig. 216. Reinterpretation of the ECORS seismic profile (Bouroz, 1989). The Brabant domain would not be overthrust and covered by the allochthonous Ardenne domain but would be subducted to the south under the Midi Fault. The Midi Fault is not a thrust but a subduction plane.

Bouroz (1989) suggests that in the Variscan geotectonic framework of the Rhenohercynian fold-and-thrust belt, it is wrong to say that the Ardenne domain overthrust the parautochthonous Brabant unit rather that the Brabant has been subducted to the south under the Ardenne. As a consequence for this type of mountain belt, the Variscan orogen has no roots. Finally, the Variscan belt in France and Belgium includes two symmetric units for which tightening and folding enable a shortening of about 600 km.

According to the shortening model for the Ardenne of Meilliez & Mansy (1990), the movement of the allochthon is controlled in theory by 3 components (Fig. 217): (1) a translation imposed at the back of the nappe that decreases towards the front, (2) a resulting thickening at the back of the allochthon and (3) simple shear of the entire domain that decreases with depth.

Before the work of Raoult & Meilliez (published in 1990), the Ardenne Allochthon is generally considered as organized initially as a single structural domain that simply and passively moves along the Midi fault plane over the Brabant Parautochthon. From 1990, Raoult & Meilliez establish a progressive Variscan deformation marked by the successive positioning from south to north of several major allochthonous thrust stacks within the main Ardenne Allochthon (Fig. 272, Yvoir Fault, section 9.8).

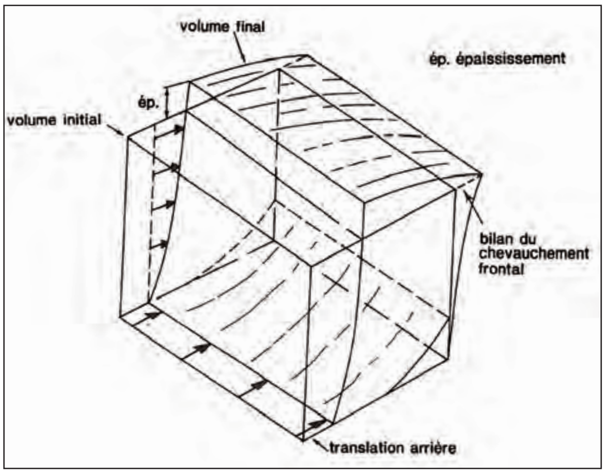


Fig. 217. Theoretical model showing the three components of the Variscan movement of the entire Ardenne Allochthon (Meilliez & Mansy, 1990).

The study also points out that the heterogeneous deformation of the Ardenne is controlled by both lithology (mainly stratified incompetent rocks) and the initial structure of the Paleozoic cover (synsedimentary structures). The total thrust displacement of the Ardenne Allochthon is actually the sum of many displacements along numerous regional and local thrusts within the allochthon, of which the Midi Fault is a part. The total thrust displacement is therefore greater in the southern inner part of the nappe and lesser to the north within the frontal part of the allochthon as the shortening is progressively accommodated by internal strain within the allochthon.

In 1992, Khatir et al. publish a S-N cross-section through the Avesnois area (Fig. 218). Fig. 218c represents the state of deformation during Namurian times and Fig. 218b the current state of the Variscan structures. The Midi Fault is considered to be a gently south-dipping shear plane that crosscuts part of the Brabant basement and its Upper Paleozoic cover during the Variscan orogeny.

The Dinant synform is a large major asymmetric and "sliced" fold made up of imbricated thrust sheets. Thanks to a low-angle south-dipping fault contact (i.e. the Midi Fault), the Dinant synform is overthrust to the north and lies over a basement broken up into tilted blocks. The basement therefore displays a particular structure inherited from a Devonian extensional stage prior to the Variscan compression. The tilted blocks are separated by synsedimentary normal faults bounding sharp variations of sediment thickness and facies.

Kathir et al. (1992) also suggest that the northward transport of the Ardenne Allochthon over the Brabant Parautochthon is made easier along the Coal Measures due to over-pressured fluids related to the transport of the nappe.

The compression within the translated Ardenne Allochthon is accommodated by an internal deformation or shortening that is composed of other thrust faults, shearing, folds and layer-parallel slips. Khatir et al. (1992) believe that their model of foreland shortening is characteristic of a notion of a "deformation sequence". The Variscan contraction progressively reduces to the north but the order in which the thrusts develop (from south to north or north to south) is complicated to deduce.

In 1992, Dejonghe et al. propose structural interpretations of several reflection seismic profiles performed in the Hainaut region in 1979. Fig. 219 shows the NNE-SSW trending "H5" seismic profile established between Mons and Erquelinnes. Three main seismic reflectors are identified on the profile: the P1 and P2 reflectors coincide with particular Dinantian Horizons and the F reflector is correlated with the Midi Fault that transport the folded Eodevonian rocks of the Dinant Synclinorium over the Upper Carboniferous "*Houiller*" rocks of the Namur Synclinorium.

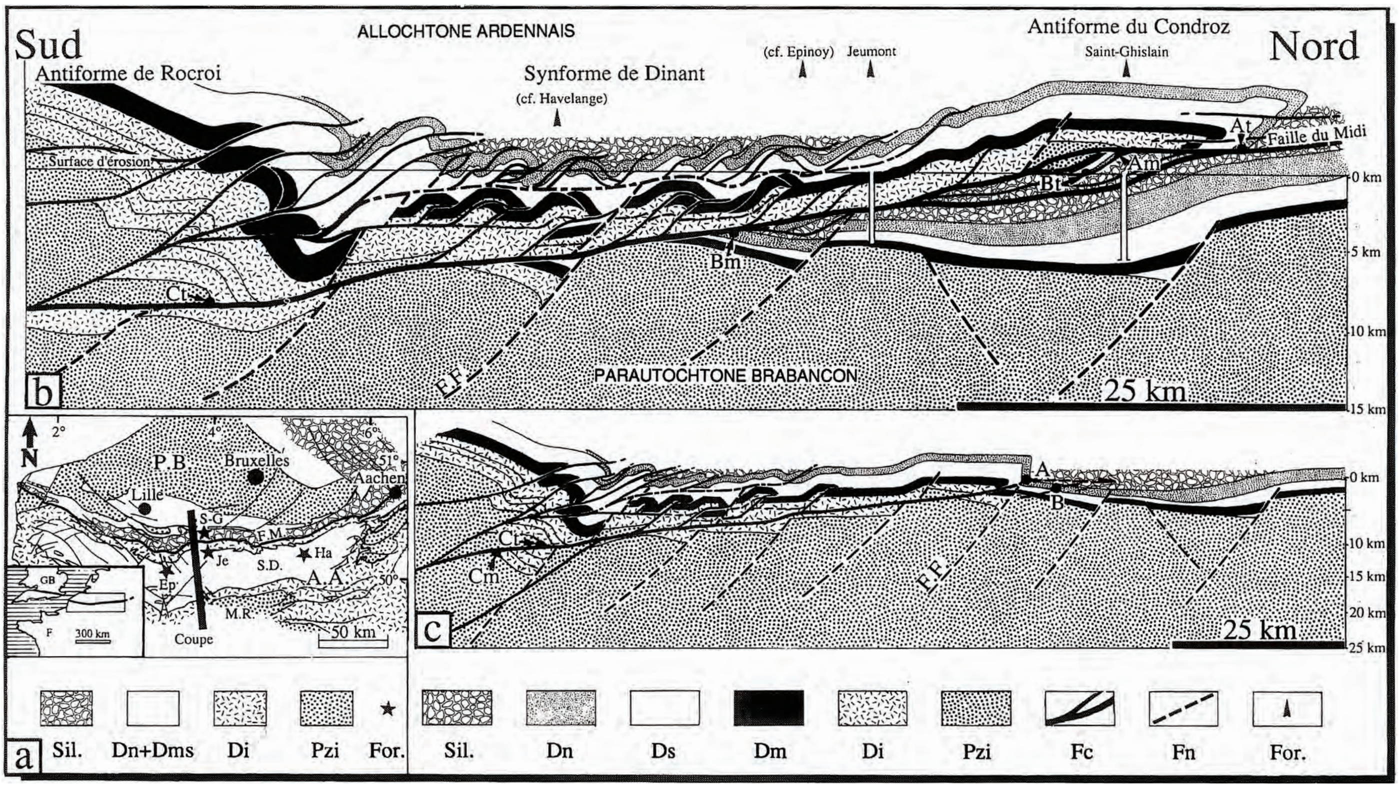


Fig. 218. a. Infra-mesozoic map of Northern France and Belgium. Sil. = Silesian. Dn+Dms = Dinantian, Upper and Middle Devonian. Di = Lower Devonian. Pzi = Lower Paleozoic. For. = boreholes: Je = Jeumont, S-G = Saint-Ghislain, Ep = Epinoy, Ha = Havelange. F.M. = Midi Fault. S.D. = Dinant synform. M.R. = Rocroi Massif. P.B. = Brabant Parautochthon. A.A. = Ardenne Allochthon. b. S-N Avesnois cross-section. Sil. = Silesian. Dn = Dinantian. Ds = Upper Devonian. Dm = Middle Devonian. Di = Lower Devonian. Pzi = Lower Paleozoic. Fc= thrust faults. Fn = normal faults. For. = boreholes. F.M. = Midi Fault. Each black dot (e. g. A) is divided by a fault in two half-dots located on the hanging wall (At) and footwall (Am). A & B = Intersection of Dinantian with a fault. C = Fault cutting the contact between basement roof and cover. c. Avesnois section during Namurian times. From Khatir et al. (1992).

Based on new microtectonic data and a revision of the ECORS deep seismic profile, Le Gall (1992) proposes a new tectonic model applied to the Variscan Front Zone. Le Gall suggests that the Rhenohercynian fold-and-thrust belt developed on a previously stretched Devonian continental margin deformed by multiple south-dipping extensional faults. Considering a palinspastic width of 120 km for the Namur-Neufchâteau basin and a timing of the Variscan crustal shortening operating from the Lower Carboniferous until at least the late Westphalian, Le Gall (1992) indicates a forward propagation rate of about 3 cm/y.

Le Gall (1992) does not observe any Caledonian deformation structures in the Cambrian rocks within the southern part of the Rocroi Massif. He therefore positions the Caledonian orogenic front within the Rocroi Massif.

From an interpretation the ECORS seismic profile, the Rocroi Massif is located about 40 km to the north of the basement footwall ramp from which it was translated to the north (Fig. 220). This suggests, therefore, a 40 km northward tectonic transport of the Lower Paleozoic basement wedge along the Dinant shallow décollement level. This major thrust is moreover consistent with a strata shortening of about 40% for the front of the Devonian detached cover.

Consequently, the Cambrian Rocroi Inlier is viewed as a "far-travelled basement wedge" just like the Stavelot-Venn Massif for which a similar tectonic model can be proposed. Le Gall (1992) correlates the "arcuate Rocroi-Libramont-Stavelot basement-cored zone" with the northern limit of a large Lower Paleozoic allochthonous basement unit made up of thrust sheets that have roots further to the south along a Devonian normal fault (Fig. 220).

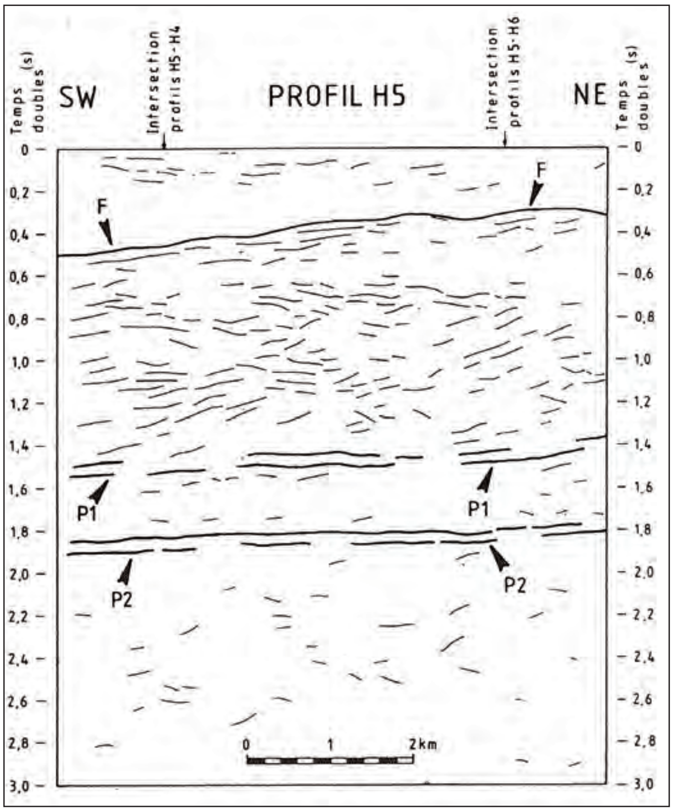


Fig. 219. "H5" reflection seismic profile (Dejonghe et al., 1992). F = Midi Fault. See the text above for a description.

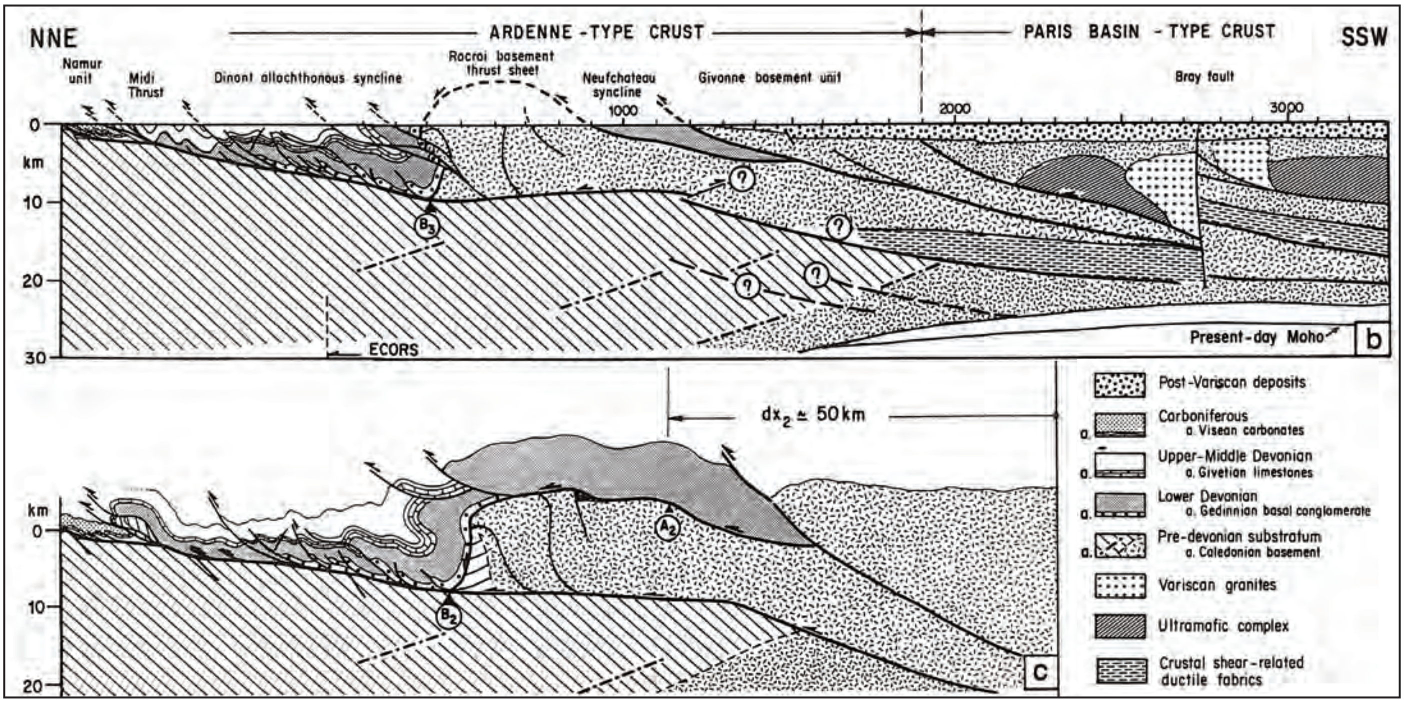


Fig. 220. NNE-SSW cross-sections showing the deep structure of the Variscan Rhenohercynian fold-and-thrust belt (Le Gall, 1992).

Fig. 221 from Le Gall (1992) shows the tectonic model and sections of the Variscan frontal thrust derived from the following observations:

- the Namur Syncline is sharply truncated to the south by the Condroz Inlier that is located in the immediate hanging wall of the Midi Thrust;
- between the thrust emergence line and the Wépion borehole that intersects the Midi Fault at a depth of 450 metres, the Midi Thrust has a moderate S dip of about 40°;
- the top of the Brabant Caledonian basement probably displays a staircase geometry below the Upper Carboniferous Namur coal-basin and probably with reactivation during the Variscan shortening as a N-directed shear plane.

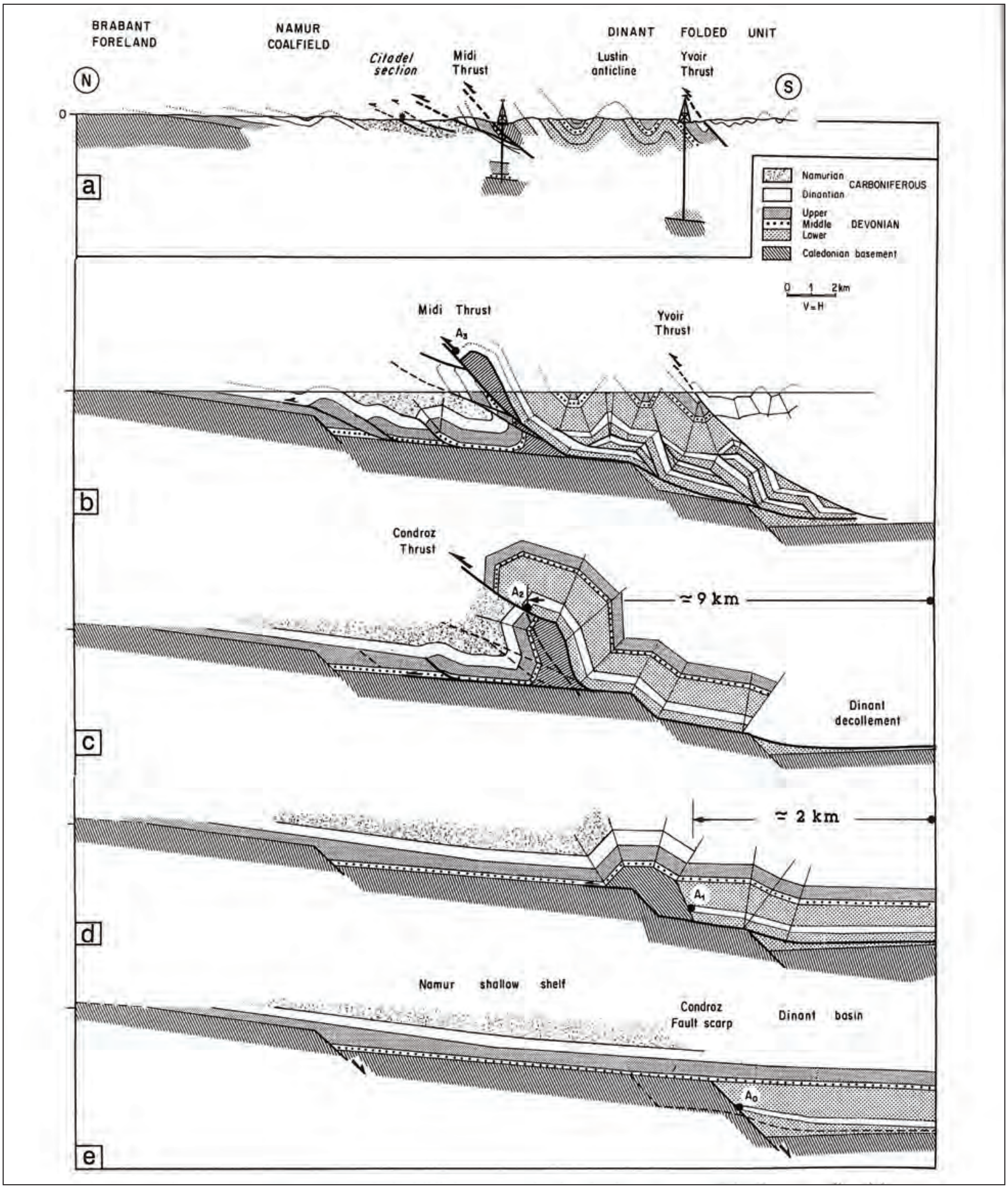


Fig. 221. N-S sections through the Variscan Midi frontal zone (Le Gall, 1992). Geometrical data (a, Graulich, 1961) were used to construct the balanced (b) and restored (e) sections.

For the first stage, Fig. 221e shows a palinspatic reconstruction of the Rhenohercynian continental margin. The "Namur shallow shelf" probably extended 8 km to the south of the present Midi Thrust emergence line and was separated from the "Dinant basin" further to the south by the synsedimentary normal fault of the "Condroz fault scarp".

The narrow Ordovician-Silurian Condroz Inlier, here called the "Condroz thrust sheet", would have been detached in depth from the basement during the rapid ramping of the basal Dinant décollement (Fig. 221d) then incorporated within the frontal thrust and transported towards the surface (Fig. 221c). The Condroz basement wedge would have been transported for 6-7 km to the north. During the shortening, the whole hanging wall was uplifted and translated to the north along the basal thrust (the "Dinant décollement", Fig. 221c). The "Condroz fault scarp" was reactivated in the opposite direction in the "Condroz Thrust" which enables the overthrust of the "Dinant clastic wedge" (the Ardenne Allochthon) over the "Namur coalfield" (the Brabant Parautochthon, Fig 221c). The Midi Thrust is viewed as an out-of-sequence minor thrust reactivated from the rotated pre-existing basement-cover surface during the final compression. Le Gall (1992) adds that despite the generally prominent character of the Midi Fault on the Rhenohercynian fold-and-thrust belt tectonic map, the Midi Thrust is probably responsible for only a minor uplift of the pre-existing major Condroz Thrust. The Variscan shortening of the "Namur shallow shelf" is estimated to be about 50% (an original width of 18 km reduced to 8 km).

Briefly, according to Le Gall (1992), a distinction is made between the major Condroz Thrust and the minor Midi Thrust. The Condroz Thrust, located between the Condroz Inlier and the Dinant Nappe, would coincide with the Dinant décollement and the regional Variscan tectonic transport of the entire allochthonous area. The Midi Thrust would only result from minor reverse movement along the basement-cover contact between the Brabant Caledonian basement and the Condroz Inlier. In other words, the Midi Fault is not the emergence line of the Dinant décollement along which the Ardenne Allochthon is translated.

In 1992, Fielitz establishes cross-sections through the Cambrian-Ordovician Stavelot-Venn Inlier.

The geometry of the thrust nappe in this frontal part of the Rhenohercynian fold-and-thrust belt and the established NNE deflection of the late tectonic transport direction are related to the proximity of the Brabant Massif to the north. The Brabant area actually displays a cratonic behaviour and acted as an obstacle to the regional northward progression. The propagation of the Variscan deformation was therefore halted to the west but was effective to the east beyond the eastern termination of the rigid Brabant Massif basement. Actually, the western blocked domain is separated from the eastern domain (which has advanced further to the north) by a "transfer zone" to which the early- to syn-orogenic and sinistral Monschau shear zone belongs (see also the Xhoris Fault in Cambier & Dejonghe, 2010).

Fielitz (1992) also proposes a section through the Stavelot-Venn Anticlinorium and through the Theux Window (Fig. 222). Four main faults constitute the Variscan Front Thrust in the vicinity east of Liège. The separation between the Rhenohercynian autochthonous and allochthonous domains is not therefore a sharp boundary but a transition zone in which several thrusts are distributed. The most northern and frontal thrust is here called the "Aachen-Midi Thrust". Three other discontinuities, the Eilendorf-Soiron Thrust, the Venn Thrust and the Monschau Shear Zone-Xhoris Thrust probably join at depth with the main basal décollement level of the Midi Fault.

In 1994, Dittmar et al. establish a partitioning of the deformation within the Rhenohercynian fold-and-thrust belt. The shortening of about 16 to 27 % in the northern Rhenish Massif increases in the internal zones to about 51%. The net orogenic shortening within the upper crust of the fold-and-thrust belt is about 42%.

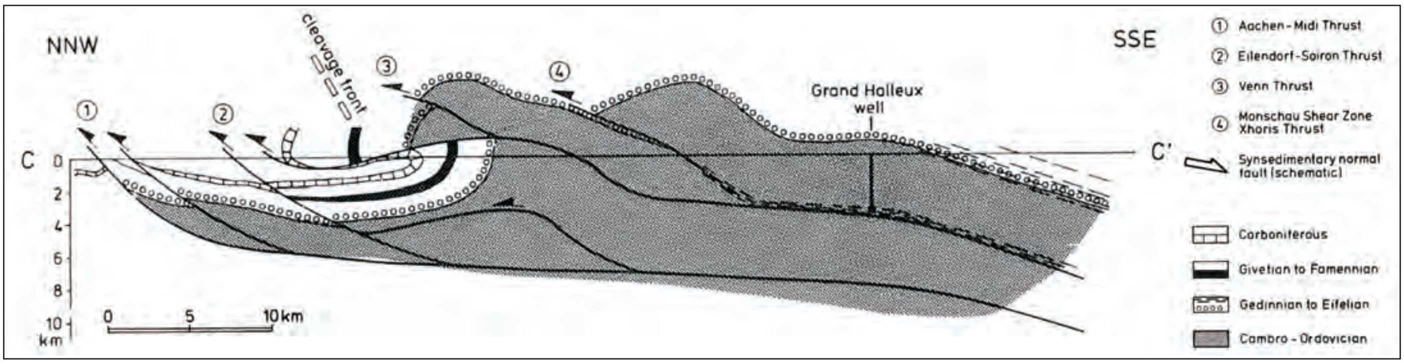


Fig. 222. NNW-SSE deep section across the Stavelot-Venn Inlier and the Theux Window (Fielitz, 1992).

