INSTITUT ROYAL DES SCIENCES NATURELLES DE BELGIQUE

KONINKLIJK BELGISCH INSTITUUT VOOR NATUURWETENSCHAPPEN

ROYAL BELGIAN INSTITUTE OF NATURAL SCIENCES

MEMOIRS OF THE GEOLOGICAL SURVEY OF BELGIUM
N° 62 - 2015

FIELD GUIDE TO THE GEOLOGY OF THE BRABANT MASSIF

THE OUTCROPS OF THE DYLE AND SENNE BASINS

ALAIN HERBOSCH & JACQUES VERNIERS

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Field guide to the geology of the Brabant Massif: the outcrops of the Dyle and Senne basins

by Alain HERBOSCH[1] and Jacques VERNIERS[2]

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40 pages, 37 figures

Cover illustration: Sequence of turbidites from the Rogissart Member of the Tubize Formation. Stop 9 along the road Braine-le-Château to Clabecq, Hain valley.

Maps in annex page 4:
top: Bouguer anomaly map of the northern part of Belgium (Everaerts & De Vos, 2012).
bottom: Aeromagnetic map of the northern part of Belgium (Geological Survey of Belgium, 1994).

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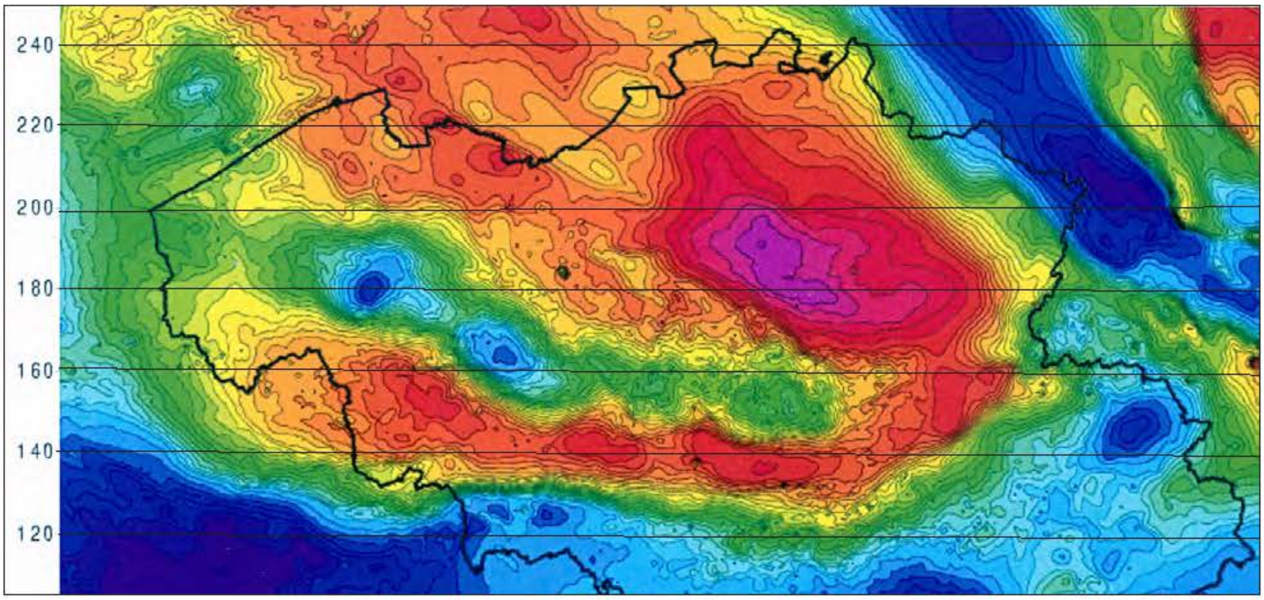
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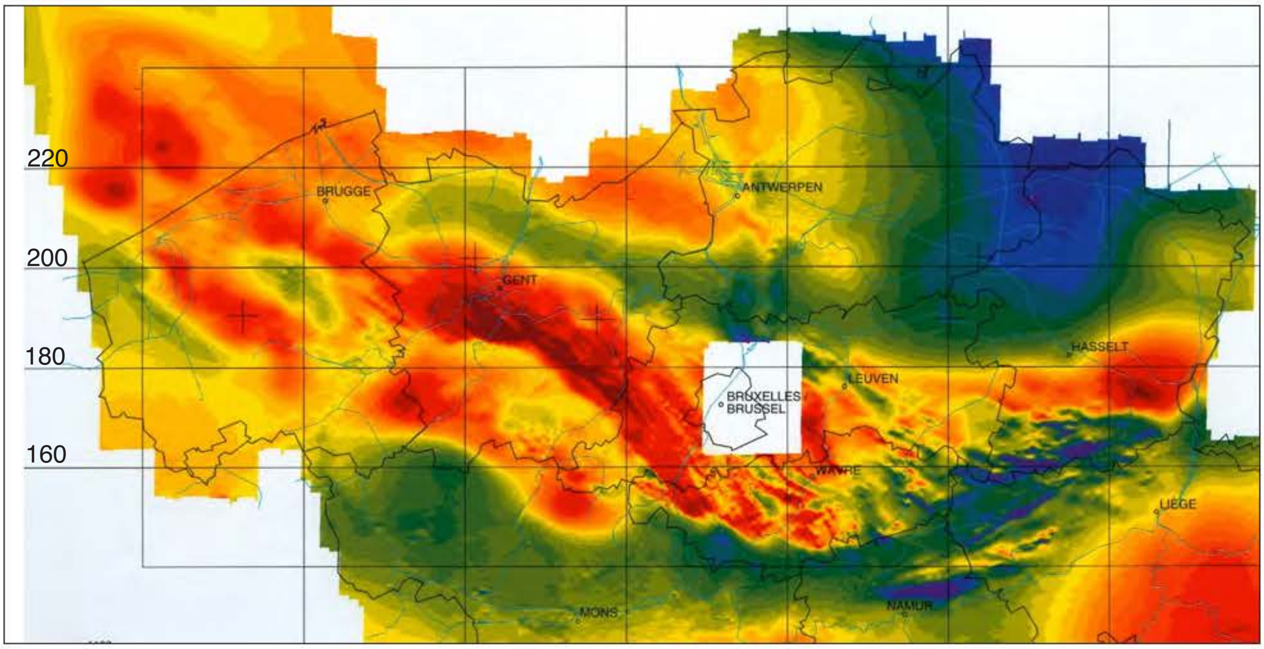
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FIELD GUIDE TO THE GEOLOGY OF THE BRABANT MASSIF: THE OUTCROPS OF THE DYLE AND SENNE BASINS

