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Stratigraphy of the Lower Palaeozoic of the Brabant Massif, Belgium. Part II: The Middle Ordovician to lowest Silurian of the Rebecq Group

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ABSTRACT. Multidisciplinary researches in the last 25 years and recent geological mapping of the Brabant Massif have completely changed our knowledge about one of the most poorly known parts of Belgian geology. The sedimentary succession is now surprisingly complete compared to what was written in the literature before the 1970s. It ranges in age from the lower Cambrian to the top of the Silurian, and is very thick (>13 km). This highlights the need to produce an up-to-date stratigraphy. In this second part about the Middle Ordovician to lowest Silurian, we describe in detail the formations, which are combined into a new group, how the description of the units evolved through time, their lithology, sedimentology, boundaries and contacts, thickness, fossil content and type sections or most typical outcrop areas. The new Rebecq Group comprises 10 formations from the Abbaye de Villers to the Brutia formations. The sedimentation in Megasequence 2 begins in a shelf environment with the Abbaye de Villers and Tribotte formations. The Rigenée Formation marks a rapid deepening that leads to the slope and/or deeper water deposits of the Rigenée, Ittre, Bornival, Cimetière de Grand Manil, Huet and Fauquez formations. Thereafter, an abrupt change of bathymetry marks the top of Megasequence 2 and leads to the shallow shelf deposits of the Madot and Brutia formations. The igneous activity, represented by interbedded volcanic to volcano-sedimentary rocks and magmatic intrusions, reached a peak in the Madot Formation, which forms the base of Megasequence 3. This group shows a moderately thick, between 1500 to 2000 metres, mostly siliciclastic succession. Some shelly facies in the Upper Ordovician show the rapid drift of Avalonia to low latitude and the warm Boda Event that precedes the Hirnantian glaciation. A chronostratigraphic comparison with the Central Condroz Inlier shows that the succession there is also almost complete from the Middle Ordovician (Huy Formation) to the lowermost Silurian (Bonne-Espérance Formation) and that a short stratigraphic hiatus marks the top of Megasequence 2. This comparison shows that, since the early Cambrian, the sediments of the Brabant Massif and the Condroz Inlier were deposited in the same Brabant-Condroz sedimentary Basin.

KEYWORDS: Brabant Massif, stratigraphy, Avalonia, Ordovician, graptolite, chitinozoan

1. Introduction

A brief introduction to the regional geology can be found in Part I of our previous paper: "The Cambro-Ordovician from the Halle and Ottignies groups" (Herbosch & Verniers, 2013). Figure 1 of this paper gives a synthetic history of the Cambro-Ordovician stratigraphic evolution of the Brabant Massif from the pioneering work of Malaise (last version 1911) to the more recent synthesis of Verniers et al. (2001). A complete history of the geological discoveries in the Brabant Massif can be found in Debacker (2001). It should also be noted that before the 70s the stratigraphic nomenclature was different in the outcrop area of each river basin and the upper part of the Ordovician formations was particularly poorly known.

The purpose of this paper is to present the new stratigraphic terminology and to describe the new groups, formations and members, and recount how the description of the units evolved through time, their lithology, sedimentology, boundaries and contacts, thickness, fossil content and, where possible, type sections or most typical outcrop areas. This new stratigraphic scheme has been recently accepted by the National Commission of Stratigraphy (sub-commission Lower Palaeozoic; NCS, 2012). In this second part, we continue with the description of the 10 formations that cover the Middle and Upper Ordovician and the lowest part of the Silurian. This time interval covers all of the Megasequence 2 and the lowest part of Megasequence 3 recognized in the Brabant Massif (Vanguestaine, 1992; Verniers et al., 2002a; Linnemann et al., 2012). Fig. 1 gives a schematic view of the geological subcrop map of the Brabant Massif (after De Vos et al., 1993; Debacker et al., 2004a) with the northern limit of the outcrop zone and the position of some other figures.

The chronostratigraphy and time scale by Gradstein et al. (2012) are followed throughout the paper. A figure showing the geographical locations of the rivers, towns, villages, highways and railways referred in the text is given at the end of the paper (Fig. 13).

2. The Rebecq Group

We introduce here formally a new group with a new name. Contrary to the Ardennes Inliers, the stratigraphical succession of the Brabant Massif has not been subdivided in groups or only for simplification purpose (as in De Vos et al., 1993). The latter terminology was not often used and abandoned in the stratigraphic overview by Verniers et al. (2001).

We include in the Rebecq Group ten formations from bottom to top: the Abbaye de Villers, Tribotte, Rigenée, Ittre, Bornival, Cimetière de Grand-Manil, Huet, Fauquez, Madot and Brutia formations. They all essentially comprise terrigenous sediments: sandstone, siltstone and often dark coloured slate: grey, dark grey, black and more rarely grey-brown or grey-green. Some shelly facies appear in the upper part (Huet and Madot formations) with also interstratified volcanic, volcano-sedimentary and intrusive magmatic rocks (especially in Madot and Brutia formations). The name of the new group comes from the town of Rebecq, in the Senne valley where several of the type sections of these formations are situated. The overlying Orneau Group is not composed of a very contrasting facies as it is also formed by terrigenous sediments of different colour, but it is mostly turbiditic with some hemipelagic to pelagic facies.

2.1. Abbaye de Villers Formation

2.1.1. Definition

The name Abbaye de Villers (N of the town of Villers-la-Ville, Thyle valley) was adopted after Anthoine & Anthoine (1943): « *Quartzophyllade siliceux de Villers* » Ar3 de l'Assise de Villers-la-Ville (Fig. 13). This formation was also described in the Senne valley as a part of the "quartzophyllades zonaires de Quenast" by Beugnies (1973) and as the lower part of the "Quenast Formation" by André et al. (1991) and Servais et al. (1993). The oldest definition was formally adopted by Verniers et al. (2001).

2.1.2. Description

The formation is formed by grey or dark grey fine-grained sandstone to mudstone, with a regular to irregular, lenticular to wavy plane parallel cm-scale lamination (Plate 1, photo 1). Locally, some pyrite and also siderite were observed that corroborate the suboxic nature of the sedimentary environment. Characteristically, the fine-grained sandstone laminae have rather diffuse limits. Frequent and abundant bioturbations, mostly parallel to the bedding planes (Plate 1, photo 2), disturb or even erase the laminar structure. Curious plurimetric structures occur frequently in the formation, some have been interpreted as

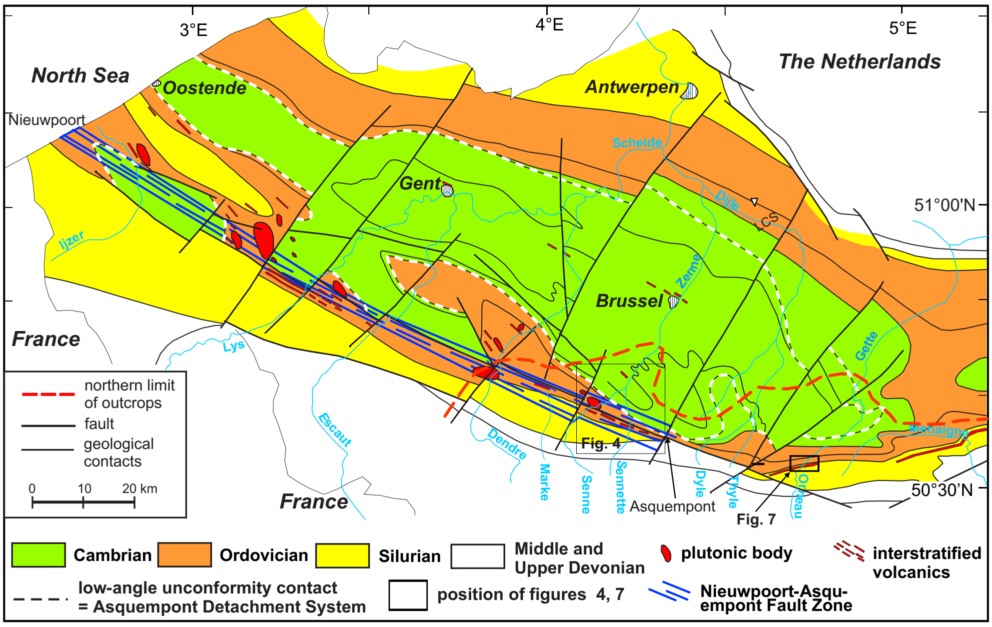

Figure 1: Geological subcrop map of the Brabant Massif (after De Vos *et al.*, 1993; Debacker et al., 2004a) with the location of the Figs 4, 7.

large-scale bedforms (megaripple and/or sandwave; Herbosch & Verniers, 2002 fig. 17) and others as synsedimentary slumps. The latter structures have been studied in detail in the Sennette valley (Debacker, 2001; Debacker et al., 2003), in the Dyle valley (Beckers, 2004; Beckers & Debacker, 2006) and also in the Senne valley in the particularly demonstrative outcrops near Quenast (Debacker & De Meester, 2009) (Plate 1, photo 3). These slumps are present throughout the formation and also in the lower half of the Tribotte Formation (see below § 2.2.2). These authors observed that in most cases these folds pre-date the cleavage and can be interpreted as slump folds, with a southwards oriented slump direction. This interpretation implies that during the Middle Ordovician (Abbaye de Villers to Tribotte formations) a regionally persistent S-dipping palaeoslope existed in this part of the Brabant Massif (see also Debacker, 2012). This direction is the opposite of that observed in the Chevlipont Formation which shows a N-dipping palaeoslope (Beckers, 2004).

At Thy (Dyle valley, Fig. 13) in outcrops along a sunken path facing the Thy castle, de Magnée & Lambot (1965) described a 2 metres thick phosphatic and manganiferous conglomerate interstratified with the sandstone of the Tribotte Formation. During the survey for the Nivelles-Genappe map (Herbosch & Vanguestaine, 1994; Herbosch & Lemonne, 2000) more conglomeratic levels were found and the outcrop could clearly be re-assigned to the Abbaye de Villers Formation. More recently, the outcrop was studied by Debacker et al. (2009) who described an alternation of fine sandstone and silty mudstone with many folds and faults. Four conglomeratic levels occur in the western part of the section where they follow the bedding disposition. The conglomerates are fine-grained (clast< 2 cm) and range from clast-supported to matrix-supported. The clasts are sub-parallel to bedding and include in decreasing order of abundance: 1) black, rounded to subangular, ellipsoidal phosphate nodules with abundant porphyroblasts of spessartite garnets; 2) angular, elongate clasts of siltstone to sandstone similar to the surrounding beds; 3) subangular monocrystalline quartz fragments (de Magnée & Lambot, 1965). Unfortunately there were no recent mineralogical and geochemical investigations at these levels. Sedimentologically the conglomerates are clearly of intraformational origin. Detailed structural analysis shows the presence of pre-cleavage folds and faults that are interpreted as slump-related deformation features. In this respect, the conglomerate beds may well correspond to debris flow deposits. Statistical analysis of slump transport directions suggests slumping from 320°-005° towards 140°-185° (Debacker et al., 2009 p. 1536). The observations from the synsedimentary slumps and the deformation of the conglomerates confirm the presence of a S-dipping palaeoslope of regional extent during the Middle Ordovician in the southern part of the Brabant Massif (cf. § 2.2.2).

The Abbaye de Villers Formation is more resistant to alteration and erosion than other Ordovician formations. As a result, when a river incises through this formation, the valley newly created will be narrow forming a kind of defile. This is particularly visible in the Senne valley about 750 m north-east of Quenast, in the Coeurq valley (Fig. 13) where a waterfall 4-5 m high is present at the old Coeurq mill (Herbosch et al., Ittre-Rebecq map 2013) also in Ways (Genappe) in the Dyle valley (Fig. 13) and in the type locality in this valley around the site of the medieval Abbaye de Villers (Herbosch & Lemonne, Nivelles-Genappe map, 2000).

The contact with the underlying Chevlipont Formation could nowhere be observed by lack of outcrops showing the boundary. However there is a very distinct sedimentological change from a deep marine environment (continental slope) in the Chevlipont Formation (Herbosch & Verniers, part 1 of this paper, 2013) towards an outer to mid-shelf environment in the Abbaye de Villers Formation. This break in sedimentary environment corresponds to a stratigraphic hiatus of about 14-15 Ma (Fig. 2) and is interpreted as the result of an uplift possibly coeval with volcanism (in the Chevlipont Formation, Linnemann et al., 2012). The event is contemporaneous with the drift of Avalonia away from Gondwana (Cocks & Fortey, 2009).

In the Sennette valley, along the canal, just south of the Asquempont fault (Debacker et al., 2003 fig. 12) a problematic member is present with decimetric grey sandstone beds, often showing convolute lamination, alternating with dark to medium grey siltstone and slate beds, interpreted as quite distal turbidites of the Bouma-type. The stratigraphic position is still problematic because of its lower and upper contacts are faulted. This unnamed member is interpreted (Verniers, unpublished) on the basis of acritarch and chitinozoan biozonations as either a lateral facies of the upper part of the Abbaye de Villers Formation or as a unit in between the Abbaye de Villers and the Tribotte formations. However on sedimentological grounds this interpretation is difficult to follow because of the quickly changing sedimentary evironment. Because of its turbiditic facies the member resembles more the younger Ittre Formation (Herbosch, unpublished).

2.1.3. Extent of the outcrop zones and thickness

The Abbaye de Villers Formation has a thickness of about 150 to 180 m, more than 130 m exposed in the section along the road Abbaye de Villers to La Roche (Beckers & Debacker, 2006 fig. 2). But this latter section does not show the base of the formation. It is one of the most widely outcropping formations: it crops out in the Senne Basin (comprising the Senne, Coeurq and Sennette valleys, Figs 4, 13) and in the Dyle Basin (comprising the Dyle, Thyle valleys and some of their tributaries, Herbosch & Verniers, 2013 fig. 10) and the problematic member crops out in the Sennette valley (Asquempont).

2.1.4. Bio- and chronostratigraphy

No macrofossils have been found in the formation. Amongst the chitinozoans in the lower third (in fact about the middle third after the geological map of Herbosch & Lemonne, 2000) of the formation in the Dyle Basin *Eremochitina brevis* is present (Samuelsson & Verniers, 2000). It is a type species present in a N-Gondwana assemblage belonging to the "middle" Arenig (Paris, 1990). The species *Lagenochitina obeligis*, *Euconochitina vulgaris* and *Cyathochitina* cf. *dispar,* found in the upper parts of the formation (Dyle and Senne basins), are less indicative, but suggest an age between middle Arenig and early Llanvirn (Samuelsson & Verniers, 2000). Martin (1976), Vanguestaine et al. (1989) and Vanguestaine (*in* André et al., 1991) reported the presence of the acritarchs *Frankea sartbernardensis* var. A., *Adorfia firma*, *Frankea hamata* var. A, etc… The genus *Frankea* is considered to have its first appearance in the upper Arenig.

A review of the acritarchs from the Senne Basin (Vanguestaine, 2008) and a renewed study on old and new samples from the Dyle basin (with samples carefully selected from the entire formation by A. Herbosch) by Vanguestaine & Wauthoz (2011) show a relatively homogeneous assemblage of species corresponding to the *Frankea hamata*-*Striatotheca rarrigulata* Zone of the English Lake District (Cooper et al., 2004). The zone is characterized by the co-occurrences of *F. hamata*, *F. breviuscula* and *S. rarrigulata*. *Adorfia firma* and *Arbusculidium filamentosum* are also typical species of this assemblage dated as upper Fennian, it corresponds in the most recent chronostratigraphy to the lower Darriwilian (Cooper & Sadler, 2012 fig. 20.1). In more detail, the lowest part of the formation yields *Striatotheca rarrigulata* that clearly indicates an upper Arenig age and the middle and upper parts show most of the species of the *Frankea hamata*-*Striatotheca rarrigulata* Zone of the Lake District dated as upper Fennian (upper Arenig). According to the acritarch evidence, the middle part of the Abbaye de Villers Formation is placed somewhat higher than it is according to the chitinozoans evidence. Giving more confidence to acritarch evidences, an upper Dapingian to earliest Darriwillian age (*circa* 468-465 Ma, Cooper & Sadler, 2012 fig. 20.1) is usually indicated for the formation (Fig. 2), but a lower Dapingian is not totally excluded.

An important time gap, of about 14 Ma, is thus proven to be present between the Chevlipont and the Abbaye de Villers formations (Fig. 2). This gap is more important than previously thought (Verniers et al., 2002a). In the Wépion borehole, a 5 cm basal conglomerate (Graulich, 1961; Vanmeirhaeghe, 2006a; Owen & Servais, 2007) between the Chevlipont and Huy formations corresponds with a slightly more important hiatus (from lowest Tremadocian to mid Darriwilian, *formosa* Zone, *circa* 16 Ma). This observation confirms the presence of an unconformity at that level present in all the Brabant-Condroz Basin (see under chap. 4 and fig. 11).

In the unnamed member the acritarch assemblage is similar to that of the Abbaye de Villers Formation and indicates an Arenig-Llanvirn age (Vanguestaine, 1978). The chitinozoans contain *Desmochitina ornensis* and *Conochitina pseudocarinata,* an assemblage occurring in Brittany above the Armorican Quartzite Formation (Paris, 1981) indicating a mid Arenig age (Samuelsson & Verniers, 2000).