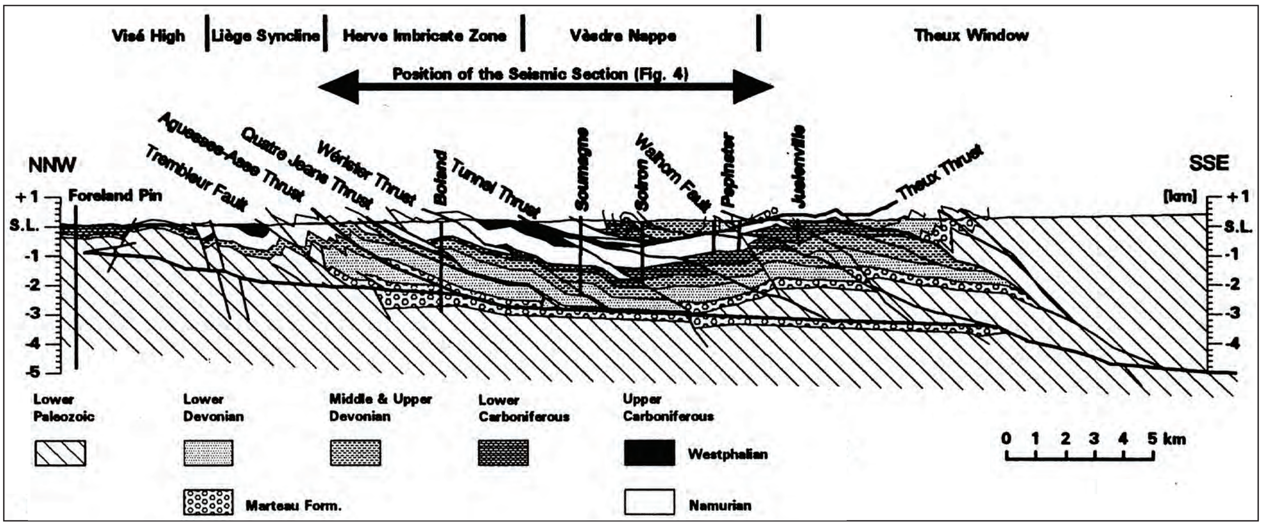


Fig. 223. N-S cross-section through the northwestern foreland of the Stavelot-Venn Massif (Hollmann & Walter, 1995).

As previously indicated, Hollmann & Walter (1995) consider the Aguesses-Asse Fault as a segment of the Midi-Aachen Thrust that subdivides into multiple thrust branches in the vicinity of Liège. The northern front of the Variscan Rhenohercynian fold-and-thrust belt between Liège and Aachen is marked by major in-sequence low-angle thrust faults parallel to structure and stratigraphy; out-of-sequence thrusts would be very rare.

The allochthonous complex (including the Theux Window, Vesdre Nappe and Herve Imbricate Zone) is separated from the parautochthonous units to the north (Liège Syncline and Visé High) by the Midi-Aguesses-Asse Thrust (cross section in Fig. 223). The section also shows that the Herve Imbricate Zone and the Theux Window combine to form a major thrust complex transported along the Midi-Aguesses-Asse detachment while the Vesdre Nappe between is the highest thrust sheet that represents the first event in the sequence of thrust movements.

The Aguesses-Asse Thrust therefore acted as a foreland detachment that typically characterizes the thin-skinned tectonics of the Variscan deformation front. The basal detachment cross-cuts the Lower Devonian terrain, propagates within the Lower Paleozoic Caledonian basement of the Brabant Massif and probably dies out above a blind thrust system in the Visé High (Fig. 223).

The horizontal displacement of the southern part of the Vesdre Nappe (i.e. originally the southernmost thrust sheets) is about 27 km while the northern part of the nappe is only transported for 18 km towards the foreland to the north. Actually, the sole thrust of the Vesdre Nappe is the Theux-Tunnel Thrust that displays a relative offset of 10.4 km. Many other internal thrust faults contribute to the total displacement and these branch off from the main deep Midi-Aguesses-Asse detachment. The horizontal tectonic transport distance of the Herve Imbricate Zone – Theux Window complex increases from 3.4 km at the northern limit of the Aguesses-Asse Thrust hanging wall to 17.7 km for the thrust sheets of the Theux Window to the south. Hollman & Walter (1995) estimate a total Variscan shortening of nearly 50% applying to the tectonic unit above the foreland detachment of the Midi Fault at a depth of 3-3.5 km.

In 1997, Mansy et al. publish an interpretation of the M146 seismic profile made at the longitude of Valenciennes in 1981 (Fig. 224). The main objectives of this study were an understanding of the formation and structure of the "Namur Synclinorium" and the kinematic history of the Ardenne Allochthon. The seismic profile (given below in Fig. 226) shows two continuous reflectors: a first upper reflector corresponding to the Midi Fault that emerges at surface near Valenciennes, and a second deeper seismic reflector probably corresponding to a particular seismic response of the Givetian and Frasnian rocks (Raoult, 1988) and therefore to the contrast of speeds between the siliclastic Givetian rocks and the carbonate Frasnian rocks. Both reflectors join at depth to the south (funnel-shaped); the average vertical separation between the two reflectors is about 700 m.

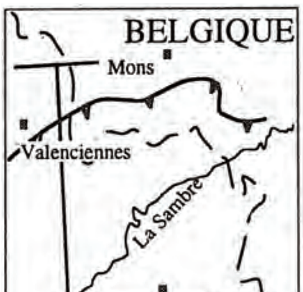


Fig. 224. Location of the M146 seismic profile (Mansy et al., 1997).

Structural interpretations of the M146 seismic profile (Fig. 226) are given in Fig. 225 (Mansy et al., 1997). The main results are (1) the flexion of the Lower Paleozoic basement under the "*Houiller*" basin and the main overthrust front (correlated with a Bouger anomaly), (2) the clear continuity of multiple seismic reflectors, (3) the presence of Silurian rocks within the overturned tectonic wedges between the Midi and the Barrois faults implying a deep origin for these wedges and a minimal thrust displacement of 80 km, and (4) the presence of Lower Paleozoic fragments within the Ardenne Allochthon removed from its basement during the main transport phase.

Mansy et al. (1997) also distinguish two units within the Upper Carboniferous ("*Houiller*") coal-basin: a southern unit within the hanging wall of the Barrois Fault overthrust onto a northern unit within the footwall of the Barrois Fault that is relatively less affected by the Variscan deformation. The concept of the "Namur Synclinorium" becomes obsolete as the two "limbs" of the "syncline" structure do not belong to the same entity. Moreover, the sedimentary facies within the overturned tectonic wedges between the Barrois and the Midi faults are more similar to the facies of the northern border of the Ardenne Allochthon than to those of the coal-basin to the north.

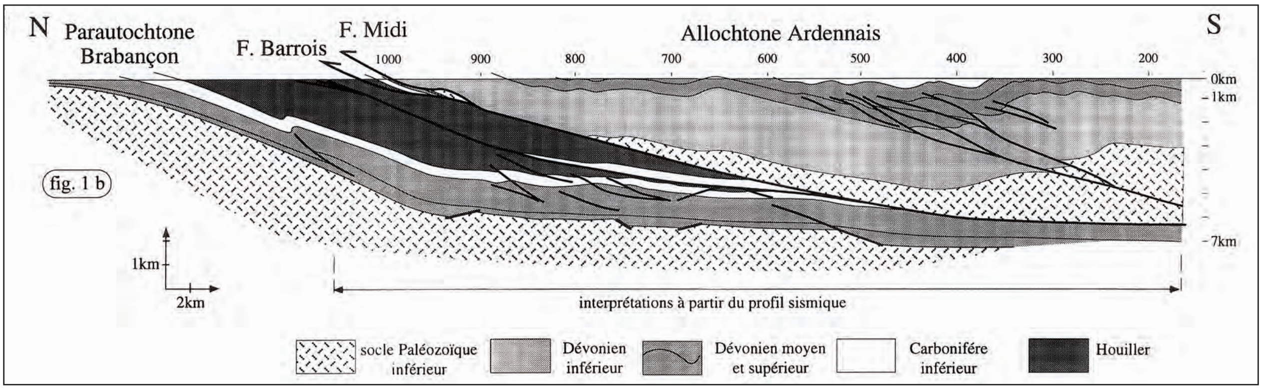


Fig. 225. N-S section of the Ardenne Allochthon and the Brabant Parautochthon based on the M146 seismic profile (Fig. 226) according to Mansy et al. (1997).

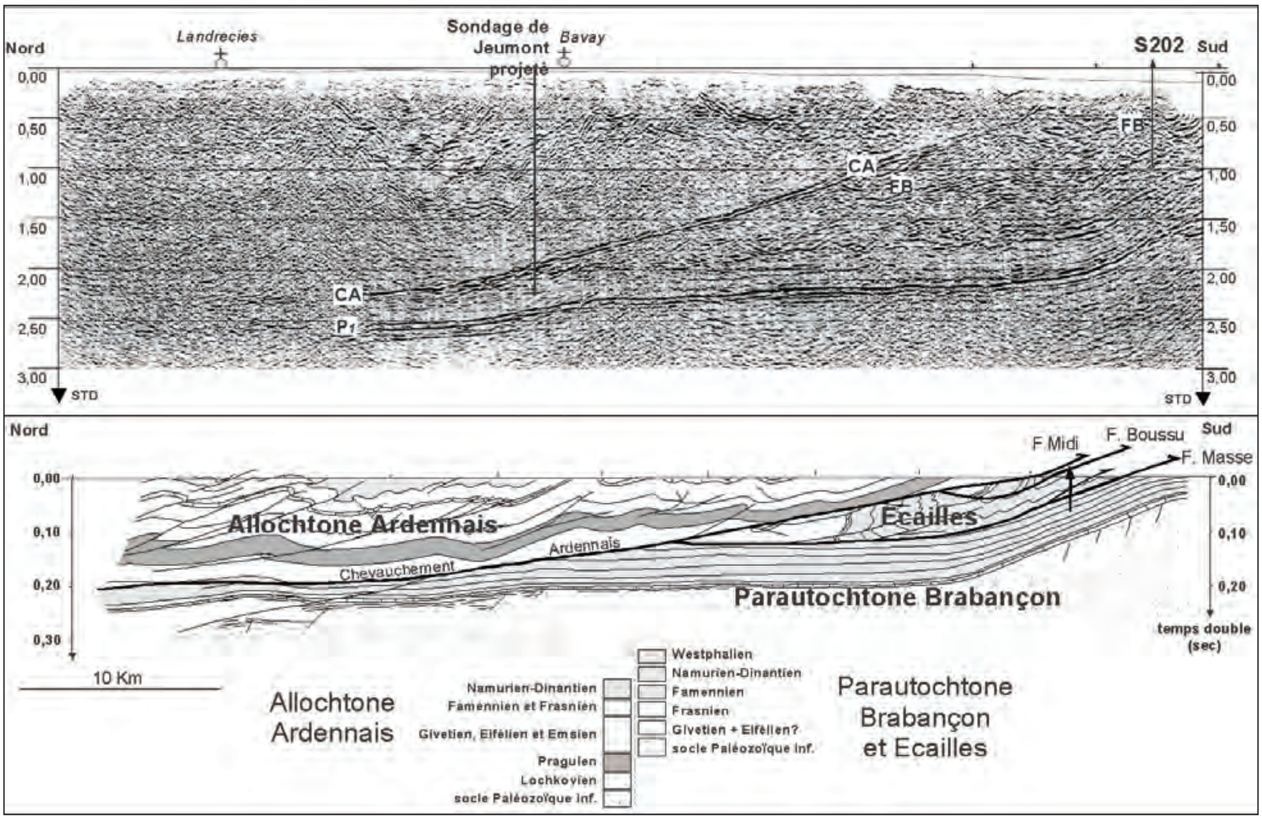


Fig. 226. M146 seismic profile and geological interpretations (Mansy & Lacquement, 2006). Two major seismic reflectors are recognized: an upper one coinciding to the décollement level of the allochthonous domain (*"CA"* for *"Chevauchement Ardennais"* or Midi Fault) and a deeper one (*"P1"*) corresponding to the Givetian and Frasnian terrains. FB = Boussu Fault. On both sections, North and South indications are shown incorrectly and must be reversed.

Based on a revised interpretation of the M146 seismic profile, Lacquement et al. (1999) propose new ideas regarding the structural geometry and total transport at the Variscan front zone. The seismic profile and structural interpretations are given in Fig. 226.

A major reflector is correlated to the Masse Fault that separates two structural units: a first southern upper allochthonous, faulted and folded unit and a second northern lower autochthonous less deformed and gently (15°) south-dipping unit. The lower unit corresponds to the Brabant Parautochthon and the upper unit to a "sliced" and faulted synform structure made up of imbricated thrust sheets (Fig. 226). Between the Masse and the Midi faults, another reflector is correlated with the Boussu Fault for which the hanging wall is the overturned Boussu "Massif" partly of Lower Paleozoic age. In other words, according to Lacquement et al. (1999), the Upper Carboniferous ("*Houiller*") coal-basin is formed from two distinct tectonic units: the Brabant Parautochthon to the north and the "parautochthon intermediate thrust sheets" (noted "*Ecailles*" in Fig. 226) to the south therefore making a transition zone with the Ardenne Allochthon further to the south. The Masse Fault represents the basal décollement level while the Boussu Fault supports overturned and "sliced" thrust sheets. This structural geometry is similar to that observed in the French Pas-de-Calais coal-basin where the Barrois Fault performs the same role as that of the Boussu Fault (Mansy et al., 1997).

An attempt to estimate the total transport of the allochthonous unit is given next by Lacquement et al. (1999). The "parautochthonous intermediate thrust sheets" in the hanging wall of the reverse Masse Fault contain Lower Paleozoic terrains which must have had a distant southward origin. The minimal apparent translation vector for the Masse Fault is therefore estimated at more than 50 km. This significant thrust located in the middle part of the coal-basin argues for the invalidity of the concept of the "Namur Synclinorium". Indeed, the two "limbs" no longer belong to the same tectonic unit but come from the amalgamation of two initially distant series. Regarding the Midi Fault, a minimal apparent thrust displacement of 20 km is envisaged, thus giving a minimum overthrust of 70 km at the front of the parautochthonous intermediate thrust sheets. The Midi Fault is therefore not considered as the main major overthrust anymore but as the last out-of-sequence thrust.

Lacquement et al. (1999) propose a new kinematic evolution of the Variscan front (Fig. 227). Thanks to the flexure of the Brabant Parautochthon (Fig. 226), the "sliced" and folded tectonic wedges progressively accumulated at the foot of the flexure and were subsequently cross-cut by the south-dipping Midi Fault. Further to the south, the top of the deep parautochthonous unit is quite horizontal meaning that the flexure observed to the north may correspond to a faulted zone. These former faults probably delimited a major crustal block formed during a Givetian extensional tectonic regime. The normal offset would have not been compensated for during the Variscan shortening but would have been amplified due to the overload induced by the Ardenne Allochthon being transported to the north. Fig. 227 shows the progressive translation of the "intermediate thrust sheets" and the allochthonous unit to the north. Indications regarding the km-scale displacement are given.

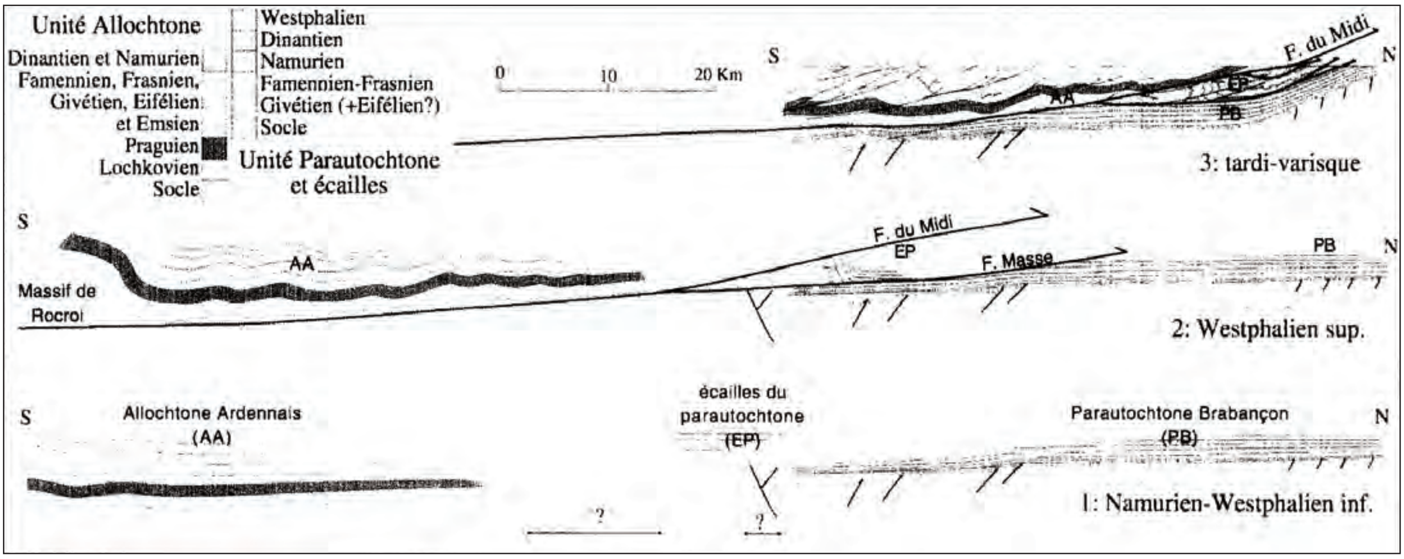


Fig. 227. Tectonic model showing the relationships between the thrust sheets and the formation of the allochthonous unit (Lacquement et al., 1999).

The note of Hance et al. (1999) presents a Variscan deformation history in northeastern Belgium. Two complementary models are envisaged: an in-sequence thrusting model and a preferred out-of-sequence thrusting model (Fig. 228) that is described below. Four phases are distinguished:

- phase 1: responding to the first stages of the Variscan shortening, folding of the terrains occurred. This model highlights the importance of the overturned forelimb of a large anticline;
- phase 2: subsequently, as a result of continued Variscan contraction, the large anticline would have been truncated and transported to the north. The faults numbered 1, 2 and 3 in Fig. 228, phase 2, could have occurred in any order;
- phase 3 & 4: a major out-of-sequence fault (fault number 4), known as the Theux-Tunnel Fault (a segment of the Midi-Aachen Thrust), modifies and crosscuts the previous structural framework.

The structural framework of the Rhenohercynian fold-and-thrust belt front zone is set out by Hance et al. (1999) based on the connective function of the Tunnel Fault between the Midi Fault to the west and the Aachen Fault to the east. In northern France and in western Belgium, the Midi-Aachen fault zone separates a major nappe, the Ardenne Allochthon, to the south, from the Brabant foreland resting on the Brabant Massif to the north (Fig. 229). In eastern Belgium, the Liège and Herve units are the eastward continuation of the Brabant foreland and the Vesdre Nappe is the northeastern continuation of the leading edge of the Ardenne Allochthon. The Theux Window is a complex structure made up of imbrications.

Following the lead of previous papers, Hance et al. (1999) considers that the 40 km northward thrust of the allochthon along the French part of the Midi Fault decreases eastward to about 10 km. Actually, the reduction of the offset in eastern Belgium is explained by both the Brabant Massif acting as an obstacle and by the distribution of movements along multiple faults (e.g. Xhoris and Eupen faults). Fig. 229 shows a comparison of the Variscan Front Zone in eastern Belgium and in northern France.

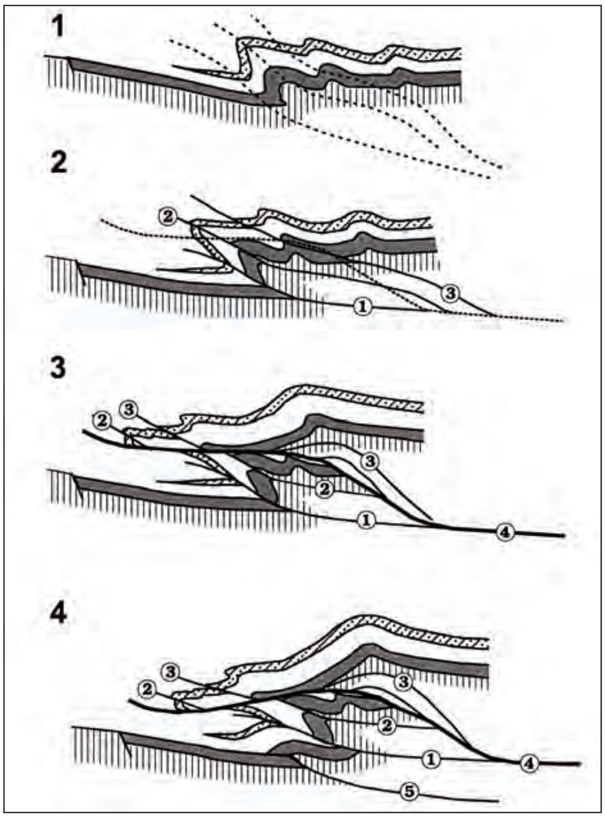


Fig. 228. Variscan shortening in northeastern Belgium according to the out-of-sequence fault propagation model (Hance et al., 1999). Shaded layers = Lower Devonian and Dinantian. Dotted lines = trajectories of future faults. See the text above for a detailed description.

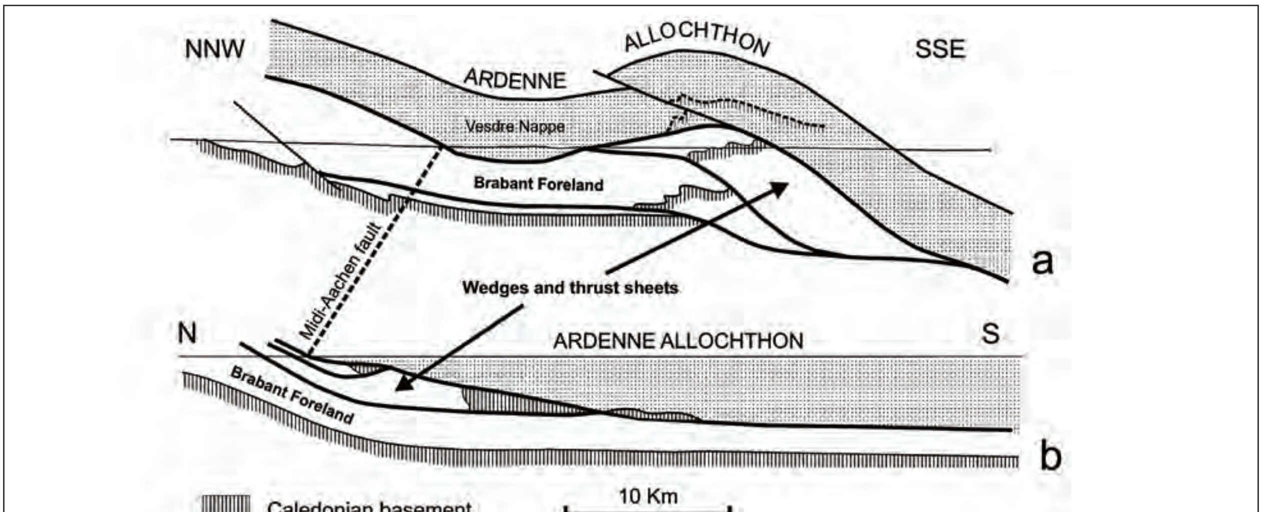


Fig. 229. The Variscan Front Zone in eastern Belgium (a) and in northern France (b) (Hance et al., 1999). See the text above for explanations.

Based on revisions of earlier Meuse cross-sections (the Rheno-hercynian fold-and-thrust belt sections of, for example, Raoult & Meilliez, 1985, 1987; Le Gall, 1992; Meilliez et al., 1991), Adams & Vandenberghe (1999) propose a new structural profile (Fig. 230). The authors actually produce a "*reconstruction of the predeformational sedimentary basin wedge geometry*". This study is based on geostatistical analysis of the thickness distribution of the stratigraphic units. It has finally shown that the profile is marked by a single, major detachment plane coinciding with the Midi-Condroz Thrust. The calculated offset of the Midi Fault is between 20 and 30 kilometres.

The principal new elements of the profile (Fig. 230) are: (1) the entire Meuse section is subdivided into three geologically homogeneous domains, from north to south respectively the Namur, Yvoir and Dinant units, (2) the old concept of the "Namur Syncline" is reassessed as an overturned and overthrust foreland, (3) the major detachment plane to the south of the Yvoir Unit is located exclusively in pre-Devonian basement at a depth of 10 km, (4) the presence of north-dipping faults are interpreted as backthrusts, and (5) the introduction of a shallow detachment plane within the Dinant Unit is justified by the tectonic imbrication observed in deep wells.

Former versions of the Meuse profile require either a major synsedimentary fault or a hidden Variscan "massif" under the detachment plane to be balanced. The latest proposed cross-section does not need such structural elements and is therefore less complicated than previous versions.

Based on Bouger anomalies, gravity gradients, aeromagnetic structural lineaments and mapping data, Mansy et al. (1999) propose new insights on the origin of the complex structures of the Variscan fold belt. The authors indicate that the particular structural trends of the Variscan Orogen in Belgium and northern France result from: (1) an oblique convergence between the Ardenne Allochthon and the Brabant area, producing variable offsets from east to west, (2) sediment thickness variations where thicker sediments initiate thrusts closer to the foreland, and (3) the Brabant Massif basement that acts as a rigid crustal block and therefore an obstacle to the general transport of the Ardenne Allochthon to the N-NW with decreasing influence to the SW.

According to Oncken et al. (1999), the Midi Fault, or "Aachen-Midi detachment" acts as a basal décollement of a thin- to thick-skinned orogenic wedge (Fig. 231) during the Variscan shortening. The Lower Devonian paleogeographic framework prior to the shortening includes: (1) the Lizzard-Giessen-Ostharz narrow oceanic basin bordered to the north by a 350 km wide passive continental margin, and (2) the prevalence of extensional tectonics with the development of a symmetric (failed) rift that includes the Eifel basin and the Mosel graben (Fig. 231).