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Abstract. This field guide will provide a good insight into the Lower Palaeozoic geology of the Dyle (first day) and Senne (second day) basin areas, which are among the most important and extensive outcropping zones of the Brabant Massif. Fifteen selected outcrops are visited. They cover all the stratigraphic range observed in the Brabant Massif from the Lower Cambrian (Blanmont Formation) to the upper Silurian (Ronquières Formation) and also the Brabantian unconformity. An up-to-date geological introduction syntheizes the most recent publications and the results of the recent 1/25,000 scale mapping of the Brabant. In each stop, a detailed description is provided of the location, stratigraphy, lithology and structural architecture, followed by interpretations. The observations and their implications are placed in the broader context of the Cambrian to Devonian odyssey of the Brabant Massif within the wandering history of the Avalonia microplate.

Keywords: regional geology, Belgium, Cambrian, Ordovician, Silurian

1. Excursion programm

1.1. First day: outcrops of the Dyle Basin (Fig. 1)

Stop 1: near the old mill Al Vau (Blanmont), Orne valley. Massive quartzite and greenish sandstone of the **Blanmont Formation**. Upper part of Terreneuvian (?) based on ichnofossil *Oldhamia radiata*.

Stop 2: Mont-St-Guibert church, Orne valley. Rhythmic alternances of sandstone, siltstone and slate of the lower part of the **Tubize Formation**. Stage 3 (lower Cambrian) based on ichnofossil *Oldhamia radiata*.

Stop 3: Beaurieux highway bridge, Orne valley. Very compact green slate with magnetite of the **Tubize Formation**. Same age.

Stop 4: old quarry of Franquenies (Mousty), Dyle valley. Graphitic, pyritic and manganiferous slate with some chert (phtanite) lenses from the lower part of the **Mousty Formation**. Furongian (upper Cambrian) based on acritarchs. Unconformity of Eocene sands over the Caledonian basement, only some sandstone and black slate pebble.

Stop 5: railway trench near the old mill of Chevlipont, Thyle valley. Rhythmic alternances of silt and mudstone laminae with wavy bedding of the **Chevlipont Formation**. Lower Tremadocian based on the graptolites *Rhabdinopora flabelliformis* and acritarchs.

Stop 6: W side of the Thyle valley 300 m to the N of the abbey. Dark brown bioturbated siltstone and argillaceous sandstone from the **Abbaye de Villers Formation**. Upper Dapingian to the lowermost Darriwilian based on acritarchs and chitinozoans.

Stop 7: wall of the church of Villers-la-Ville, Thyle valley. Very heavily bioturbated siltstone and argillaceous sandstone of the upper part of the **Tribotte Formation**. Lower half of the Darriwilian (upper Arenigian to lowermost Llanvirnian) based on acritarchs and chitinozoans.

Stop 8: old sunken road to Rigenée, Thyle valley. Rapid transition to the dark siltstone (slate) of the **Rigenée Formation**. Second half of the Darriwilian to lowermost Sandbian (Llanvirnian to lowermost Caradocian) based on acritarchs and chitinozoans. Short walk to the old castle Le Chatelet (XIII century).

1.2. Second day: outcrops of the Senne Basin (Fig. 1)

Stop 9: Rogissart, road Braine-le-Chateau to Clabecq, Hain valley. Rhythmic alternances of sandstone (greywacke), siltstone and slate of the middle member of the **Tubize Formation** (Rogissart Mbr.). Series 2 (lower Cambrian) based on the ichnofossil *Oldhamia*.

Stop 10: Asquempont bridge over the Brussels-Charleroi canal, Sennette valley. East side of the canal trench:

a) Greenish slate without stratification of the upper member of the **Oisquercq Formation** (Asquempont Mbr.). Upper part of the Stage 4 to lower part of the Stage 5 (lower to middle Cambrian boundary) based on acritarchs.

b) The massive greenish slate changes progressively to green stratified slate. At the 40.12 km of the canal, the major Asquempont Fault divides the Cambrian core of the Brabant Massif from its Ordovician and Silurian rim.

Stop 11: sunken road of Fauquez, Sennette valley. Black, pyritic and graptolitic slate of the **Fauquez Formation**. Very short interval in the middle Katian (uppermost Onnian to lowermost Pushgillian) based on graptolites and chitinozoans.

Stop 12: “Bois des Rocs” in a small valley to the W of Fauquez. Dacitic underwater flows form towers by erosion of the surrounding slate of the **Madot Formation**. Upper part of the Katian (lower Pushgillian to middle Rawtheian) based on chitinozoans.

Stop 13: Ronquières, east side of the Brussels-Charleroi canal. Rhythmic alternations of siltstone, argillaceous siltstone and slate from the **Ronquières Formation**. Gorstian (lower Ludlow) based on graptolites and chitinozoans.

Stop 14: Ronquières, northern slope of the Samme river valley. Red conglomerate of the base of the Bois de Bordeau Formation (Les Mautiennes Mbr.). Givetian (Brabant Parautochthon).

Stop 15: Inclined shiplift of Ronquières, Brussels-Charleroi canal. View of the **Brabantian unconformity** that superposes the Givetian conglomerates of the Bois de Bordeaux Formation over the Gorstian Ronquières Formation. The stratigraphic hiatus is about 30 Ma.