2.1.5. Stratotype

The best outcrops are in the Thyle valley just E of the old Abbaye de Villers, in the railway cutting from km 38.7 (50°35'35.28" N/ 4°31'52.83" E) to km 39.0 (50°35'22.77" N/ 4°31'52.46" E). Good discontinuous outcrops have been studied in detail by Beckers & Debacker (2006 fig. 2) along the road from Abbaye de Villers to La Roche, from a curve in the road at: 50°35'35.28" N/ 4°31'52.83" E to an old quarry at the rear of the "Café de la Forêt" at: 50°35'28.35" N/4°31'36.98" E (Fig. 13). We propose that these outcrops to be recognized as the stratotype of about the upper two-third parts of the Abbaye de Villers Formation. The lower part of the formation is only partly visible in small outcrops about 100 to 150 m E of the railway in the Bois de l'Hermitage (Thyle valley, Fig. 13).

2.2. Tribotte formation

2.2.1. Definition

The Tribotte Formation was defined from the locality of Tribotte south of Villers-la-Ville in the Thyle valley (Fig. 13). Anthoine & Anthoine (1943): «*Psammites de Tribotte*» Ar2 de l'Assise de Villers-la-Ville. Later, Michot (1977) grouped in the «Assise de Tribotte » the «Psammite de Tribotte» and the «grès et psammites de Strichon» of Anthoine & Anthoine (op. cit.). The latter definition was adopted by Verniers et al. (2001). In the Senne valley, the formation forms part of the "quartzophyllades zonaires de Quenast" of Beugnies (1973) and the upper part of the "Formation de Quenast" of André et al. (1991) and Servais et al. (1993).

2.2.2. Description

In the Dyle Basin, this formation shows three main lithofacies easily recognizable in the field. A first facies, in the lower third, contains brownish grey, clayey fine sandstone and siltstone with coarse laminations and is strongly bioturbated. Some beds show metrical scale oblique stratification interpreted as large-scale bed forms like those in the Abbaye-de-Villers Formation (Herbosch & Verniers, 2002 fig. 17). Thin section examination shows abundant potassic feldspar and plagioclase in the sandstone (Plate 1, photo 4) at the base of the formation in the Thyle valley (Jodart, 1986; Herbosch *in* André et al., 1991). The upper two thirds contain a second facies with yellowish grey to dark grey sandstone and siltstone, clearly more clayey than the underlying facies. Bioturbation is rather strong with predominance of oblique and vertical burrows ("fucoides" of the older literature). This upper part shows typical structures such as flaser bedding, vertical bioturbation with spreiten indicating an intertidal environment of deposition (Plate 1, photo 5). A third facies can locally be observed near Strichon ("grès et psammites de Strichon" of Anthoine & Anthoine, 1943) in the upper part of the formation: it is a rather mature, bioturbated, light grey to white sandstone. This sandstone facies is probably a lateral variation of the more frequently occuring intertidal facies.

In the Senne valley the outcrops are not so well exposed as in the Dyle valley and the facies look sandier. At Quenast, on the northern slope of the Senne valley, a long outcrop shows numerous slump and large-scale bed forms that have been studied in detail by Debacker & De Meester (2009). They concluded that in most cases these folds are pre-cleavage and can be interpreted as slump folds (Plate 1, photo 3). This interpretation shows that during the Middle Ordovician (Abbaye de Villers to Tribotte formations) a regionally persistent S-dipping palaeoslope existed in the southern part of the Brabant Massif (cf. § 2.1.2).

The lower and upper boundaries of the formation have both been observed. However exactly pinpointing the lower limit of the formation is very difficult because of the gradual transition from the Abbaye de Villers to the Tribotte formations. The transition shows an upward increasing sand fraction and a change of colour from dark grey to brown.

2.2.3. Extent of the outcrop zones and thickness

The formation crops out in the Senne (badly, Figs 4, 13) and Dyle basins (Herbosch & Verniers, 2013 fig. 10), while in the Orneau valley only temporary excavations have been observed (N of Gembloux; Delcambre et al., 2002; Herbosch, unpublished). The thickness is estimated at 250 to 300 m in the Dyle Basin and 150 to 200 m in the Senne Basin.

2.2.4. Bio- and chronostratigraphy

No macrofossils were observed. A poor chitinozoan assemblage containing *Euconochitina vulgaris* indicates a mid Arenig to early Llanvirn age (Verniers et al., 1999; Samuelsson & Verniers, 2000). The assemblage of acritarchs recorded by Vanguestaine & Wauthoz (2011) from the lower part of the Tribotte Formation in the Dyle Basin is very similar to that of the Abbaye de Villers Formation and belongs to the *Frankea hamata*-*Striatotheca rarrigulata* Biozone of the English Lake District (Cooper et al., 2004). An early Darriwilian (late Arenig) age is indicated. The uppermost part of the formation yields *Frankea sartbernardensis* and possibly *Vogtlandia multiradialis* possibly indicating a

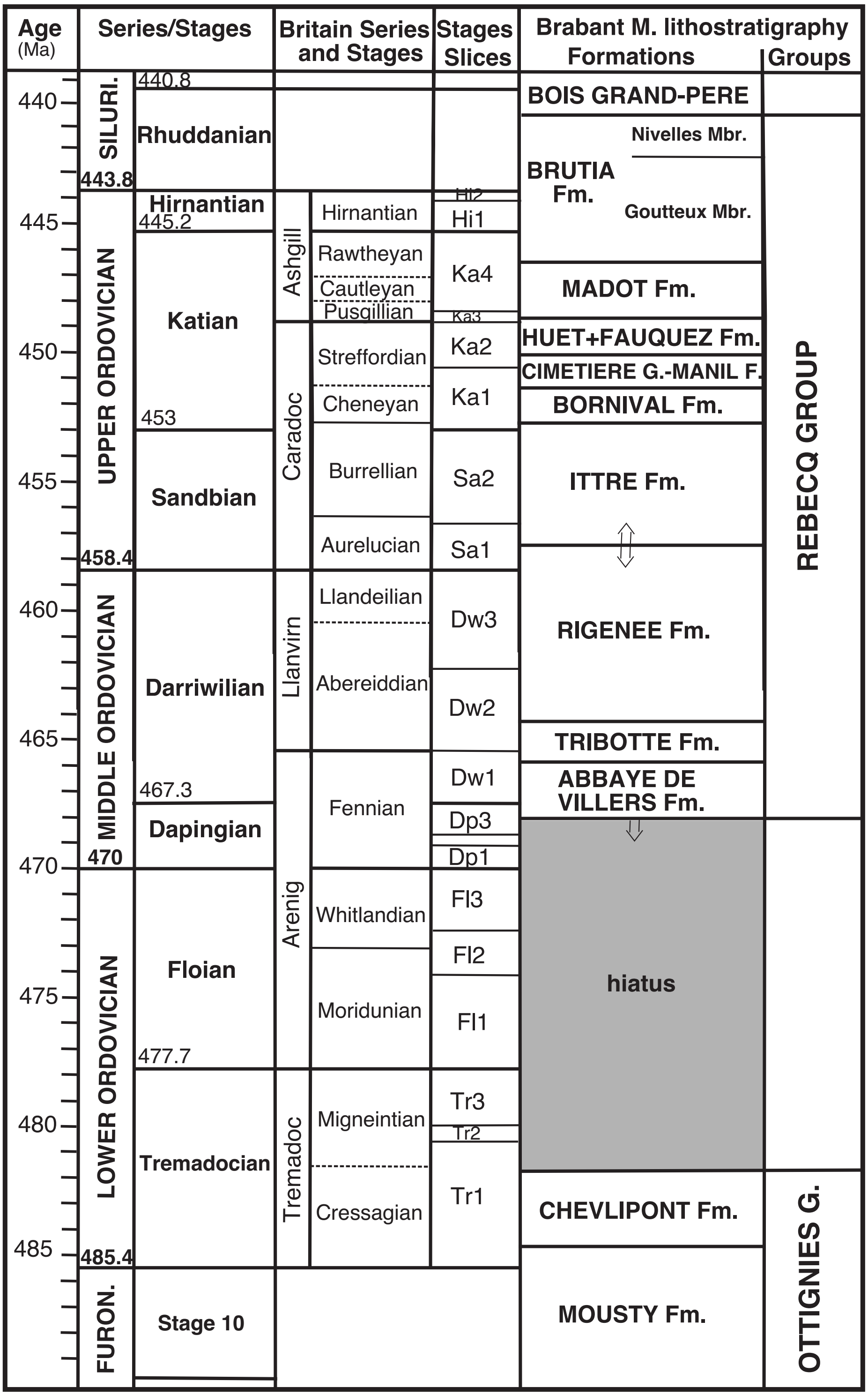

Figure 2: Chronostratigraphic position of the Ordovician and lowest Silurian lithostratigraphic units in the Brabant Massif (chronostratigraphy after Cooper & Sadler (2012). See text for explanations and references.

latest Arenig to earliest Llanvirn age (Vanguestaine & Wauthoz, 2011 fig. 4). In conclusion and taking into consideration the age of the Abbaye de Villers Formation, the Tribotte Formation encompasses a small time interval from the late Arenig to the earliest Llanvirn i.e. lower Darriwilian (Fig. 2).

2.2.5. Stratotype

No stratotype could be defined, but the type area is in the Thyle valley between the old Abbaye of Villers to the N, Villers-la-Ville in the centre, Rigenée to the SW (Vallon des Goutailles) and Gentissart to the SE (Ri de Gentissart valley) (Fig. 13). Good outcrops of the lower part of the formation could be observed in a small valley that faced the Abbaye de Villers to the west (about 50°35’243 N/4°31’34”E). A good section of the upper part (and the transition to the Rigenée Formation) can be observed along the disused railways from Court-St-Etienne to Fleurus between km 42.3 and 42.7 (50°34’11.80” N/4°33’16.09 E to 50°33’59.02 N/4°33’28.00” E) (Servais, 1991, 1993; Herbosch & Lemonne, Nivelles-Genappe map, 2000).

2.3. Rigenée Formation

2.3.1. Definition

This formation is defined after Malaise (1909): «*Assise de Rigenée Sl2b*». The hamlet of Rigenée is situated in the Thyle valley 1.5 km south of Villers-la-Ville (Fig. 13). It was used with the same meaning by Servais (1993) and Verniers et al. (2001). This formation was also named “Unité D” by Martin & Rickards (1979) and “Formation de La Tourette” by Lenoir (1987) and Servais et al. (1993) in the Senne Basin (Fig. 13).

2.3.2. Description

The Rigenée Formation comprises monotonous dark grey to bluish grey slate (mudstone and siltstone), vaguely or coarsely laminated (Plate 1, photo 6) or without any stratification, bearing locally pyrite. The formation was never studied in detail lithostratigraphically except for the Gembloux section and the section through the lower boundary in the Dyle Basin (Servais,

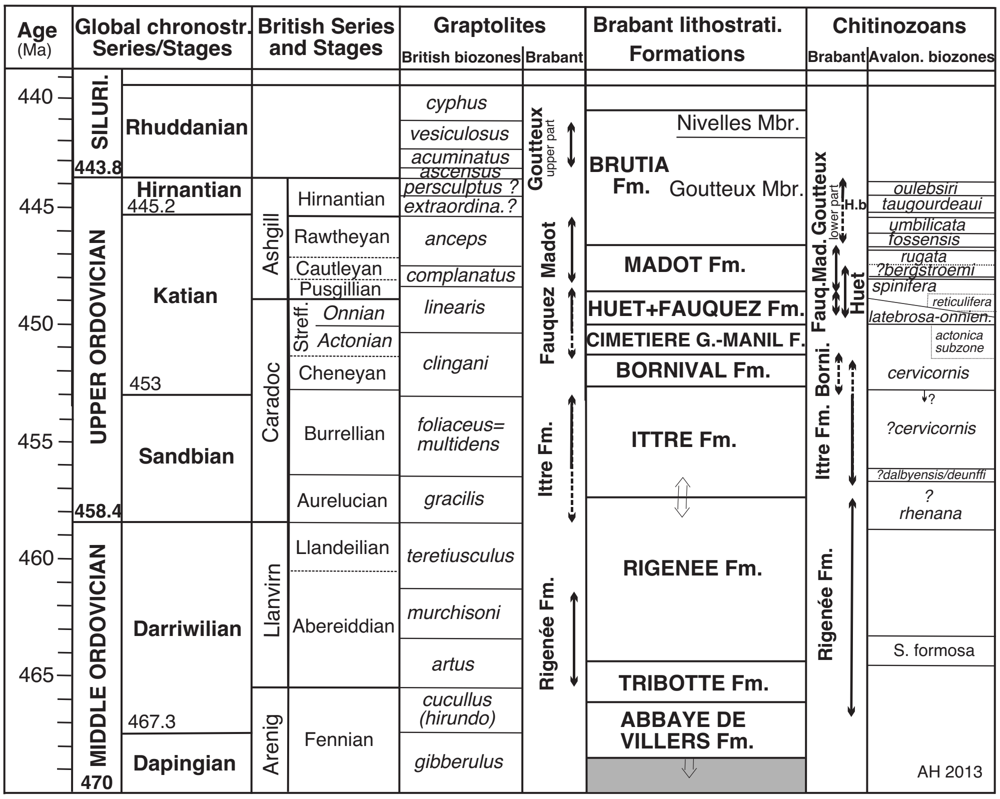

Figure 3: Comparison of the chronostratigraphic position achieved with the graptolites and the chitinozoan occurrences from the Middle Ordovician to lowest Silurian formations of the Brabant Massif. Chronostratigraphy of the global Series/Stages, of the local British Series/Stages and of the British graptolite biozones after Cooper & Sadler (2012). Composite Avalonian chitinozoan biozonation after Vandenbroucke (2008) except *S. Formosa* Zone after Cooper & Sadler (2012). Abbreviations: Avalon. :Avalonian; H.b: Harelbeke borehole; Streff.: Streffordian; *extraordinari.*: *extraordinarius*. See text for explanation and references.

1991, 1993). In the Dyle Basin, the lower boundary is marked by a transition over 5 to 10 metres and a distinct change from a pale clayey siltstone in the upper part of the Tribotte Formation to dark slate in the Rigenée Formation. The transition is so abrupt that Anthoine & Anthoine (1943) and Michot (1977) indicated a fault. The transition is clearly visible in the sunken road to Rigenée (Servais, 1991, 1993; Herbosch & Lemonne, Nivelles-Genappe map, 2000) and along the old vicinal south of Tilly (Ri de Gentissart valley; Servais, 1991; Delcambre et al., Chastre-Gembloux map, 2002). The work of Servais (op. cit.) and recent mapping show that no fault is present (Herbosch & Verniers, 2013 fig. 10). This abrupt change of lithofacies records a regional or even global sea level rise well known in North Gondwana and Baltica as the « Formosa flooding Event » (Paris et al., 2007). Indeed the Tribotte-Rigenée transition is located according to a combination of acritarchs and chitinozoans biostratigraphic data, in the lower Llanvirn *Siphonochitina formosa* Biozone, Stage slice Dw2 (Cooper & Sadler, 2012 fig. 20.9) (Figs 2, 3).

The formation was also studied in the Senne basin (Figs 4, 13): in the Bief 29 section the lower contact is faulted and the upper contact with the Ittre Formation clearly visible (Debacker et al., 2003 figs 7, 24). The formation is poorly exposed along the Virginal disused railway section (Debacker et al., 2003 figs 9, 24) and along the canal Brussels-Charleroi canal at km 39.840-39.7 (Debacker et al., 2003 figs 15, 24) where its upper and lower limits are present. The upper transition into the turbidites of the Ittre Formation can be observed in the Ri des Goutailles, just W of Rigenée (Herbosch & Lemonne, Nivelles-Genappe map, 2000).

2.3.3. Extent of the outcrop zones and thickness

The formation crops out in the Dyle and Senne basins and in the Orneau valley, within the town of Gembloux (Fig. 13). Its thickness is about 175 m in the Thyle valley (Delcambre et al., Chastre-Gembloux map, 2002), > 200 m in the Sennette valley along the bief 29 (Debacker et al., 2004b fig. 8) and estimated at about 150 to 200 m in the Orneau valley (Servais, 1993).

2.3.4. Bio- and chronostratigraphy

Graptolites from the lithological “unit D” in the Sennette valley, now attributed to the Rigenée Formation (Debacker et al., 2001), were placed by Martin & Rickards (1979) at the top of the *Didymograptus bifidus* Biozone (now *D. artus* Biozone). Study of old and new collections from the canal section at km 39,775 by Maletz & Servais (1998 fig. 4) shows that the fossils are too poorly preserved to differentiate between the *Didymograptus artus* and the *Didymograptus murchisoni* zones. The two zones correspond in the British chronostratigraphy to the Abereiddian (*circa* 466-461 Ma, Cooper & Sadler, 2012).