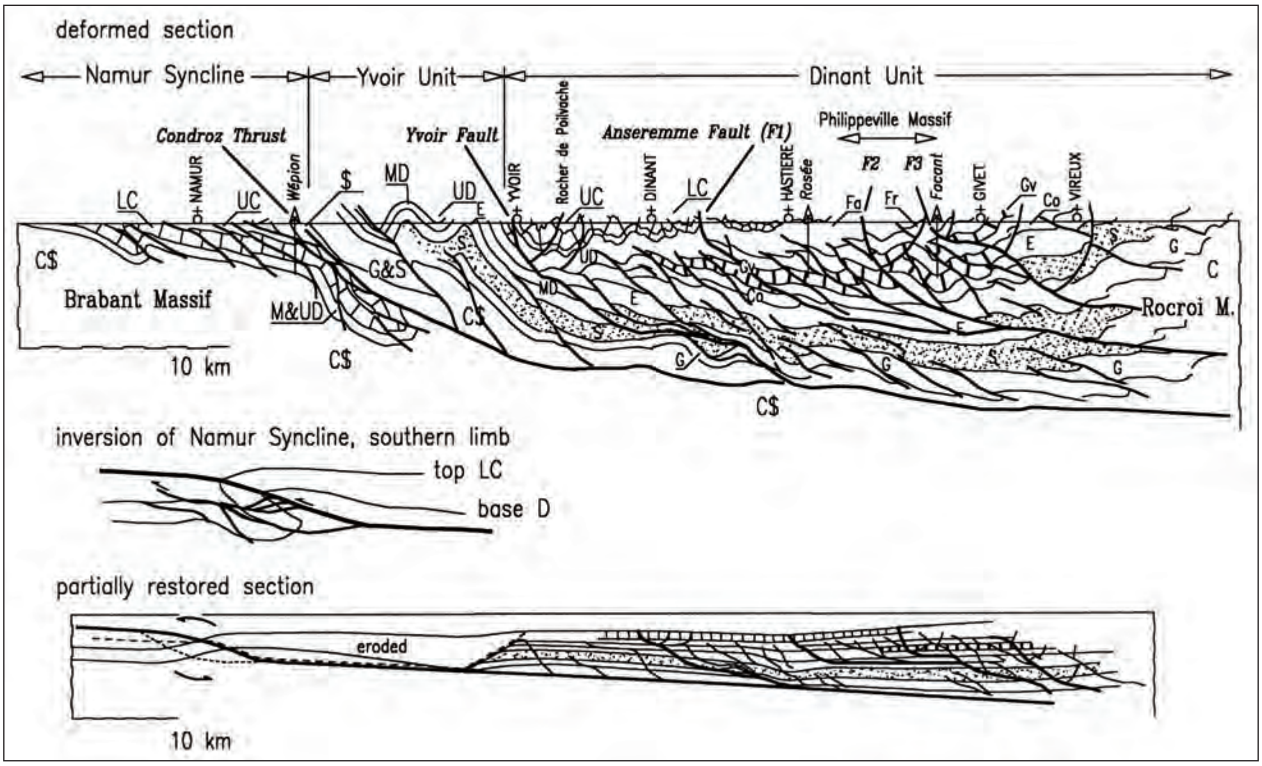


Fig. 230. N-S cross-section through the Variscan thrust front in the Meuse valley (Adams & Vandenberghe, 1999). The restored section is also given. The small section shows the inversion of the southern limb of the "Namur Synclinorium". C = Cambrian; $ = Silurian; D = Devonian; G = Gedinnian; S = Siegenian; E = Emsian; MD = Middle Devonian; Co = Couvenian; Gv = Givetian; UD = Upper Devonian; Fr = Frasnian; Fa = Famennian; LC = Lower Carboniferous; UC = Upper Carboniferous.

The Variscan deformation begins in the late Lower Carboniferous with continental subduction and continues until the late Upper Carboniferous. The highly segmented (rifted) passive continental margin is then accreted and the sedimentary cover of the margin is detached. The crust above the basal detachment is shortened by 52% (180 km). The shortening is made by folding and by the superposition of imbricated systems with an average frontal offset of 10-20 km. Actually, the limited shortening of 10-30% at the northern deformation front increases to the south to about 60-70% within the southern inner margin of the belt.

The Rhenohercynian lower plate probably encountered a large lithospheric flexural bending in response to the tectonic load resulting from the movement and advancement of the Saxothuringian upper plate wedge (Fig. 231). The lithospheric bending was responsible for a "ductile failure" that probably initiated the Aachen-Midi detachment. The décollement level therefore propagated progressively into the passive margin under the load of the advancing upper plate (Oncken et al., 1999).

The basal Aachen-Midi detachment undercuts all the basin fill including the rifted structures. The detachment was probably controlled by the softening of quartz which is dependent on temperature. Indeed, the décollement level probably traces a 300-400°C paleoisotherm that coincides with the softening of quartz and that separates a brittle upper crust from a ductile midcrustal layer (Fig. 231).

Based on paleomagnetic data, Márton et al. (2000) try to explain the origin of the arcuate shape of the northern front of the Variscan Orogen. Indeed, the Variscan front zone typically displays major fold and fault axes varying from N110°E in the Boulonnais (France) to N70°E in the Ardenne. Considering that the Midi Fault is an out-of-sequence thrust cutting pre-existing folds and faults (even of Upper Carboniferous age) and that the offset varies greatly along the arc (about 20 km in the Boulonnais, 80 km in the Hainaut, a few km near Aachen), Márton et al. (2000) propose 3 models to explain the arcuate shape of the Variscan front (Fig. 232).

- The first model ("oroclinal bending", Fig. 232a) constitutes the collision of an originally linear Ardenne Allochthon and a wedge-shaped Brabant margin. This margin acted as an obstacle on which the linear Ardenne Allochthon is progressively moulded. Actually, this model is inconsistent with the palaeomagnetic data which show no significant rotation of the western (clockwise) and eastern (counter clockwise) branches of the Variscan front.
- The second model ("inherited shapes", Fig. 232b) includes the collision of an originally arcuate allochthon with an autochthon showing a similar arcuate shape. The model does not imply rotations in either branch of the Variscan front and therefore matches better with the lack of major palaeomagnetic rotations.
- The third model (Fig. 232c) also includes an originally arcuate Ardenne Allochthon that, during the progressive collision, shows a decreasing curvature. This implies clockwise and counter clockwise movements in the eastern and western branches respectively. The palaeomagnetic data are fully consistent with the clockwise rotation of the eastern Ardenne branch of the Variscan front. This model is also supported by the increasing offset established along the Midi Fault between the eastern and central segments.

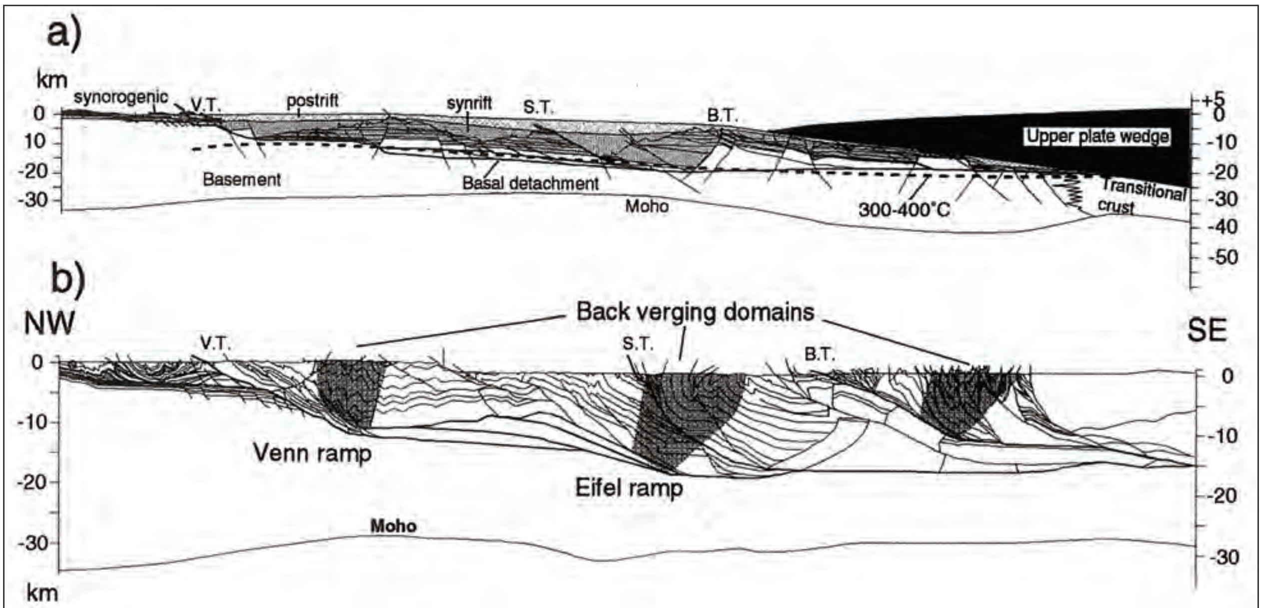


Fig. 231. (a) Reconstruction (N-S section) of the former Rhenohercynian passive margin (Oncken et al., 1999). The flexural bending relating to the advancing upper plate, the trajectory of the Aachen-Midi detachment and the trace of the 300-400°C isotherm (dashed line) are represented. (b) N-S cross-section across the Rhenohercynian fold-and-thrust belt. The areas with out-of-sequence thickening (shaded) and the major ramps are shown. VT = Venn thrust. ST = Siegen thrust. BT = Boppard thrust.

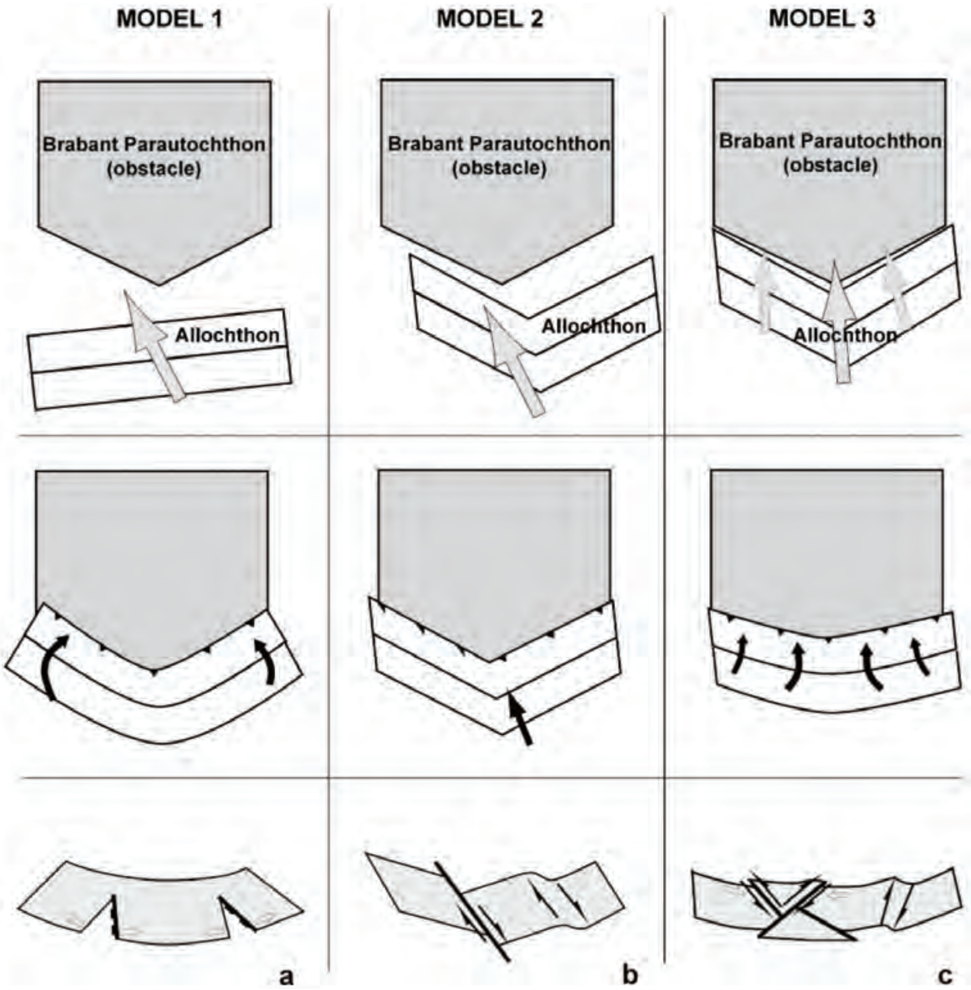


Fig. 232. Hypothetical geodynamic models explaining the curved shape of the Variscan front (Márton et al., 2000). See the text for descriptions.

As indicated in the descriptive data sheet of the Tombe Fault, Delmer (1997, 2004) believes that the allochthonous "*Grand Massif Superficiel*" in the footwall of the Midi Fault has not been transported and overthrust from the south (like the Ardenne Allochthon) but was initially situated over the Brabant Massif and glided by gravity towards the south (see Fig. 256, Tombe Fault, section 9.6). Dissolution of deep evaporites would be the cause for this movement. Based on a northern origin of the "*Grand Massif Superficiel*", Delmer (2004) suggests that the Condroz Inlier has no individual structural significance as it belongs entirely to the Ormont segment of the "*Grand Massif Superficiel*"

In 2004, Averbuch et al. present a deformation history for the Boulonnais part of the Variscan Thrust Front. Briefly, the Variscan shortening of the southern margin of the Old Red Sandstone Continent (i.e. the Ardenne domain) is made by a "slicing" or an in-sequence imbricated thrust sheet which finally butts against the Brabant Massif to the north (Fig. 233). Actually, the Brabant Massif acted as a major "buttress" that hindered the northward progression of the tectonic wedges and that induced the subsequent out-of-sequence deformation of the Variscan thrust front. The progressive accumulation of imbricated tectonic wedges induced a major flexure of the buried autochthonous domain that occurred at the northern extremity of the Upper Carboniferous ("*Houiller*") coal-basin (see the "locking point" on Fig. 233d). Moreover, at the foot of the sharp flexure, the reactivation of a Caledonian north-dipping thrust created a southerly transported wedge. This thrust located in the Caledonian basement was in the opposite direction to the main Variscan NNE-directed stresses and has also hindered the regional northward progression (Fig. 233c).

Averbuch et al. (2004) consider the Midi Thrust Zone (MTZ) as the emerging part of the major basal thrust underlying the allochthonous Variscan wedge and propose a significant out-of-sequence initiation of the MTZ. Compared to the Variscan front in the Ardenne, the Boulonnais thrust belt displays a shorter allochthonous displacement along the "Midi Thrust Zone" and a greater degree of truncation of the MTZ footwall thrust sheets. Taking into consideration the folding and thrusting, a shortening of 52% is estimated for the external part of the former northern Rheno-hercynian margin.

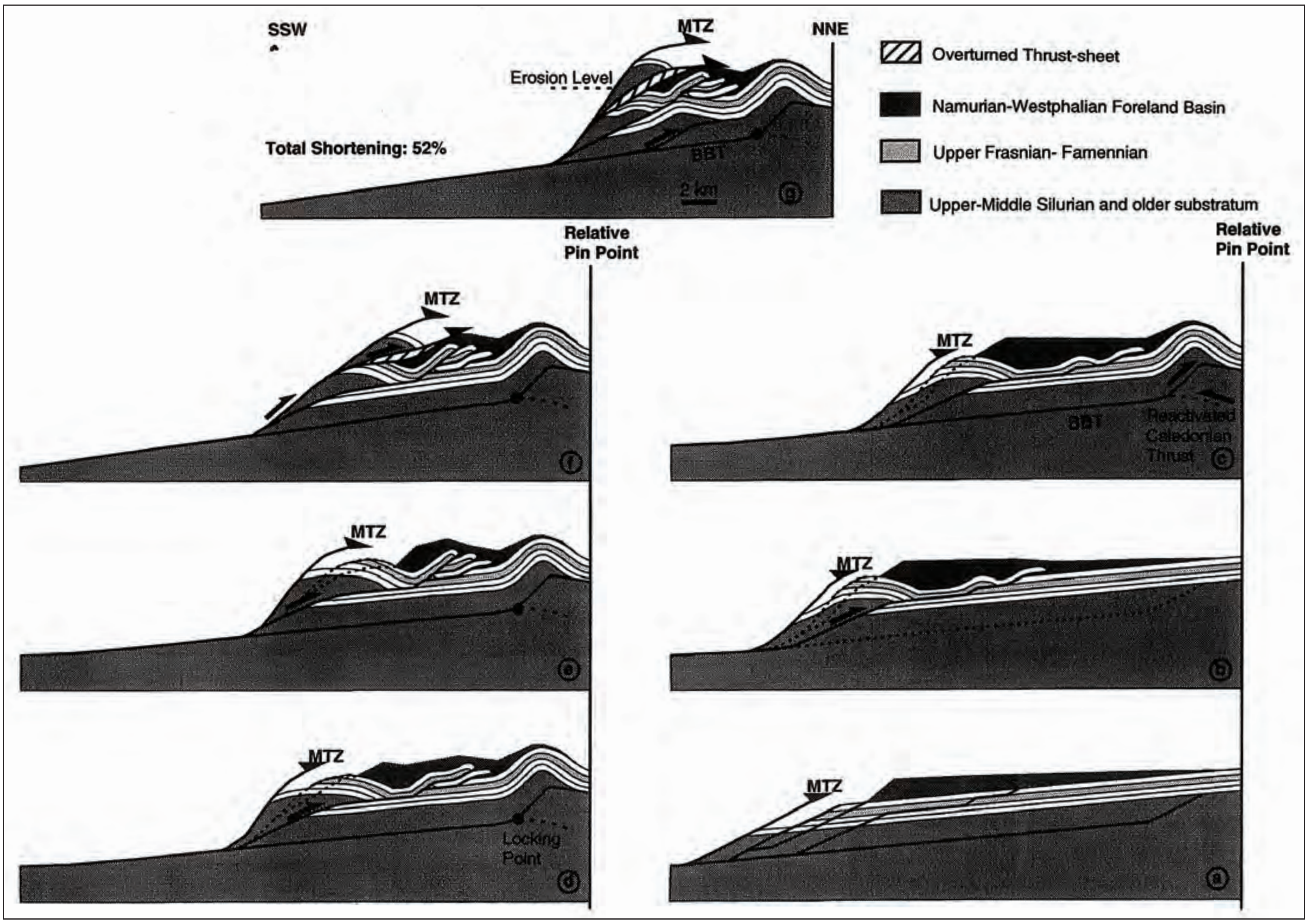


Fig. 233. Sequential thrust development in the Boulonnais thrust belt (Averbuch et al., 2004). (a) state of the continental margin before the Variscan diastrophism; Mid-Westphalian. (b) and (c) piggy-back propagation of thrusts and the development of a frontal thrust wedge in response to the reactivation of a Caledonian south-dipping thrust. (d), (e), (f) and (g) locking of the forward thrust transport with the formation of a localized thrust stack, flexure of the underthrust unit and the resulting out-of-sequence thrusting. MTZ = Midi thrust zone. BBT = Boulonnais basal thrust.

Averbuch et al. (2004) also propose a geotectonic evolution model for the SE England and Boulonnais frontal segments of the Variscan belt. The closure of the Lizard – Rhenohercynian oceanic basin is effective during Late Devonian – Early Carboniferous times. The resulting contractional deformation progressively moves through the northern margin of the basin from the Visean until the Early Stephanian. Finally, an inversion of the northern margin (also called the Brabant margin) of the basin during Late Westphalian – Lower Stephanian times produces the Variscan folds and thrusts (Fig. 234a).

Averbuch et al. (2004) also suggest that the dextral wrench Bray fault zone of Variscan age and located within internal thrust units argues for NW-SE-striking transpressional tectonics. These major strike-slip faults are correlated with the strain partitioning during an oblique collisional process. The link between the frontal Boulonnais thrust zone and the hinterland Bray wrench fault zone would result from strain partitioning of a NNW-directed convergence along an oblique NW-SE-striking continental margin (itself inherited from the geometry of the Devonian Lizard – Rhenohercynian oceanic basin; Fig. 234a). Fig. 234b indicates that the SE England-Boulonnais thrust belt is an oblique transfer zone between the frontal belts of the Ardenne-Rhenish and SW England domains. Averbuch et al. (2004) consider the northern Variscan thrust front as a highly segmented belt that developed from a pre-existing Early Devonian basin structure. The authors also propose the rotation of the shortening direction from a nearly N-S orientation during Variscan times to a nearly E-W orientation in Late Variscan times (Fig. 234b).

The paleotectonic history of Belgium and northern France during the formation of the Variscan belt integrated in a regional/global framework is very complex and the subject of much discussion. For information, many geodynamic reconstructions have been proposed, for example those by Ziegler (1990), Matte (1991, 2001), Blakey (1999), Stampfli & Borel (2002), von Raumer et al. (2003), Nance & Linneman (2008), Sintubin (2008) and von Raumer & Stampfli (2008).

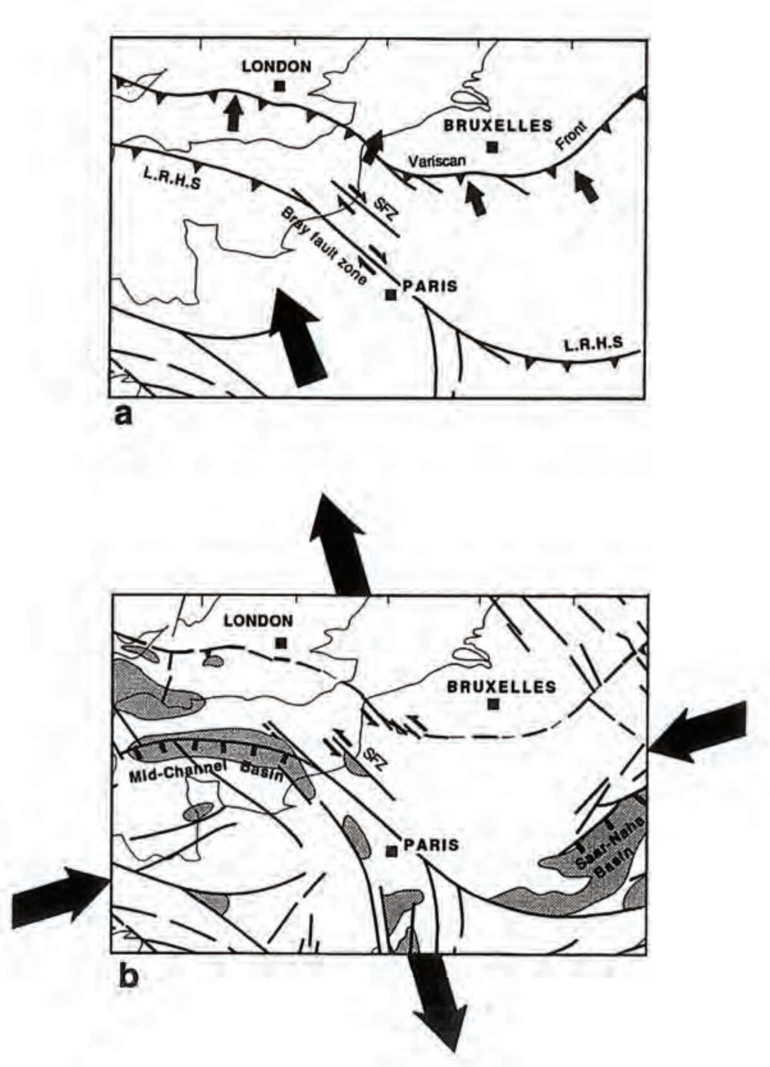


Fig. 234. Tectonic evolution model for the Variscan front and hinterland in England, France and Belgium (Averbuch et al., 2004). (a) Late Westphalian – Lower Stephanian: Variscan NNW-directed convergence and strain partitioning. (b) Upper Stephanian – Middle Permian: Late Variscan nearly E-W-trending shortening and N-S-trending extension.

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9.5. Opont(-Carbonnière) Fault

Location

The Opont Fault is introduced in the literature by Asselberghs in 1944 although the fracture is already identified and recognized in 1940 when Asselberghs (1940b) considers it as a segment (i.e. the eastern extremity) of the Vencimont Fault. We refer the reader to the Vencimont Fault described in section 9.7 for further details. The Opont Fault does not appear on the geological maps at 1:40 000 of Malaise (1900, 1901) and Dormal (1897).

The Carbonnière Fault is identified in 1937 by Waterlot in the vicinity of Deville (in the Rocroi Massif; French department of the Ardennes) but its existence is suspected from at least 1842 when Sauvage & Buvignier remark on the abnormal contacts between slate veins within the Cambrian of the Rocroi Massif. The junction between the Opont and the Carbonnière faults is envisaged by Beugnies in 1983.

Sensu Asselberghs (1944), the Opont Fault strikes over a distance of nearly 30 km within the axial zone of the Ardenne Anticlinorium from north of Bièvre in the west to the region of the Cambrian Serpont Inlier and Libramont in the east.

Stratigraphy and lithology of the country rocks

Asselberghs (1940b) observes anomalous stratigraphic contacts both in the Our and Lesse valleys in the area northwest of Libramont. The discontinuities are associated with the presence of a longitudinal fracture that brings into contact the Saint-Hubert Formation (the former “G2b” of Upper Lochkovian age) in the south with the Oignies Formation (the former “G2a”, Lochkovian in age) and the Early Lochkovian (formerly “G1”) in the northwest and northeast respectively. The Saint-Hubert Formation is made up of green shales and quartzites and the Oignies Formation and Early Lochkovian are composed of various siliciclastic rocks, mainly slates, shales, silty slates and micaceous sandstones.

Geometry

The Opont segment:

Asselberghs identifies in 1940 a longitudinal fracture between Libramont in the east and Opont in the west. The fracture is initially correlated with a segment of the Vencimont Fault (see section 9.7) that explains the anomalous stratigraphic contacts between Lochkovian rocks. From a geometrical point of view, the Vencimont Fault may be differentiated into 3 segments (see Vencimont Fault, Fig. 258 in section 9.7): two northern and southern segments both striking E-W but at different latitudes and a third transverse N-S-trending segment connecting the other two. Fig. 235 shows the southern E-W-striking segment that will later become the Opont Fault (see below).

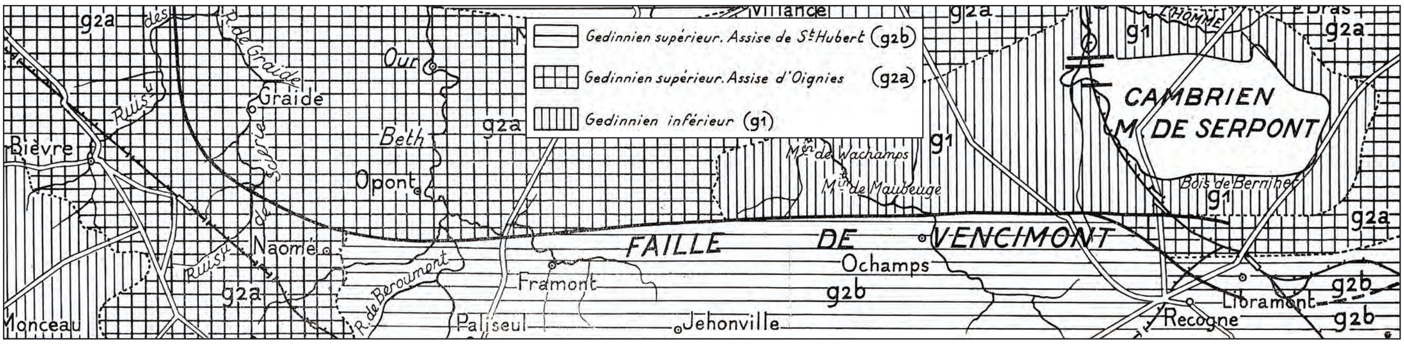

Fig. 235. Extract of the geological map of the Lower Devonian in the vicinity of the Serpont Massif (Asselberghs, 1940b). The trace of the eastern part of the Vencimont Fault is represented.