This field trip has been organized for the **Groupe Français du Paléozoique** September 23th and 24th, 2013.

A third major outcropping area of the Brabant Massif, the Gette Basin (Jodoigne), has been visited by an excursion organized by Geologica Belgica, September 19th, 2009. The guidebook was published by Debacker & Herbosch (2011).

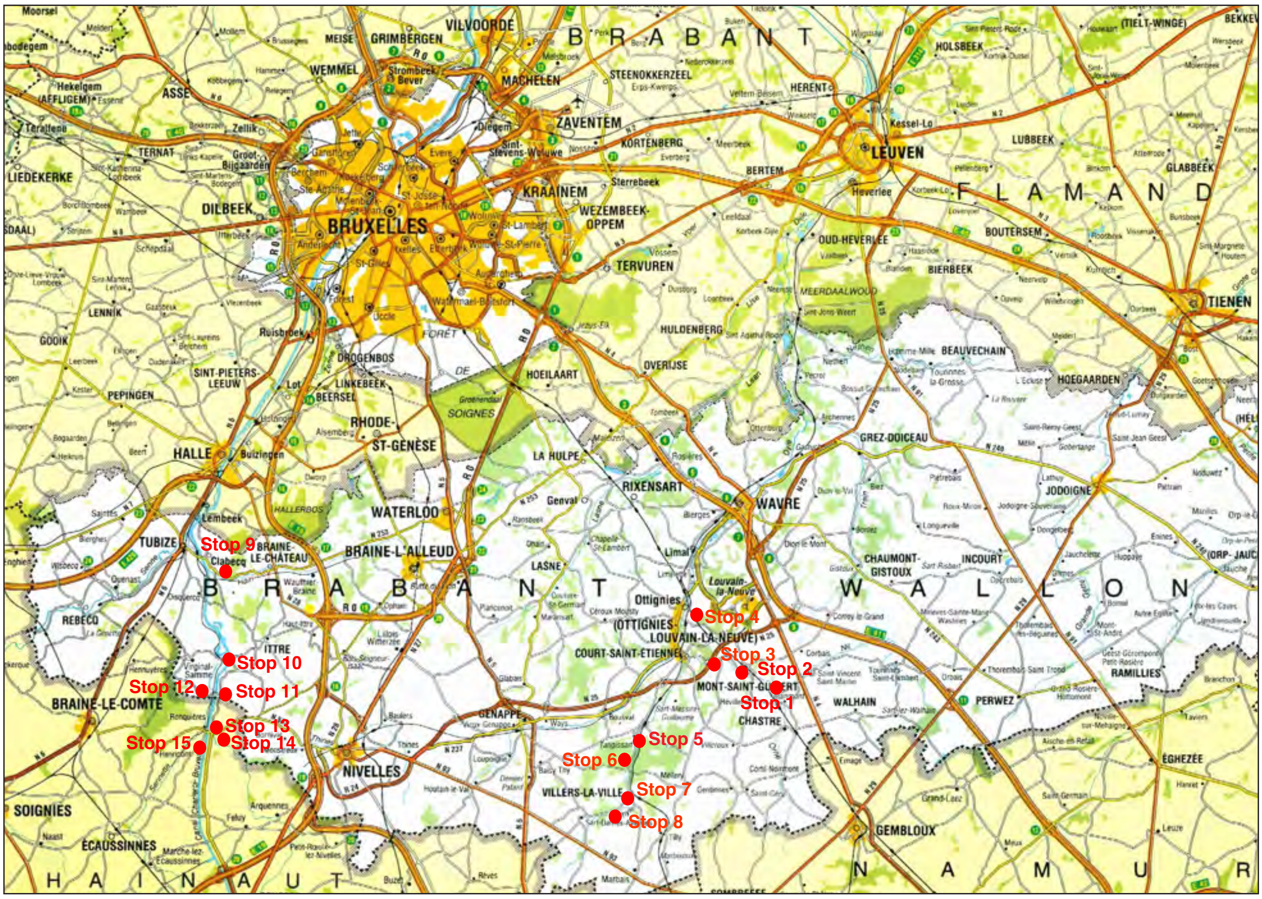


Figure 1. Geographical location of the stops in the “Brabant Wallon” Province.

2. Introduction to the geology of the Brabant Massif

2.1. General overview

The Brabant Massif (BM) consists of a largely concealed WNW-ESE directed fold belt developed during Early Palaeozoic times, documented in the sub-surface of central and north Belgium (Fourmarier, 1920; Legrand, 1968; De Vos *et al.*, 1993; Piessens et al., 2006). At first sight, it appears as a gently ESE-plunging broad anticlinal structure, with a Cambrian core flanked on both sides by Ordovician to Silurian strata (Fig. 2A). To the S, SW and SE, it is unconformably overlain by the Devonian to Upper Carboniferous deposits of the Brabant Parautochthon (Fig. 3; Mansy et al., 1999; Belanger et al., 2012). Its southeastern border is marked by the Condroz Inlier, a 70 km long to 1 to 3 km large strip, elongated W-E along the Sambre and Meuse rivers. These Upper Palaeozoic thrust sheets belong to the BM (Verniers et al., 2002). The Brabant Parautochthon is tectonically overlain by the Ardenne Allochthon along the Midi Fault System that marks the Variscan Thrust Front (Fig. 3, 4). To the NW, the massif continues beneath the North Sea and links up with the East-Anglian Basin (Fig. 2B), (Woodcock & Pharao, 1993). Both areas form part of the Anglo-Brabant deformation belt (ABDB; Pharaoh *et al.*, 1993, 1995; Van Grootel *et al.*, 1997), the eastern branch of a predominantly concealed slate belt moulded around the Neoproterozoic Midlands Microcraton. The ABDB belongs to the Avalonia microplate (Figs. 2B, 4).