Acritarchs (*Frankea sartbernardensis* and *Striatotheca principalis*) from the Dyle-Thyle valleys indicated an (early) Llanvirn age for Martin (1969). Servais (1991, 1993) observed the successive appearances of *Frankea sartbernardensis*, *Arkonia virgata* and *Frankea hamulata* in the lowermost, lower and middle parts of the formation, species that are Llanvirn markers and never recorded from Arenig strata. Several proposals have been made to identify the Arenig-Llanvirn boundary based on the appearance of *F. sartbenardensis* (*in* Vecoli et al., 1999), *A. virgata* (*in* Servais, 1997) or *F. hamulata* (*in* Cooper et al., 2004). For Vanguestaine & Wauthoz (2011 fig. 4), taking all available data into account, the boundary seems to be within the Rigenée Formation (at the appearance of either *A. virgata* or *F. hamulata*). The acritach diversity in the Senne Basin is very low compared with the Rigenée Formation at Rigenée (Servais, 1991; Vanguestaine, 2008).

A poor assemblage of chitinozoans with *Lagenochitina obeligis* and *Cyathochitina calix* indicates a large age bracket from late Arenig to early Caradoc (Verniers et al., 1999; Samuelsson & Verniers, 1999, 2000 p. 114).

Taking into account all these results and the stratigraphic position of the underlying Tribotte Formation, the Rigenée Formation can be dated as entirely Llanvirn excepting its base, which corresponds to the upper two thirds of the Darriwilian (Figs 2, 3).

2.3.5. Stratotype

The most complete section lies in the Thyle valley in the Vallon des Goutailles, N of Rigenée: 50°34’15.64”N/4°31’16.63” E to 50°34’04.74” N/4°31’15.65” E (Fig. 13). Good sections of the Tribotte-Rigenée transition are also seen near Gentissart along the railway km 42.8-42.9 and along the abandoned vicinal (50°33’58.48” N/4°33’33.97” E). We propose these outcrops to be recognised as stratotype of the formation.

2.4. Ittre Formation

2.4.1. Definition

The Ittre Formation was named after the *« Assise d’Ittre »* (Beugnies, 1973 fig. 9) and subsequently formalized by Servais

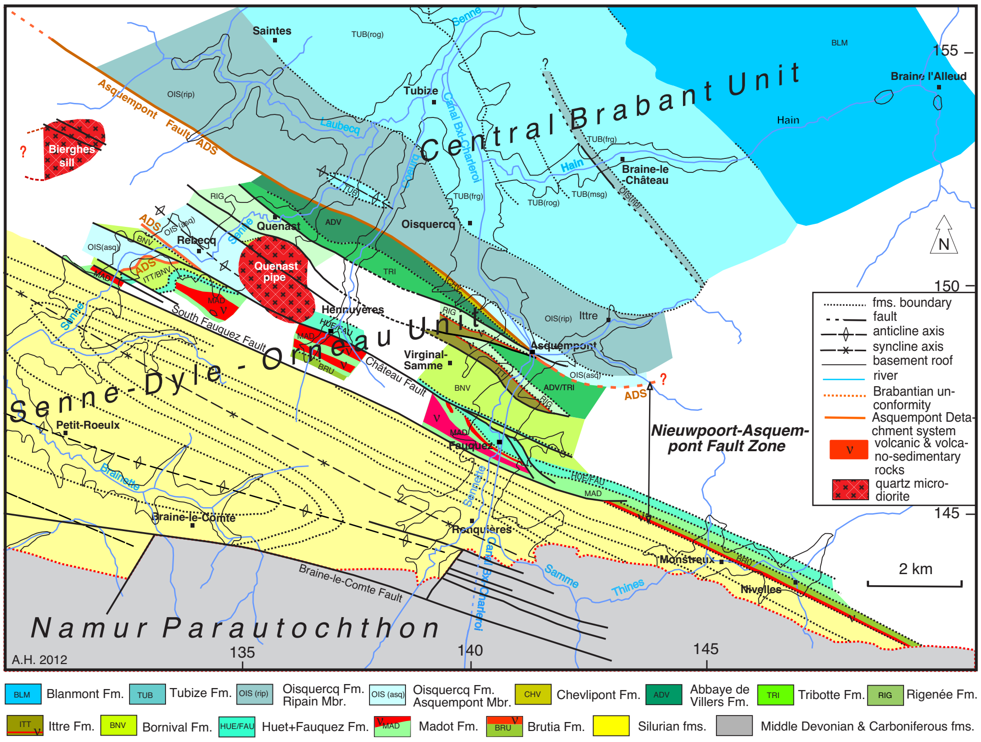

Figure 4: Sub-crop geological map of the central part of the Senne Basin. The Asquempont Detachment System (ADS), a low-angle extensional detachment system that formed prior to folding and cleavage development (Debacker et al., 2001, 2004b, 2005), separates the Central Brabant Unit formed by Cambrian formations from the Senne-Dyle-Orneau Unit essentially formed by Ordovician and Silurian formations. The Ordovician formations (in green colours) are cut by numerous faults that belong to the Nieuwpoort-Asquempont fault Zone (Legrand, 1968). This zone strongly contrasts with the northern Central Brabant Unit and with the southern Silurian formation gently folded and weakly affected by faults. The Silurian formations are globally coloured in yellow as they are not the aim of the paper. New extended map based on personal field observations and modified after the maps of Hennebert & Eggermont (2002), Debacker et al. (2003 fig. 24), Verniers et al. (2005 fig. 10), Debacker et al. (2011) and Herbosch et al. (2013). Location inside the Brabant Massif reported in Fig. 1.

(1991) and Servais et al. (1993). The town of Ittre is situated in a tributary of the Sennette valley (Fig. 4, 13). It is equivalent to the "unité F" of Martin & Rickards (1979).

2.4.2. Description

This formation comprise rhythmically alternatiing of beds, with sharp bedding planes, of light grey fine sandstone, medium grey siltstone, and dark grey mudstone. The sandstone beds are 5 centimetres thick up to pluridecimetres thick and show parallel, oblique and convolute laminations, fining upward graded bedding, with load casts and other current marks at their base. The siltstone shows parallel lamination and the mudstone no stratification. They are interpreted as quite distal turbidites of the Bouma-type (Servais, 1991). Vanmeirhaeghe et al. (2005) confimed the presence of Bouma sequences with Tc, Td and Te intervals but only in the lower part. In the rest of the formation the sequences look like the fine-grained turbidites *sensu* Stow (Fig. 5; Stow & Piper, 1984; Piper & Stow, 1991).

In the Lessines borehole the lithofacies are more diverse than seen in outcrops and three lithofacies have been described (Unité II, Herbosch et al., 1991; Herbosch et al., 2008). A first facies shows pluricentimetric rhythmic sequences that begin with irregular or wavy basal laminae of siltstone (T0) followed by an alternation of fine silt laminae and mudstone (T1, T2,… T5) and end with dark graded to ungraded mudstone (Plate 2, photo 7). The basal laminae (T0) are frequently affected by load structures and the top laminae (T8) by ball and pillow structures. The succession of structures and the centimetric scale of the sequences correspond quite well with the low-density turbidite model of Stow (Fig. 5; *op. cit.*). The second lithofacies corresponds to the decimetric turbidites of Bouma-type observed in outcrops. The third facies shows fine sandstone and siltstone with parallel, oblique and convolute laminations that correspond alternately to low- and high-density turbidites (Plate 2, photo 8). All three lithofacies are interpreted as turbidites of different density that have been deposited in a deep-marine environment (slope to basin plain). These former observations were confirmed by the field observation of Storme (2004 ms *in* Vanmeirhaeghe, 2006a).

Fine-grained volcanic levels were observed in the lowermost part of the formation in the Sennette valley by Corin (1963) and Martin & Rickards (1979). They were also observed recently in other places (Debacker, 2001; Debacker et al., 2003 figs 7, 9, 15, 23, 24) (Figs 4 in red, 13). On the E side of the canal at Asquempont several interstratified layers are present, having a thickness of about 1 metre in the lower part and thinning in the succeeding layers upwards. They are also present on the W side of the canal. The volcanic levels can also be observed at km 8.58 of the northern Virginal old railways section (Debacker et al., 2003 fig. 9), several metres thick showing mudstone fragments in its lowest part. In the "Bief 29" section centimetric tuff layers were observed near the base of the Ittre Formation (Debacker *et al.,* 2003 fig. 7).

A megaslump within the Ittre and the Bornival formations has been described along the E side of the Brussels-Charleroi canal cut (Debacker et al., 2001, 2003 figs 15, 23). It occurred between the deposition of the lower and the middle members of the Bornival Formation (Debacker et al., 2001) in Cheneyan times (Fig. 2; Vanmeirhaeghe, 2006a). A comparison of the cleavage/fold relationship and the stratigraphic polarity shows that a

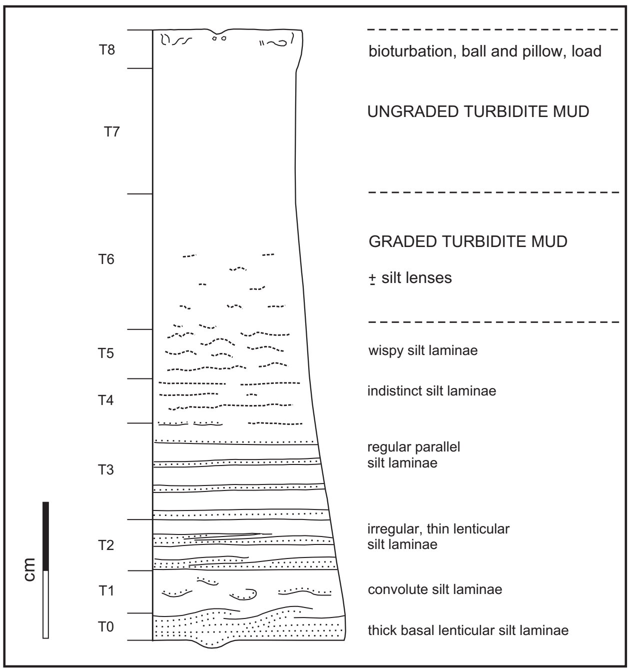

Figure 5: Standard sequence of structures or model of fine-grained turbidite (modified from Piper & Stow, 1991 fig. 1). Nine structural divisions are described from T0 to T8.

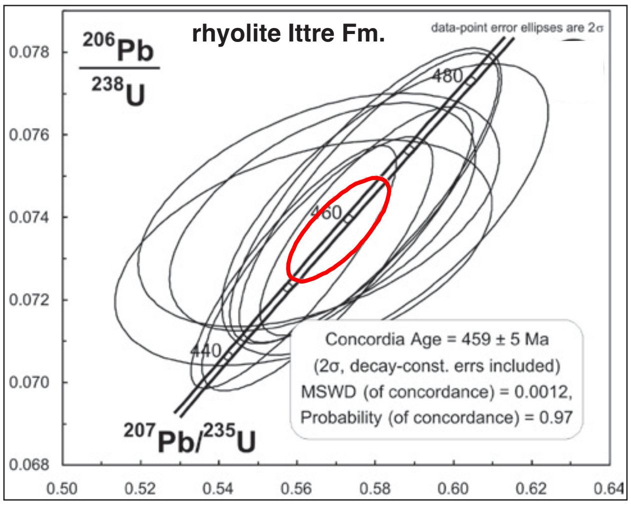

Figure 6: Zircon ages of the interstratified volcanic rocks level observed near the base of the Ittre Formation (from Linnemann et al., 2012). Virginal railway section km 8.59 (Debacker et al., 2003 fig. 9). Position: N 50°38'44.56"; E 4°13'41.50.

mass of turbidites at least 200 m thick in the core of a synform was overturned prior to the tectonic deformation. This overturn is attributed to a large-scale slumping occurring in the lower Katian (*circa* 452 Ma). The most likely trigger mechanism for this event is a major earthquake (Debacker et al., 2001 pp. 349-350), resulting from a tectonic instability linked to the northward drift of Avalonia with subduction of the Tornquist Ocean and before the soft docking with Baltica, that occured by the mid-Katian (Samuelsson et al., 2002a; Cocks & Torsvik, 2002; Cocks & Fortey, 2009).

The lower boundary of the formation is observed in the "Vallon des Goutailles" (Rigenée, Thyle valley, Fig. 13) and also in the "bief 29" section of the Sennette valley (Debacker et al., 2003 fig. 24). The upper boundary is always faulted (Debacker et al., 2001; Debacker et al., 2003 fig. 24).

2.4.3. Extent of the outcrop zones and thickness

The Ittre Formation crops out almost exclusively in the Senne Basin (Fig. 4). All sections are incomplete and/or bounded by faults. The lowermost part is visible in the Dyle basin and a temporary excavation exposed it in the Orneau valley. The formation is also present in the Lessines borehole where it is also bounded by faults (Herbosch et al., 1991, 2008). The interlayered volcanic levels observed near the base of the formation were only observed in the Senne Basin. The thickness of the formation is estimated at >180 m in the Sennette valley (Debacker, 2001; Debacker et al., 2001) and >82 m in the Lessines borehole in the Dender valley (Herbosch et al., 1991).

2.4.4. Bio-and chronostratigraphy

Graptolites from the Brussels-Charleroi canal trench studied by Legrand (1967, exact location unknown) and by Martin & Rickards (1979 "unité F") belong to the *Nemagraptus gracilis* or the *Diplograptus multidens* biozones that indicate an early or middle Caradoc age. The fauna collected in the field or stored in museums were restudied by Maletz & Servais (1998): they indicate the *N. gracilis* or the *D. multidens* biozones with a preference for the latter (presence of *Climacograptus bicornis)*. These two biozones cover the entire Sandbian (Zalasiewicz et al., 2009; Cooper & Sadler, 2012).

The chitinozoans, moderately well preserved and diverse, with the presence of *Belonechitina* cf. *robusta* indicate a Burrellian to early Streffordian age (Samuelsson and Verniers, 2000). A later study of the chitinozoans by Vanmeirhaeghe et al. (2005) and Vanmeirhaeghe (2006a) showed an assemblage with *D. junglandiformis*, *D. cocca*, *B. robusta*, *B. micracantha*. The presence of *D. junglandiformis* coincides with the major part of the index fossil *Spinachitina cervicornis* on Baltoscandia (Novak & Grahn, 1993). This allows us to conclude that the Ittre Formation corresponds to an interval situated within the *Spinachitina cervicornis* Biozone. This biozone corresponds on Avalonia (Vandenbroucke, 2008) to a Burrelian age, but a Cheneyan age cannot be excluded.

In summary, all the constraints of the three biozonations indicate a mid Sandbian age, possibly extending to the lowest Katian (Figs 2, 3).

Zircons present in the tuffs from the lower part of the formation in the Sennette valley (km 8.585 of the Virginal railway section, Debacker et al., 2003 fig. 9) have recently been dated by LA-ICP-MS (Linnemann et al., 2012). Eleven spot measurements gave a concordant age of 459 ± 5 Ma (Figs 2, 6), in good agreement with the stratigraphic age of the base of the formation (Fig. 2; *circa* 457 Ma).

2.4.5. Stratotype

The best outcrops are in the Sennette valley, along the E side of the canal cutting south of Asquempont from km 39,7 to 39,400 (50°38'09.29" N/4°14'03.21" E) (Fig. 13; Debacker et al., 2003 figs 15, 23, 24). In the absence of other sections we propose this outcrop as the stratotype for the formation.

2.5. Bornival Formation

2.5.1. Definition

The Bornival Formation was first informally mentioned as «…*laminated siltstone and mudstone of Bornival Formation*» by Van Grootel et al. (1997) in the «Rue de Bornival» at Fauquez (*sic*), Sennette valley (Fig. 13). It was formalized and briefly described *in* Verniers et al. (2001). More recently, the Member 1 was described in detail by Vanmeirhaeghe et al. (2005) and the members 2 and 3 by Verniers et al. (2005), both in the Sennette valley.

2.5.2. Description

The formation is formed by centimetric alternation of dark grey mudstones and dark grey to blackish claystones with gradual transitions. Abundant coarse silt grains are distributed throughout the clayey part. The bedding planes are rarely sharp and mostly gradual. Occasional very fine sandstone beds, laminated or finely obliquely stratified and never >10 cm thick were observed. In the Sennette valley, the formation can be subdivided into (at least) three members.

The unnamed Member 1 (also called informally the "laminite member") shows a centimetric alternation of light to medium grey fine-grained sandstone, siltstone to mudstone with dark grey to black mudstone (Plate 2, photo 9). The member contains centimetric to millimetric light grey mudstone «lenses», irregularly distributed in the dark grey mudstone. Brachiopods and ichnofossils only are present in this member. These sediments are interpreted as distal turbidites with individual sequences 1 to 4 cm thick, and are frequently bioturbated (Debacker et al., 2001; Van Noten *in* Vanmeirhaeghe et al., 2005).

The unnamed Member 2 is characterized by a centimetric alternation of black to dark grey mudstone and claystone with visible quartz silt grains. Some of the bedding planes show small-scale ripples that are crenulated by a cleavage. Some isolated patches of millimetric pyrite crystals occur parallel to the bedding.

In the unnamed Member 3, the sediments are finer grained than the underlying Member 2: a very faint centimetric lamination, caused by a varying proportion of fine silt grains, is present in the dark grey to black slate. In the Senne valley, the upper Member 3 seems to crop out to the west of Rebecq and presents a black lustrous facies unseen in the Sennette valley (Herbosch, 2005 fig. 3 outcrops a to e; Herbosch et al., Ittre-Rebecq map, 2013).