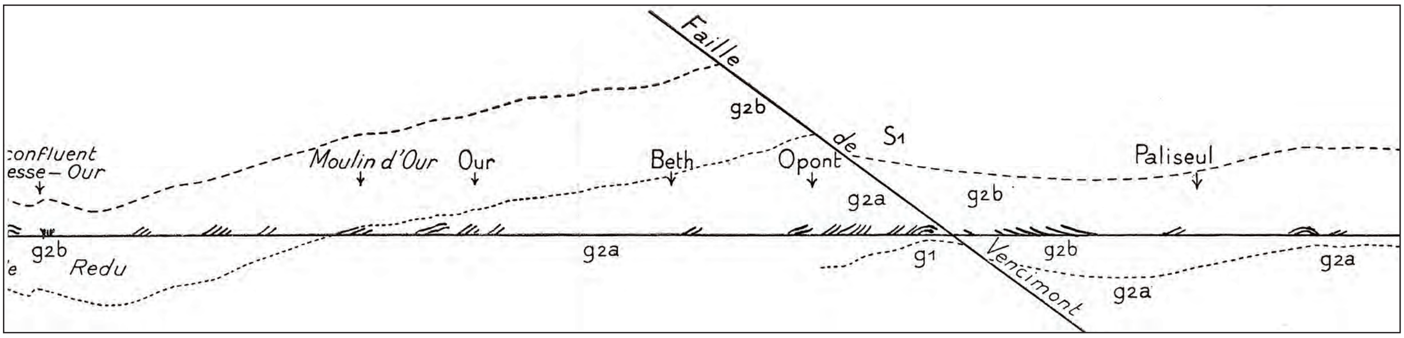

Fig. 236. N-S cross-section at the longitude of Paliseul (Asselberghs, 1940). The 35° south-dipping Opont Fault (considered in 1940 as a segment of the Vencimont Fault) shows an apparent normal displacement of about 1600 metres. G1 = Early Lochkovian. G2a = Oignies Formation (Lochkovian). G2b = Saint-Hubert Formation (Upper Lochkovian).

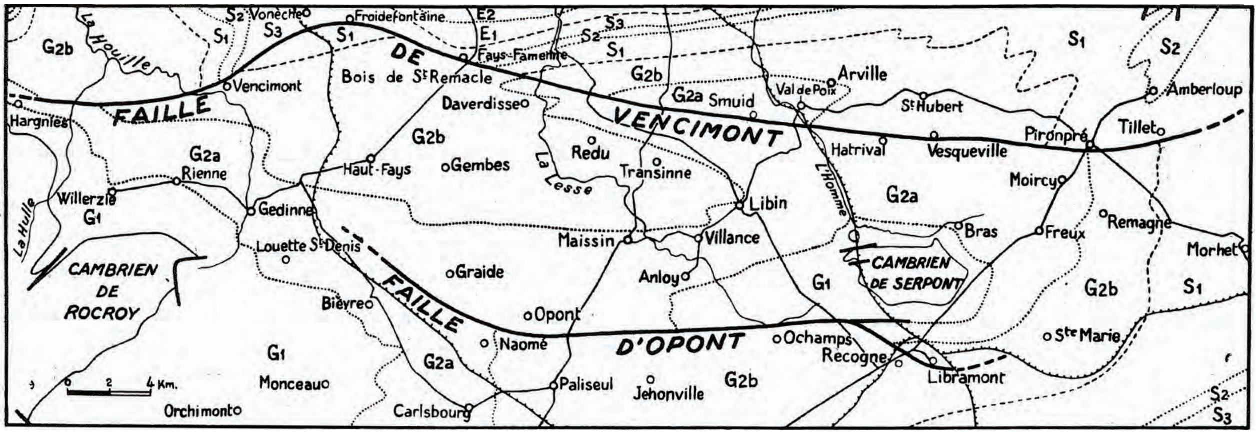

Fig. 237. Traces of the Opont and Vencimont faults (Asselberghs, 1944). G1 = Lower Lochkovian. G2a = Oignies Formation (Lochkovian). G2b = Saint-Hubert Formation (Upper Lochkovian). S1 = Lower Pragian. S2 = Middle Pragian. S3 = Upper Pragian. E1 = Lower Emsian. E2 = Middle Emsian. E3 = Upper Emsian.

The southern segment of the Vencimont Fault dips moderately (~35°) to the south. The northern block is composed of older rocks of Lower Lochkovian age relative to the southern hanging wall block that comprises younger rocks of the Upper Lochkovian. The hanging wall is therefore downthrown relative to the northern footwall block. The apparent offset is normal and can be measured (on the cross-section in Fig. 236) at approximately 1600 m. Despite this apparent displacement, the Vencimont Fault (*sensu* Asselberghs, 1940) is considered to be a reverse fault.

The Vencimont Fault, and in particular its southern longitudinal segment located in the area southwest of the Cambrian Serpont Inlier, is re-interpreted differently in 1944 by Asselberghs who establishes another eastward extension of the northern segment of the Vencimont Fault (see section 9.7). In other words, the Vencimont Fault, in the region of Vonèche and Froidefontaine, conserves its E-W strike instead of adopting a N-S direction. Asselberghs therefore defines two major and distinct longitudinal faults in the anticlinal zone of the Ardenne: the Vencimont Fault to the north and the newly named Opont Fault to the south (Fig. 237). The ground surface trace of the Opont Fault reaches 30 km long. No new ideas concerning the Opont Fault appear in the famous 1946 work of Asselberghs (see Fig. 159).

The Carbonnière segment:

In 1842, Sauvage & Buvignier publish observations on slate veins in the vicinity south of Deville and Monthermé (France). Following their ideas, the disposition of the veins cannot be coherent without the presence of a major longitudinal fault. No further information is given. Gosselet (1888) indicates that no field evidence for this major fault has been found and therefore he does not believe in his existence.

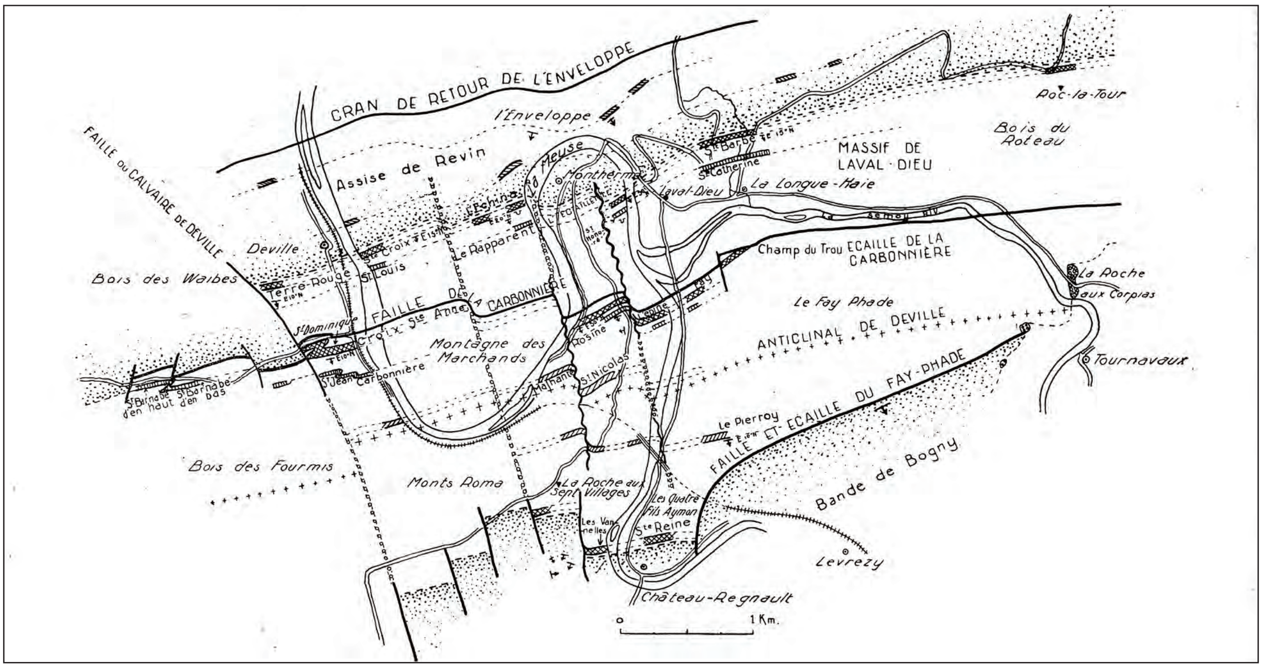

Fig. 238. Structural map of the Deville area showing the positions of slate veins and the main faults (Waterlot, 1937).

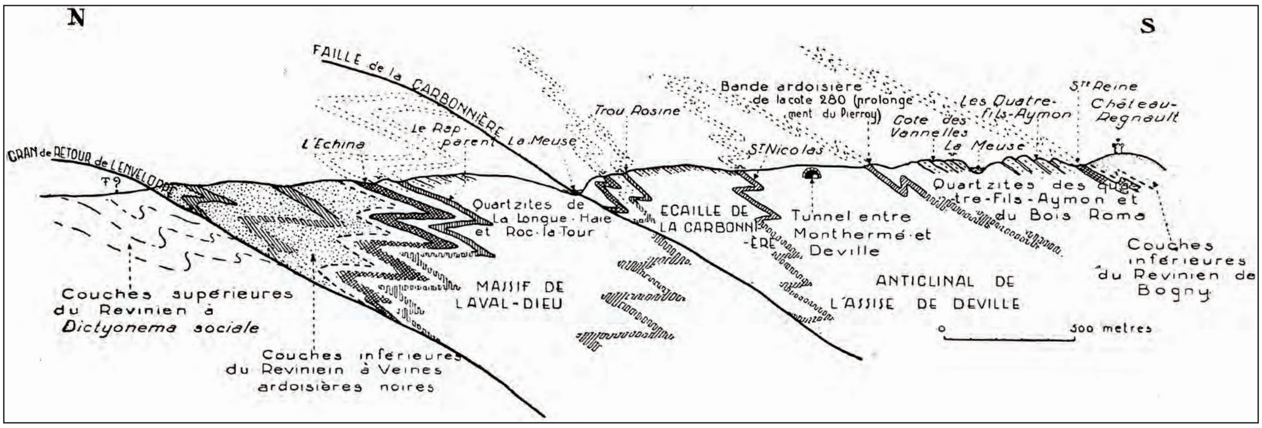

Fig. 239. N-S cross-section through Monthermé and Château-Regnault (Waterlot, 1937).

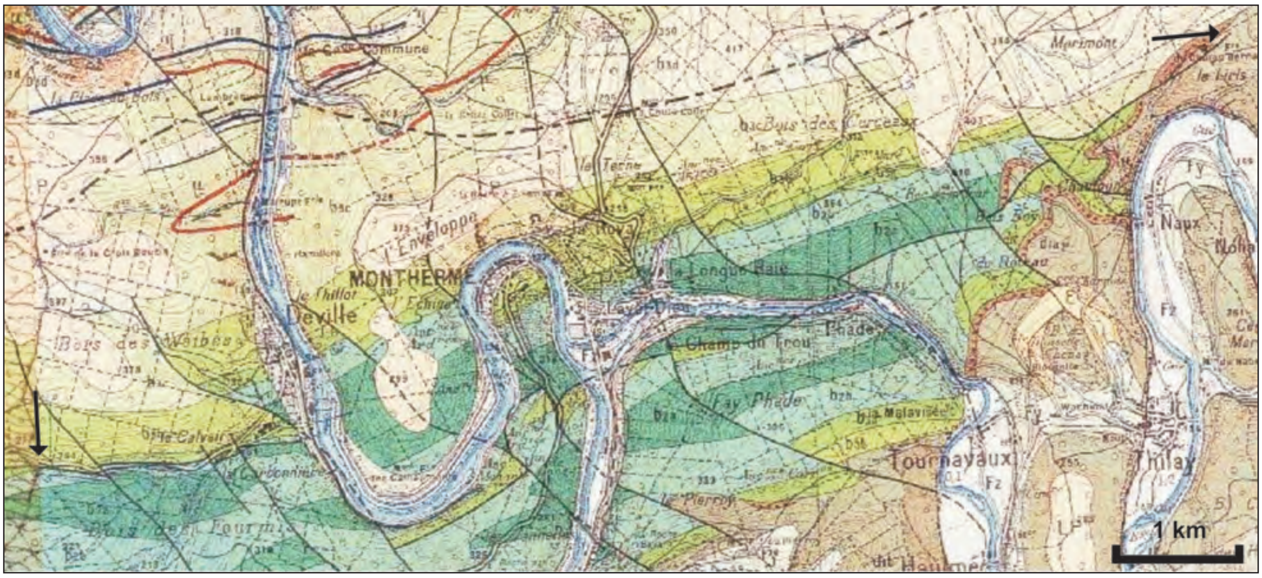

Fig. 240. Extract of the French geological map of Fumay (n°53; Beugnies et al., 1965). Arrows indicate extremities of the Carbonnière Fault.

Waterlot (1937) proposes a detailed study of the French part of the Cambrian Rocroi Massif. Based on the hypothesis of Sauvage & Buvignier, he proposes the location of the trace (Fig. 238) of a reverse longitudinal fracture, named the Carbonnière Fault, responsible for the repetition of the "Echina" and "Rapparent" slate veins. The Carbonnière Fault is traced over a distance of nearly 8 km and is frequently disrupted by transverse faults with either dextral or sinistral offset.

The Carbonnière Fault has a moderate dip to the south of about 30-35° (Fig. 239). The southern hanging wall block, called the Carbonnière tectonic stack, is thrust in a northerly direction over the Laval-Dieu Massif. The reverse displacement measured on the cross-section in Fig. 239 reaches 580 metres.

The French geological map of Fumay (the version released in 1965, Beugnies et al.) displays the trace of the Carbonnière Fault for a distance of about 11 km. The fracture strikes from an E-W to SW-NE direction (Fig. 240).

The Opont-Carbonnière Fault:

Beugnies (1983) proposes a connection between the Opont Fault in the anticlinal area of the Ardenne in the east and the Carbonnière Fault in the Lower Palaeozoic Rocroi Inlier in the west. The resulting Opont-Carbonnière Fault is recognized over a distance of 59 km. Fig. 241 shows the eastern part of the fracture.

In the vicinity of Naux, the Opont-Carbonnière Fault has a quite gentle dip of 20-30° to the SE. The anomalous stratigraphic contact along the lineament is generally sharp (no transition or brecciated zone) and can be associated with quartz filling. Beugnies also indicates that the southern hanging wall block is made up of younger rocks than those located in the northern footwall. As a consequence, Beugnies does not believe in a major north-directed thrust component but in a dextral strike-slip and normal fault (see below).

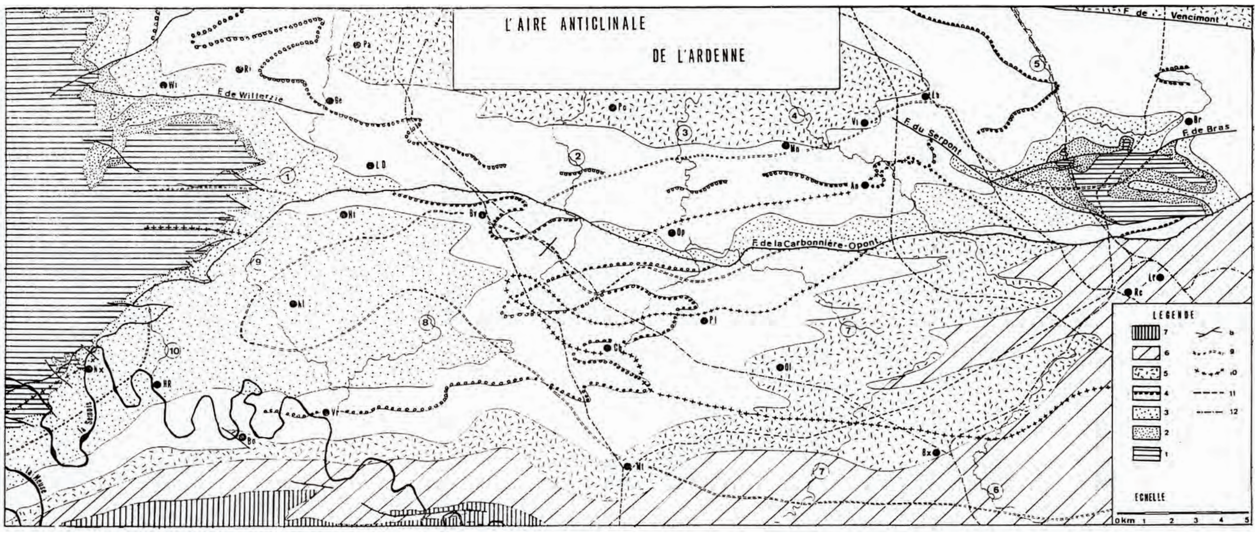

Fig. 241. Geological map of the anticlinal area of the Ardenne Massif (Beugnies, 1983). 1. Cambrian. 2. Lower Lochkovian (“G1a”). 3. Lower Lochkovian (“G1b”). 4. Upper Lochkovian (“G2a”). 5. Upper Lochkovian (“G2b”). 6. Lower Pragian (“Sg1”). 7. Middle Pragian (“Sg2”). 8. Faults. 9&10. Metamorphic zone limits. 11. Roads. 12. Railways. See Beugnies (1983) for the complete legend.

Beugnies (1983, 1985) considers the Opont-Carbonnière Fault as separating two important units of the anticlinal area of the Ardenne: the Opont Unit to the north (but located to the south of the Vencimont Fault) and the Carlsbourg Unit to the south. In 1986, Beugnies adds that the Opont Fault is recognized for a distance of 93 km. It has a low-angle dip to the south. In its eastern part, the fault runs through the areas of Bastogne, Mardasson, Wardin, Bras and reaches Luxembourg territory. There, he proposes a connection between the Opont Fault in the west and the Malsbenden Fault (see Cambier & Dejonghe, 2010) in the east. In this case the total length of the Opont-Malsbenden Fault would reach more than 180 km. However, we believe this connection to be highly unlikely considering the latest interpretations for the two fractures: the dextral strike-slip (and normal) offset of the gently south-dipping Opont Fault is hard to associate with the reverse offset of the north-dipping Malsbenden backthrust.

Interpretations

The Opont segment:

Asselberghs (1940b, 1944) gives little information on how to interpret the Opont Fault. Initially considered as a segment of the reverse Vencimont Fault (1940), in 1944 the Opont Fault is again envisaged as a reverse fracture despite the fact that rocks found in the hanging wall are younger than those of the older footwall block.

The Carbonnière segment:

Waterlot (1937) considers the reverse Carbonnière Fault as resulting from the Caledonian shortening. The Caledonian Orogony has affected the Cambrian Rocroi Massif and produced several north-verging overturned folds disrupted by a few thrust-type fractures of which the Carbonnière Fault is one. Beugnies (1962) also interprets the Carbonnière Fault as a reverse fault with a northerly thrust.

The explicative note attached to the French geological map of Fumay (Beugnies et al., 1965) considers the Variscan shortening as playing a major role in the genesis of the Carbonnière Fault. Two stages are distinguished:

- a first ductile stage forming regionally three large longitudinal folds (from north to south: the Fépin Anticlinorium, the Willerzie Synclinorium and the Louette-Saint-Pierre Anticline); and
- a second brittle stage cross-cutting the folds. In this case the Willerzie Synclinorium is disrupted by two major thrusts of which the Carbonnière Fault is one.

Caledonian shortening has also affected Cambrian formations of the Rocroi Massif but the subsequent Variscan Orogeny is considered to be responsible for the current structural disposition.

The Opont-Carbonnière Fault:

Beugnies (1983) specifies a Variscan origin for the Opont-Carbonnière Fault. The generation of the fault is before the transverse strike-slip faults that affect it but after the formation of the schistosity, the metamorphism and the folds within the Devonian cover. These elements enable Beugnies to relate the Opont-Carbonnière Fault to late-Variscan tectonics.

As indicated before, Beugnies gives up the hypothesis of a northerly thrust fault (which was proposed by previous geologists) and proposes a major right-lateral displacement of up to 18 km. Applying this dextral strike-slip hypothesis, the Carlsbourg Unit (located to the south of the Opont-Carbonnière Fault) would correspond to the cover of the

Opont Unit. This cover would have glided from north to south along the normal and wrench Opont-Carbonnière Fault. The normal offset is supposed to be about 2-3 km. In 1985, Beugnies adds to the hypothesis an increase in dextral and normal offsets from east to west.

References

Asselberghs, 1940b.
Asselberghs, 1944.
Asselberghs, 1946.
Beugnies, 1962.
Beugnies, 1983.
Beugnies, 1985.
Beugnies, 1986.
Beugnies et al., 1965.
Cambier & Dejonghe, 2010.
Dormal, 1897.
Gosselet, 1888.
Malaise, 1900.
Malaise, 1901.
Sauvage & Buvignier, 1842.
Waterlot, 1937.

9.6. Tombe Fault

Location

Belgian geologists and engineers introduced the Tombe Fault on the “Carte générale des mines de Belgique”. The fault, located in the area west of Charleroi, is introduced in the literature by Smeysters in 1880 and 1887 and later by Briart in 1894(a). Briart already interpreted the fracture as limiting a thrust sheet at the front of the Condroz Nappe.

The Tombe Fault limits the base of the “Fontaine-l’Évêque – Landelies Massif”, also known as the “Tombe Massif”. The “Tombe Massif” is a special structure of the Sambre-and-Meuse coal-basin which appears as an isolated unit of mainly Upper-Devonian - Lower Carboniferous rocks “lost” in the Westphalian coal-basin. Fig. 242 presents the location and stratigraphy of the “massif”[1] (Delcambre & Pingot, 2000).

1 The term “massif” has been widely but sometimes improperly used in the early Belgian literature. It should be replaced by “tectonic unit” or “thrust sheet” when it does not apply to the basement. However, in this work, we will keep this term between inverted commas to facilitate understanding of older Belgian papers.

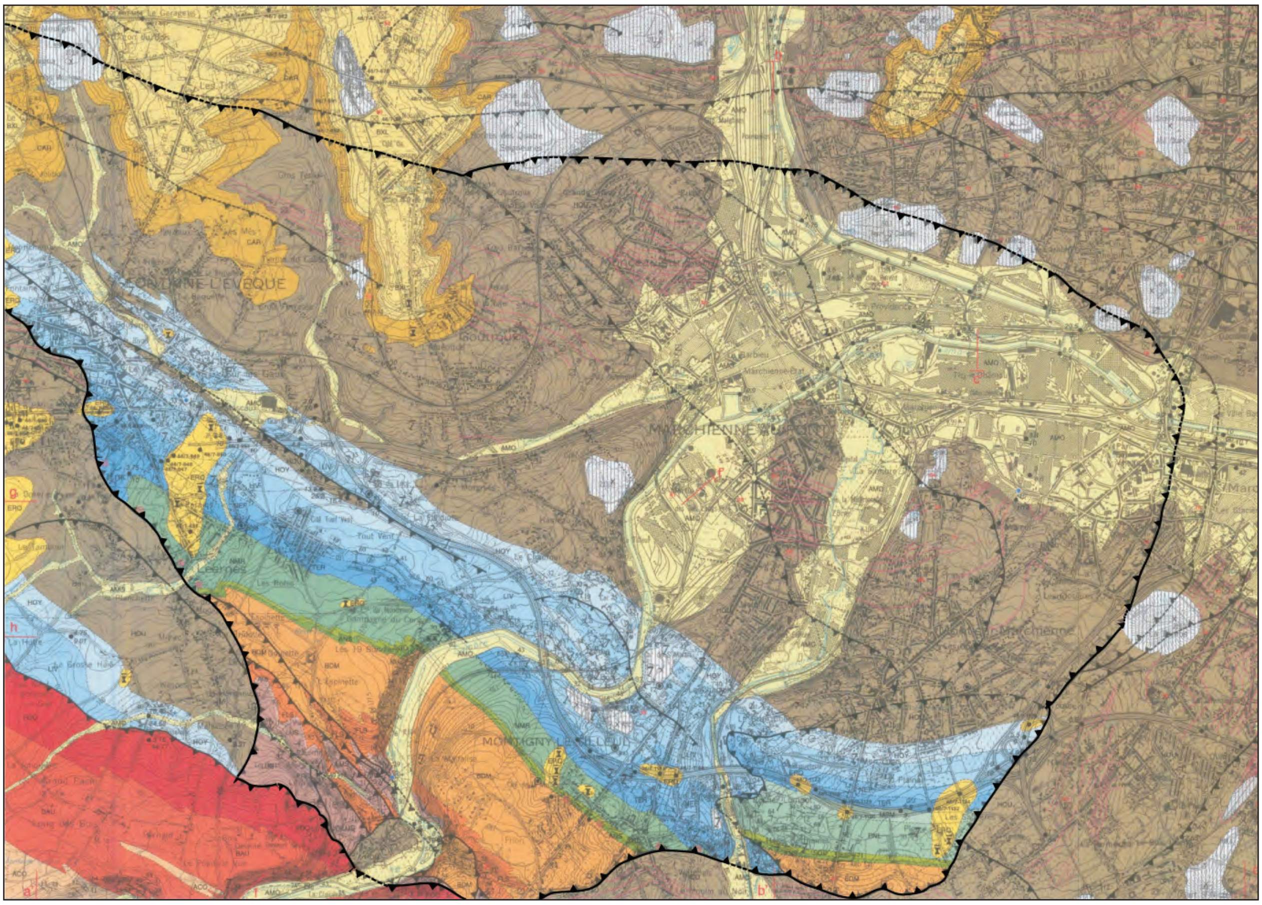

Fig. 242. Extract of the geological map of Fontaine-l’Évêque – Landelies (Delcambre & Pingot, 2000). The work presents the latest state of knowledge of the Devono-Carboniferous “Tombe Massif”.

Lithology and stratigraphy of the country rocks

The recent geological map of Fontaine-l'Évêque – Charleroi (Delcambre & Pingot, 2000) provides a detailed and revised stratigraphy of the "Tombe Massif". From SW to NE, Frasnian-Famennian, Dinantian and Lower Silesian terrains are successively observed (Fig. 242).