The BM is poorly exposed and is almost completely covered by Meso-Cenozoic deposits (Fig. 2; Fourmarier, 1920; Legrand, 1968). Along its southeastern rim, incising rivers provide narrow outcrop areas that were recently mapped at the 1/25000 scale (10 maps) in the framework of the new geological map of Wallonia. A substantial part of the data is based on boreholes (Legrand, 1968; Herbosch *et al.*, 1991, 2008b) and the interpretation of gravimetric and aeromagnetic surveys (cover figures) (BGS, 1994; Mansy et al., 1999; Sintubin & Everaerts, 2002; Everaerts & De Vos, 2012).

The BM displays a very thick siliciclastic, often turbiditic sequence, ranging from the upper part of the lower Cambrian in the core to the upper Silurian and even to the lowermost Devonian along the rims (Fig. 5). Twenty years of stratigraphic research (Vanguestaine, 1991, 1992; André *et al.*, 1991; Maletz & Servais, 1996; Herbosch & Verniers, 2002; Verniers *et al.*, 2001, 2002; Vanmeirhaeghe *et al.*, 2005; Herbosch *et al.*, 2008 a, b; Owen and Servais, 2007; Debacker & Herbosch, 2011) and mapping data suggests that the sedimentary record is continuous, with the exception of an important hiatus in the Lower-Middle Ordovician. The total thickness of this sedimentary pile exceeds 13 km, with over 9 km for the Cambrian only (Herbosch *et al.*, 2008a). The Brabant rocks show a low-grade metamorphic overprint, ranging from epizone in the core to anchizone and diagenesis in the rim (Van Grootel *et al.*, 1997).

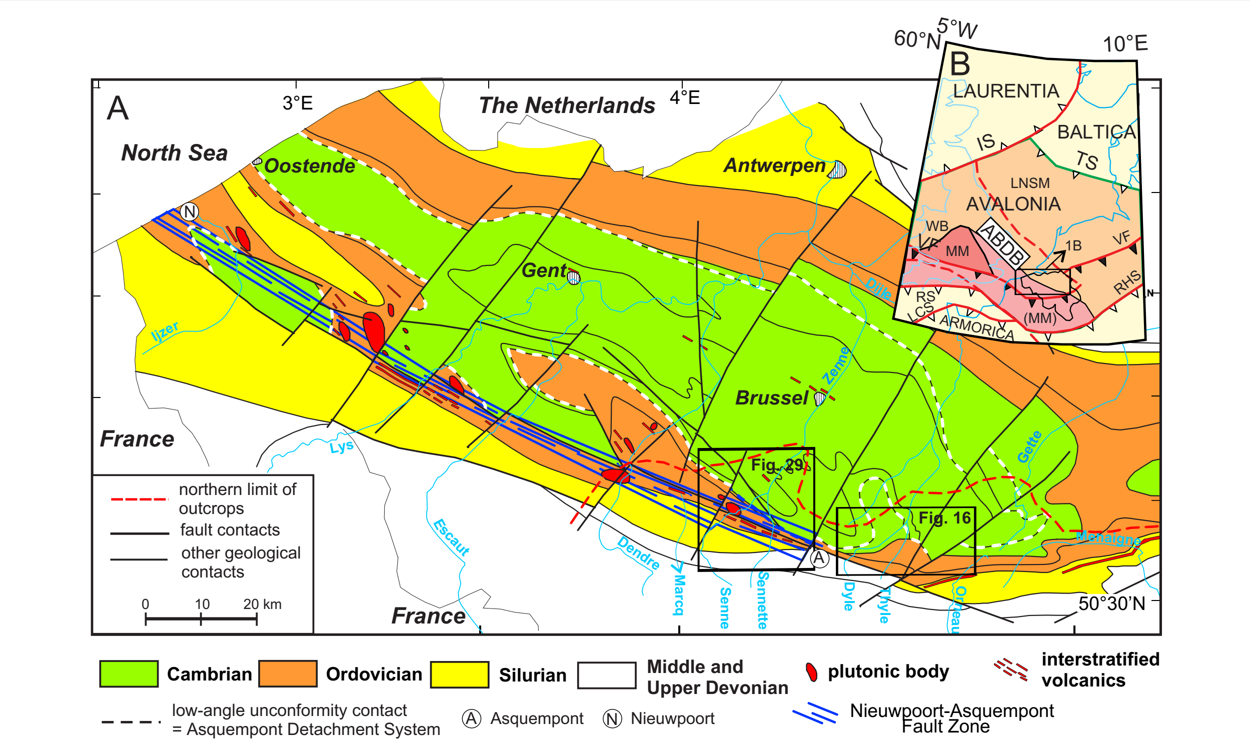


Figure 2. (A) Geological subcrop map of the Brabant Massif (after De Vos et al., 1993; Debacker et al., 2004b) (B) position of the BM within the Anglo-Brabant Deformation Belt (ABDB) along the NE-side of the Midlands microcraton (MM). IS: Iapetus suture; LCS: Le Conquet suture; LNSM: Lüneburg-North Sea microcraton; TS: Tornquist suture; RHS: Rhenohercynian suture; RS: Rheic suture; VF: Variscan front; WB: Welsh Basin (after Sintubin et al., 2009).