2.5.3. Extent of the outcrop zones and thickness

The lower and upper boundaries of the formation and the boundaries between the three members are always faulted (Debacker et al., 2003 fig. 24). This formation crops out only in the Sennette and Senne valleys (Figs 4, 13). The total thickness is estimated in the Sennette valley to be at least 265 m, with the lower member >116 m, the middle member >85 m and the upper member >64 m (Van Noten *in* Vanmeirhaeghe, 2006a).

2.5.4. Bio- and chronostratigraphy

Scarcely any macrofossils have been found in this formation except for rare specimens of brachiopods and ichnofossils in Member 1 (Verniers et al., 2005). Acritarchs from Member 1 were described by Martin & Rickards (1979) who inferred a ?late Arenig to Llanvirn age. Chitinozoans from all the members were found by Van Grootel et al. (1997) and Samuelsson & Verniers (2000) to contain *Lagenochitina dalbyensis*, *Belochitina hisuta* and *B.* cf. *robusta*, which permitted them to deduce a Burrelian age. A restudy of more samples by Vanmeirhaeghe et al. (2005 fig. 3) and Vanmeirhaeghe (2006a) could not confirm the identification of *L. dalbyensis* and *B. hisuta*. The lack of *D. juglandiformis* and *L. baltica* argues for a rather high position in the *S. cervicornis* Zone, but still below the *F. spinifera* Zone. As this age is deduced in the absence of the index species, it remains speculative. Hence, Vanmeirhaeghe (2006a) suggested a Cheneyan age that corresponds to a lowest Katian age (*circa* 452,5-451 Ma, Cooper & Sadler, 2012). The authors added also that the chitinozoan assemblages of the Ittre and Bornival formations do not differ much from each other even though the Ittre Formation can clearly be assigned to the *cervicornis* Biozone.

2.5.5. Stratotype

The type section area lies in the Sennette valley, for Member 1 along the southern half of the large canal section south of Asquempont km 39.318 to 39.210 (see Debacker et al., 2001). For Member 2 in the crags at the crossing of the Ry de Fauquez and the Rue de Bornival (see Verniers et al., 2005 fig. 3; 50°37'26.49" N/4°14'06.85" E) and for Member 3 along the abandoned railway section north of the abandoned Huet quarry (see Verniers et al., 2005 fig. 7; 50°37'42.74 N/4°13'42.96" E) (Fig. 13). In the absence of other sections we propose these outcrops to be recognised as the stratotype for the formation.

2.6. Cimetière de Grand-Manil Formation

2.6.1. Definition

This lithostratigraphic unit was first described by Michot (1980) as « …schistes compact verts affleurant au N du Moulin de Grand-Manil »(Fig. 13). He grouped this unit with the overlying shelly slate (now Huet Formation) in the Moulin Formation. More recently, Delcambre & Pingot (2002 fig. 15) grouped within the Bornival Formation two very poorly outcropping units: the « Unité de la Chapelle Sainte Adèle: schistes silteux verts, argileux, assez homogènes et micacés » and the « *Unité du Cimetière de Grand-Manil: siltites gréseuses compactes, riches en pyrite* ». Herbosch (2005) correlated the « green slate » from Grand-Manil with other « green slate » that crops out near Rebecq in the Senne valley under the new name Hospice de Rebecq Formation. The study of the magnetic properties of the Brabant slate (Debacker et al., 2010) strongly suggests that the latter correlation was erroneous and that the green rocks of Rebecq belong to the Asquempont Member of the Oisquercq Formation. After new investigations, Debacker et al. (2011) concluded that the "green slate" of Grand-Manil is indeed different from the green slate of Rebecq and proposed the name « *Cimetière de Grand-Manil Formation* » for this new formation.

2.6.2. Description, extent of outcrop zone and thickness

The formation shows faintly stratified greenish-grey to often olive-green slate (mudstone to siltstone). More silty slates with pluricentimetric sequences of planar, oblique and convolute laminations were frequently observed and are interpreted as very distal turbidites of Bouma type (Plate 2, photo 11). In outcrop, weathered and oxidized nodules parallel to the bedding were also frequently observed (Debacker et al., 2011 Plate 1). The formation poorly crops out in the Orneau valley (Fig. 7; Herbosch, 2005 fig. 4 outcrops 1 to 14) and possibly also in the Sennette valley near the old castle of Fauquez (Debacker et al., 2011). Its thickness is estimated at about 100 m. The upper and lower contacts have not yet been observed.

2.6.3. Bio- and chronostratigraphy

No macrofossils and no microfossil have yet been investigated. Based on its geometrical position between the Cheneyan (lower Katian) Bornival Formation and the lower Onnian (middle Katian) Huet Formation (see below), this formation should be placed in the lower half of the Streffordian, mid Katian Stage (Figs 2, 3). A biostratigraphical hiatus already recognized at this level by Verniers et al. (2001) was confirmed by Vanmeirhaeghe (2006a), who emphasized the strongly differing chitinozoan

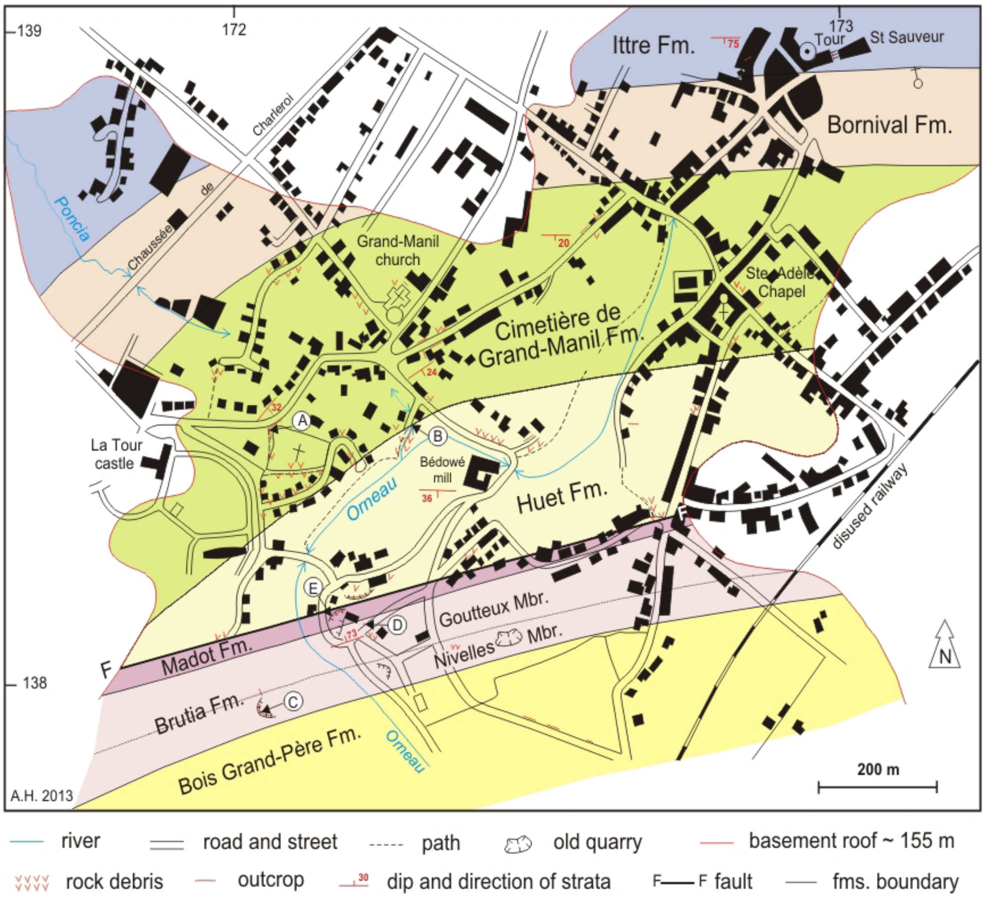

Figure 7: Geological map of the region of Grand-Manil (Gembloux), Orneau valley. Modified from Herbosch (2005). A: best outcrop of Cimetière de Grand-Manil Fm.; B: outcrop of the base of the Huet Fm.; C: outcrop of the Nivelles Mbr. ("Eurite de Grand-Manil") of Brutia Fm.; D: outcrop of the bioturbated mudstone of the Goutteux Mbr. of the Brutia Fm.; E: outcrop of the volcano-sedimentary rocks of the lower part of the Madot Fm. Position in the Brabant Massif reported in Fig. 1.

assemblages of the Huet Formation compared to the Ittre and Bornival formations.

2.6.4. Stratotype

No stratotype could be designated but discontinuous small outcrops occur around Grand-Manil, in the Orneau valley (Figs. 6, 13). The best are along a path to the NW corner of the cemetery (Fig. 7 outcrop A) (50°33'19.16" N 4°40'48.48" E); Herbosch, 2005 fig. 4 outcrop number 3, 5, 7).

2.7. Huet Formation

2.7.1. Definition

The formation is named after the abandoned quarry owned by Mr. Huet, north of the hamlet of Fauquez, on the western side of the Sennette valley (Fig. 13). The fossiliferous rocks that belong to this formation have been known since Dumont (1848) and were previously described as « grauwacke fossilifère de Fauquez » (Leriche, 1920), informally mentioned by Van Grootel et al. (1997), described in Verniers et al. (2001) and formally defined in Verniers et al. (2005). In the older literature, authors might have confused the macrofossil levels in the Madot Formation with those in the Huet Formation.

2.7.2. Description

The Huet Formation is formed of greenish to grey mudstone, siltstone and fine sandstone, poorly sorted, with characteristic orange-yellow alveoli of decalcified fossil fragments (Plate 2, photo 10). It has heterogeneous sub-metrical beds, with a faint fining upward granulometry, a slightly undulating base and sometimes faint oblique stratification. Some levels are rich in (often decalcified) macrofossils (bryozoans, brachiopods, corals, cephalopods, crinoids, cystoids and trilobites, Mailleux, 1926) that indicate a shallow shelf environment of formation (not of deposition). These fossils are fragmented and concentrated mostly at the base of some beds. For Vanmeirhaeghe et al. (2005) all the features indicated a tempestite deposit on the shelf above storm wave base. This postulated rather shallow environment between the Cimetière de Grand-Manil and Fauquez formations, both interpreted to be deep water and is difficult to explain. Either the interpretation as tempestite should be replaced by the very similar looking distal turbidites or it could be explained by an eustatic drop in sea level postulated at this period of time (Vanmeirhaeghe, 2006a).

It is the first occurrence of calcareous fossil debris in the Brabant Massif since the beginning of the sedimentation in the lower Cambrian (Fig.11 formations named in red). It is the record of the rapid northward drift of Avalonia to lower latitudes that permitted the development of a warm water ecosystem and/or the warming episode that preceded the Hirnantian glaciation ("Boda event", Fortey & Cocks, 2005; Lefebvre et al., 2010).

2.7.3. Extent of the outcrop zones and thickness

The formation crops out in the Sennette, Senne and Thisnes valleys (Figs 4, 13; Verniers et al., 2002b, 2005; Vanmeirhaeghe et al., 2005) and in the Orneau valley (Fig. 7). The base of the formation could only be observed in the Orneau valley (Fig. 7 outcrop B). The top is exposed in the sunken road at Fauquez (Sennette valley) and along the abandoned disused railway cut north of the Huet quarry. The transition upwards to the Fauquez Formation is quite rapid (Verniers et al., 2005 figs. 2, 4). The thickness is more than 60 m in the Sennette valley (Van Grootel et al., 1997; Verniers et al., 2005).

2.7.4. Bio- and chronostratigraphy

The brachiopods suggest a Caradoc or Ashgill age for the formation (Maillieux, 1926). The cystoids from the Sennette and the Orneau valleys studied by Regnell (1951) indicate a (upper) Caradoc age. The trilobites from the Orneau valley at the Lefèvre quarry indicated a (mid) Caradoc age (Maillieux, 1926). Richter & Richter (1951) collected trilobites in the Orneau valley which indicated an (early) Ashgill age. The location of this new collecting site is unclear and it may have been sampled from the Madot Formation.

The chitinozoan assemblages containing abundantly *Lagenochitina baltica, L. prussica, Belonechitina robusta* and *Tanuchitina bergstroemi* indicate a late mid-Caradoc age, hence an early Cheneyan age is concluded to be the oldest possible age in relation to the Baltoscandian stages (Van Grootel et al., 1997). In a new study of the chitinozoans, Vanmeirhaeghe et al. (2005) and Vanmeirhaeghe (2006a) placed the Huet Formation in the *F. spinifera* Biozone, inferred by the accompanying species, particularly *Lagenochitina baltica* that appears in the basal part of the formation. On British Avalonia, Vandenbroucke (2008 fig. 7) showed that the *F. spinifera* Zone occurs from the upper Streffordian (Onnian, uppermost Caradoc) to the lowermost Cautleyan (lower Ashgill). Hence, the Huet Formation could encompass the interval from the Onnian to the lowermost Cautleyan but, after the age assignment of the Cimetière de Grand-Manil and Madot formations (cf. § 2.6.4 and 2.8.4), its age should be restricted to the lower half of the Onnian that corresponds to a very small time interval (< 1 Ma) in the lower Katian (Fig. 2; *circa* 450 Ma Cooper & Sadler, 2012).

2.7.5. Stratotype

The best sections are in the Fauquez area (Sennette, Fig. 13), along the abandoned railway section and in the abandoned Huet quarry (50°37'42.63" N/4°13'42.58" E), 230 to 300 m north of the railway bridge in Fauquez (Verniers et al., 2005, figs 2, 7). We proposed these outcrops to be recognised as the stratotype of the Huet Formation.

2.8. Fauquez Formation

2.8.1. Definition

This formation was named after the locality of Fauquez, Sennette valley (Fig. 13). Informally known in the literature since Malaise (1873), the Fauquez Formation has been described as « *schistes noirs de Fauquez* » or «*grauwacke fossilifère de Fauquez* » by Maillieux (1926). It was used as *« Formation de Fauquez* » by Herbosch et al. (1991 fig. 6) and more formally defined by Van Grootel et al. (1997).

2.8.2. Description

The Fauquez Formation is graptolitic black slate; specifically it is a centimetric and rhythmic alternation of pyritic siltstone laminae and dark grey to black slates (claystones) laminae (Plate 2, photos 12; Plate 3, 13). The base of the sequences is often marked by millimetric pyrite crystals that are completely oxidised in outcrops (Plate 3, photo 13). From about 4 m above the base abundant graptolites occur in the darker claystone laminae (Plate 3, photo 13). The rhythmicity is enhanced by the numerous pyrite levels and the sequence of sedimentary structures (particularly visible in borehole, Plate 2, photo 12; see also Herbosch et al., 1991 plate 1) points to an interpretation as low-density turbidites. Compared with the Stow model (Fig. 5; Stow & Piper, 1984; Piper & Stow, 1991) these black slates are interpreted as base-cut Stow sequence or mud turbidites. The dark colour, the abundance of graphite and pyrite and the preservation of the fine sedimentary structures indicate anoxic conditions. A deep marine anoxic environment of deposition, probably the slope, has been postulated.

2.8.3. Extent of the outcrop zones and thickness

The lower boundary is a rapid transition, with a distinct change in grain size, colour and bed thickness, changing from decimetric in the Huet Formation to centimetric in the Fauquez Formation. The upper boundary with the Madot Formation is nowhere visible or faulted. The formation crops out in the Senne and Sennette valleys (Figs 4, 13; Verniers et al., 2005; Herbosch et al., Ittre-Rebecq map, 2013) and in quarries in the Dender valley (Legrand & Mortelmans, 1948; Lecompte, 1950; Legrand, 1965; "Unité I" in the Lessines borehole, Herbosch et al., 1991, 2008). It is not present in the Orneau valley as the result of faulting (Fig. 7; Herbosch, 2005 fig. 4). It is >35 m thick in the Sennette valley (Verniers et al., 2005) and >58 m thick in the Lessines borehole (Dender valley, Herbosch et al., 1991, 2008).

2.8.4. Bio- and chronostratigraphy

Malaise described very early on, in 1873, graptolites in many places: initially in the famous "chemin creux" of Fauquez Sennette valley, then on the bottom of the "porphyres" quarry in the Dender valley, near Lessines, and in the canal Brussels-

Charleroi canal cutting near Fauquez, and more recently in the Lessines borehole (see Maletz & Servais, 1998 for a complete history). Mailleux (1926) first placed this assemblage in the *Pleurograptus linearis* Biozone. After Miss Elles studyied the specimen, she later noted the difficulties of attributing the fauna to either the *D. clingani* or the *P. linearis* Biozone (Mailleux, 1930b). Bulman (*in* Lecompte, 1950) concluded that the fauna of the "carrière de la Dendre" should most probably be assigned to the *D. clingani* Biozone rather than the *P. linearis* Biozone. Legrand (1967) found graptolites during the digging of the canal Brussels-Charleroi canal, and attributed it to the *P. linearis* Biozone. In the Lessines borehole, Degardin (*in* Herbosch et al., 1991) found two levels with abundant graptolites that were attributed, at the 159,66m level, to the *D. clingani* Biozone, and at the 204,4 m level, probably to the *D. multidens* Biozone. Maletz & Servais (1998) restudied all the old material found in the collections and new faunas collected in the Fauquez and Lessines boreholes. They could not observe the nominal species of the biozones, but concluded that the assemblage might belong to the *P. linearis* Zone with the possibility that the fauna may belong to the upper part of the *D. clingani* Zone. That corresponds to the Streffordian - mid Pusgillian interval of the British chronostratigraphy and to the middle Katian of the global chronostratigraphy (about 451-448 Ma, Cooper & Sadler, 2012 fig. 20.1).