The Presles, Lustin, Falisolle and Bois des Mouches formations constitute the Frasnian-Famennian terrains. Lithologies are various but are mainly shales, limestones and sandstones. Dinantian formations are those of Anseremme, Station de Gendron, Namur, Pont-à-Nôle, Mont-sur-Marchienne, Terwagne, Neffe, Lives and Hoyoux. Typically the Visean part comprises limestones and dolostones, whereas shales are more abundant in the Tournaisian. Finally, the "*Houiller*" Group of shales, siltstones and sandstones, with interbedded coal seams, make up the Lower Silesian terrains.

Geometry

In 1894(a), Briart draws the Tombe Fault for a distance of 10.8 km from north of Fontaine - l'Évêque to SE of Mont-sur-Marchienne (Fig. 243). The fault bounds a major unit called the "*lambeau houiller de Marchiennes*" to the north and the "*lambeau carbonifère de la Tombe*" to the south. The southern unit consists of a tectonic stack thrust northwards over the large (autochthonous) coal-basin and constitutes the northernmost thrust of the "Tombe Massif". The fault has a very gentle dip to the south and is connected to other fractures related to the Midi Fault. The Fontaine-l'Évêque and the Leernes faults are also represented on the map. These fractures bound other tectonic stacks within the "massif" (see the interpretations section). Later, Briart confirms the trace of the Tombe Fault on the 1/40 000 scale Belgian geological map (Briart & Bayet, 1904).

On the basis of the numerous colliery wells drilled in the eastern part of the Charleroi coal-basin, Smeysters (1905) proposes an extension of the "*Houiller*" terrain under the "Tombe Massif" without discontinuity between the two. Both units would be separated from each other by the Tombe Fault, which would have a northern dip of about 10-12° at the southern limit of the "massif". Smeysters adds that depending on position within the "Tombe Massif", the Tombe Fault has a variable dip. For example, the fault acquires a southward inclination in the area the north of the "massif".

In 1912 and 1919a, Fourmarier confirms the trace of the Tombe Fault of Briart (1894a). He considers the delimitation of the main "Tombe Massif" as correct but raises doubts about the faults located within the "massif" (see the interpretations below, Fig. 253).

Stainier (1922) subdivides the "Tombe Massif" into two tectonic stacks: the "*lambeau de la Tombe*" in the east and the "*lambeau de Saint-Martin*" in the west. These units are delimited at their base by the Tombe and the Saint-Martin faults respectively (Fig. 244).

In 1947, Kaisin Jr. subdivides the coal basin of Charleroi into 3 groups of "massifs". He makes a distinction between the "massifs d'entraînement" (or subautochthonous "massifs"; Delmer, 2004) to the north and the thrust nappes to the south. The "Tombe Massif" is considered as a "disparate tectonic feature" that overlies the thrust nappes at the southern limit of the basin.

More recently, Beugnies (1976) has greatly contributed to the structural understanding of the "Tombe Massif". The "massif" is composed of two distinct (upper and lower) units with particular sedimentary and tectonic features. The Gaux Fault, a listric gently south-dipping fracture introduced by Fourmarier in 1912, separates the two units from each other (see the interpretations, Fig. 255). The upper unit, composed of Frasnian to Namurian rocks, is bounded to the south by the Midi Fault; while the lower unit, made up Visean to Lower Westphalian rocks, everywhere overlaps the Westphalian substratum by means of the Tombe Fault. The lower unit of the "Tombe Massif" is therefore delimited by two fractures, the Gaux Fault separating the upper unit from the lower unit) and the Tombe Fault (separating the substratum from the lower unit (Fig. 245).

Cross-section "B" on Fig. 246 illustrates an apparent folding of the Tombe Fault, which therefore demarcates two subunits: the Fontaine-l'Évêque thrust stack to the north and the Wespes thrust stack to the south. Both structures form the lower unit of the "Tombe Massif".

Delcambre & Pingot (2000) revise the geological map of Fontaine-l'Évêque – Charleroi and consider the local tectonic structure to be composed of three major units, from north to south (Fig. 247): the parautochthon "massifs" of Namuro-Westphalian rocks, the thrust "massifs", subdivided into several tectonic stacks (including the "Tombe Massif") and the "Midi Massif" of Caledonian and Lower Devonian terrains. Note that compared to the ideas of Beugnies (1976), the Wespes tectonic stack is no longer bounded by the Tombe Fault but by the Wespes Fault and the relationship between the fractures is not given. Moreover, the Wespes stack no longer belongs to the "Tombe Massif" but is correlated with another.

Delcambre & Pingot (2000) subdivide the "Tombe Massif" into 4 units (Figs. 248 & 249): the Monceau Unit (bounded by the Monceau Fault), the Forêt Unit (bounded by the Monceau and the Forêt faults), the Conception Unit (bounded by the Forêt and the Tombe faults) and the Mont-sur-Marchienne units (bounded by the Mont-sur-Marchienne Fault).

Fig. 243. Geological map of the Fontaine-l'Évêque and Landelies area (Briart, 1894a). Arrows indicate the extremities of the fault.

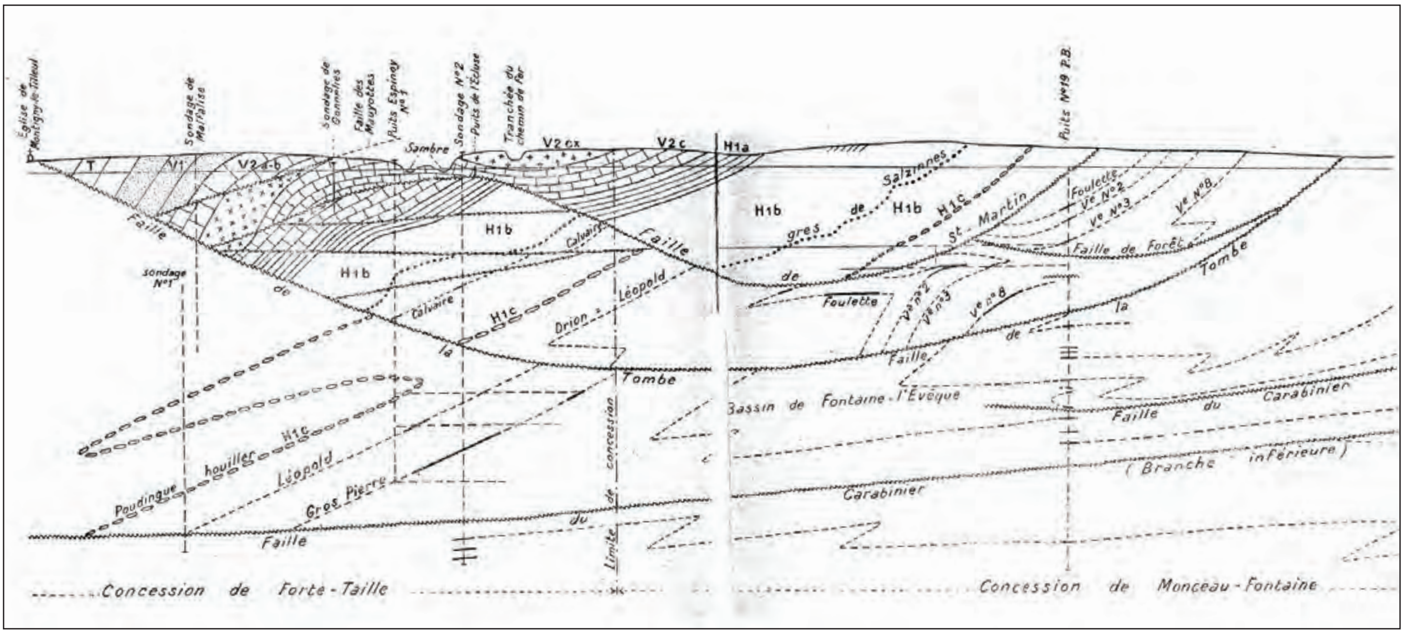

Fig. 244. S-N cross-section through the "Tombe Massif" in the vicinity of Monceau-Fontaine (Stainier, 1922). T = Tournaisian. V1 = Lower Visean. V2 a-b, V2c & V2cx = Upper Visean. H1a, H1b & H1c = Lower "*Houiller*".

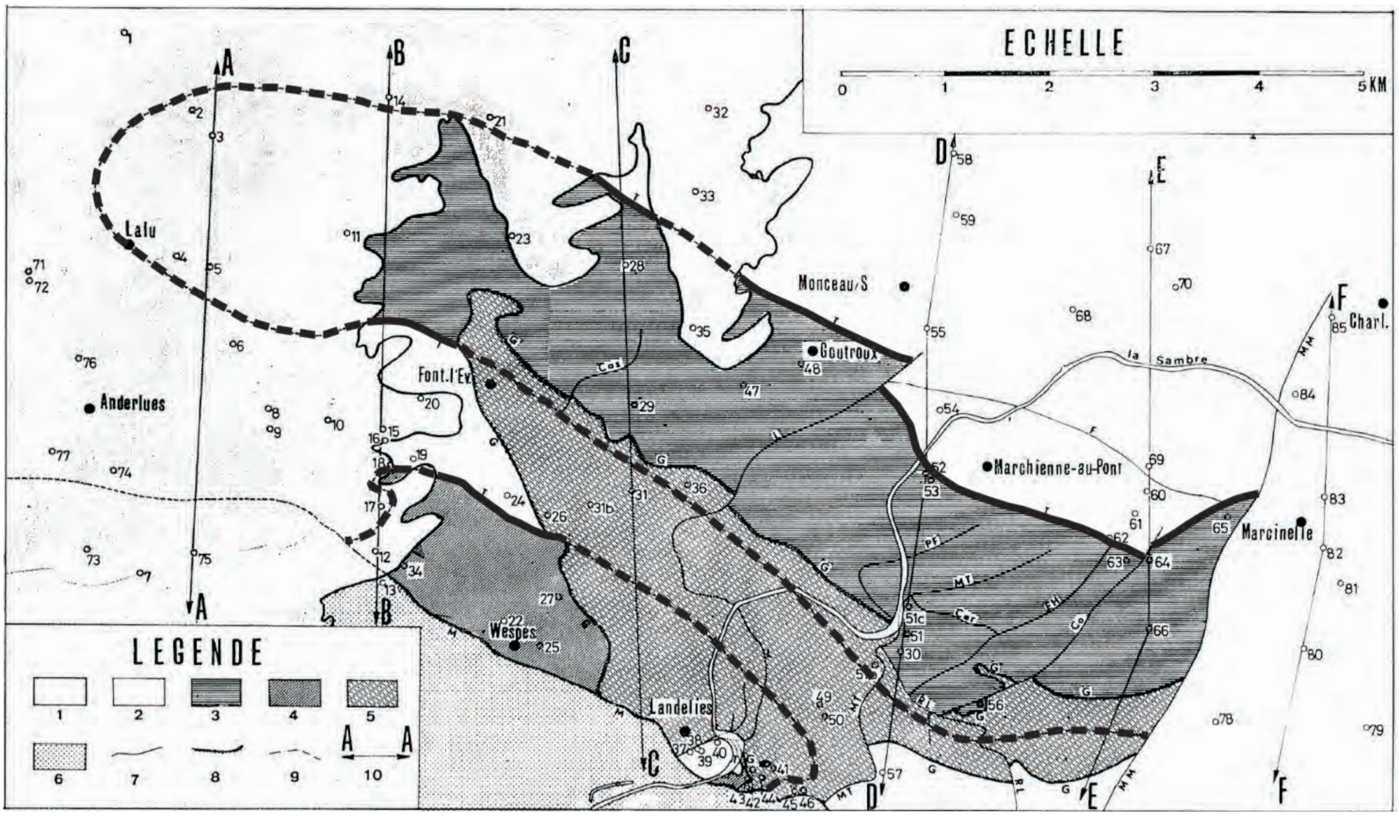

Fig. 245. Structural map of the "Tombe Massif" (Beugnies, 1976). The Tombe Fault is highlighted. Cross-sections B and C are given below. 1 = Cenozoic cover; 2 = Westphalian substratum; 3 = Fontaine-l'Évêque unit; 4 = Wespes unit; 5 = Gaux unit; 6 = "Midi Massif"; 7 = transverse faults; 8 = thrust faults; 9 = faults under tectonic stacks; 10 = cross-sections.

The cross-section of Delcambre & Pingot (2000) in Fig. 249 presents the four tectonic stacks of the "Tombe Massif". On this particular cross-section, the "massif" is delimited almost completely from its Silesian substratum by the Tombe Fault, the thrust plane of which coincides with other faults (the Forêt and the Monceau faults) in its southern part.

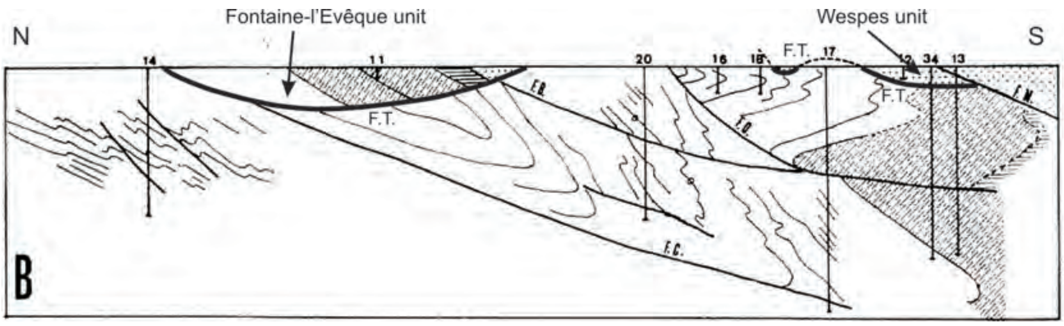

Fig. 246. Cross-section ("B") of the "Tombe Massif" to the west of Fontaine-l'Évêque (Beugnies, 1976). Fig. 245 shows the location of the section. F.B. = Beaulieusart Fault; F.C. = Carabinier Fault; F.M. = Midi Fault; F.O. = Ormont Fault; F.T. = Tombe Fault.

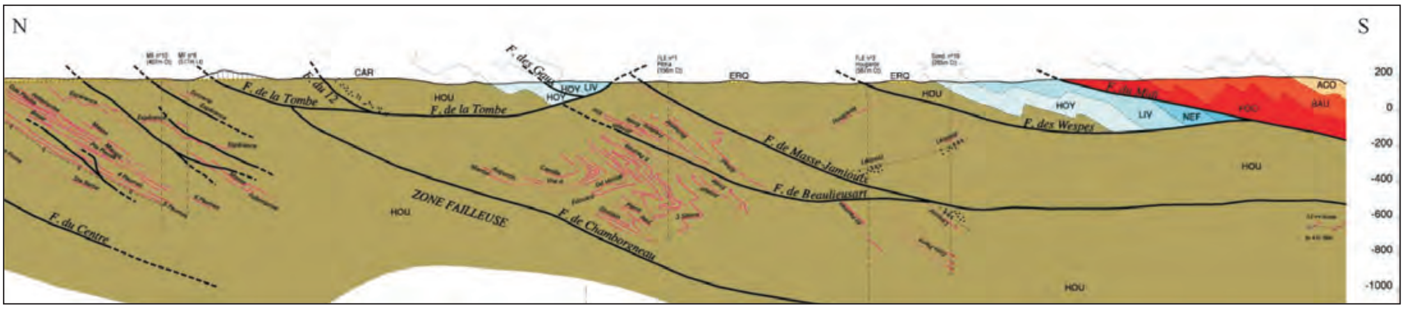

Fig. 247. Extract of the N-S cross-section "a-a'" at the meridian line of Fontaine-l'Évêque (Delcambre & Pingot, 2000). FOO, BAU & ACO = Fooz, Bois d'Ausse and Acoz formations (Lower Devonian). NEF, LIV & HOY = Neffe and Lives formations and Houyoux Group (Visean). HOU = "*Houiller*" Group (Namurian-Westphalian).

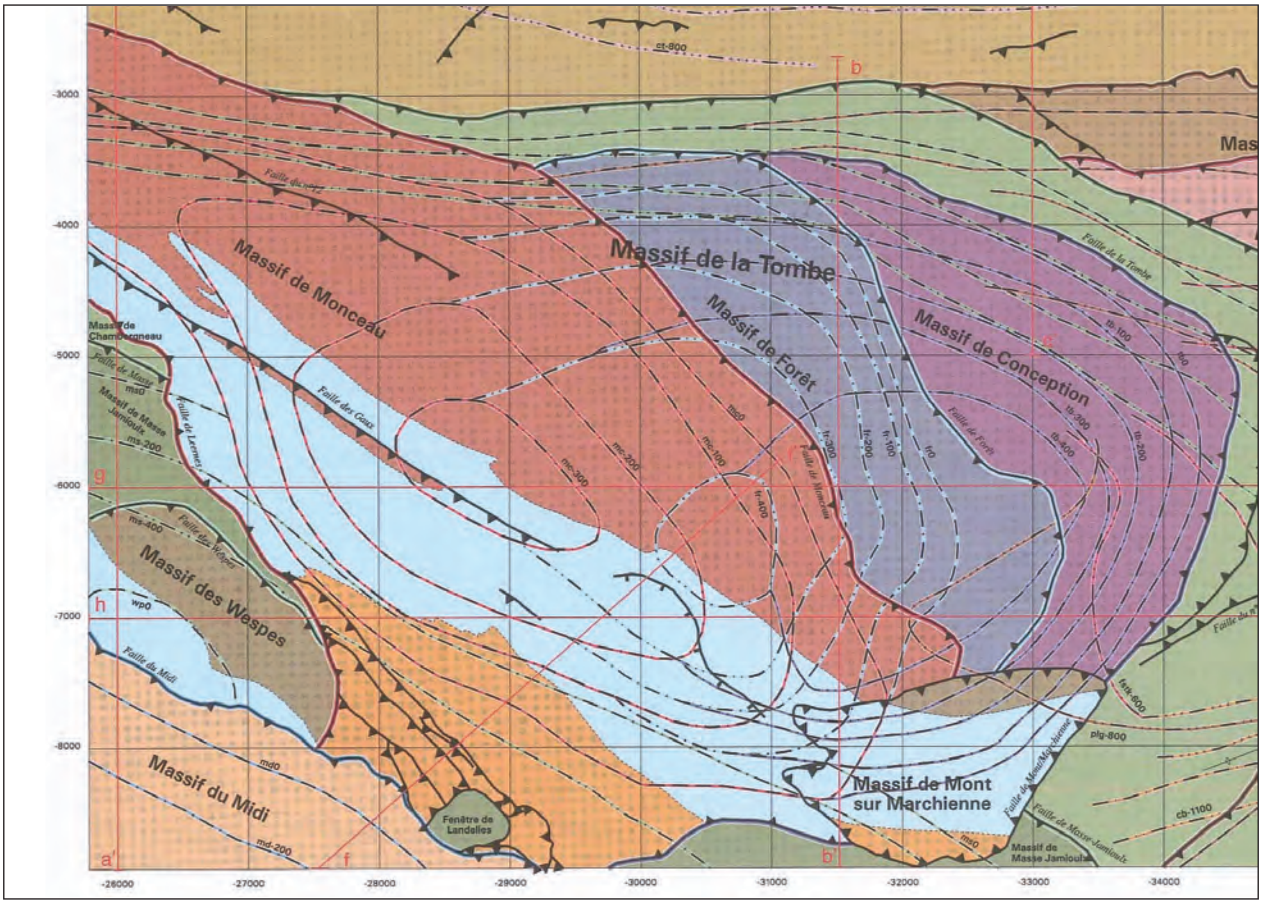

Fig. 248. Extract of the structural map of Fontaine-l'Évêque – Charleroi (Delcambre & Pingot, 2000).

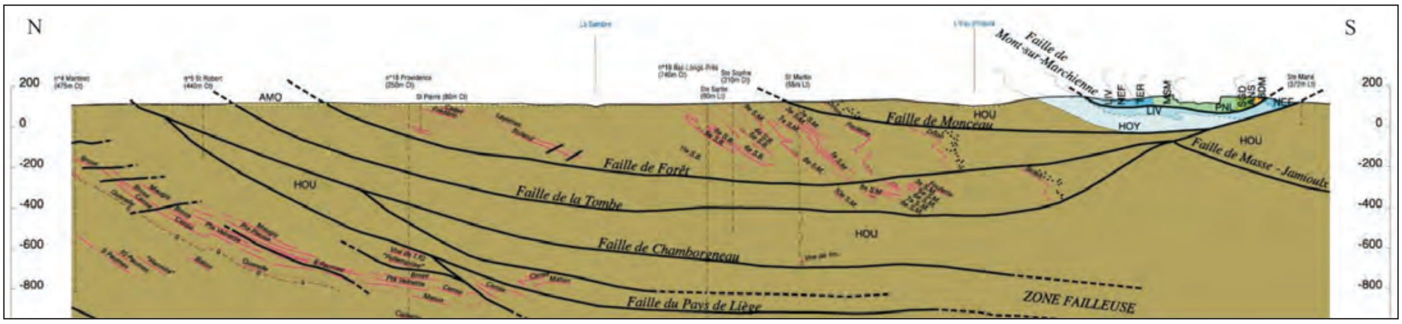

Fig. 249. N-S geological cross-section at the meridian line of Marchienne-au-Pont (Delcambre & Pingot, 2000).

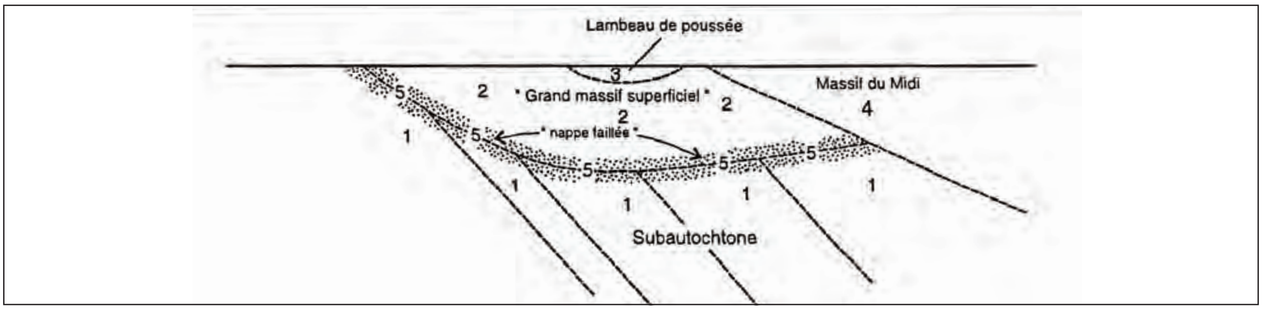

Fig. 250. N-S schematic cross-section through the Hainaut coal-basin (Delmer, 2004). See the explanations in the text.

Just like Kaisin Jr. in 1947, Delmer (1997, 2004) observes two main types of tectonic "massifs" in the Hainaut coal-basin. The northern and deepen parts of the basin are the subautochthonous (imbricate) massifs (n°1 on Fig. 250) and are overlain by the "Great Superficial Massif" (n°2 on Fig. 250). The latter unit can be associated with tectonic stacks ("*lambeaux de poussée*", n°3 on Fig. 250) of which the "Tombe Massif" is one.

Interpretations

In 1894, Briart supposes that the "Tombe Massif" is formed by three superimposed tectonic stacks, of which the most northerly, the Marchiennes-Tombe "*lambeau*", is bounded by the Tombe Fault (Fig. 251). This fracture is described in French as a "*faille de refoulement*" and is interpreted as the first major thrust movement involved in the progressive establishment of the "massif". From this, Briart suggests that the "Fontaine-l'Évêque Massif" was not thrust during a single episode but is the result of three successive thrusts. The northernmost tectonic stack (and the Tombe Fault) would represent therefore the first expression of the thrust phenomenon, while the Midi Fault, bounding the "massif" to the south, would be the last thrust event deforming the "Tombe Massif".

Smeysters (1905) indicates that the Tombe Fault results from tectonic stresses acting from SW to NE. He also indicates that the "*lambeau de poussée de la Tombe*" is a thrust of an "exceptional tectonic significance". The "Tombe Massif" constitutes an important thrust stack overthrust on the underlying and faulted "*Houiller*" (Namurian-Westphalian) coal-basin.

Brien (1905a,b) points out that the southern segment of the Tombe Fault crosscuts two Frasnian anticlines. The thrust is locatedwithin strata without being influenced by folds or by the strata inclination. Brien believes, therefore, that the fault cannot result from the emphasis of a large anticline and its break-up. Contrary to the ideas of Briart (1894a), the thrusts located in the Landelies vicinities would result from two compressive stages that correspond to the formation of the Tombe and the Midi faults respectively. This idea will be taken up later, in 1912, by Fourmarier.

The cross-section in Fig. 252, of Brien (1905a), represents the aspect of the Tombe Fault after the work of Smeysters. To the north, the Tombe Fault would coincide with the thrust plane of the Carabinier Fault; and to the south, the Tombe Fault would be connected at depth to the Midi Fault. The fault affects an overturned fold. Famennian strata (to the south) are vertical and "*Houiller* strata" (to the north) are in a normal position.

Contrary to the ideas of Briart (1894a), Fourmarier (1912) believes that the "Fontaine-l'Évêque –Landelies Massif" is less complex than previously envisaged. The "massif" is not composed of three distinct superimposed thrust stacks (separated by low-angle faults) but forms a single and unique unit overlying the "*Houiller*" and separated by a sole fracture that is the Tombe Fault (Fig. 253). The "massif" has rather few internal faults of local and secondary significance. These faults dislocate the "massif" and probably appear during its thrust to the north. The Landelies region was therefore not to be the location of successive fold events.

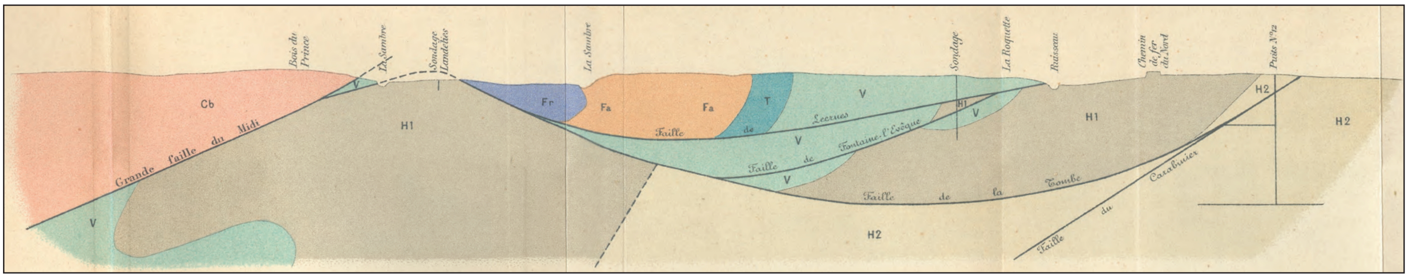

Fig. 251. N-S cross-section through Landelies (Briart, 1894a). Cb = "*Coblencien*" (i.e. Lower Devonian). Fr = Frasnian. Fa = Famennian. T = Tournaisian. V = Visean. H1 = Lower "*Houiller*". H2 = Upper "*Houiller*".