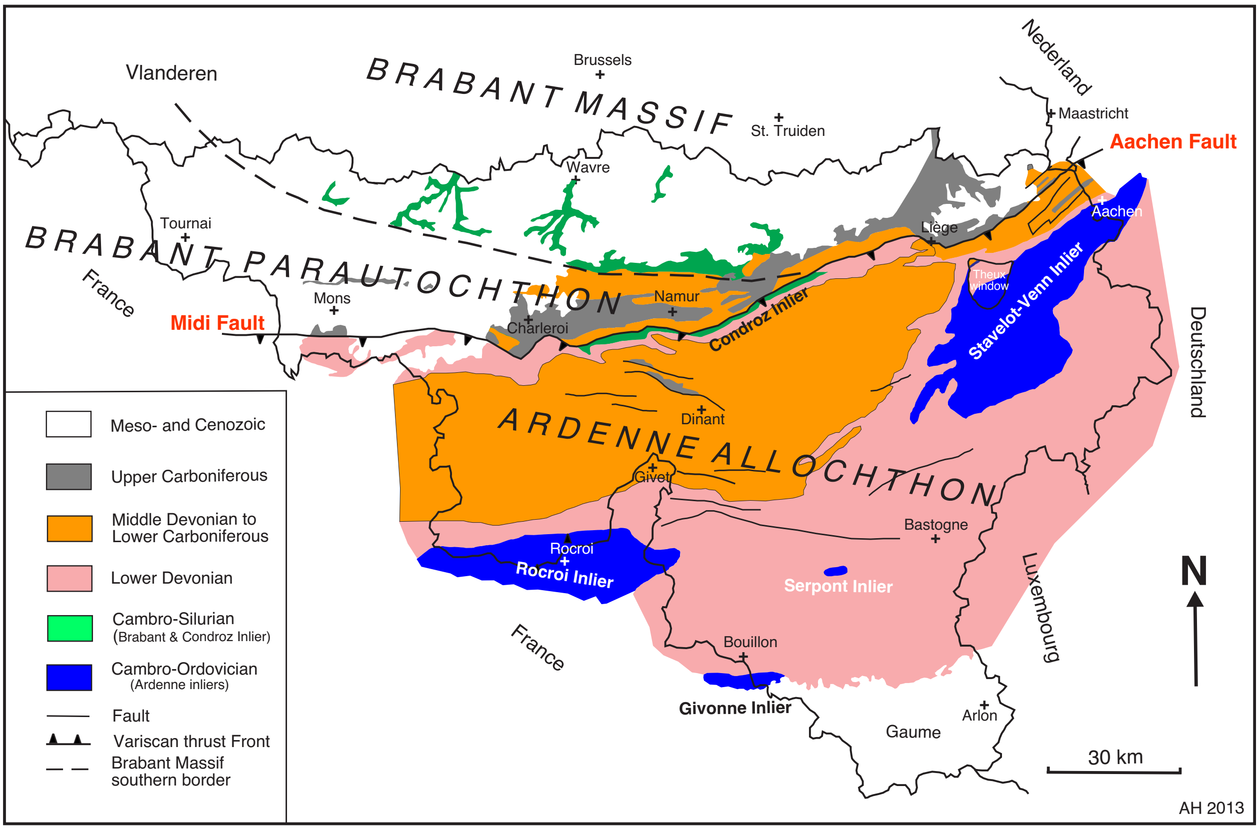


Figure 3. Caledonian basement and main geological structures of southern Belgium and surrounding countries.

Position in the Central European Variscides

The BM belongs to the Avalonia microplate lying in its southern central part (Fig. 4) that corresponds to the Gondwana side during the Cambrian and the passive margin during the Ordovician and Silurian (Fig. 6). The BM was not affected by the Variscan Orogeny, the orogenic front of which is about ten to thirty km to the S along the Midi Fault (Fig. 3).

In Belgium and surrounding countries the Caledonian basement is represented:

- by the BM and the Condroz Inlier, which were deposited in the same basin, the Brabant-Condroz Basin (Verniers et al., 2002; Herbosch & Verniers, 2014);
- by the four Ardenne inliers: Venn-Stavelot, Rocroi, Givonne and Serpont massifs (Fig. 3) that were also affected by the Variscan Orogeny (Ardenne Parautochthon).

2.2. Stratigraphy, sedimentology and basin evolution

The three sedimentary megasequences bounded by basin-wide unconformities recognized in the Welsh Basin by Woodcock (1990) have also been found in the Brabant Massif (Woodcock, 1991; Vanguestaine, 1992; Verniers et al., 2002; Herbosch & Verniers, 2013). The basement of the sedimentary sequence is not known at the surface neither in boreholes. It has been speculated that the Brabant lies above a Neoproterozoic block (André, 1991) or alternatively on the boundary of two lithospheric blocks, namely the Midlands microcraton to the SW and the Lüneburg-North Sea microcraton to the NE (Fig. 9; Sintubin & Everaerts, 2002, Sintubin et al., 2009).

2.2.1. Megasequence 1: a > 9 km Cambrian rift sequence

Megasequence 1 begins with the >1.5 km thick *Blanmont Formation* which consists of massive sandstone and quartzite with sparse slate intercalations (Fig. 5). The lithology, the important thickness and the sedimentary structures (cross stratification, dewatering structures) indicate a high rate of sedimentation in a shallow and rapidly subsiding basin suggesting a rift environment (Debacker & Herbosch, 2011). This is confirmed by the very thick remaining part of the megasequence (>7.5 km) that was continuously deposited in a deep-sea marine environment. This deep-sea sequence comprises 5 formations spanning all the Cambrian from the base of Stage 3 to the lowermost Ordovician. Complete sections and contacts between these formations are nowhere observed, (micro-) fossils are scarce and their stratigraphical succession has been firmly established only recently (Herbosch *et al.*, 2008a; NCS, 2011).

Figure 4. Geology of the Central European Variscides with location of the BM and Fig. 2A (after Ballèvre et al., 2009).