Earlier chitinozoan studies (Van Grootel, 1995; Van Grootel et al., 1997; Samuelsson & Verniers, 2000) mentioned the presence of *Lagenochitina baltica, L. prussica, Belonechitina robusta* and *Tanuchitina bergstroemi* indicating a late Vormsi to early Pirgu in terms of Baltoscandian stages. Restudy of samples from Fauquez and Lessines boreholes (Vanmeirhaeghe, 2006a) showed that the formation belongs to the *A. reticulifera* subzone of the *F. spinifera* Biozone of probable late Onnian to lowermost Pusgillian age in British zonation (Vandenbroecke, 2008 fig. 7). This confirms the global dating by means of graptolites. The Fauquez Formation should be placed in a very short time interval in the late Onnian to lowermost Pusgillian from the middle Katian Stage (Figs 2, 3; *circa* 448 Ma, Cooper & Sadler, 2012).

2.8.5. Stratotype

The stratotype lies in the sunken road « rue de Fauquez » at Fauquez , on the east side of the Sennette valley (Figs 4, 13) (from 50°37'28.55" N/4°13'57.73" E to 50°37'28.96" N/4°13'57.84" E). A detailed lithological log of this outcrop is reported *in* Verniers et al. (2005 fig. 4).

2.9. Madot Formation

2.9.1. Definition

Named after the Madot hill on the E side of the canal, SE of the Fauquez canal bridge (Figs 4, 13). It was informally mentioned by Van Grootel et al. (1997), formally defined by Verniers et al. (2001) and described in detail by Verniers et al. (2005). Volcanic rocks of the "Bois des Rocs" were first described by Omalius d'Halloy (1808, 1828) and later by Dumont (1848) and Malaise (1873). A detailed historical overview can be found in Verniers et al. (2005 pp. 160-164).

2.9.2. Description

The Madot formation contains many volcanic rocks (volcano-sedimentary rocks, volcanic breccias containing shale fragments), interstratified with greenish-grey heterogeneous coarse siltstone and mudstone rich in macrofossils such as bryozoans, brachiopods, crinoids, trilobites, rugose corals and pelmatozoans. Black, dark grey to bluish graptolitic slate and fine siltstone that often contain dispersed single crystals of volcanic origin are also interstratified. It is subdivided in the Sennette valley into 7 members (Verniers et al., 2005 fig. 10). Member 1: dark grey to bluish shale and fine siltstone (slate), with no stratification observed and characteristic presence of large, coarse sandy grains, single or in clusters up to 25 cm long, interpreted as volcanoclastic material incorporated in the mud matrix. Member 2: at least 7 fining or coarsening upwards sequences of poorly sorted grey greenish siltstone, sandstone and conglomeratic levels of volcano-clastic origin with a thin marine shale interval. The conglomerate contains centimetric to sub-centimetric shale clasts. Member 3: homogeneous black to dark grey slate: the contact with the underlying member 2 is clear and with the overlying member abrupt. Member 4: greenish-grey heterogeneous coarse siltstone with a 10 cm thick coarse basal breccia-like unit with clasts of slate or volcanoclastic rock. No stratification was observed. Numerous macrofossils occur, such as bryozoans, brachiopods, crinoids or trilobites. There is a gradual transition over 10 m to the next member. Member 5: dark grey to black slate dotted with clusters of light grey grains. In the upper part of the member a coarse, sandy bed is present. Within this unit a large boulder (>1 m) of volcanic origin can be observed. There is an abrupt transition to the coarse volcano-sedimentary rock of the overlying member. Member 6: dacite, dacitic volcanic breccia, sometimes containing slate fragments, volcano-sedimentary rock and black slate. The volcanic complex of the "Bois des Rocs" (Plate 3, photo 14), part of Member 6, is described in more detail by André *in* André *et al.* (1991). Member 7: dark grey to black silty slate dotted with brownish to orange alveoli, the remnants of dissolved calcareous fossil fragments.

In the Senne and Coeurq valleys (Figs 4, 13), numerous small outcrops of volcanic and volcanosedimentary rocks occur (Herbosch, 2005 fig. 3 outcrops m to t; Ittre-Rebecq map; Herbosch et al., 2013). The only good section described in detail by Mortelmans (1952) at Hennuyères (Coeurq valley) shows about 180 metres of tuffs (Plate 3, photo 15) interstratified with slate and shelly slate. The presence of cross-stratification in the tuffs indicates a shallow shelf environment of deposition. This is also demonstrated by the observation of colonial corals that have grown *in situ* on dacitic submarine flows (Van den Haute, 1974; L. André, pers. comm.). The upper parts of the formation crop out also in the Orneau valley (Fig. 7 outcrop E), but the "tuffoides kératophyriques" described by Mathieu (1905) are no longer visible (see Delcambre & Pingot, 2002 pp. 28-30).

The Lichtervelde 53W57 borehole contains Upper Ordovician black graptolitic slates, dark grey argillaceous sandstones and volcano-sedimentary beds (Michot & de Magnée, 1936; Legrand, 1964; Corin, 1965). The rocks from the interval 291,00 m to 415,10 m (borehole bottom) were assigned by chitinozoans to the Madot Formation (Van Grootel et al., 1997). Graptolites belonged according to Legrand (op. cit.) to the *D. anceps* Zone from the base up to 308 m depth. A restudy of the graptolites by Maletz & Servais (1998) allowed recognition of the *D. complanatus* Zone at 336,70 m and 318,75 m (Pusgillian-Cautleian; Fig. 3) and the *D. anceps* Zone at 311,00 m (Cautleian-Rawtheian, Fig. 3). Restudy of the chitinozoans by Vanmeirhaeghe (2006a) with the identification of the *T. bergstroemi* Zone corroborates the chronostratigraphical position deduced from graptolites. The Ordovician of the Lichtervelde well is unconformably overlain at 291,00 m by Silurian black to dark grey slates and sandstones, with no fault reported between the two units (Michot & de Magnée, 1936). This slate belongs most probably to the late Rhuddanian to Aeronian (Van Grootel, 1990) implying the absence of the Brutia Formation, due to a hiatus or faulting.

As in the Huet Formation, the presence of abundant warm water macrofossils records also the rapid drift of Avalonia to lower latitudes. But Fortey & Cocks (2005) have also shown that a late Ordovician global warming episode, the Boda Event, has affected the high-latitude Gondwana in the Cautleyan-Rawtheyan stages of the upper Katian. The movement of benthic faunas such as trilobites and brachiopods to progressively higher latitude and the abrupt appearance of limestone formations during the Cautleian-Rawtheian stages are observed not only in the Gondwana palaeocontinent but also in Baltica (Boda limestones of Sweden) and Avalonia (N Wales). The Madot Formation corresponds exactly to this stratigraphic interval (see Figs 2, 3, 11 and § 2.9.4).

The transition from the Fauquez to the Madot formation marks an abrupt and important change of bathymetry. In fact, the Fauquez Formation is formed by mud turbidites interpreted as a deep marine environment (most probably slope; § 2.8.2) and the Madot Formation shows numerous levels of shelly facies and volcano-sedimentary facies formed in a shallow shelf environment. This event cannot be explained only by a sea-level change only (Vanmeirhaeghe, 2006a) and it could easily be linked to the soft docking of the Avalonia microplate with the Baltica continent that occurs at about the same time (Cocks &

Torsvik, 2002; 2005; Cocks & Fortey, 2009; Torsvik & Cocks, 2011; Linnemann et al., 2012).

2.9.3. Extent of outcrop zones and thickness

The lower and upper boundaries of the Madot Formation are nowhere visible in outcrops. The formation is observed in the Senne, Sennette, Coeurq, Thisnes (Nivelles) and Orneau (only in temporary outcrops) valleys (Figs 4, 13). The Lichtervelde 53W57 borehole (291 to 415,10 m) shows a similar range (Legrand, 1964, 1966 *in* Martin, 1969; Vanmeirhaeghe, 2006a p. 264). The thickness of the Madot Formation is about 215 m on the east side of the Sennette valley and about 290 m on the west side of the valley (Verniers et al., 2005) due to a marked thickening of the volcanic rocks of Member 6 (only 10 m in the E and estimated at >130 m in the "Bois des Rocs" to the W). The thickness is estimated at more than 180 m at Hennuyères in the Senne valley (Mortelmans, 1952).

2.9.4. Bio- and chronostratigraphy

Macrofauna and flora include crinoids, bryozoans, brachiopods, trilobites, corals and pelmatozoans. The trilobites determined by Richter & Richter (1951), possibly deriving from this formation in the Orneau valley, indicate an early Ashgill age (middle Katian). Acritarch studies by Martin & Rickards (1979) indicate a broad Caradoc to Llandovery interval. The chitinozoans from the Madot Formation where first studied by Samuelsson & Verniers (2000) with the presence of *Tanuchitina bergstroemi*, *Lagenochitina baltica*, *L. prussica* and *Belonechitina robusta* indicated to them a late Vörmsi to earliest Pirgu age of the Baltoscandian stages (late Streffordian to early Pusgillian, Ka2 to Ka3). A more detailed study of the chitinozoans from the Sennette valley (Vanmeirhaeghe et al., 2005; Vandenbroucke et al., 2005; Vanmeirhaeghe, 2006b) permits the formation to be placed in the *spinifera* and possibly also in the *bergstroemi* biozones. They proposed an Onnian (latest Caradoc) age for the lower part of the Madot Formation (Member 1 to Member 4 *pro parte*) and a Pusgillian to Cautleyan age (early to mid Ashgill) for the upper part of the formation (Member 4 *pro parte* to members 5 to 7).

Very recently, investigations in the uppermost part of the Madot Formation outcrops at Hennuyères in the Coeurq valley (Herbosch *et al.,* Ittre-Rebecq map, 2013; Mortier *et al.,* 2012) show the presence of *Lagenochitina baltica*, *L. prussica* and *Conochitina rugata* species. The presence of *C. rugata*, the index fossil of the *rugata* Biozone, indicates a possible late Cautleyan to early Rawtheyan age (Vandenbroucke, 2008), which is a little younger than the upper member of the type section in the Sennette valley (Member 7). In conclusion, all the results are quite consistent and the Madot Formation covers the entire upper half of the Katian. But as the Huet and Fauquez formations belong to the Onnian and the lowermost Pusgillian, the Madot Formation was more probably deposited in lower Pusgillian to mid Rawtheian time (Figs 2, 3).

The tuffs of the same outcrop at Hennuyères (Plate 3, photo 15) have recently been dated by LA-ICP-MS on zircons (Linnemann et al., 2012): seven spots give a concordant age of 445 ± 2 Ma (Fig. 8) which is in excellent agreement with the late Cautleyan to early Rawtheyan chitinozoan biostratigraphic age (447.5-446.5 Ma, Cooper & Sadler, 2012 fig. 20.9). The dacite of the "Bois des Rocs" (Member 6) has also been dated using zircons (op. cit.): 6 spots give a concordant age of 444 ± 6 Ma (Fig. 9) that is also in excellent agreement with a Pusgillian to Cautleyan biostratigraphic age.

For Legrand (1965) the age of the Lessines sill should be youger than the *P. linearis* Zone found under the sill (Legrand & Mortelmans, 1948; Lecompte, 1950) and older than the *P. acuminatus* Zone found at the top of the sill. This time interval corresponds to the Madot and lower part of the Brutia formations (Fig. 2). The very similar Bierghes sill could be the same age.

2.9.5. Stratotype

There is no stratotype but the best outcrops are in the Sennette valley, south of the Fauquez bridge, close to the hill of Madot, in sections along the east side of the canal between km 38.06 and 37.81 (from 50°37'32.47" N/4°13'00.60" E to 50°37'18.27 N/4°13'43.50" E), and on the west flank of the Sennette valley in the hamlet of Fauquez and in large outcrops in a side valley at the "Bois des Rocs" (from about 50°37'32.47" N/4°13'00.60" E to about 50°37'31.55" N/4°13'15.02" E) (Figs 4, 13).

2.10. Brutia Formation

2.10.1. Definition

Named after a locality south of Gembloux, in the Orneau valley (Delcambre & Pingot, Chastres-Gembloux map 2002). The Brutia Formation is subdivided in two members: a lower one herewith named the Goutteux Member and the upper volcanic Nivelles Member (Delcambre & Pingot, 2002 p. 31). The hamlet of Goutteux (1 km to the SSW of Hennuyères railway station) was described by Leriche (1924 p. 24) who mentioned the "tranchée du Goutteux" and a small quarry in the hamlet (GSB 115W68). The volcanic horizon has already been described by Dumont (1848) and is often cited in the old literature as "eurite" (see Corin, 1965). The "Eurite de Nivelles" observed in the Thisnes valley and the "Eurite de Grand-Manil" observed in the Orneau valley (Fig. 13), are at the same stratigraphic level (Corin, 1965; Verniers et al., 2001).

2.10.2. Description

The lower Goutteux Member (new name) contains medium to dark grey slate (mudstone), compactly bedded that it becomes darker in its upper part. The first third of the member is a grey mudstone that shows dark grey tubular lenses (about 5 mm wide and 2-3 cm

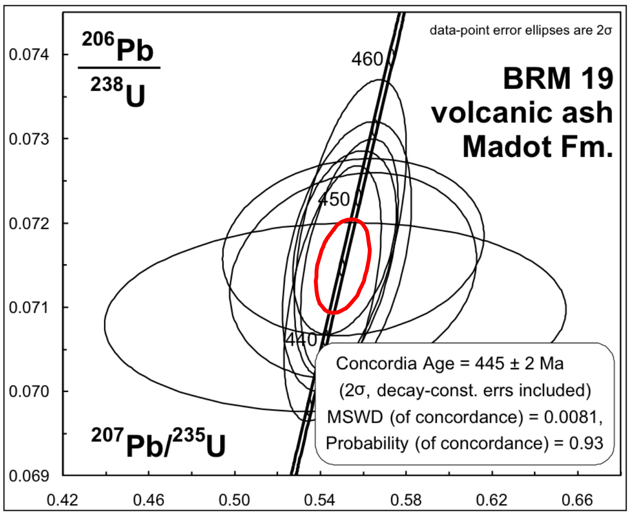

Figure 8: Zircon ages of the rhyolitic tuffs (Plate 3, photo 15) observed near the top of the Madot Formation at Hennuyères, Coeurq valley (from Linnemann et al., 2012). Position: N 50°38'40.43"; E 4°10'13.22.

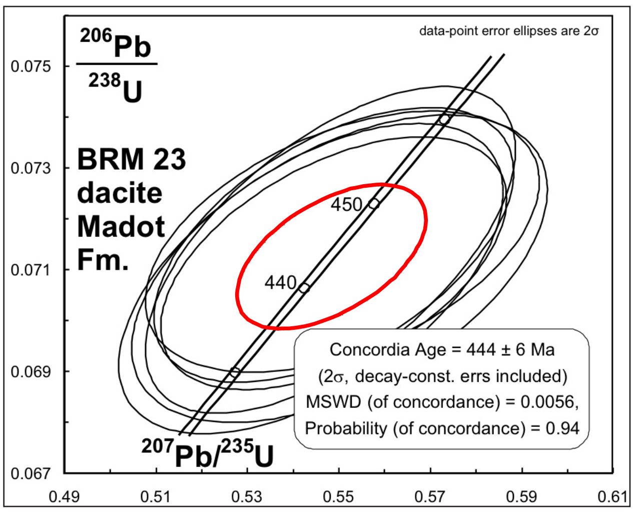

Figure 9: Zircon ages of the dacite of the Bois des Rocs (Plate 3, photo 14), W of Fauquez, small brook tributary of the Sennette (from Linnemann et al., 2012). Position: N 50°37'32.11"; E 4°13'09.71".

long; Plate 3, photos 16, 17). These very characteristic structures are interpreted as bioturbations (*Chondrites* sp. or "*fucoïdes*" of authors) and are observed in the Orneau valley (Fig. 7 outcrop D) (Delcambre & Pingot, Chastres-Gembloux map, 2002; Herbosch, 2005 fig. 4 outcrop y and z) and the Coeurq valley (Hennuyères; Herbosch et al., Ittre-Rebecq map, 2013). The upper part of the formation, named the Nivelles Member, consists of a 40 to 50 m thick volcanic (metarhyolite or ignimbrite?) interstratified layer. It is formed by fine-grained quartzitic tuffs, very hard, white, light pink or yellow quartzitic tuffs (Corin, 1965; Linnemann et al., 2012 analysis Table 3 sample Brab-16). The volcanic rocks contains some black slate centimetric inclusions coming from the underlying slate of the lower member. It has been mined near Nivelles for kaolinite by the ceramic industry (Ladeuze, 1990).