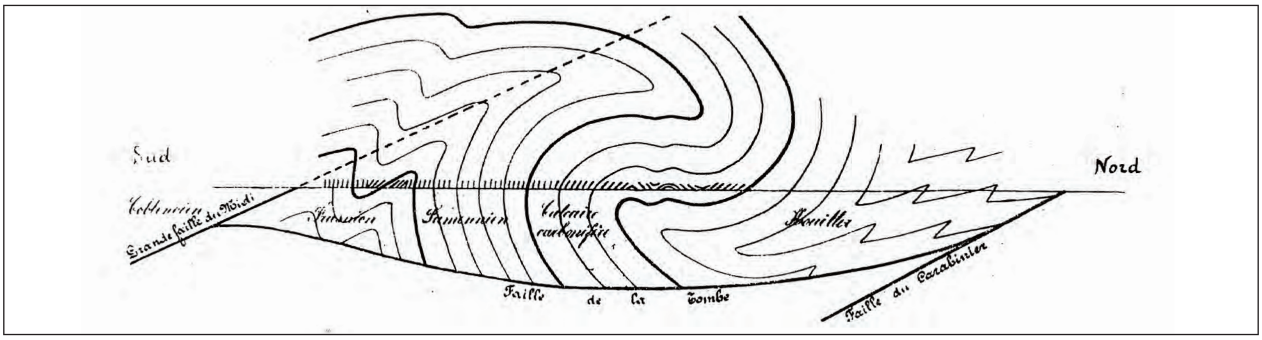

Fig. 252. S-N cross-section through the "Tombe Massif" (Brien, 1905a)

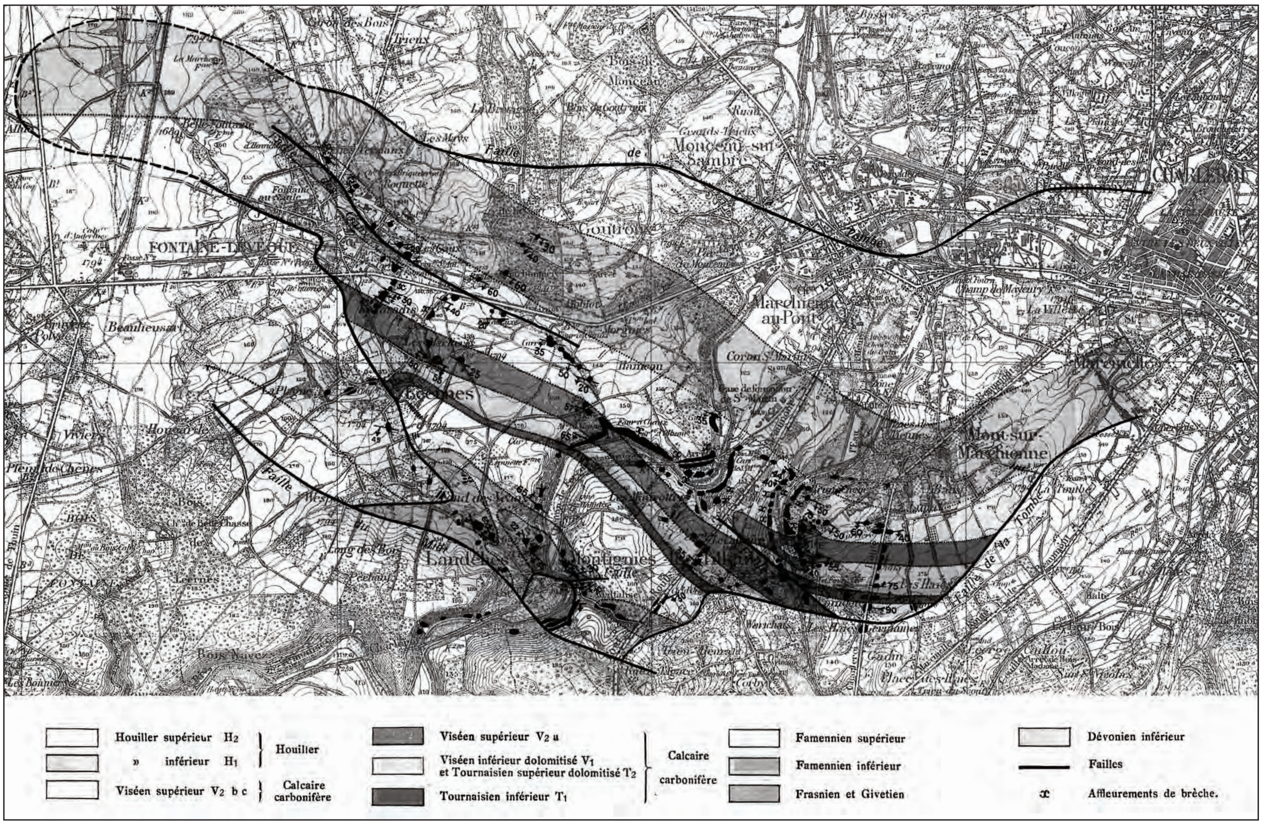

Fig. 253. Map of the "Fontaine-l'Évêque – Landelies Massif" (Fourmarier, 1912).

In his famous publication of the Belgian geology, Fourmarier (1954) proposes a simple and schematic geological map and cross-section of the Fontaine-l'Évêque – Landelies or Tombe "Massif" (Fig. 254). His observations are unchanged: the "Tombe Massif" is composed of Devono-Namurian terrains that step forward into the Westphalian of the Charleroi basin. The "massif" consists of a sole tectonic stack delimited at its base by the thrust-type Tombe Fault.

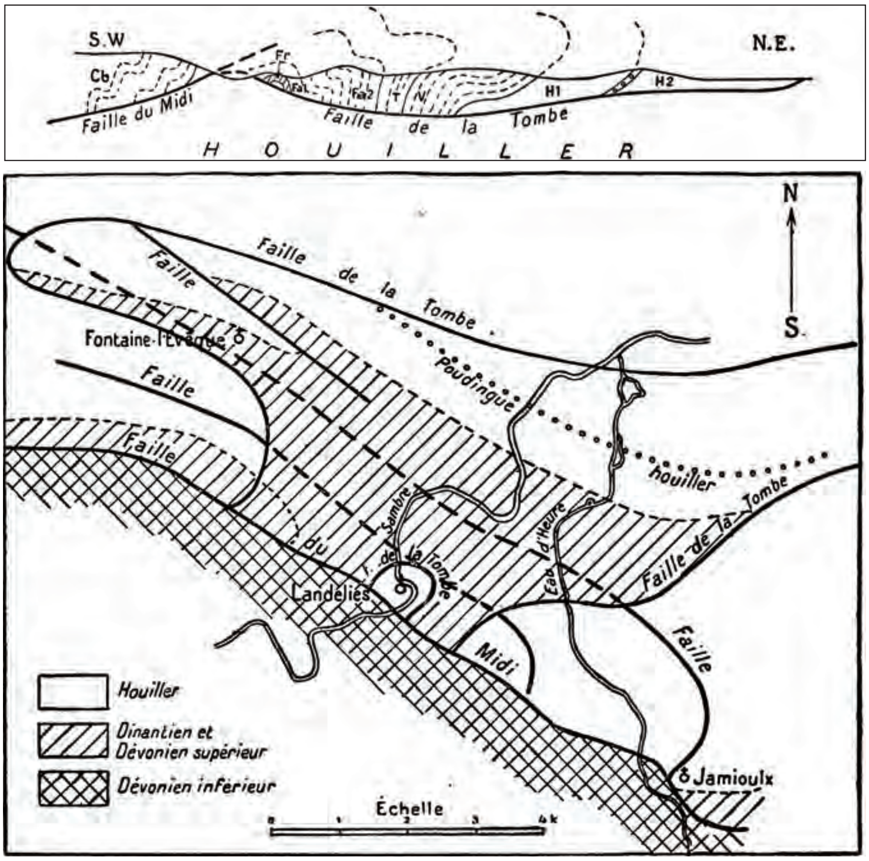

Fig. 254. Geological map and cross-section of the "Tombe Massif" (Fourmarier, 1954). Cb = Lower Devonian. Fr = Frasnian. Fa1 = Lower Famennian. Fa2 = Upper Famennian. T = Tournaisian. V = Visean. H1 = Namurian. H2 = Westphalian.

Cross-section C (Fig. 255) presents the Tombe and the Gaux faults that, according to Beugnies (1976), appeared simultaneously during the main (Asturian stage) deformation event of the Variscan orogeny. Beugnies interprets the structure of the "Tombe Massif" as involving two tectonic stages: a first stage of folding and overturning of the reverse limbs and a second faulting stage. He suggests that the Fontaine-l'Évêque and Wespes tectonic stacks (i.e. the lower unit) were initially grouped together before being separated by the Gaux upper unit. Following this idea, the Gaux unit and fault would have pushed away the Fontaine-l'Évêque stack to the north for about 3 km. Beugnies estimates the total displacement of the "Tombe Massif" (along the Tombe Fault) to be at least 11 km. The Tombe and the Gaux faults connect southwards with the Midi Fault, which crosscuts and therefore postdates both fractures.

In 2004, Delmer put forward a particular tectonic interpretation of the Hainaut coal-basin. The "Great Superficial Massif" (Fig. 256) would not originally have been located to the south before being thrust northwards but would have come from the Brabant Massif to the north (over which the "massif" was deposited). The "Great Superficial Massif" is not therefore thrust from south to north but has glided under the influence of gravity from north to south (Fig. 256). This major movement would be initiated by the dissolution of evaporites at depth.

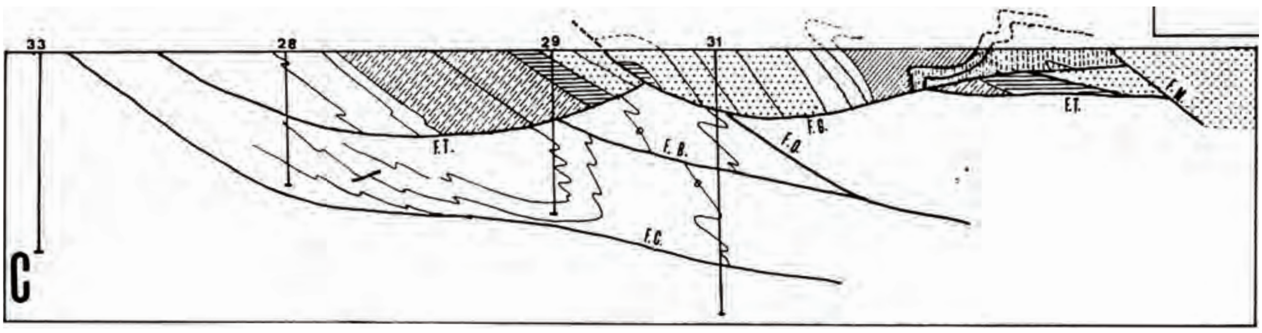

Fig. 255. Cross-section of the "Tombe Massif" to the east of Fontaine-l'Évêque (Beugnies, 1976). Fig. 245 shows its location. F.B. = Beaulieusart Fault; F.C. = Carabinier Fault; F.G. = Gaux Fault ; F.M. = Midi Fault; F.O. = Ormont Fault; F.T. = Tombe Fault.

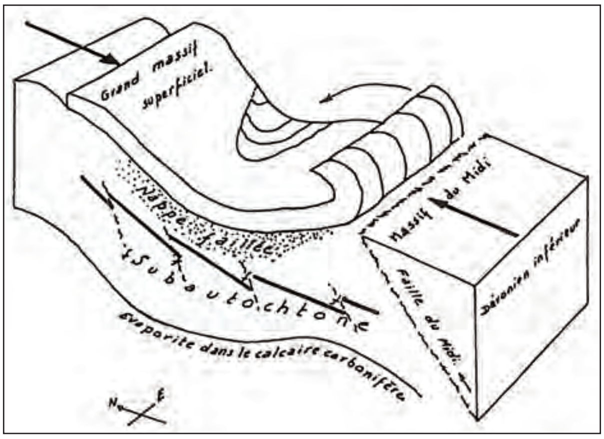

Fig. 256. Schematic representation of the gravity-induced movement of the “Great Superficial Massif” (Delmer, 2004). The “Midi Massif” coming from the south would have cut the southern terrains of the glided “massif” forming few tectonic stacks (e.g. the “Tombe Massif”).

The later northward progression of the “Midi Massif” along the Midi Fault would have involved the detachment of some parts of the terrains located at the southern limit of the “Great Superficial Massif”; these parts already being in reverse position as a result of the southward gravity-induced movement. Delmer believes that the latest stage could be the “back-glide” (or “*retroglissement*” in French) of these detached parts from south to north creating multiple tectonic stacks (“*lambeaux de poussée*”) such as the “Tombe Massif”.

References

Beugnies, 1976.
Briart, 1894a.
Briart & Bayet, 1904.
Brien, 1905a.
Brien, 1905b.
Delcambre & Pingot, 2000.
Delmer, 1997.
Delmer, 2004.
Fourmarier, 1912.
Fourmarier, 1919a.
Fourmarier, 1954.
Kaisin Jr., 1947.
Smeysters, 1880.
Smeysters, 1887.
Smeysters, 1905.
Stainier, 1922.

9.7. Vencimont Fault

with references to the Coblencienne, Vesqueville, Tillet, Baronville… faults.

Location

The Vencimont Fault, named by Asselberghs (1940b), was first introduced as the “*Faille Coblencienne*” in 1896 by Forir. Many papers have focused on the fault, which has become more and more significant over the years. The fault disrupts the axial zone of the Ardenne Anticlinorium and is considered to be one of its most significant structural discontinuities. The geographic position of the trace can be established (e.g. Asselberghs, 1946; Fig. 159), running roughly, from west to east, through the localities of Hargnies, Vencimont, Vonèche, Froidefontaine, Fays-Famenne, Smuid, Hatrival, Vesqueville and Tillet (i.e. over a distance of 84 km).

Stratigraphy and lithology of the country rocks

Asselberghs (1940b) observes three lithostratigraphic units to the south of the fault: (1) the Oignies Formation (formerly “G2a”, Lochkovian in age), (2) the Saint-Hubert Formation (formerly “G2b”, Upper Lochkovian in age) and (3) the Mirwart Formation (formerly “Sg1”, Lower Pragian in age). The three formations are composed of various siliciclastic rocks, mainly green shales, quartzites, sandstones, siltstones and slates.

The difficulty in following the eastern termination of the Vencimont Fault (Dejonghe, 2008) resides in the significant thickness of the Mirwart Formation (reaching 1050 m on the SW border of the Stavelot Inlier; Dejonghe & Hance, 2001). The absence of a lithologic marker prevents the detection of any tectonic discontinuity in this formation. The eastern termination of the Vencimont Fault therefore remains hypothetical (see below).

Geometry

The geological mapping of the Felenne-Vencimont sheet (Forir, 1896a, n°193, 1:40 000) has enabled the identification of an E-W-striking fault (Fig. 257) disrupting Lower Devonian rocks of the southern border of the Dinant basin. The fault causes the disappearance of some “coblencian” strata (i.e. Pragian and Lower Emsian terrains) to the south of Vencimont and was therefore given the name “*Faille Coblencienne*” (Forir, 1896b).

Forir (1897) makes further observations of the Coblencian Fault during the mapping of the Prondrôme-Wellin sheet (n°194, 1:40 000). The Coblencian Fault *sensu* Forir (1896a, 1897) is located between a point 2800 m west of Vencimont and a point located between the Lesse and Lhomme valleys. The fault reaches 24.5 km in length. Forir (1896b) summarises that the fault has a south dip and an offset that is comparable to that of the Eifelian Fault. The Coblencian and Eifelian faults, both of major and regional significance, are therefore very similar. However, two years later, Forir (1898) considers the Coblencian Fault to be less important than he previously envisaged (i.e. as of local significance).

During his work on the Lower Devonian of the axial zone of the Ardenne Anticlinorium, Asselberghs (1940b) correctly recognizes the Coblencian Fault to the north of Vencimont where Forir identified it in 1896. However, Asselberghs does not succeed in following the fault trace of Forir and proposes different ideas. As the geometrical and interpretative views of Asselberghs (1940, Fig. 258) are very different from those of Forir (1896), Asselberghs (1940) renames the Coblencian Fault as the Vencimont Fault.

The Vencimont Fault is characterized by three main segments (from west to east, Fig. 258): (1) a first part or western termination, striking in an east-west direction, observed between the Hugne valley on the French-Belgian border and Froidefontaine; (2) a second part or eastern termination (later connected to the Opont Fault), again striking east-west but located farther to the south where it can be traced between Naomé and Libramont; and (3) a final third segment with a N-S strike that connects the other two segments.

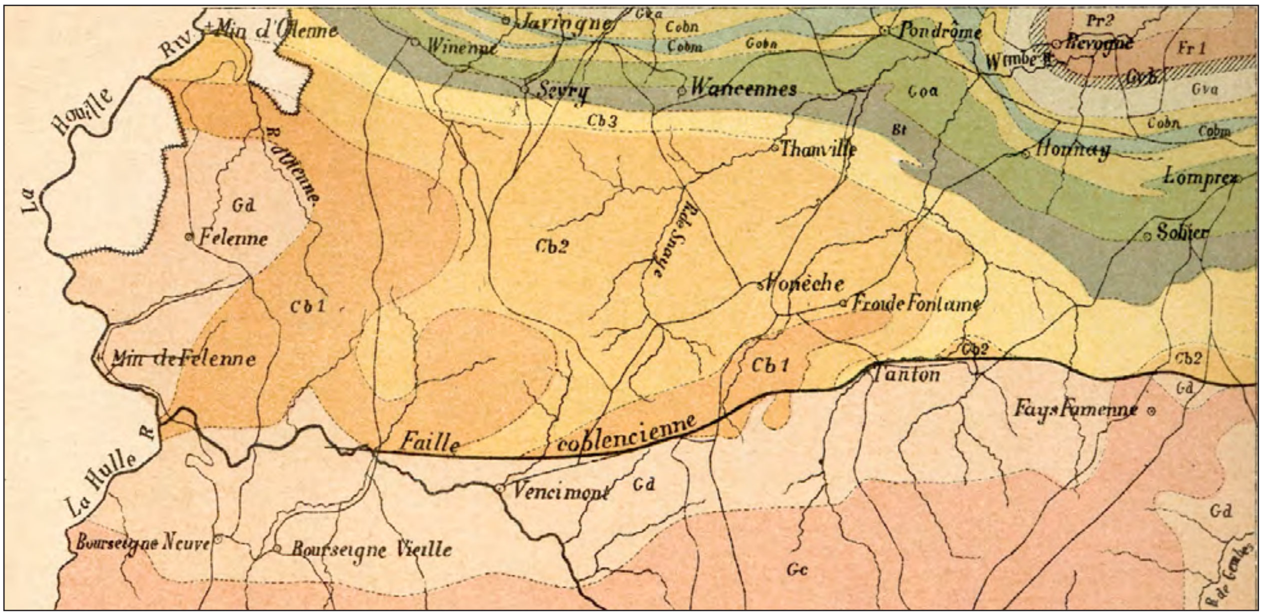

Fig. 257. Geological map of the Felenne - Vencimont area showing the trace of the Coblencienne Fault (Forir, 1896b). Gc & Gd = "*Gedinnien*" (i.e. Lochkovian). Cb1, Cb2 & Cb3 = "*Coblencien*" (i.e. Lower Devonian). Bt = "*Burnotien*" (i.e. Upper Emsian-Eifelian). Coa & Cob = "*Couvinien*" (i.e. Eifelien). Gva & Gvb = Givetian. Fr1 & Fr2 = Frasnian.

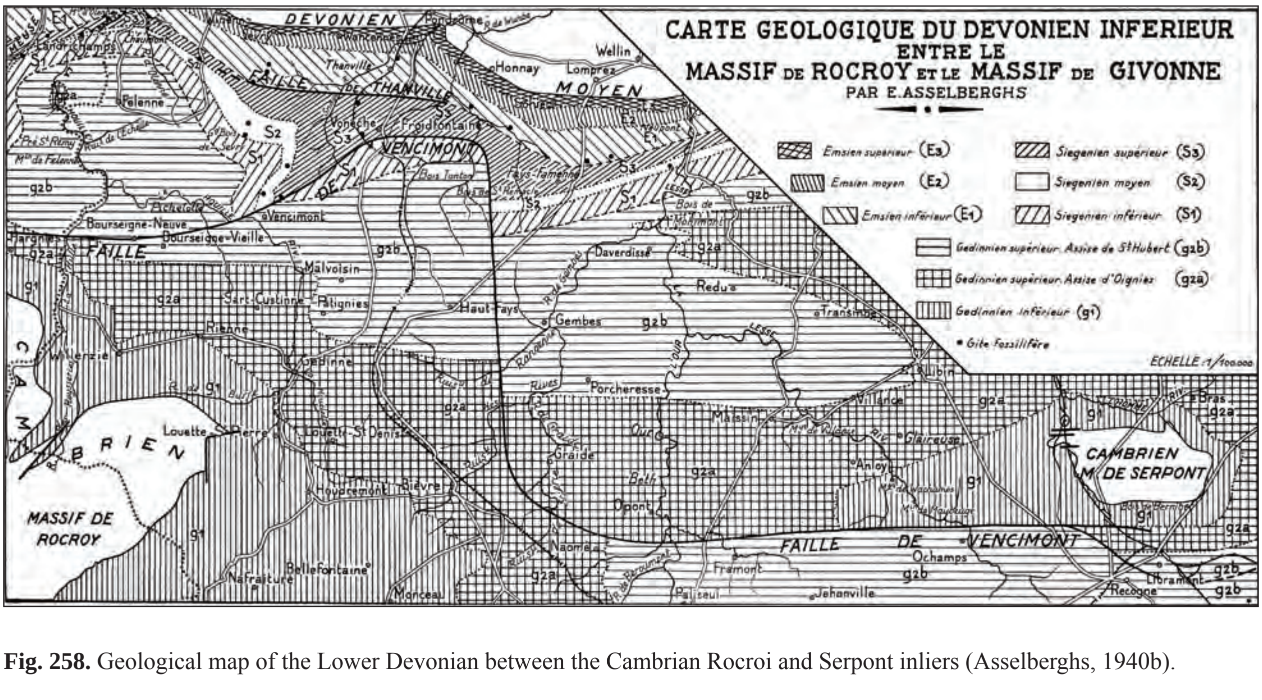

Fig. 258. Geological map of the Lower Devonian between the Cambrian Rocroi and Serpont inliers (Asselberghs, 1940b).

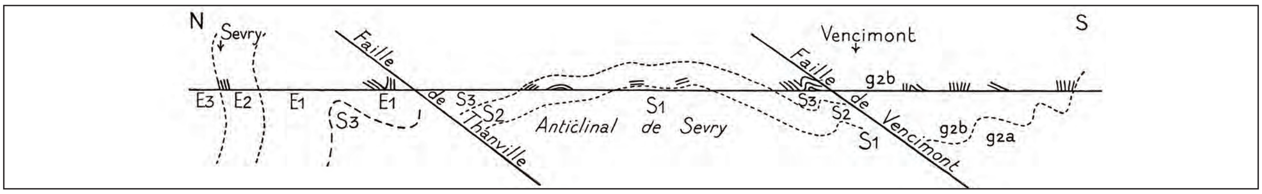

Fig. 259. Cross-section between Sevry and Vencimont (Asselberghs, 1940b). G2a = Oignies Formation (Lochkovian). G2b = Saint-Hubert Formation (Upper Lochkovian). S1 = Lower Pragian. S2 = Middle Pragian. S3 = Upper Pragian. E1 = Lower Emsian. E2 = Middle Emsian. E3 = Upper Emsian.

The cross-section (Fig. 259) at the longitude of Vencimont indicates a moderate southerly dip as well as an upward movement of the southern hanging wall block. Asselberghs (1940) estimates the reverse (thrust-type) offset to be about 2000 metres. The anomalous stratigraphic contact consists of Upper Lochkovian rocks thrust to the north over the Upper Pragian terrain.

In 1944, Asselberghs suggests a truly different geometrical point of view in which the Vencimont Fault (*sensu* Asselberghs, 1940) is subdivided into two major and distinct longitudinal fractures: the Vencimont Fault to the north and the Opont Fault to the south (see Fig. 237, Opont Fault, section 9.5). The arguments that enable Asselberghs to extend the Vencimont Fault in an easterly direction beyond Froidefontaine to Tillet (instead of being directed in a southward curve firstly to Naomé and then to the south of the Cambrian Serpont Inlier) are as follows:

- Fourmarier (1911) identifies a fracture in the area southwest of Saint-Hubert. The fault is unnamed and not well understood. Fourmarier draws its trace over a distance of nearly 10 km through the localities of Hatrival and Vesqueville. The fault marks a discontinuity between the Oignies Formation to the south and the Saint-Hubert Formation to the north; both Lochkovian in age. A dip to the south is envisaged.
- In 1943, Fabry proposes a revision of the trace of the unnamed fault of Fourmarier (1911). The fracture is again recognized from Vesqueville (in the east) to Hatrival (in the west) and is, moreover, extended to the west as far as the Lhomme valley (Fig. 260). The fault is therefore traced for a distance of 20 km.
- Between 1940 and 1943, Asselberghs (1944) focuses on the fault of Fourmarier (1911). The lineament is identified over a distance of 23 km from Smuid in the west to Tillet in the east and is given the name of the Vesqueville Fault.
- Taking into account the work of Fabry (1943), Asselberghs (1944) indicates that the Vesqueville Fault continues farther westward along the fault of Fabry. Beyond Smuid, the fracture continues to the Lhomme valley where a distance of only 7 km separates it from the Vencimont Fault farther again to the west. Asselberghs establishes, therefore, a connection between the first segment of the Vencimont Fault (*sensu* Asselberghs, 1940b) and the fault of Fourmarier (1911) and Fabry (1943). The complete fracture keeps the name of Vencimont Fault.