The > 2 km thick *Tubize Formation* consists of green slate, siltstone, (arkosic) sandstone and greywacke interpreted as turbidites and hemi-pelagic to pelagic sediments. Magnetite is commonly observed in the siltstones inducing high relief in the aeromagnetic map (Chacksfield *et al.*, 1993; Sintubin, 1999; Sintubin and Everaerts, 2002). This deep-seated sedimentation continues with the *Oisquercq Formation*, ~1.5 km thick, formed by very fine and homogeneous slates, purple at the base and green at the top. The Blanmont, Tubize and Oisquercq formations are dated (*Oldhamia* or acritarchs) from the same age interval between Cambrian Stage 3 and three-quarter of Stage 5 (Vanguestaine, 1992; Herbosch and Verniers, 2011). The *Jodoigne Formation*, >3 km thick, comprise slate, siltstone, sandstone and massive quartzite mostly of black colour. It is interpreted as a sequence of turbidites and hemi-pelagic to pelagic sediments deposited in an anoxic, deep-sea environment. In the absence of any fossils, this formation is supposed to be middle Cambrian (Series 3) in age (Herbosch *et al.*, 2008a, NCS, 2011). The deep, anoxic sedimentation continued during the *Mousty Formation*, with a minimum thickness estimate of 1 km. The Mousty sediments are mainly black graphitic and pyritic slates with episodic low-density turbidites. Frequent enrichments in Mn (up to 5 % MnO, André *et al.*, 1991) are shown by black coatings in outcrops and by spessartine and Mn-ilmenite porphyroblasts in thin sections (de Magnée & Anciaux, 1945; André *et al.*, 1991). The Mousty Formation ranges at least from the lower part of Furongian (Vanguestaine, 1992) to the lowest Tremadocian (graptolite *Rhabdinopora flabelliformis*; Lecompte, 1948). It passes gradually to the *Chevlipont Formation* formed by wavy bedded gray siltstone interpreted as silt turbidite. This formation, about 150 m thick, is dated from the Early Tremadocian by graptolites and acritarchs (Lecompte, 1948; André *et al.*, 1991; Vanguestaine, 1992; NCS, 2011; Herbosch & Verniers, 2013) and shows a shallowing trend.

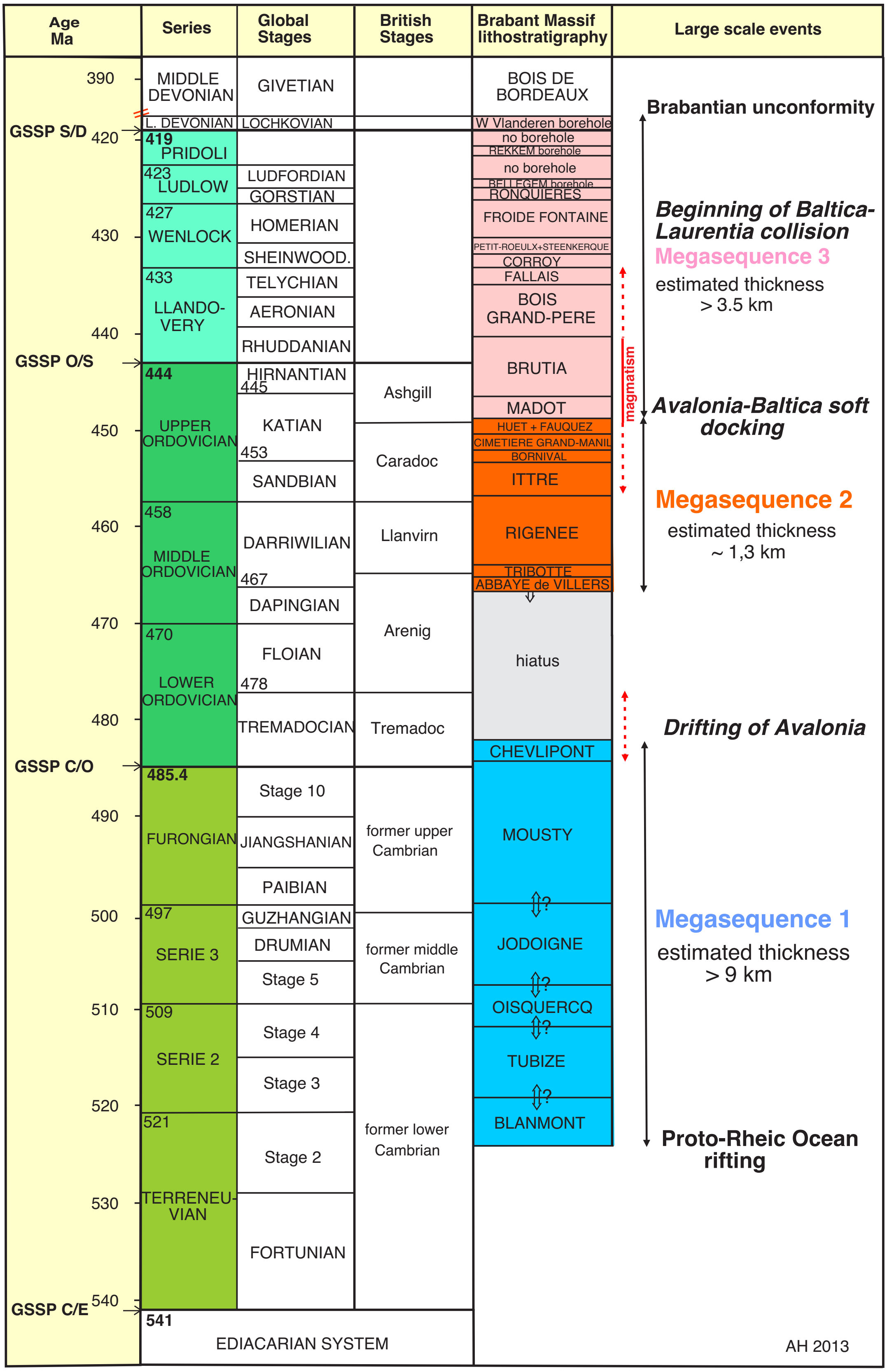


Figure 5. Chrono- and lithostratigraphy of the Brabant Massif sedimentary registration (chronostratigraphy after Gradstein et al., 2012).

The thickness of Megasequence 1 is thus estimated at a minimum of 9 km, which is a huge value, only recently well established (Herbosch *et al*., 2008a; Herbosch & Verniers, 2013). Such a thickness and the inferred deep-sea setting that point to a deep basin environment have also recently been depicted in the Cambrian of the Harlech Dome (Wales) and Meguma (Nova Scotia) terrane (Waldron et al., 2011).