2.10.3. Extent of the outcrop zones and thickness

Outcrops of the Brutia Formation are mostly small and poor: they are located in the Orneau (Grand-Manil), La Ligne (Sombreffe), Thines (Monstreux-Nivelles) and Coeurq (Hennuyères) valleys (Fig. 13; Verniers et al., 2002a, b; Herbosch, 2005; Hennebert & Eggermont, Braine-le-Comte - Feluy map, 2002; Delcambre & Pingot, Fleurus-Spy map, 2008; Herbosch et al., Ittre-Rebecq map, 2013; Mortier et al., 2012). It is absent in the Sennette valley, removed by the South Fauquez Fault (Fig. 4). The lower and upper boundaries are unknown. Its thickness in the Orneau valley is between 80 and 100 m: about 40 to 60 m is the lower member and 40 m is the Nivelles Member ("Eurite de Grand-Manil"). The "Eurite de Nivelles" in the Thisnes valley is about 50 m thick (Herbosch & Lemonne, Nivelles-Genappe map, 2000).

2.10.4. Bio- and chronostratigraphy

Graptolites described in black slate below the "Eurite de Grand-Manil" in the Orneau valley belong to the *Coronograptus cyphus* Biozone (Elles *in* Maillieux, 1930a). But graptolites found in the same level by Gerlache (1956) and determined by Bulman as *Climatograptus scalaris* indicate the *acuminatus* Biozone (second graptolite zone of the Rhuddanian; Melchin et al., 2012). The latter determination seems more probable as the graptolites described from within the "Eurite de Nivelles" indicate undoubtedly a *Cystograptus vesiculosus* Zone (Rickards *in* Verniers & Van Grootel, 1991) which is the third graptolite zone of the Rhuddanian (Melchin et al., 2012). The determinations from both valleys indicate an upper Rhuddanian (Early Llandovery) age for the Nivelles Member.

The chitinozoans from the mottled mudstone of the lower member are dominated by *Belonechitina* cf *gamachiana*, indicating a mid or late Ashgill, or possibly Hirnantian age (Samuelsson & Verniers, 1999, 2000). In the Harelbeke borehole (GSB 83E446) a Hirnantian chitinozoan fauna was discovered by Van Grootel (1995). Vanmeirhaeghe (2006a) confirmed that the chitinozoans are well correlated with the *Spinachitina taugourdeaui* Zone, which is situated in the lower *N. extraordinarius* graptolite Zone of the Hirnantian Stage (Figs 2, 3). Recently, the study of a newly discovered outcrop of the upper part of the Goutteux Member SSW of Hennuyères (Herbosch *et al.,* Ittre-Rebecq map, 2013; Mortier *et al.,* 2012) showed it to contain the index fossil *Spinachitina oulebsiri* together with *Herchochitina* spp., *Belochitina* spp. *and Spinachitina verniersi*. This implies that the Goutteux Member is not younger than Ordovician in age (Paris et al., 1999) and the presence of *S. oulebsiri* indicates an upper Hirnantian age (Vandenbroucke *et al.,* 2009). In conclusion the Brutia Formation crosses the Ordovician-Silurian boundary and belongs to the interval between the uppermost Katian (?) and the Hinantian to upper Rhuddanian (Figs 2, 3).

2.10.5. Stratotype

The stratotype has not yet been designated. For the lower Les Goutteux Member the more representative outcrop is SW of Hennuyères in a small quarry on the E-side of the Brussels-Mons railway at 24,600 km (50°38'39.61" N/4°10'13.52" E). For the "Eurite de Grand-Manil" the type area is in an abandoned quarry at Grand-Manil, in the Orneau valley (Fig. 7 outcrop C; Delcambre & Pingot, 2002 fig. 17) (50°33'39.61" N/4°40'49.92" E). For the "Eurite de Nivelles" the type area is in the Thines valley, between Monstreux and Nivelles (Figs 4, 13; Verniers et al., 2002; Herbosch & Lemonne, Nivelles-Genappe map, 2000; Hennebert & Eggermont, Braine-le-Comte – Feluy map, 2002), but a representative outcrop is an old quarry along the E-side of the Brussels-Mons highway (50°36'09.57" N/4°17'48.30" E).

3. Subsidence history

New estimates of the sedimentary thicknesses, especially for the Cambrian and Ordovician formations (Herbosch et al., Ittre-Rebecq map 2013; Jodoigne-Jauche map, accepted; NCS 2012) allow us to construct the new cumulative thickness curve (figure 7). This curve is not a subsidence curve, because compaction and water column are not corrected for. Nevertheless, it reveals major changes in the rate of sedimentation. A similar curve has already been constructed by Van Grootel et al. (1997) and by Debacker (2001) based on of the thicknesses of Verniers et al. (2001) and repeatedly used (Verniers et al., 2002a; Debacker et al., 2005; Sintubin et al., 2009) as evidence for an extensional basin for Megasequence 1 and the onset of a foreland basin for Megasequence 3 but never used to explain the Ordovician subsidence in Megasequence 2. If we examine the curve for the Rebecq Group (Fig. 10) we see that, after the hiatus, Megasequence 2 (in red) begins with a low sedimentation rate from the Abbaye de Villers to the Rigenée formations interpreted as shelf deposits. This subsidence rate progressively increased from the Ittre to the Fauquez formations, recording the initiation of deeper marine turbiditic and pelagic sedimentation. This change from shelf to deep-sea environment marks the beginning of a long period of tectonic instability (megaslumping) and magmatism. From the Madot to the Brutia formations, the rate of subsidence is slowing down corresponding to a drastic change of bathymetry from a slope environment (Fauquez Formation) to a shallow shelf with a peak of volcanic and magmatic activity (Madot and Brutia formations). This later event marks the end of Megasequence 2 (Linnemann et al., 2012) and could be explained by the soft docking of Avalonia with Baltica that occured since the mid-Katian (Winchester et al., 2002; Cocks & Torsvik, 2002; Cocks & Fortey, 2009) which is also the age of the base of the Madot Formation (Figs 2, 3).

4. Comparison with the Condroz Inlier

The Condroz Inlier ("Ride du Condroz", "Bande de Sambre-et-Meuse", "Bande condrusienne" of authors) is a narrow strip about 65 km long and 0.5 to 4 km wide composed of Ordovician and Silurian rocks, emplaced as a series of tectonic wedges along the Midi Overthrust Fault. The inlier is flanked unconformably to the north by the Middle Devonian rocks of the Brabant Para-autochthon, and to the south by the Lower Devonian rocks of the Ardennes Allochthon (Michot, 1980, Vanmeirhaeghe, 2006a; Debacker & Vanmeirhaeghe, 2007; Belanger et al., 2012). Three main parts, with each a different stratigraphic registration and tectonic history, could be recognized (Michot, 1980; Hance et al., 1992; Verniers et al., 2002): (1) a central and main part situated north of the Midi Fault that belongs to the structural unit "Haine-Sambre-Meuse Overturned Thrust sheets" (Belanger et al., 2012); (2) a northern part situated north of the Midi Fault

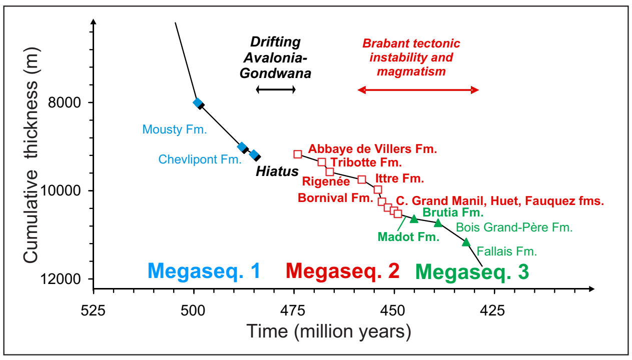

Figure 10: Cumulative sediment (by formation) thickness curve for the Rebecq Group plotted against the stratigraphic age (time scale Gradstein et al., 2012). Reported thicknesses are estimated mean thicknesses (NCS, 2012).

at Ombret (Michot, 1978; Steemans, 1994; Hance et al., 1992); (3) a southern part, found mostly at the Pointe de Puagne (SW) and at the Fond d'Oxhe (NE), situated south of the Midi Fault and belonging to the Ardennes Allochthon. The tectonic origin of these thrust sheets is controversial, but during the basin formation they belonged to the Brabant Basin, as termed by Verniers & Van Grootel (1991) or later called the Condroz-Brabant Basin (Vanmeirhaeghe, 2006b). Here we prefer to name it the "Brabant-Condroz Basin".

Given its position within the Variscan front zone, the scarcity of macrofossils and its poor degree of exposure, the stratigraphy of the Condroz Inlier is more difficult to establish. Detailed mapping (Delcambre & Pingot, 2000a, b, in press) and recent research (Vanmeirhaeghe et al., 2005; Vanmeirhaeghe, 2006a, b; Owens & Servais, 2007) show that the sedimentary record in the Condroz Inlier is more complete than previously thought (Michot, 1954, 1980; Verniers et al., 2001). In the Central Condroz Inlier, which is the main and most extensive of these thrust sheets, the record extends quite continuously from the Middle Ordovician (Huy Formation) to the lowermost Silurian (Bonne-Espérance Formation) and higher in the Silurian. The careful palaeontological investigations of these authors and their bio-and chronostratigraphy permit establishment of a new stratigraphical correlation chart with the Brabant Massif (Fig. 11). This latter is slightly different from the chart proposed by Vanmeirhaeghe (2006a figs 73, 74) essentially due to progress in the Brabant Massif stratigraphy. Stratigraphical comparison with the two other parts of the Condroz Inlier are less interesting for our purpose because their sedimentary record is too short or discontinuous.

The important stratigraphic hiatus observed in the Lower Ordovician of the Brabant Massif is also observed in the Central Condroz Inlier with a slightly larger time interval (between the Chevlipont and Huy formations; Fig. 11). Two younger gaps are also present: the first in the mid-Darriwilian between the Huy and the Chevreuils formations and a very small second one at the Caradoc-Ashgill boundary between the Vitrival-Bruyère and the Fosses formations (Fig. 11). The first gap is most probably a hiatus of observation due to the extremely poor conditions of exposure and the second is quite definitely a stratigraphic hiatus due to an emergence as suggested by Vanmeirhaeghe (2006a p. 202). This emergence occurs in the Onnian (upper half of the Streffordian) of the British stages that is about 1 Ma older than the lower Pusgillian base of the Madot Formation (Fig. 11). As we have seen previously (§ 2.9.2 and chap. 3), the base of Madot Formation is marked by a drastic change in bathymetry from slope to shallow shelf marking the end of Megasequence 2 in the Brabant Massif. The repercussion of the same event in the Central Condroz Inlier leads to an emergence because this region is globally shallower than the southern Brabant Massif (Vanmeirhaeghe, 2006a).

The old debated question of the Cocriamont conglomerate observed at the base of the Fosses Formation by Michot (1931, 1932, 1934) belongs to the same topic. This conglomerate was interpreted as an unconformity resulting from the Ardenne Deformation Phase (Michot, 1980). After careful

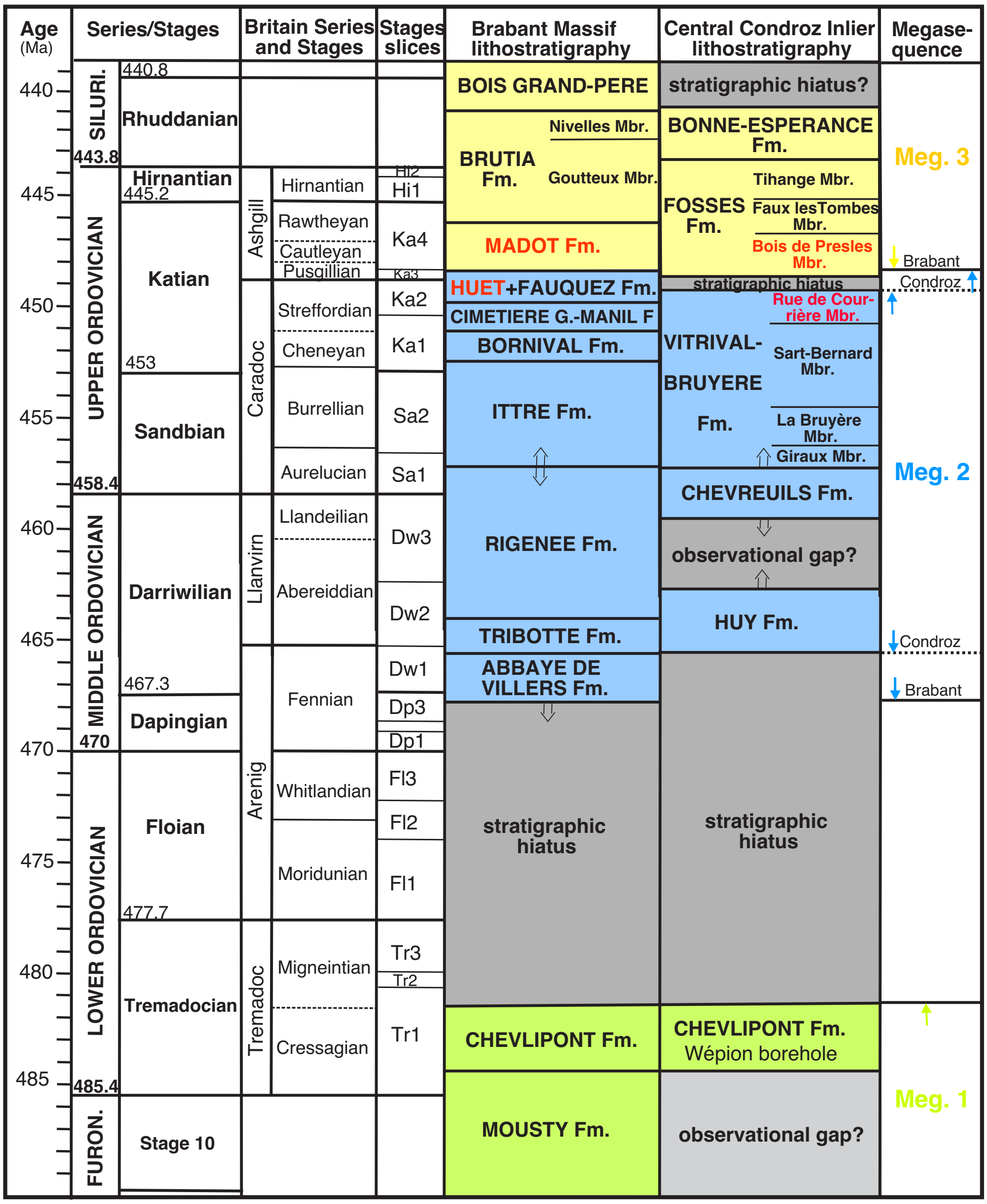

Figure 11: Stratigraphic comparison between the Ordovician lithostratigraphic units of the Brabant Massif (this paper Fig. 3) and of the Central Condroz Inlier (modified after Vanmeirhaeghe, 2006a Figs. 73, 74). The name of the formations and members are *sensu* Vanmeirhaeghe (2006a) and are partially different from those of Delcambre & Pingot (2000a, b, in press), Owen and Servais (2007) and older literature. In particular the Sart-Bernard Member of the Vitrival-Bruyère Formation corresponds to the Sart-Bernard Formation of previous literature. The formation names in red show the first appearance of shelly facies. See text for explanation. Abbreviations: Meg. : Megasequence.

field investigations, Vanmeirhaeghe (2006a pp. 107-117) demonstrated that the angular unconformity does not exist and that the supposed basal conglomerate of the Fosses Formation is the result of reworking at the base of the transgression that followed the Solvang Lowstand Event (Figs 3, 11; *spinifera* Biozone, Webby et al., 2004).

Interestingly the stratigraphycally lowest shelly facies (containing brachiopods, corals, trilobites,…) appears faintly, as lag deposits at the base of tempestites, in the Rue de Courrière Member of the Vitrival-Bruyère Formation (Vanmeirhaeghe, 2006a p. 120) and very clearly in the Bois de Presles Member of the Fosses Formation (Michot, 1934; Lespérance & Sheehan, 1988; Servais et al., 1997; Vanmeirhaeghe, 2006a; Owen & Servais, 2007). The appearence of these levels at the end of the Caradocian (mid Katian) is synchronous with similar shelly levels observed in the Brabant Massif (Fig. 11, name of the formations in red). This confirms the palaeolatitude interpretations of the Brabant Massif (§ 2.9.2) showing a rapid drift of Avalonia to lower latitudes and the influence of the Boda warming Event.

5. Palaeobiogeographical implications of the palaeontologic investigations

The palaeogeographic reconstructions of Cocks & Torsvik (2002, 2005) show a very rapid northward drift of the Avalonia microcontinent to lower latitudes and its approach to Baltica between the Lower and Upper Ordovician. This is confirmed in the Brabant-Condroz Basin by the onset of shelly facies from the upper Caradoc upwards (Fig. 11; § 2.7.2, 2.9.2 and 4) and by the change of detrital zircon sources (Linnemann et al., 2012 fig. 22). This is also confirmed by the palaeontological investigations. In the Brabant-Condroz Basin, chitinozoan assemblages have northern or peri-Gondwana affinities until mid-Darriwilian time (Figs 3, 11), whereas a Baltoscandian signature becomes clear from late Sandbian time until the end of the Ordovician (Verniers et al., 2002a; Vanmeirhaeghe, 2006a fig. 74; Vandenbroucke, 2008). Avalonian chitinozoan endemicity possibly occured during the late Darriwilian to early Katian interval, as suggested by the occurrence of the fauna of the Chevreuils Interzone and that of the early Katian of British Avalonia (Vanmeirhaeghe, 2006a p. 228).