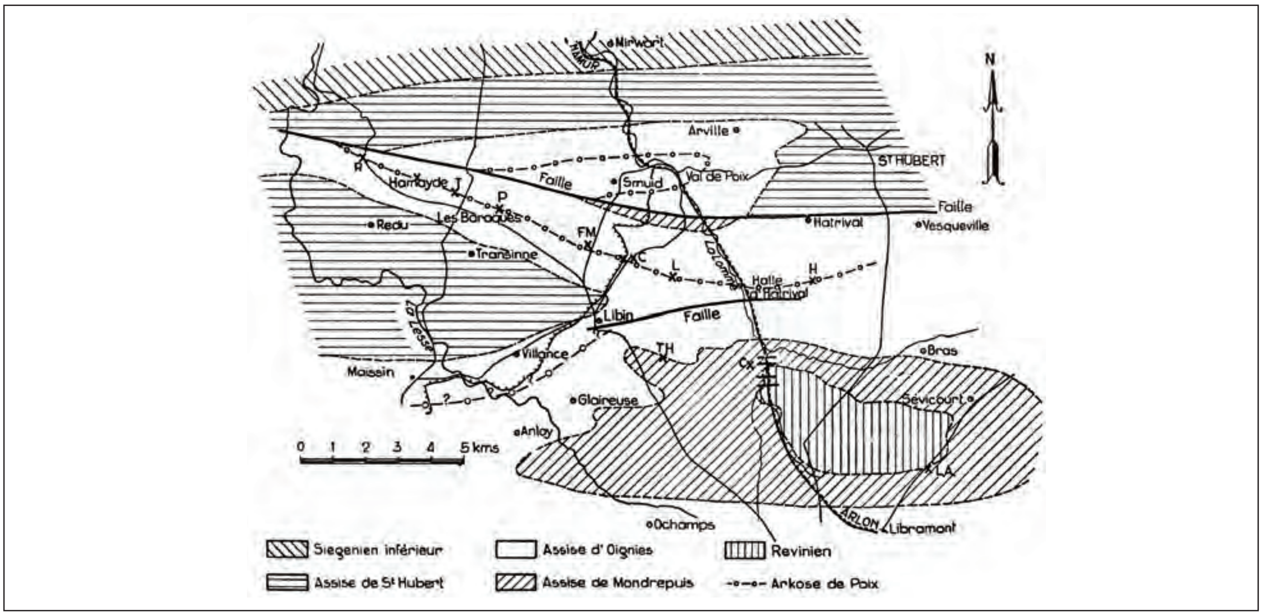

Fig. 260. Geological map of the Ardenne in the Libin region (Fabry, 1943). The distribution of kaolin deposits is presented. The interesting feature of the Fabry's work is the revised trace of an unnamed fault discovered by Fourmarier in 1911.

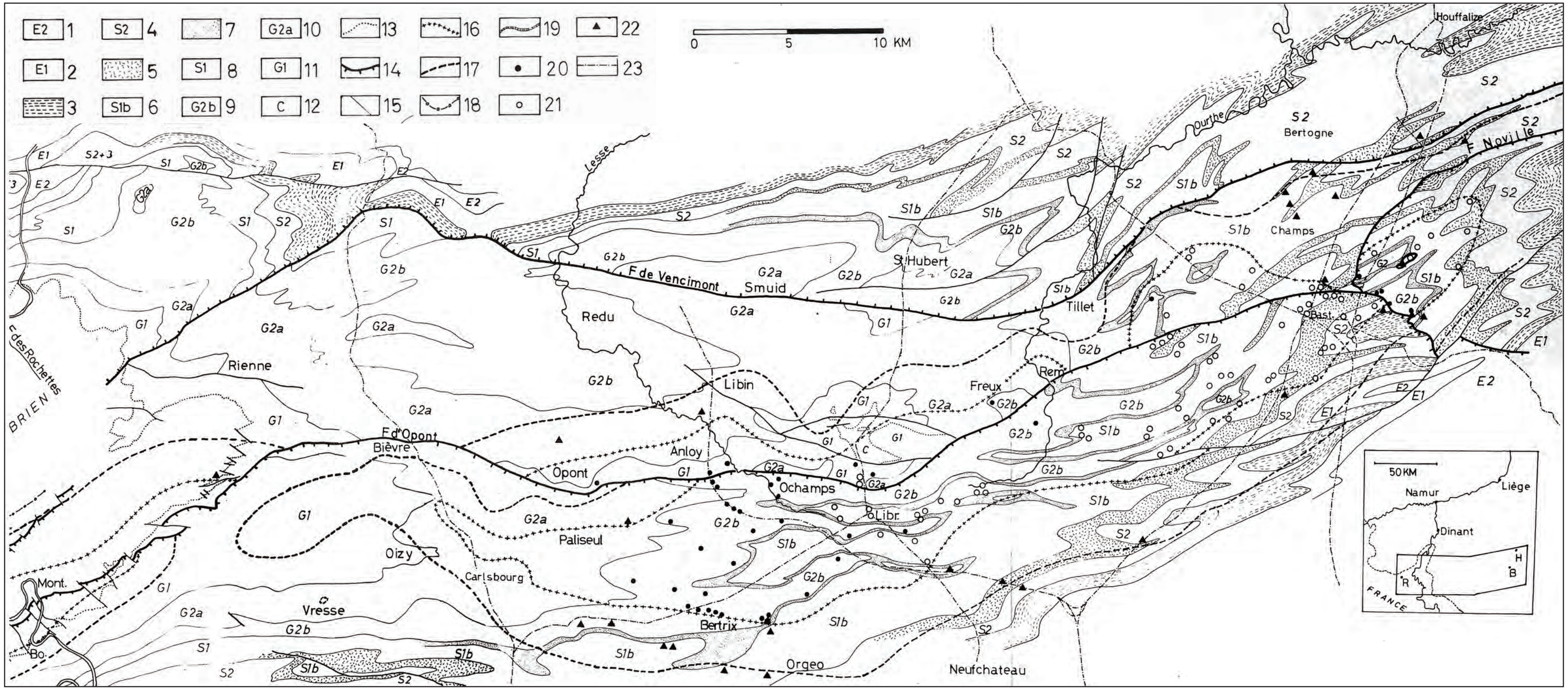

Fig. 261. Locations of the Vencimont and Opont faults (Beugnies, 1986a). 1. Upper Emsian or Winenne Formation. 2. Lower Emsian or Vireux Formation. 3. Upper Siegenian (Pragian) or Neufchâteau Formation. 4. Middle Siegenian (Pragian) or Longlier Formation. 5. Lower Siegenian (Pragian) or Alle Formation. 6. Lower Siegenian (Pragian) or Mohret Formation. 7. Lower Siegenian (Pragian) or Verlaine Formation. 8. Undifferentiated Lower Siegenian (Pragian). 9. Upper Gedinnian (Lochkovian) or Saint-Hubert Formation. 10. Upper Gedinnian (Lochkovian) or Oignies Formation. 11. Lower Gedinnian (Lochkovian) or Mondrepuits Formation. 12. Undifferentiated Cambrian. 13. The great unconformity of Ardenne. 14. Subhorizontal fault. 15. Subvertical fault. 16. Limit of the internal zone of metamorphism. 17. Limit of external zone of metamorphism. 18. North limit of the extent of amphibole-diabases. 19. Diabase sill or dyke. 20. Corneite. 21. Amphibolite and associated garnet rocks. 22. Pyrophyllite rock. 23. Main road.

Following the view of Asselberghs (1944, 1946), the E-W striking and south-dipping Vencimont Fault is recognized for a distance of 60 km from Hargnies in the west to Tillet in the east (see Fig. 237, Opont Fault, section 9.5). It has a maximum reverse offset in the region between the Lesse and Lhomme valleys that decreases laterally (Asselberghs, 1940b, 1944). From west to east, the horizontal component of the thrust is estimated to be 2000 m near Vencimont, 4000-4500 m near Daverdisse and 3200 m near Pironpré.

In 1966, Brühl suggests the existence of the Tillet Fault, located between the Vencimont Fault(s) *sensu* Asselberghs (1940) and *sensu* Beugnies (1986a, see below) and running from Tillet to Rachamps. The NE-striking fracture has a south dip and a trace of 20.5 km long. However, the fault is insignificant as Dejonghe (2008) shows that it coincides with a simple facies variation within the thick Mirwart Formation.

Beugnies (1986a) proposes for the anticline area of the Ardenne two major longitudinal fractures: the Vencimont and Opont-Carbonnière faults (Fig. 261). The more northerly Vencimont Fault is mapped for 84 km and the sinuous trace, located 3-4 km to the north of the Vencimont Fault *sensu* Asselberghs (1944), indicates a low-angle north dip. The Opont Fault (section 9.5) is a low-angle, south-dipping fracture.

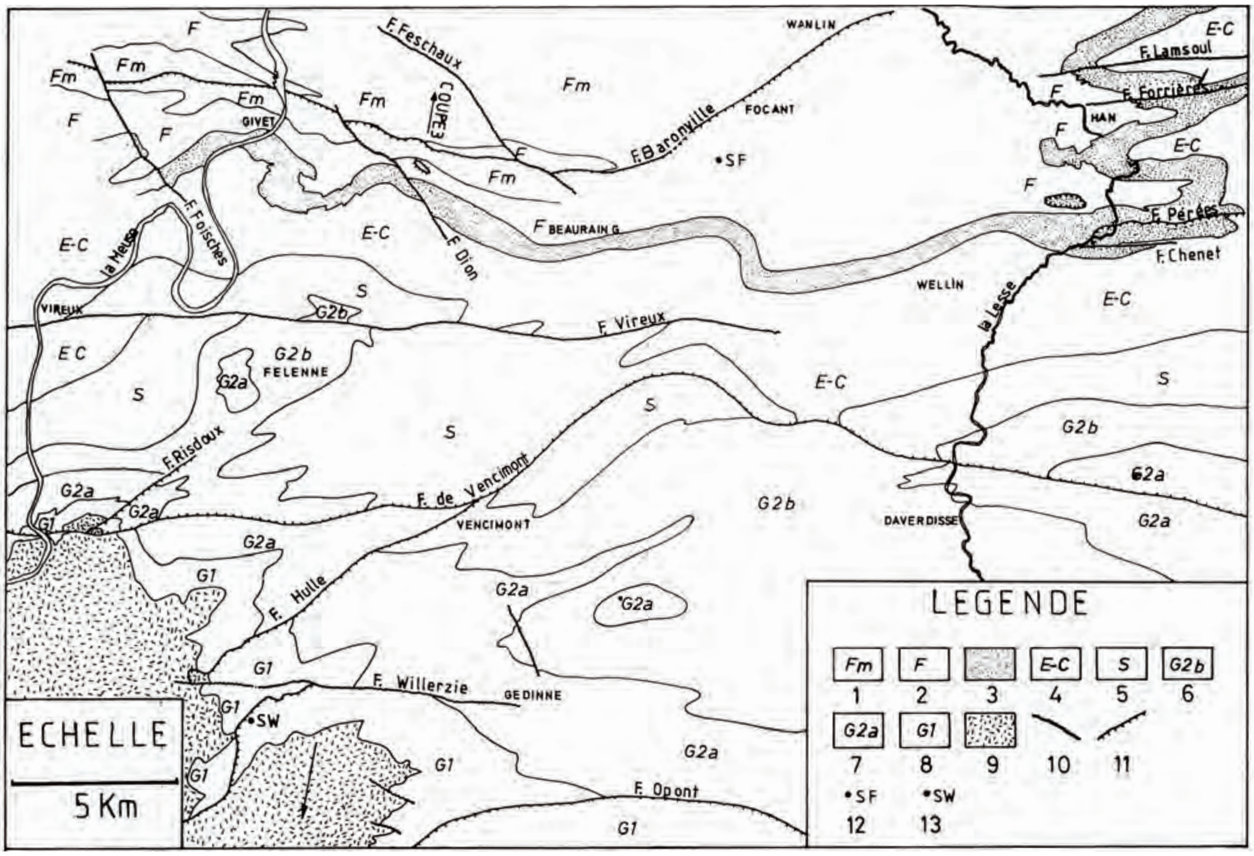

Fig. 262. Geological map of the southern border of the Dinant Synclinorium and the Ardenne Anticlinorium between the Meuse and Lesse rivers (Beugnies, 1988). 1 = Famennian. 2 = Frasnian. 3 = Givetian. 4 = Eifelian-Emsian. 5. Pragian. 6 = Upper Lochkovian (Saint-Hubert Formation). 7 = Lochkovian (Oignies Formation). 8 = Lower Lochkovian. 9 = Cambrian-Ordovician. 10 = Subvertical faults. 11 = Subhorizontal faults. 12 = Martouzin-Neuville (or Focant) borehole. 13 = Willerzie borehole.

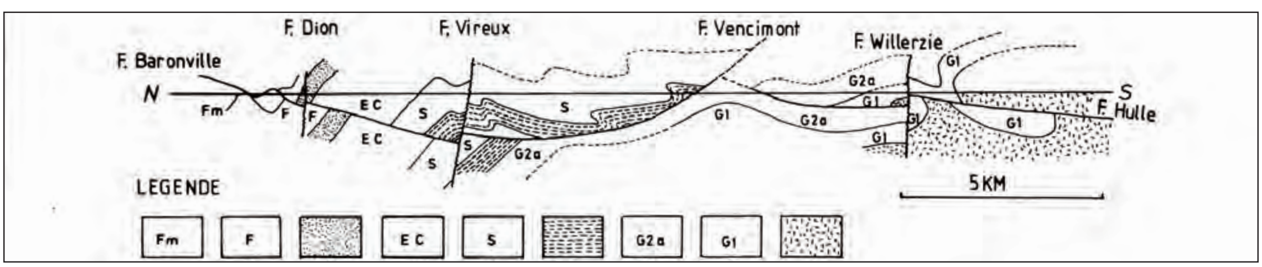

Fig. 263. Cross-section between Feschaux and Willerzie (“coupe 3” on Fig. 262 above, Beugnies, 1988). 1 = Famennian. 2 = Frasnian. 3 = Givetian. 4 = Eifelian-Emsian. 5. Pragian. 6 = Upper Lochkovian (Saint-Hubert Formation). 7 = Lochkovian (Oignies Formation). 8 = Lower-Lochkovian. 9 = Cambrian-Ordovician.

The two faults enables three tectonic units to be distinguished, which are (1) the Saint-Hubert Unit to the north of the Vencimont Fault; (2) the Opont Unit between the two fractures and (3) the Carlsbourg Unit located to the south of the Opont-Carbonnière Fault (Fig. 261). The geometrical characteristics of the Vencimont Fault are very different to those envisaged by Asselberghs (1940). The fracture shows an oblique-slip: a reverse thrust-type offset of a few hundred metres (here, the northern hanging wall block has moved upward to the south) and a sinistral strike-slip of between 2 and 12 km (Beugnies, 1985, 1986a,b).

Again in 1986b, Beugnies considers the Vencimont Fault as a major tectonic structure of the Ardenne. He indicates that no outcrop provides any direct observations of the fault but its existence can be deduced from the very different structures present on either side. The low-angle south-dip is only deduced from the aspect of the fault trace on the ground surface.

In 1988, Beugnies introduces in the literature a longitudinal fracture identified in the military domain of Baronville that marks an anomalous stratigraphic contact between Lower Famennian rocks to the south and Middle Frasnian rocks to the north. The so-called Baronville Fault is recognized on the southern border of the Dinant Synclinorium from north of Romerée (10 km west of Givet) to the vicinity east of Wanlin, i.e. over a distance of at least 26 km (Fig. 262). A distinction is made between the Feschaux and Givet units, located to the north and south of the Baronville Fault respectively. The fault plane has a low-angle dip to the south or is subhorizontal. Beugnies (1988) interprets the fracture as a reverse fault thrusting the Givet Unit in a northerly direction over the Feschaux Unit. A small tectonic window located to the north of the Baronville Fault argues for this point of view. The minimum reverse thrust-type offset is estimated to be 1 km. The Baronville Fault also displays a major dextral strike-slip component of 2.5 km.

Beugnies (1988) proposes a connection (at depth) between the Vencimont and Baronville faults that dip gently to the north and to the south respectively (Fig. 263). The latter would be the northern re-appearance of the Vencimont Fault.

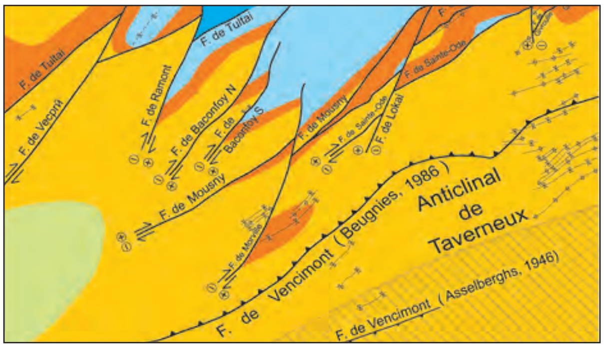

Fig. 264. Geological map of Amberloup-Flamierges (N°60/5-6) (Dejonghe, 2008). Locations of the Vencimont Fault sensu Beugnies (1986a) and sensu Asselberghs (1946) are shown.

However, Lemonne & Dumoulin (1998), who propose a revised geological map of the Agimont-Beauraing region, consider the Baronville Fault quite differently. The fracture has a steep dip to the north and a reverse offset (consequently from north to south) of 700 metres. These current geometrical considerations no longer allow a connection between the Vencimont and Baronville faults as envisaged previously by Beugnies in 1988.

Fig. 264 gives the relative positions of possible traces of the eastern extremity of the Vencimont Fault. Currently, there is no field evidence east of Tillet to confirm the presence of a major structural discontinuity. The eastward continuation of the fault beyond Tillet therefore remains hypothetical (Dejonghe, 2008).

Interpretations

Faulting within the axial zone of the Ardenne Anticlinorium is generally attributed to the contractional tectonics that acted during the Variscan Orogeny.

Beugnies (1985, 1986a) considers the Vencimont and Opont faults as representing a single strike-slip fault, which would have been subjected to bulging. The bulging of the fault would come from a late uplift movement of the axial zone of the Ardenne Anticlinorium and would also be responsible for several culmination areas such as those of Rocroi and Serpont (Cambrian). The Vencimont-Opont wrench fault is later than the folds, the schistosity and the metamorphism and results from east-west stresses.

Dejonghe (2008) does not agree with the theory of bulging of the strike-slip fault; this view being inconsistent with the sinistral wrench component assumed by Beugnies.

References

Asselberghs, 1940b.

Asselberghs, 1944.

Asselberghs, 1946.

Beugnies, 1985.

Beugnies, 1986a.

Beugnies, 1986b.

Beugnies, 1988.

Brühl, 1966.

Dejonghe, 2008.

Dejonghe & Hance, 2001.

Fabry, 1943.

Forir, 1896a.

Forir, 1896b.

Forir, 1897.

Forir, 1898.

Fourmarier, 1911.

Lemonne & Dumoulin, 1998.

9.8. Yvoir (or Hun) Fault

Location

The Yvoir Fault (or Hun Fault[2]) is located in the Yvoir vicinity. Straddling the Meuse river, it cuts the northern limb of the Namurian ESE-trending Anhée Syncline in the central part of the Dinant basin. The fault constitutes the boundary between the "Condroz Sedimentation Area" ("ASC" on Fig. 265) and the "Dinant Sedimentation Area" ("ASD" on Fig. 265) located to the north and to the south respectively (Hance et al., 2001; Fig. 265).

Stratigraphy and lithology of the country rocks

The geological map of Bioul – Yvoir (N°53/3-4, n°166) covering the Anhée basin has not yet been revised on the 1:25 000 scale. The most recent and detailed cartographic document is that of Soreil et al. released in 1908 (Fig. 266). In the Yvoir vicinity, Lower Carboniferous rocks are disrupted by the fault: the Visean in the southern block shows an anomalous contact with the Upper Famennian in the northern block. To the west, where the fault is supposed to extend (Kaisin, 1936b), the Upper Visean is thrust onto the Belgian Coal Measures.

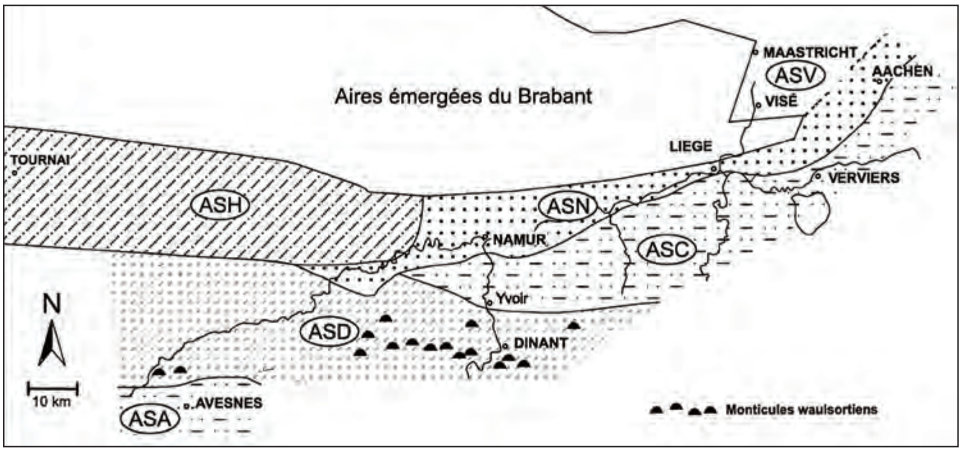

Fig. 265. Sedimentation areas of the Namur-Dinant basins during the Dinantian (Hance et al., 2001). "ASH" = Hainaut sedimentation area. "ASN" = Namur sedimentation area. "ASV" = Visé-Maastricht sedimentation area. "ASC" = Condroz sedimentation area. "ASD" = Dinant sedimentation area. "ASA" = South Avesnois sedimentation area.

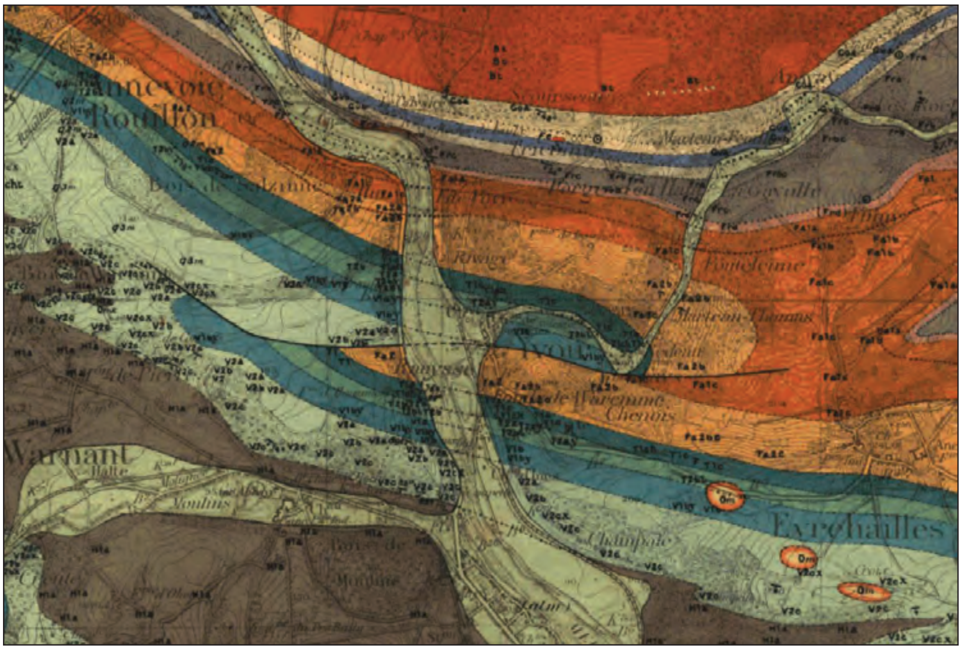

Fig. 266. Geological map of the Yvoir area (Soreil et al., 1908). Scale is given by the 4.5 km-long fault.

2 The term Yvoir Fault, preferentially used by Soreil, seems to have priority over the term Hun Fault, which is preferentially employed by de Dorlodot (Kaisin, 1936b).

Geometry

In 1901, the Yvoir Fault was already known for a long time to occur on the western bank of the Meuse river. Soreil & de Brouwer (1901) envisage its continuation on the eastern bank and again further eastwards. The main anomalous stratigraphic contact observed is of Famennian siliciclastic rocks thrust over Visean carbonate rocks. The eastern segment of the fracture has a southern dip and a reverse offset (Fig. 267).

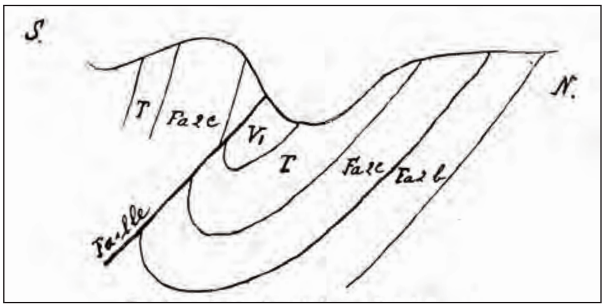

Fig. 267. Cross-section in the vicinity east of Yvoir. The eastern end of the Yvoir Fault is shown (Soreil & de Brouwer, 1901). Fa2b & Fa2c = Upper Famennian. T = Tournaisian. V1 = Lower Visean.

Fourmarier (1907) draws the Yvoir Fault over a distance of nearly 5 km (Fig. 268). In the Meuse valley the fault is characterized by an anomalous contact between Upper Famennian and Visean rocks. Fourmarier also indicates that the Anhée Carboniferous basin is bounded to the north by the Yvoir Fault and to the south by the Houx Fault. The south-dipping Yvoir Fault shows a reverse movement from south to north while the north-dipping Houx Fault presents a reverse offset acting from north to south (Fig. 271).

Soreil et al. (1908) propose a detailed map of the Yvoir Fault covering both sides of the Meuse river (geological map of Bioul-Yvoir, n°166, Fig. 266). The E-W striking fault trace is approximately 4.5 km long.

In 1936b, Kaisin focuses on the western continuation of the Yvoir Fault. He points out numbers of thrust fractures located to the west of the Yvoir lineament and considers these to be continuations of it. From east to west, from the Meuse river to the area northwest of Saint-Gérard and passing through the locality of Haute-Bise, the Yvoir Fault runs for at least 12 km (Fig. 269). Kaisin envisages a probable gentle southerly dip and a relative upthrown movement of the southern hanging wall block. The reverse offset is not quantified.

The Yvoir Fault is therefore an oblique-slip fault, combining a northward thrust movement with a dextral wrench component. The relative right-lateral strike-slip of 2 km can be measured on the geological map (Fig. 266).