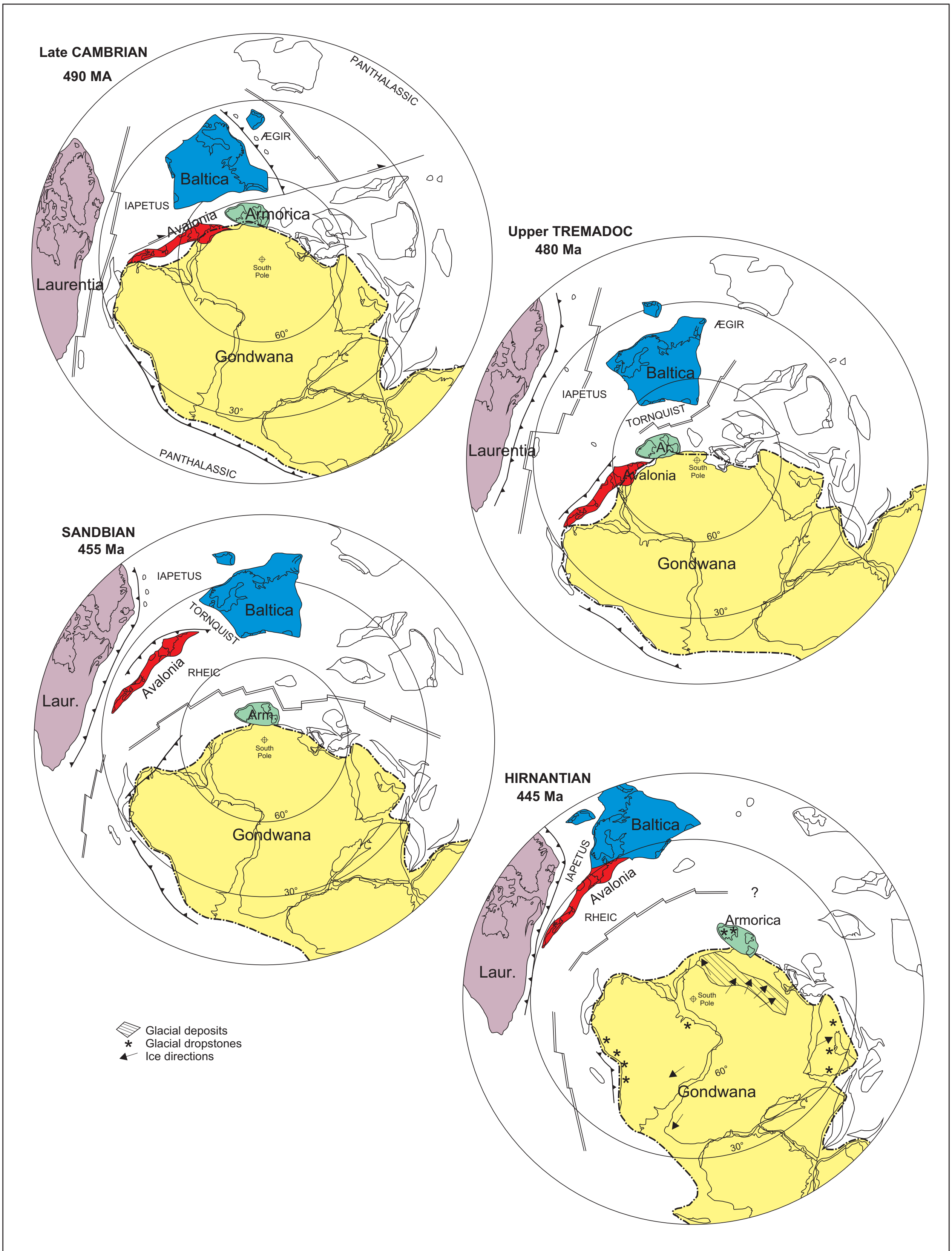


Figure 6. Cambrian to Ordovician palaeogeographic reconstruction after Cocks & Torsvik (2002).

Megasequence 1 is interrupted by a stratigraphic hiatus extending from the lower part of the Tremadocian (*circa* 483 Ma) to the upper part of the Dapingian or even the lowermost Darriwilian (c. 467 Ma) (Vanguestaine & Wauthoz, 2011; Herbosch & Verniers, 1013). This unconformity is believed to reflect the drifting of the Avalonia microcontinent from Gondwana and the opening of the Rheic Ocean (Verniers *et al*., 2002; Herbosch and Verniers, 2013, Herbosch et al., 2014). This event was dated from the end of the Tremadocian, around 480 Ma ago (Cocks and Torsvik, 2002, 2005; Cocks and Fortey, 2009).

2.2.2. Megasequence 2: Avalonia as a wandering microcontinent

The contact between megasequences 1 and 2 has not been observed in the main outcropping part of the Brabant Massif. But it was observed in the Wépion borehole (Condroz Inlier) where a 5 cm thick basal microconglomerate of lower Darriwilian age confirms the unconformity (Graulich, 1961; Vanmeirhaeghe, 2006).

The sedimentary record begins with the *Abbaye de Villers* and *Tribotte Formations* made of outer to inner shelf clayey sandstone and siltstone. The upper part of Tribotte evolves to an intertidal facies with abundant vertical bioturbation (Herbosch & Lemonne, 2000). An abrupt transition exists with the overlying *Rigenée Formation*, essentially dark siltstones to mudstones that record a regional transgression. The latter is well known in North Gondwana and Baltica as the "*Formosa* flooding Event" (Paris et al., 2007). The remaining part of Megasequence 2 corresponds to a deep marine environment. These Upper Ordovician deposits consist of more or less distal dark turbiditic or hemi-pelagic sediments (*Ittre*, *Bornival* and *Fauquez formations*), very distal green turbiditic and pelagic slate (*Cimetière de Grand-Manil Formation*; Debacker *et al.*, 2011; Herbosch & Verniers, 2014), and gray muddy sandstone with some bioclastic levels (e.g.: corals, bryozoans, crinoids; *Huet Formation*) interpreted as tempestites (Verniers *et al.*, 2002) or distal turbidites (Herbosch, 2005) (*Huet Formation*). The Huet and Madot formations, of mid to upper Katian age (Figs. 5, 8), both show shelly facies which have recorded the rapid shift of Avalonia to lower latitudes as well as the pre-Hirnantian global "Boda Event" (Fortey & Cocks, 2005).