Assemblages of trilobites from the Condroz Inlier show typical affinity with northwestern Gondwana faunas until the upper Sandbian. The Huy Formation shows typical Avalonian trilobite assemblages, almost identical to those recorded from the Skiddawn Group of northern England, and from the Ebbe and Remscheid anticlines in the Rhenisch Massif of Germany. From the Katian, these faunal assemblages indicate increasing proximity with Baltica, Armorica and the Prague Basin (Owen & Servais, 2007). Strong Bohemian affinities are noted for

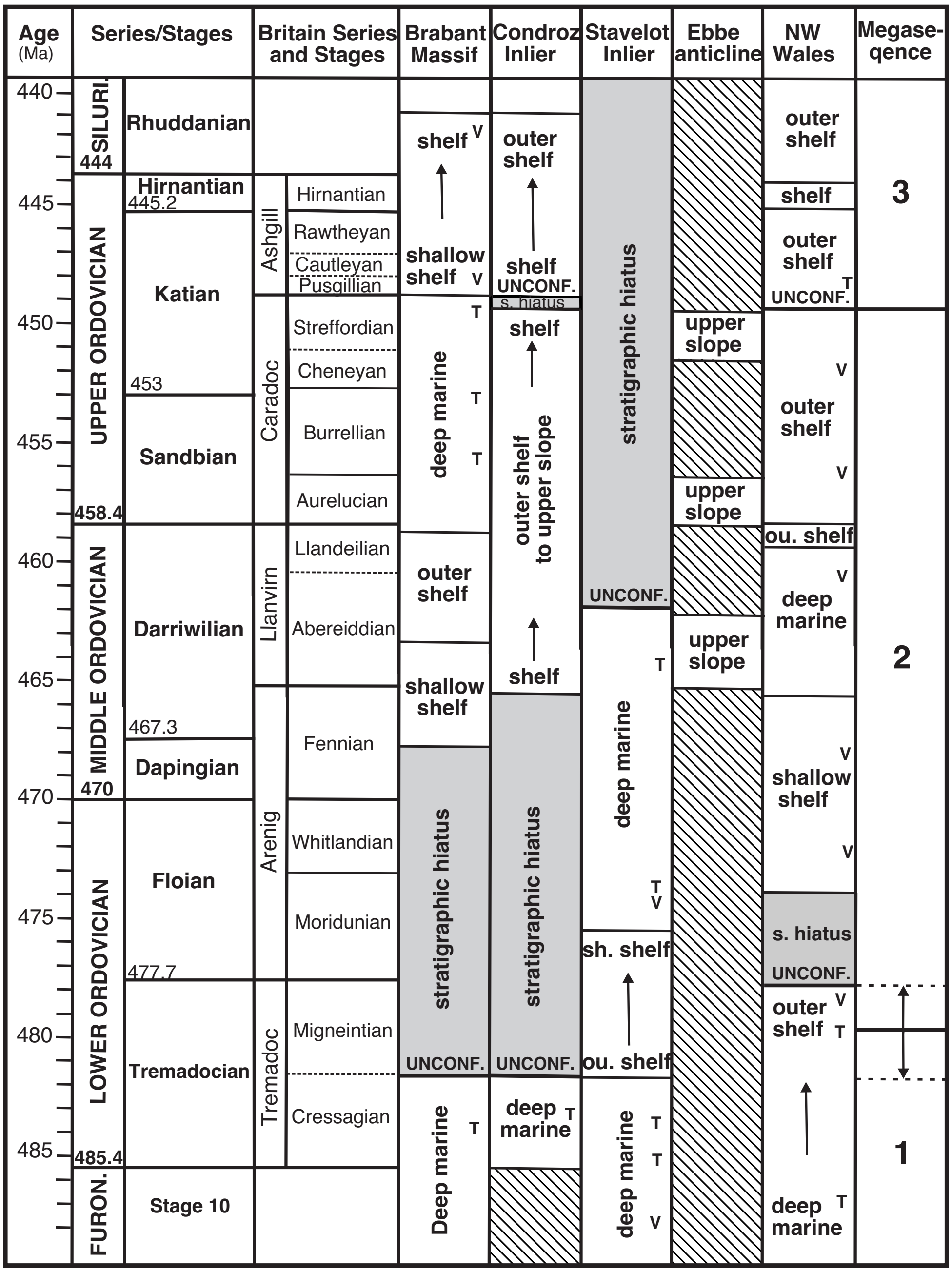

Figure 12: Comparison of the sedimentary registration and environmental interpretation of the facies for the Ordovician of the European part of the Avalonia microplate. Data from Vanguestaine (1992), Lamens (1985), Breuer & Vanguestaine (2004), Vanguestaine et al. (2004) for the Stavelot Inlier (Ardenne, Belgium); from Samuelsson et al. (2002b), Owen & Servais (2007) for the Ebbe Anticline and from Woodcock (2000), Brenchley et al. (2006 p. 46-47) for the Welsh Basin (N Wales, England). Megasequence in the meaning of Woodcock (1990). In grey: stratigraphic hiatus; oblique lines indicate lack of evidence for deposition.
Abbreviations: UNCONF. : unconformity; T: turbidite; V: volcanism; sh.: shallow; ou.: outer.

the macrofauna from the Sart-Bernard Member (Fig. 11, late Sandbian-early Katian) which is unique for Avalonia in that time interval (Mailleux, 1939; Owen & Servais, 2007 p. 290).

6. Discussion about the Brabant-Condroz Basin

Vanguestaine (1992) and Verniers et al. (2002a) proposed a threefold subdivision of the sedimentary succession of the Lower Palaeozoic in Belgium and its surrounding area. This scheme was mainly based on the three megasequences observed in the Welsh Basin by Woodcock (1990). These authors saw the Brabant megasequences as three megacycles evolving from shallow water to deep water deposits, but at that time the stratigraphical and sedimentological interpretations were incomplete and their demonstration not totally convincing. This hypothesis is now confirmed in the Brabant Massif by the most recent investigations (Linnemann et al., 2012, Herbosch & Verniers, 2013; this paper Figs 8, 11). The three megasequences begin in a shelf environment: respectively the Blanmont, Abbaye de Villers and Madot formations, and end in a deep basin: respectively the Chevlipont, Fauquez and Ronquières formations. They can be identified as a deepening upward sequence of second order (10-100 Ma; Fig. 7).

In the Condroz Inlier, the stratigraphical and sedimentological knowledge has greatly evolved since Verniers et al. (2001), particularly for Megasequence 2 and the lower part of Megasequence 3 (cf. § 4). As seen in the figure 11, Megasequence 2 (in blue) begins with the Huy Formation and ends with the stratigraphic hiatus at the top of the Vitrival-Bruyère Formation (see reference at § 4). The Huy Formation is interpreted as sediments of >100 m thick deposited on an outer shelf to upper slope (Vanmeirhaeghe, 2006a ; Owen & Servais, 2007). As the « Formosa flooding Event » (Paris et al., 2007) occurs at about the middle of the formation, we could expect that the Huy Formation sedimentation began on the shelf and continued deeper on the slope. After a gap of exposure, the sedimentation continued with the Chevreuil (>170 m) and Vitrival-Bruyère (>400 m) formations deposited on the outer shelf to upper slope. The Rue de Courrière Member, interpreted by Vanmeirhaeghe (2006a) as tempestites deposited in the mid-shelf, is followed by a stratigraphic hiatus interpreted as an unconformity. By comparison with the Brabant Massif, the environment is globally shallower since the mid-Darriwilian and more influenced by eustatic sea-level fluctuations (Vanmeirhaeghe, 2006a p. 212). So there is no evidence for a deepening upward trend in Megasequence 2, but the megasequence is clearly limited by two unconformities.

After the emergence in the lower part of Megasequence 3, the Fosses Formation, >120 m thick, comprises shelf deposits from the middle Katian to Hirnantian, whereas the Bonne-Espérance Formation (>30 m) shows a deeper environment, with outer shelf to upper slope sediments. Hence, the upper Rhuddanian to mid-Aeronian hiatus is « likely due to a tectonic event » with emergence (linked to the Ardennes Deformation Phase?; Vanmeirhaeghe, 2006a p.165). Moreover, during the Hirnantian-Rhuddanian stages the Condroz shelf was affected by rapid variations of sea level due to glaciation and deglaciation and probably also by tectonic movements that complicate the interpretation.

In conclusion, the hypothesis (Verniers & Van Grootel, 1991 ; Verniers et al., 2002a) that the sediments of the Brabant Massif and those of the Condroz Inlier were deposited in the same basin, the Brabant-Condroz Basin, is entirely confirmed by the recent investigations. The most important arguments are: 1) the unconformities observed between Megasequences 1 and 2 and between Megasequences 2 and 3 are at about the same stratigraphic level in both areas ; 2) the comparable length of the sedimentary record with the same palaeogeographical and palaeoclimatical implications. However, as pointed out by Vanmeirhaeghe (2006a p. 214) clear differences occur between the two areas. Shelf settings prevails in the Central Condroz Inlier from the upper Darriwilian to Rhuddanian while slope to basin environments prevail in the Brabant Massif from the upper Darriwilian to mid-Katian. The subsidence in the Condroz inlier is lower as is shown by the lower thickness of the sedimentation. Effectively, despite the very rough thickness estimates (Vanmeirhaeghe, 2006a) it seems that the total Ordovician thickness for the Central Condroz Inlier of about 800 m is half that of the Brabant Massif (1400 to 2000 m).

Owen & Servais (2007 pp. 278-279) showed also that the successions in the Ebbe and Remscheid anticlines (Rhenisch Massif) are lithologically and biostratigraphically very similar (Maletz & Servais, 1993; Samuelsson et al., 2002b) to that of the Condroz Inlier. They also concluded that it is probable that the Brabant-Condroz Basin extended into northern Germany during the Ordovician.

7. Conclusions

The Rebecq Group constitutes a mostly siliciclastic succession, 1400 to 2000 m thick, deposited from the Middle Ordovician to the lowermost Silurian (468-441 Ma). This thickness is very small compared to the underlying Cambrian and Lower Ordovician deposits which are more than 9 km thick (Herbosch & Verniers, 2013). The succession of the Rebecq Group was deposited during the rapid northward drift of the microcontinent Avalonia in the Iapetus Ocean, forming the Rheic Ocean at its rear (Cocks & Torsvik, 2002, 2005; Linnemann et al., 2012). The Brabant-Condroz Basin lay on this Rheic Ocean passive margin. During this period, Avalonia had probably not much relief and a relatively small area that could explain the relatively small thickness of the deposits (Verniers et al., 2002a; Vanmeirhaeghe, 2006a). The rapid drift and the subduction of the Iapetus Ocean under the northern (mainly NW) active margin of Avalonia could explain the tectonic instability (e.g. megaslump in the Bornival Formation). Hence, the soft docking of Avalonia with Baltica in the middle of the Katian (*circa* 448 Ma; Samuelsson et al., 2002a; Cocks & Torsvik, 2002; 2005; Cocks & Fortey, 2009) corresponds in the Brabant Massif to an abrupt change in bathymetry and a peak of volcanism (Madot Formation, sill of Lessines and Bierghes). This event marks the top of Megasequence 2 which forms a deepening upward sequence lasting about 19 Ma long (468-449 Ma). The new position of the end of Megasequence 2 at the top of the Fauquez Formation, is also supported by the presence of a short stratigraphic hiatus corresponding probably to an emergence in the Central Condroz Inlier (Fig. 11). The two uppermost formations of the Rebecq Group belong to Megasequence 3 and correspond to the initiation of the foreland basin that only really begins from the Bois Grand-Père Formation up (Fig. 8 in green).

The Rebecq Group contains ten formations and several members that can be traced across the entire southern outcrop area of the Brabant Massif without noticeable lateral variations. The sedimentation begins on the shelf with the Abbaye de Villers and Tribotte formations, while the Rigenée Formation marks a rapid drowning (northern Gondwana "Formosa flooding Event") that leads to the deep-sea deposits of the Rigenée (*pro parte*), Ittre, Bornival, Cimetière de Grand Manil, Huet and Fauquez formations. Then, an abrupt change of bathymetry leads to the shallow shelf of the Madot Formation and the mid to outer shelf of the Brutia Formation. The appearance of shelly facies in the Huet and Madot formations testify to the rapid northward drift of Avalonia to lower latitudes and also, for the Madot Formation, to the Boda global warming event that occurred in the late Katian just before the Hirnantian glaciation.

The sedimentary succession of the Central Condroz Inlier is quite complete from the Middle Ordovician Huy Formation to the lowermost Silurian Bonne-Espérance Formation and a precise bio- and chronostratigraphical correlation with the Brabant Rebecq Group is presented (Fig. 11). A short stratigraphical hiatus, interpreted as an emergence is observed at the Caradoc-Asghill boundary, an unconformity that marks the top of Megasequence 2. This hiatus is exactly coeval with the drastic change in bathymetry observed in the Brabant Massif. This confirms the presence of only one sedimentary basin: the Brabant-Condroz Basin. However, the sedimentation in the Central Condroz Inlier is generally shallower and in consequence more influenced by sea-level fluctuations. Its thickness (about 800 m) is about half the value that in the Brabant Massif (1700 m).

On the scale of the Belgian caledonides, comparison with the Ardennes Inliers (Ardenne Allochthon), in particular with the more complete and better known Stavelot Inlier (Fig. 12) is very difficult because of the probable absence of the Lower

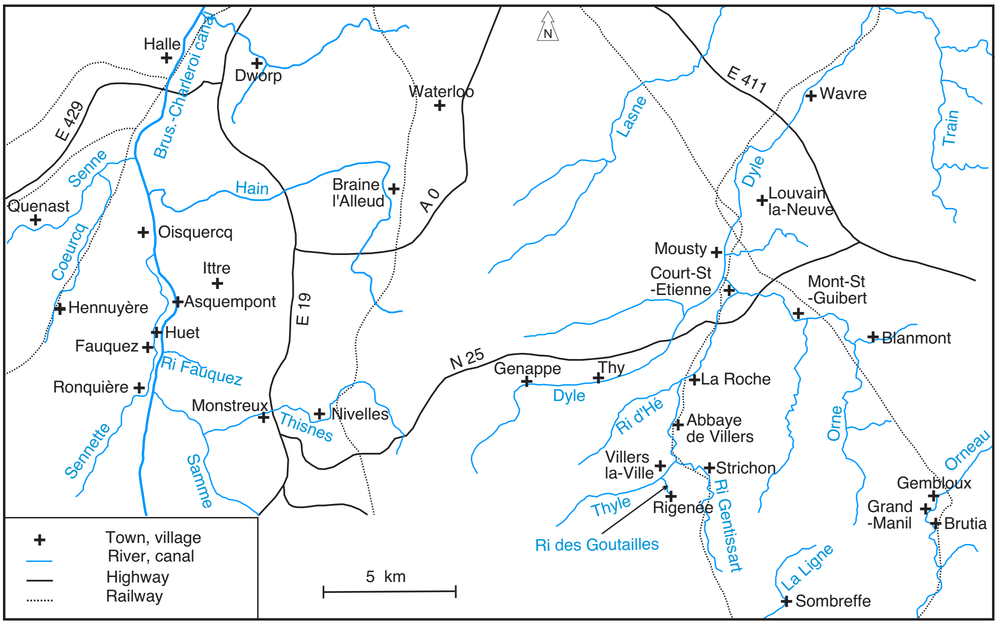

Figure 13: Geographical locations of the rivers, towns, villages, highways and railways referred in the text.

Ordovician hiatus that marks the Megasequence 1-2 boundary in the Brabant (Breuer & Vanguestaine, 2004; Vanguestaine et al., 2004) and also because the sedimentary succession ends earlier towards the upper Darriwilian to lower Sandbian (Vanguestaine, 1974, 1992). It is nevertheless tempting to suggest a prolongation certainly in the Cambrian (Linnemann et al., 2012) and probably in the Ordovician of the Brabant-Condroz Basin to the SW in the Ardenne Inliers.

Comparison at the scale of the European part of the Avalonia microplate ("E Avalonia"; Fig. 12) shows that in the Harlech Dome of North Wales the sedimentary registration of Megasequence 1 reaches the top of the Tremadocian and is marked by an angular unconformity due to an uplift and the onset of arc volcanism (Waldron et al., 2011; Brenchley et al., 2006). The Ordovician transgression of Megasequence 2 begins with early Floian shallow sediments. Then, towards the end of Arenig (lower Darriwilian), a general shallowing occurred, followed by an early Llanvirn transgression (Formosa flooding Event, see §2.3.2) causing a deep-water interval ending with a modest late Darriwilian regression. The early Sandbian "*gracilis* transgression" caused the largest Ordovician rise in sea level accompanied by extensive coastal onlap. High sea level prevailed throughout Sandbian and Katian times until the large and rapid Hirnantian glacio-eustatic fall and subsequent sea level-rise that continued into the early Silurian (Woodcock, 2000; Brenchley et al., 2006 p. 27). A persistent disconformity or angular unconformity around the Welsh Basin in the Ashgill corresponds with an uplift termed the Shelvian event. This story is quite different to that of the Brabant-Condroz Basin as this NE side of Avalonia has been an active margin since the upper Tremadocian until the mid-Caradocian. The closure of Iapetus involves oceanic subduction beneath Eastern Avalonia with formation of abundant igneous rocks. However, Megasequence 2, which corresponds to the Gwynedd Supergroup, is limited by two unconformities that are observed at the same stratigraphic levels as in the Brabant-Condroz Basin (Fig. 12).