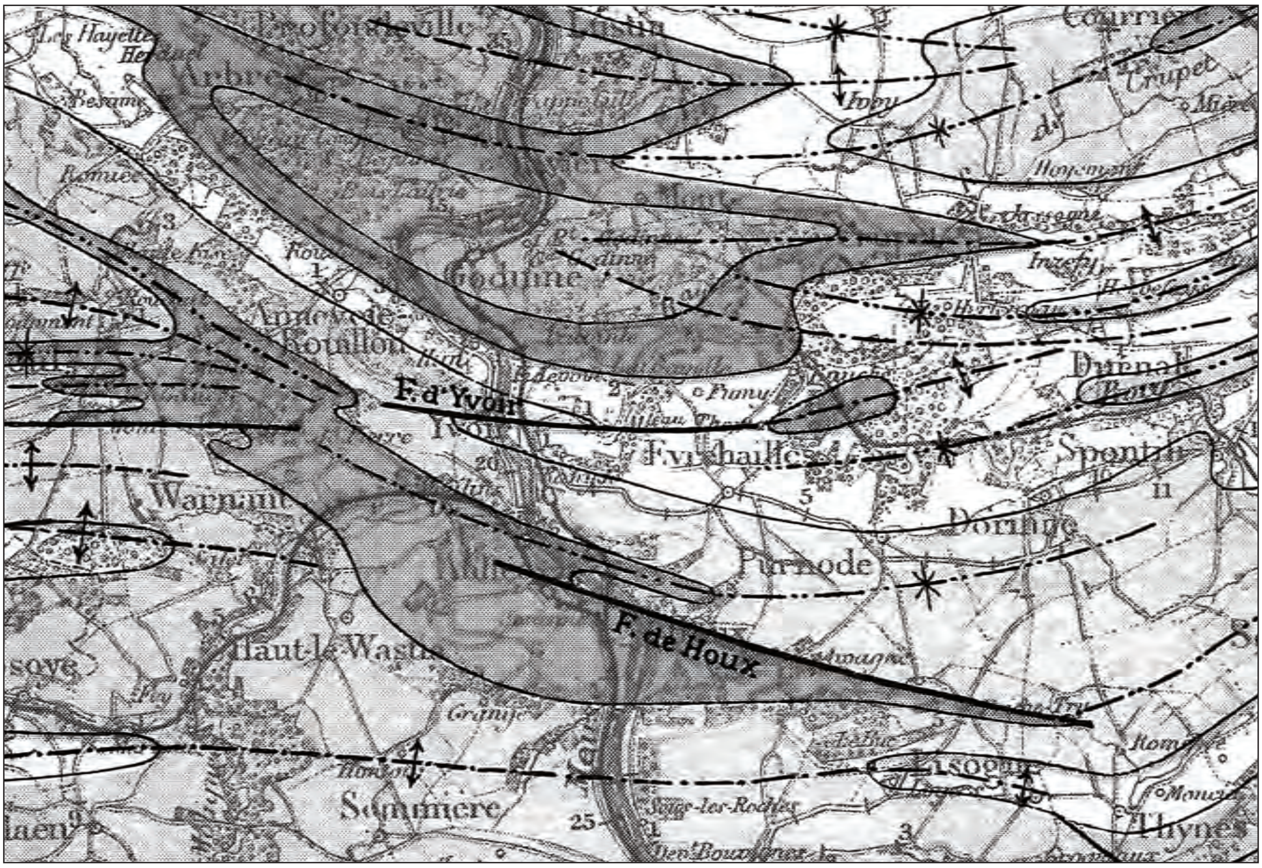

Fig. 268. Extract of the geological map of Fourmarier (1907). The Yvoir Fault strikes over a distance of 5 km. The western termination shows Carboniferous limestones on both sides of the fault, while the eastern termination displays Upper Devonian rocks on both sides.

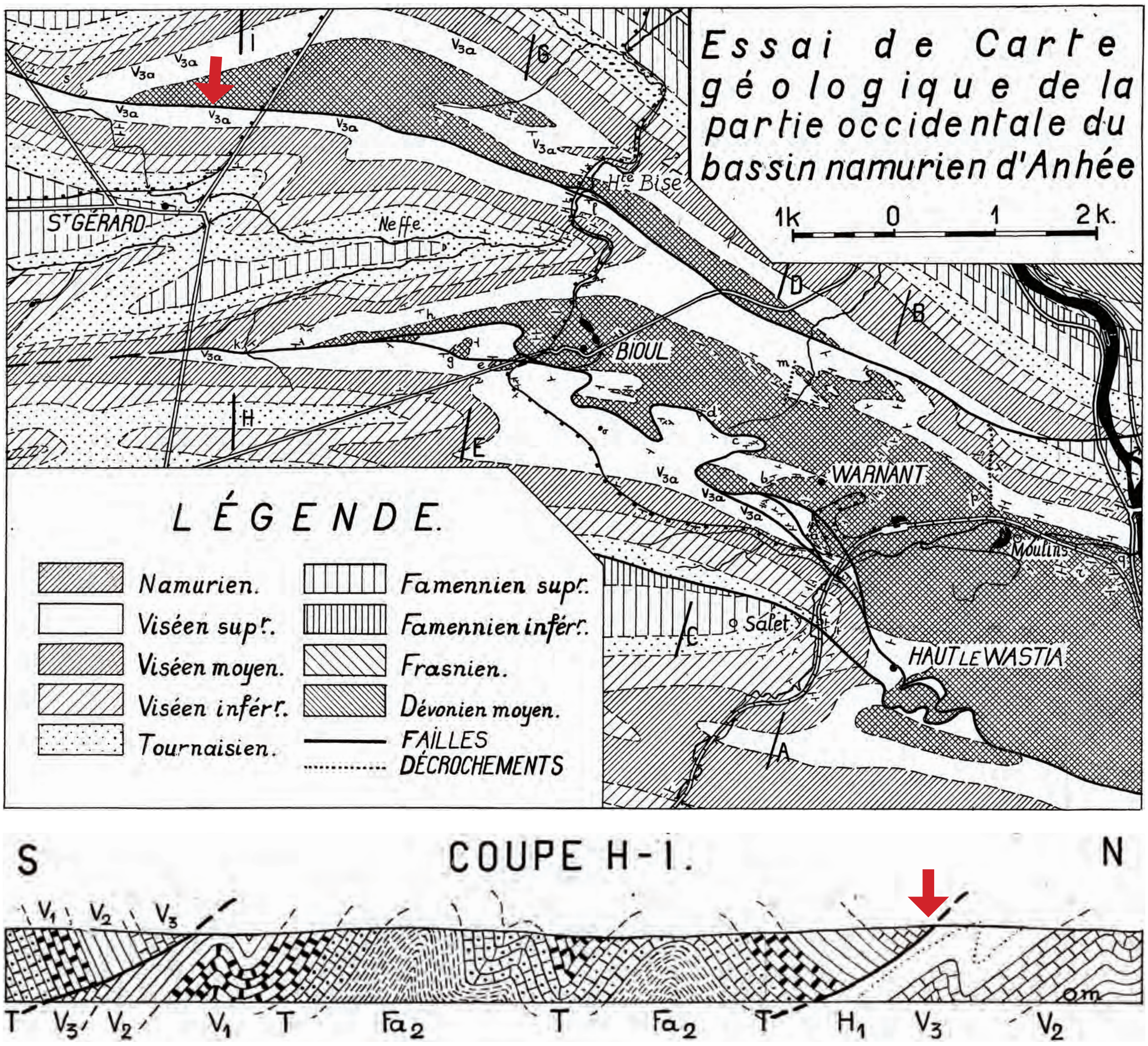

Fig. 269. Geological map of the western part of the Namurian Anhée basin and a S-N cross-section (H-I on the map) (Kaisin, 1936b). Arrows indicate the western segment of the Yvoir Fault.

In 1948, Bellière focuses on the eastern continuation of the Yvoir Fault (Fig. 270). The fault segment is 6.4 km long; the total fault distance (Kaisin, 1936b & Bellière, 1948) reaching at least 18 km. The fault dips gently to the south and displays a reverse offset. Considering the regional significance that Kaisin (1936) attributes to the fault, Bellière envisages a (hypothetical) connection between the Yvoir Fault and the major overthrusting along the Midi-Eifelian Fault.

On the basis of sedimentological arguments, Hance *et al.* (2001) consider the Yvoir Fault as being of major regional significance. The fracture strikes for a distance of at least 57 km and separates two major sedimentary domains within the Namur and Dinant basins: the "Condroz Sedimentation Area" to the north ("ASC" on Fig. 265) and the "Dinant Sedimentation Area" ("ASD" on Fig. 265) to the south.

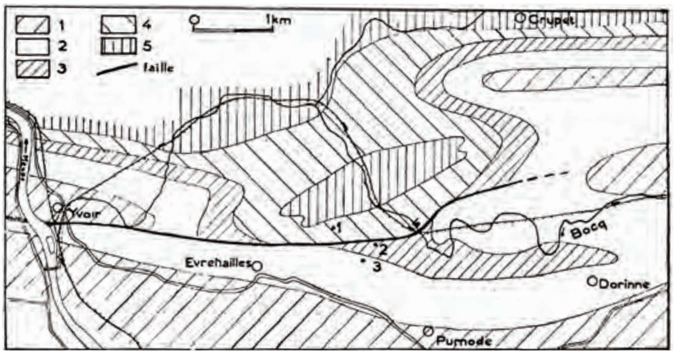

Fig. 270. Geological map of the eastern segment of the Yvoir Fault (Bellière, 1948). 1 = Carboniferous limestones. 2 = Montfort Sandstones ("Fa2a"). 3 = Esneux "Psammites" (i.e. an old Belgian term for micaceous sandstones) ("Fa1c"). 4 = Famenne schists ("Fa1ab"). 5 = Frasnian.

Interpretations

Soreil & de Brouwer (1901) consider the Yvoir Fault as resulting from the accentuation of an anticline occurring during the general (i.e. Variscan) folding of the Belgian basins.

Fourmarier (1907) also envisages the Yvoir Fault as due to the accentuation and breaking of an anticline during the northerly Variscan shortening. Keeping in mind that the Anhée basin is limited by two antithetic and mainly E-W-striking faults (the south-dipping Yvoir Fault to the north and the north-dipping Houx Fault to the south, Fig. 268 & 271), Fourmarier (1907) interprets the tectonics as pop-up structures. This interpretation assumes thrust movements along both faults with the upward extraction of a tectonic wedge (i.e. the Anhée basin). The central parts of the Dinant Synclinorium, some of the most compressed terrains, preferentially deform into pop-up structures. The tectonics are related to the Variscan Orogeny and to the major northward overthrusting of the Ardenne Allochthon.

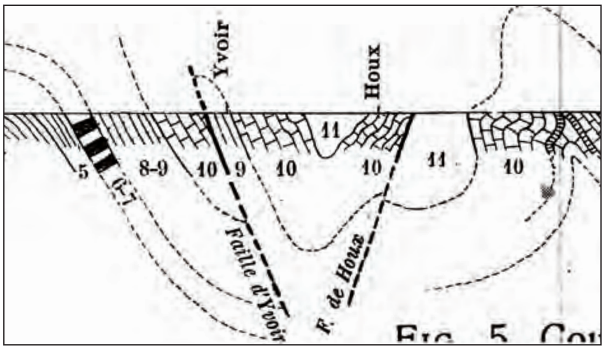

Fig. 271. Extract from the N-S geological cross-section through the Meuse river (Fourmarier, 1907). 5 = Eifelian. 6 = Givetian. 7 = Frasnian limestones. 8 = Frasnian shales and Lower Famennian. 9 = Upper Famennian. 10 = Carboniferous limestones. 11. Upper Carboniferous ("*Houiller*").

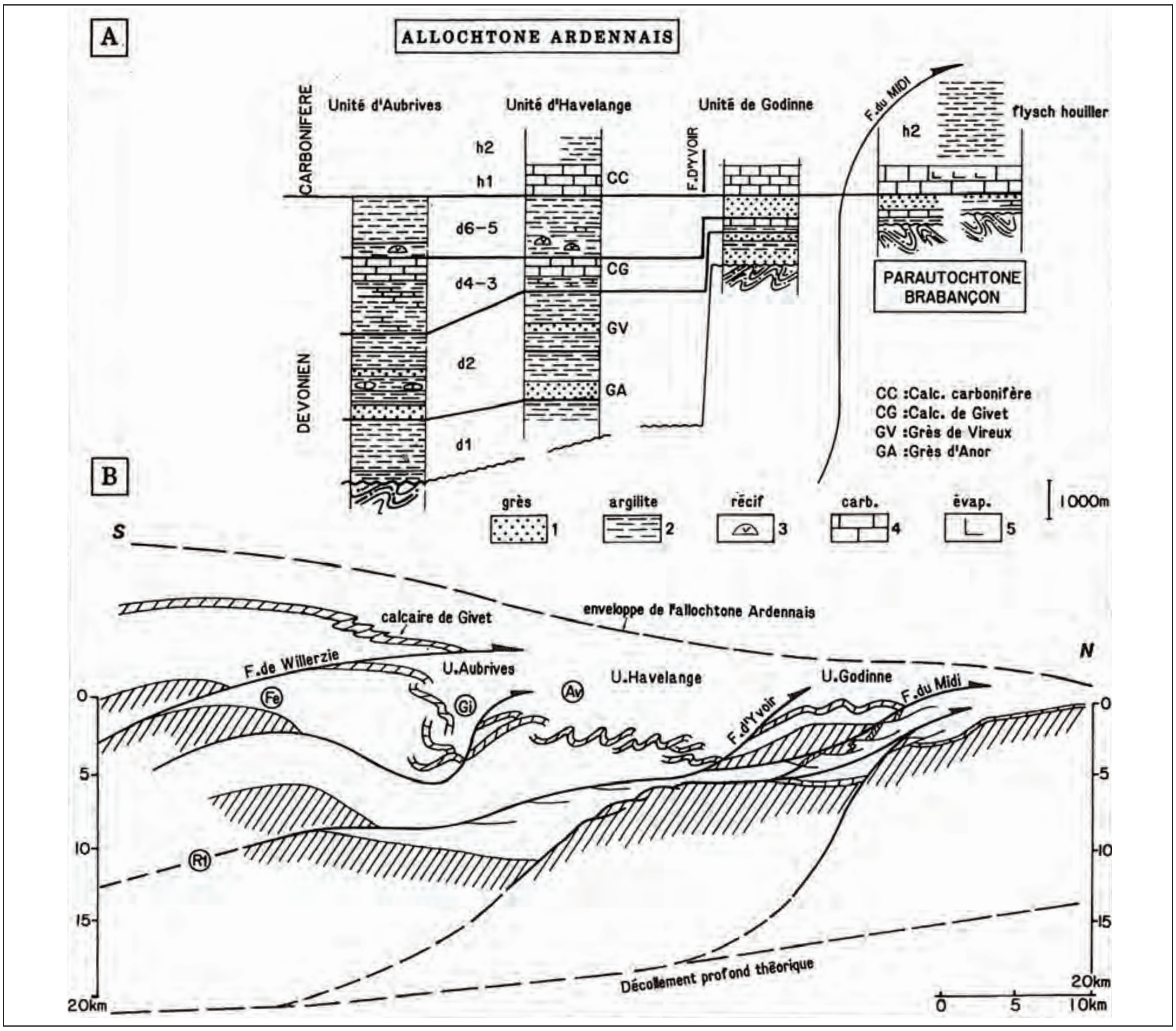

Fig. 272. Lithostratigraphy and tectonic structure (N-S cross-section) along the Meuse valley (Meilliez & Mansy, 1990). Fe = Fépin. Gi = Givet. Av = Avesnes. R1 = major reflector.

In 1985, the Hun (i.e. Yvoir) Fault is considered as limiting the Godinne Unit to the north (i.e. the frontal unit of the Ardenne Allochthon) and the Dinant Unit to the south (Raoult & Meilliez, 1985). The fault dips to the south and displays a thrust-type displacement. The authors also indicate that the reverse offset is difficult to define and could be either moderate or significant. Considering a moderate displacement, the fault would coincide with the reverse reactivation of a former palaeogeographic bank or slope limiting a downthrown block to the south; while a significant displacement would involve a westerly continuation of the Yvoir Fault within the Silurian of the "Puagne Band" and its connection with the Midi Fault.

In 1990, Meilliez & Mansy consider the Yvoir Fault as playing a major role in the progressive establishment of the Ardenne Allochthon. As already stated in the description of the Midi Fault (section 9.4), before 1990 the Ardenne Allochthon was correlated with an initial single tectonic domain which was passively transported along the Midi-Eifelian Fault. The model proposed by Meilliez & Mansy is of progressive deformation with successive positioning of major tectonic stacks from south to north. This model takes into account many thrust fractures within the allochthon of which the Yvoir Fault is one. Fig. 272B shows the Yvoir Fault as connecting at depth with the "R1" major seismic reflector that itself connects further to the south with the "theoretical deep décollement".

To the north of the Yvoir Fault (the Godinne Unit), stratigraphic succession is condensed and arenite-dominant, while to the south (the Havelange Unit), the series are thick and pelite-dominant (Fig. 272A). Meilliez & Mansy (1990) therefore interpret the Yvoir Fault as acting in a normal sense during the sedimentation then reactivated in a reverse sense during the Variscan Carboniferous shortening.

Adams & Vandenberghe (1999) have doubts regarding the synsedimentary character of the Yvoir Fault. Indeed, no obvious facies or thickness variations are observed on either side of the fault. The Yvoir Fault is interpreted as a thrust starting from a detachment plane at depth and stepping up progressively to the surface with a low-angle dip (see Fig. 230, Midi Fault, section 9.4).

Hance et al. (2001) propose the Yvoir Fault coincides with a major limit of sedimentary areas (Fig. 265) that corresponds to a former normal synsedimentary fault active during the Dinantian period. The fault is later reactivated in a reverse sense during the Variscan shortening.

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Kaisin, 1936b.

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Soreil & de Brouwer, 1901.

Soreil et al., 1908.

10. Map and table synthesis (part 2)

The map in Fig. 273 provides a synthesis of the faults described in Cambier & Dejonghe (2010) and in this volume. A summary table of the major geometric data for these faults is given in Table 4.

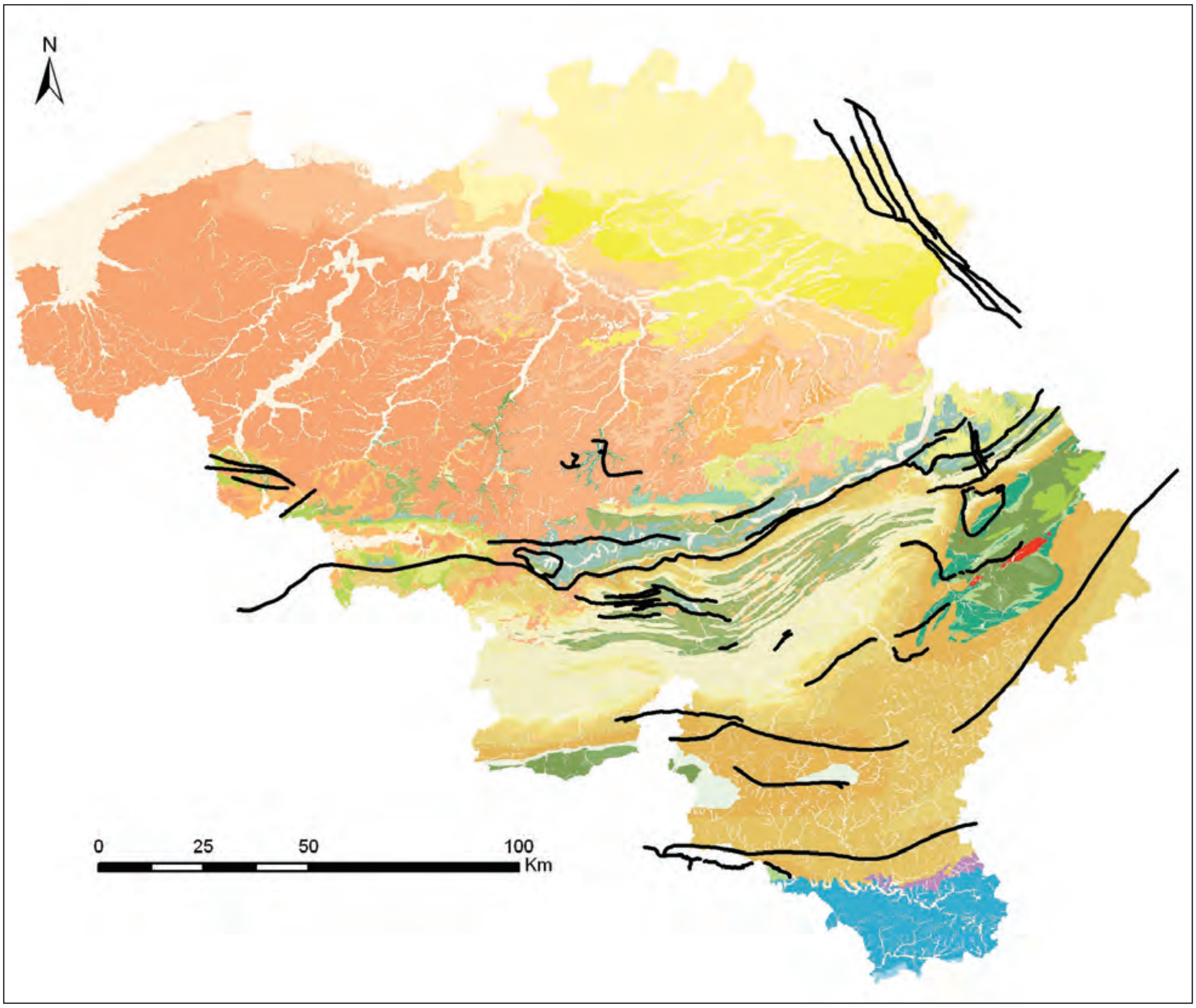

Fig. 273. Map of the faults studied in this work. The faults described by Cambier & Dejonghe (2010) are also given. The data used on this map are the most recent or the more coherent. The lithostratigraphic background is modified from http://www.onegeology.org and the legend corresponds to the International Stratigraphic Chart (http://www.stratigraphy.org).

Table 4. Summary table of the main structural features of the faults studied by Cambier & Dejonghe (2010) and described in this work. The data used are the most recent or the more coherent. The direction (strike) of the fault trace is a general trend (L = longitudinal; T = transverse). The dips constitute local observations and correspond generally for thrusts and normal faults to the minimum and the maximum values observed respectively. The strike-slip is given where this constitutes the main component of the offset.

Name	Length (km)	Strike	Dip	Nature	Dip-slip (m)	Strike-slip (m)
Aguesses-Asse	17	L, WSW-ENE to SW-NE	Ag: 30°S; As: 13°S	reverse (thrust)	1100	
Aiglemont	7	L, E-W	gentle S	reverse (thrust)	>10,000	
Amerois	14	T, NW-SE	20-30 to 45°SW	dextral, reverse	200	1200-1600
Boussale	15	L, SW-NE	gentle S	reverse (thrust)	?	
Bruyelle	13	L, WNW-ESE to E-W	subvertical or steep N	N block downthrown	220	
Centre	46	L, E-W	40° S	reverse	1200	
Court-Saint-Etienne	1.5	circular (klippe)	subhorizontal	reverse (thrust)	see Orne Fault	
Denée-Thynes	24	L, WSW-ENE to WNW-ESE	45°S	reverse	900	
Dondaine	16	L, E-W	70°N	reverse	60	
Feldbiss Fault Zone	75	NNW-SSE to NW-SE	70-85° NE	normal	various, a few meters	
Gaurain-Ramecroix	21	L, WNW-ESE to NW-SE	80°S	reverse	160-170	
Genappe	> 50	L, sinuous	5°N	reverse (thrust)	several km	
Hanzinelle-Biesmerée	16	L, E-W	50°S	reverse	225	
Hanzinne-Wagnée	20	L, E-W	45-55°S	reverse	300	
Haversin	1.5	L, SW-NE	?	reverse	?	
Herbeumont	75	L, E-W to WSW-ENE	10-15° S	reverse (thrust)	10,000	
La Roche	> 10	L, NW-SE to SW-NE	75°S	reverse	700	
Lamsoul	21	L, SW-NE to WSW-ENE	70-75°S or subvertical	normal	1000	
Landenne	14.6	L, WSW-ENE	60°N	reverse	920	
Malsbenden	90	L, SSW-NNE	N	reverse	?	
Mettet	9	L, E-W	45°N	reverse	100	
Midi	> 220	L, E-W to WSW-ENE	gentle S	reverse (thrust)	40,000 to 150,000	
Molinia	5.5	L, SW-NE	75-80°S	reverse, sinistral	50	125
Monty	13	T, N-S to NNW-SSE	subvertical or steep E	normal	90	
Mouhy	10	T, N-S to NNW-SSE	60-70°W	sinistral, normal	28	100
Opont(-Carbonnière)	28(-93)	L, E-W	20-35° S	dextral, normal	2000-3000	18,000
Orne-Noirmont-Baudecet	35 - 50	L, sinuous	5°N	reverse (thrust)	several km	
Ostende	7.5	T, N-S to NNW-SSE	60-70°W	sinistral, normal	50	300
Oster	15	L, WSW-ENE to SW-NE	70-80°S	?	?	
Scry-Bois de Neffe	13	L, E-W	45°S	reverse	50	
Soiron	14	L, WSW-ENE	25 to 45°S	reverse (thrust)	800-1200	
Theux	30	circular (window)	10-15°various	reverse (thrust)	2000-3000 to 5000	
Thozée-Responette	10	L, E-W to WSW-ENE	40-45°S	reverse	100	
Thy	1.9	L, E-W	gentle S	sinistral	?	kilometric
Tombe	7	L, sinuous	gentle S	reverse (thrust)	11,000	
Tunnel	16	L, WSW-ENE	20-25°S	reverse (thrust)	see Theux Fault	
Vaulx	10	L, E-W to WNW-ESE	60°N or steep S	dextral	12	?
Vencimont	60	L, E-W	gentle S	reverse (thrust)	4500	
Vêves	4	L, SW-NE	70-80°S	reverse	25	
Vezin	3.2	L, WSW-ENE	S	dextral, normal	weak	560
Vireux	28	L, WSW-ENE to E-W	70°S to subvertical	normal	375	
Walhorn	40	L, SW-NE to E-W	10-15°S	reverse (thrust)	900	
Xhoris	40	L, sinuous	S	reverse (thrust)	5000	
Yvoir	18	L, WNW-ESE to E-W	gentle S	reverse (thrust)	?	

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