This paper will be followed by Part III, devoted to the Orneau Group comprising Silurian and probably the lowermost Lochkovian rocks in the Brabant Massif and the Condroz Inlier.

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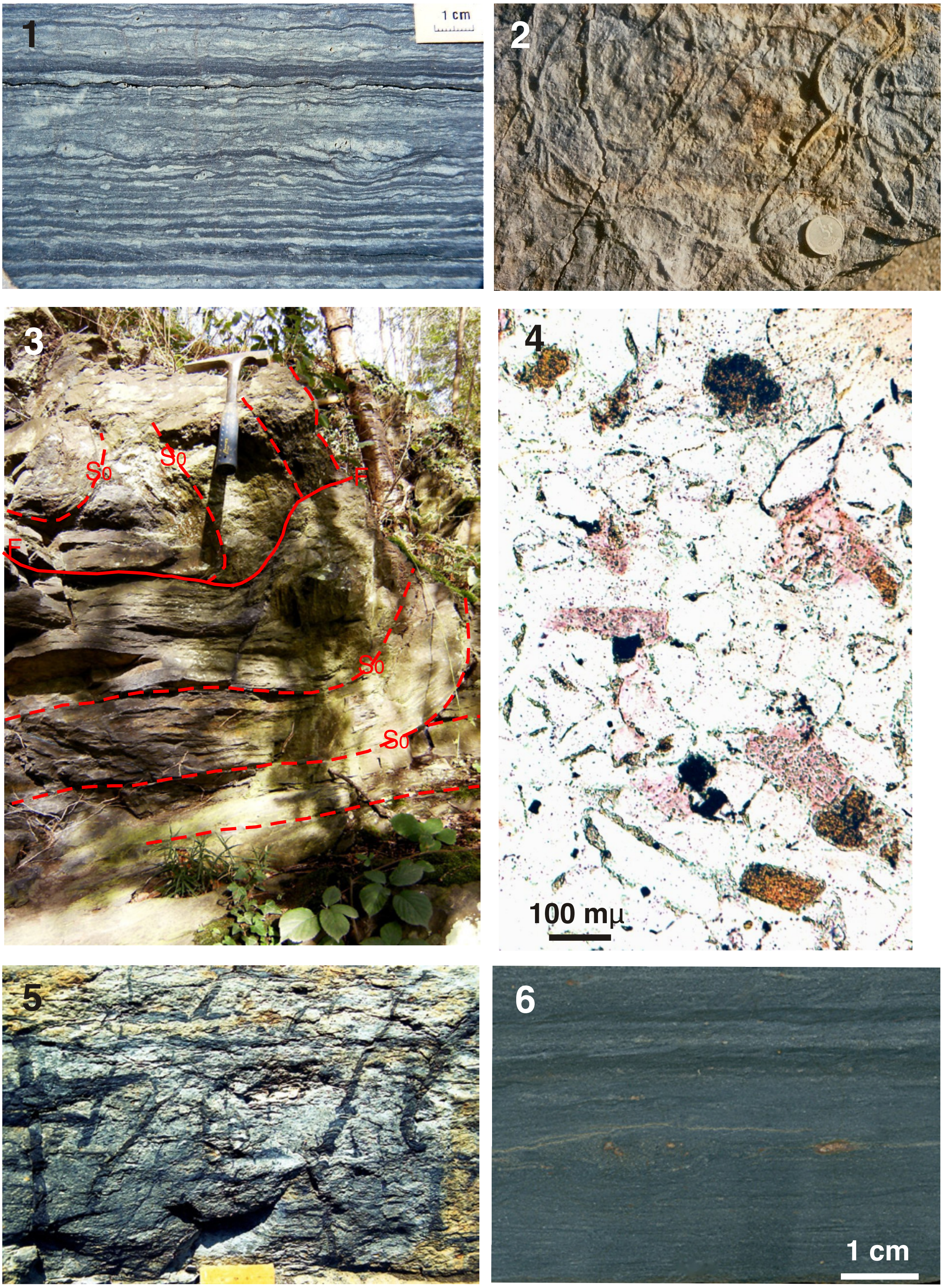

Plate 1

1: Abbaye de Villers Formation, railway cut km 38.8 at the E of the Abbaye. Macrophoto of a sawn and dressed sample. Alternation of siltstone and mudstone laminae perturbated by abundant horizontal bioturbations.

2: Abbaye de Villers Formation, railway cut km 38.7 at the E of the Abbaye. Sample fractured parallel to the stratification. Traces of horizontal bioturbations. The coin is 2 cm in diameter.

3: Upper part of the Abbaye de Villers Formation, western side of the Senne valley near Quenast (Fig. 4). Oucrop affected by low-angle fault (red line F) and folds (dashed red line S0) interpreted as slump folds (outcrop 6 of Debacker & De Meester, 2009).

4: Base of the Tribotte Formation, railway cut SE of the Abbaye just after the bridge km 39.15. Coloured thin section, normal light. Sandstone with K-feldspar coloured in yellow and plagioclase coloured in pink. Quartz crystals are colourless and their outline underlined by impurities.

5: Upper part of the Tribotte Formation, about 200 m to the N of the church of Villers-la-Ville. Macrophoto of a fractured surface. Intertidal facies with pluricentimetric vertical bioturbations with “spreiten”. The yellow scale is 6 cm long.

6: Lowermost part of the Rigenée Formation, sunken road to Rigenée. Macrophoto of a sawn and dressed sample. Bluish grey slate (mudstone) vaguely laminated. The reddish elongate dots are probably bioturbations with altered pyrite.

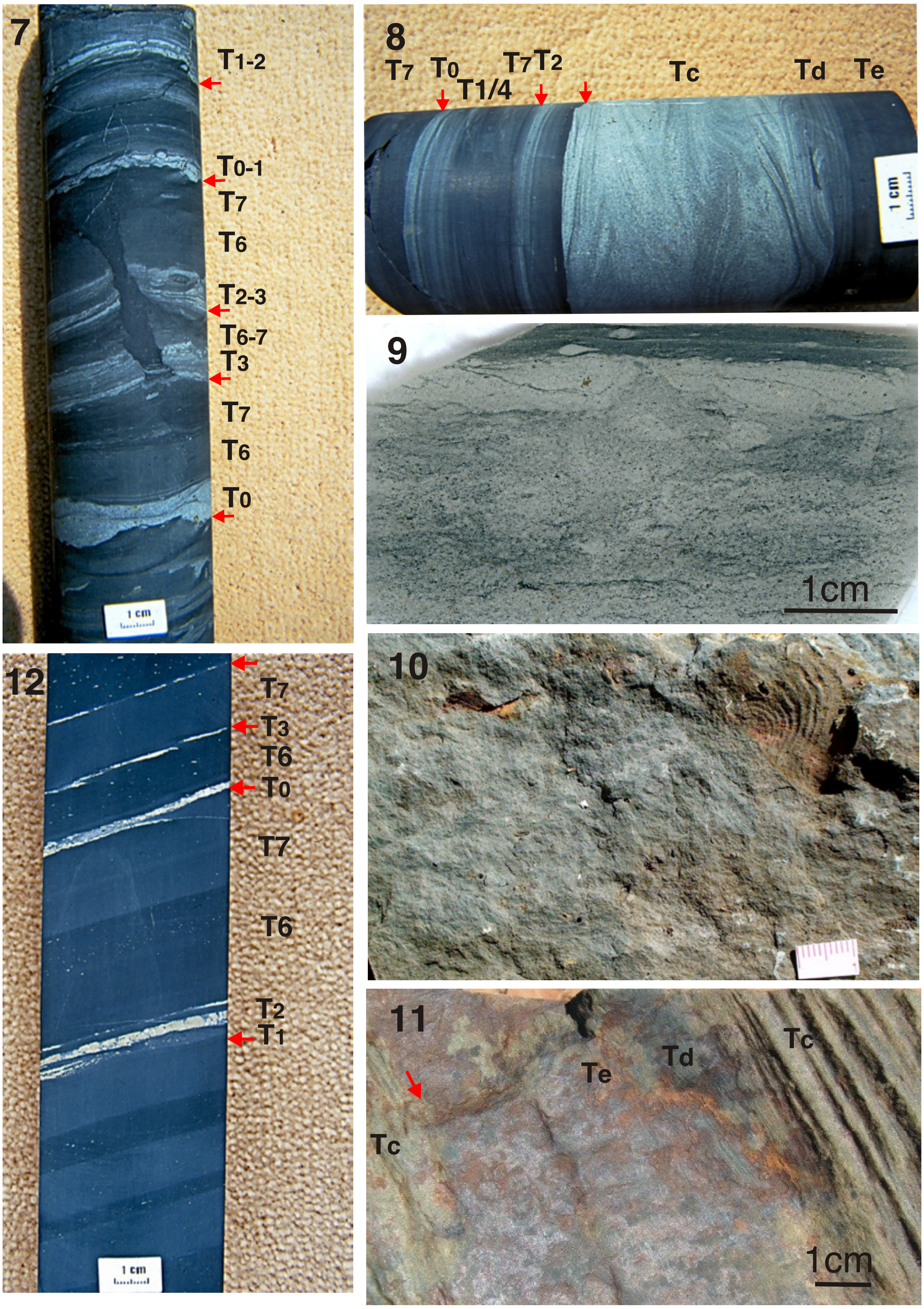

Plate 2:

7: Ittre Formation, Lessines borehole 271 m, Dender valley. Macro of a core. Rhythmic alternation of grey siltstone and black mudstone. The cm-scale, the ryhmicity and the structures succession in the sequences are typical of the low-density turbidite model of Stow (compare with Fig. 5). The base of each sequence (red arrow) begins with a wavy lamina of siltstone (T0), frequently followed by smaller silt laminae (T1, T2, T3) and subsequently by graded (T6) and ungraded (T7) mudstone. The load structure at the base of T0 and ball and pillow structures at the top of the sequences clearly show the polarity. At mid core a large bioturbation crosses two sequences.

8: Ittre Formation, Lessines borehole 213.8 m, Dender valley. Macrophoto of a core. Rythmic alternation of black mudstone and grey siltstone. The first three sequences from the left are low-density turbidites of cm-scale (Stow model see Fig. 5). The upper sequence shows distal high-density turbidite (Bouma model) which begins abruptly with oblique and convolute laminations in a fine grained sandstone to siltstone (Tc), that evolves into plane parallel laminations in a more argillaceous siltstone (Td), then ends in a compact mudstone (Te).

9: Member 1 of the Bornival Formation, E side of Bruxelles-Charleroi canal cut km 39.300 (Debacker *et al*., 2003 fig. 15). Macrophoto of a thin section. Light grey heavily bioturbated fine-grained siltstone alternating with dark grey mudstone.

10: Huet Formation, near Huet old quarry, Sennette valley (Verniers et al., 2005 fig. 7 sample GVG95-148). Macrophoto of a broken sample. Grey siltstone with characteristic orange-yellow alveoli of decalcified fossil fragments. To the right mould of a strophomenid brachiopod.

11: Cimetière de Grand-Manil Formation, Grand-Manil (Gembloux), Orneau valley (Fig. 6). Macrophoto of a sample. Greenish slate with two sequences of Bouma-type distal turbidites. The red arrow marks the summit of the right sequence Tcde.

12: Fauquez Formation, Lessines borehole 149.4 m, Dender valley. Macrophoto of a cut and polished core. Pyritic claystone (black slate) with a rhythmic alternation of pyritic siltstone and mudstone. It is interpreted as base-cut Stow sequences of low-density turbidites (mud-turbidites, op. cit.). Each sequence (red arrow) begins with a fine pyritic siltstone layer (T0) followed by graded (T6) and non-graded mudstone (T7). Compare with Fig. 5.

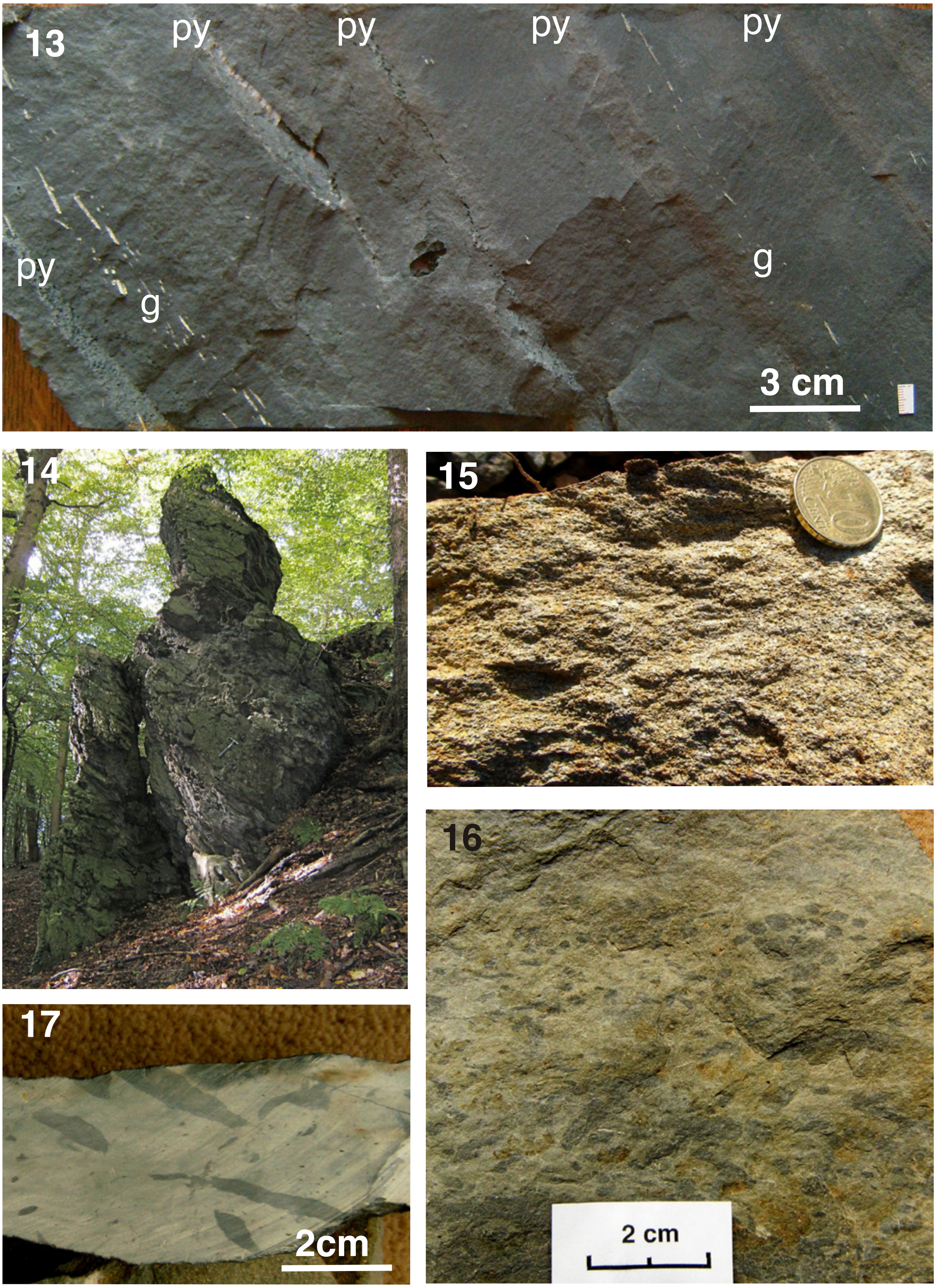

Plate 3

13: Fauquez Formation, temporary outcrop near the old railway-station of Fauquez, Sennette valley (Fig. 4). Macrophoto of a cleaved surface. Pyritic (py) and graptolitic (g) black slate. Base-cut Stow sequences like in photo 12. Small cubic holes mark the location of the pyrite crystals (py) completely dissolved in outcrop and the base of each sequence. Graptolites are poorly preserved and belong to the *Pleurograptus linearis* and/or upper part of the *Dicranograptus clingani* biozones (see text).

14: Upper part of the Madot Formation, "Bois des Rocs" at Fauquez, W side of the Sennette valley. The 5-6 m high "towers" are formed by sub-marine dacite flows that are embedded in mudstone (slate) later removed by erosion.

15: Uppermost part of the Madot Formation, Bruxelles-Mons railway cut km 24.40, Hennuyères, Coeurq valley (Fig. 4). Macrophoto of a broken surface. Volcano-sedimentary tuffs. The coin is 2,5 cm in diameter.

16: Lower Goutteux Member of the Brutia Formation, small excavation along the Bruxelles-Mons cut km 24.60, SSW of Hennuyères, Coeurq valley (Fig. 4). Macrophoto of a broken surface. Strongly bioturbated mudstone with *Chondrites* ("fucoids" of authors).

17: Same provenance as photo 16. Macrophoto of a sawn sample. Larger bioturbations in a dark mudstone.