The mean thickness of Megasequence 2 is estimated at about 1300 m (new estimate; NCS, 2011; Herbosch & Verniers, 2014); this is considerably less than Megasequence 1. Megasequence 2 is believed to correspond to the behaviour of Avalonia as an independent terrane, when the microplate drifted very rapidly to lower latitudes from Gondwana to Baltica, with which it collided at *circa* 450 Ma (Fig. 6; Cocks and Torsvik, 2005; Cocks & Fortey, 2009).

2.2.3. Megasequence 3: from Baltica to Laurentia docking

New stratigraphic investigations (Vanmeirhaeghe *et al.*, 2005; Herbosch, 2005; Debacker *et al.*, 2011) show the absence of any unconformity or hiatus in the Upper Ordovician as supposed until recently. Accordingly, we define the Megasequence 3 lower boundary at the base of the *Madot Formation* (~ 448 Ma, upper Katian), in a slightly higher stratigraphic position than Verniers *et al.* (2002) and Sintubin *et al.* (2009). Indeed, the transition documents a drastic change in palaeobathymetry, from deep-sea to shallow shelf accompanied by an important magmatic episode. The sedimentation changed abruptly from black pyritic slate (*Fauquez Formation*) interpreted as mud-turbidites deposited on the slope (Herbosch *et al.*, 1991; NCS, 2011) to shelly facies and volcano-sedimentary cross-bedded tuff (*Madot Formation*, from 100 to >220 m) deposited on a very shallow shelf (NCS, 2011; Herbosch *et al.*, 2013; Herbosch & Verniers, 2014).

Megasequence 3 records shelf deposition from upper Katian to mid Telychian (*Madot*, *Brutia* and *Bois Grand-Père formations*, about 500 m thick together) that evolved rapidly into deep-sea turbiditic deposits from mid Telychian to upper Gorstian. Thick turbidite sequences are well developed from the mid Telychian (*Fallais Formation*, 500 to 600 m thick), continued during the Sheinwoodian and Homerian (*Corroy*, *Petit Roeulx*, *Steenkerque*, *Froide Fontaine*, together about 1200 m in *circa* 6 m.y.) and lasted until the top of the Gorstian (*Ronquières Formation*, 540 to 600 m thick). These turbidites were distal at first, and more proximal from the end of Sheinwoodian onwards.

As a whole, the mean thickness of Megasequence 3 is estimated to be about 3400 m in the outcropping area (Verniers *et al.*, 2001). This significant thickness, reflecting an acceleration of the subsidence from the mid Telychian (upper Llandovery), has been interpreted as heralding the onset of a Silurian foreland basin (Van Grootel *et al*., 1997; Debacker, 2001; Verniers *et al.*, 2002). The upper part of Megasequence 3 is only known in boreholes (Verniers & Van Grootel, 1991) and ended most probably in the lowermost Lochkovian (Lower Devonian) at *circa* 417 Ma (its presence was proved in the Condroz Inlier). No younger rocks are known in the Brabant Massif and the Condroz Inlier until the deposition of the Givetian conglomerates (*Bois de Bordeaux Formation*) belonging to the Brabant Parautochthon sedimentary cover (Figs. 3, 5). The conglomerate was deposited after the "Brabantian orogeny" (tectonic inversion of the Brabant Massif; see below) responsible for the Lower-partly Middle Devonian hiatus.

2.2.4. Subsidence analysis

New estimates of the sedimentary thicknesses, especially in megasequences 1 and 2 (NCS, 2011; Herbosch et al., 2008a, 2014, Herbosch & Verniers, 2013) allow us to construct a new cumulative thickness curve (Fig. 7). This curve reveals the major changes in the rate of sedimentation. Megasequence 1 shows a very high sedimentation rate, especially from Blanmont to Jodoigne Formations (about 8 km deposited in 27 m.y.), only compatible with a rift environment. The megasequence 1 curve is concave-up, with the Mousty and Chevlipont formations showing a slower sedimentation rate, announcing the Avalonia-Gondwana drifting marked by the Lower-partly Middle Ordovician hiatus.

After this hiatus, megasequence 2 began with a low sedimentation rate (from Abbaye de Villers to Rigenée formations). This rate progressively increased from Ittre to Fauquez formations, which correspond to deep marine turbiditic sedimentation linked to incipient tectonic instability accompanied by volcanism.

Megasequence 3 marked a new slowdown of the sedimentation rate in its lower part (Madot and Brutia formations), a progressive increase from Bois Grand-Père to Fallais formations and finally a very rapid rate of sedimentation from Corroy to Ronquières formations (about 1800 m in *circa* 7 m.y.). The latter corresponds to the development of the rim foreland basins, contemporaneous with the Brabantian tectonic inversion.

2.3. Palaeontological investigations and their palaeogeographic implications

Palaeontological investigations in the BM have been essentially carried out on graptolites (Lecompte, 1948; Legrand, 1964, 1965; Maletz & Servais, 1996, 1998) and on two groups of microfossils, namely acritarchs (Martin, 1969a, b, 1976; Servais 1993; Vanguestaine, 1991, 1992, 2008; Vanguestaine & Wauthoz, 2011) and chitinozoans (Verniers, 1976, Verniers & Van Grootel, 1991; Verniers et al., 1999, 2002, 2005; Vecoli et al., 1999; Vanmeirhaeghe et al., 2005; Vanmeirhaeghe, 2006). Acritarchs are useful for the Cambrian and Ordovician while chitinozoans are helpful for the Ordovician and Silurian. These different fossil groups of fossils are complementary and permit the establishment of a complete and very accurate chronostratigraphical assignement of lithostratigraphic units, especially in the Ordovician, as it is shown in Figs. 8 and 28.

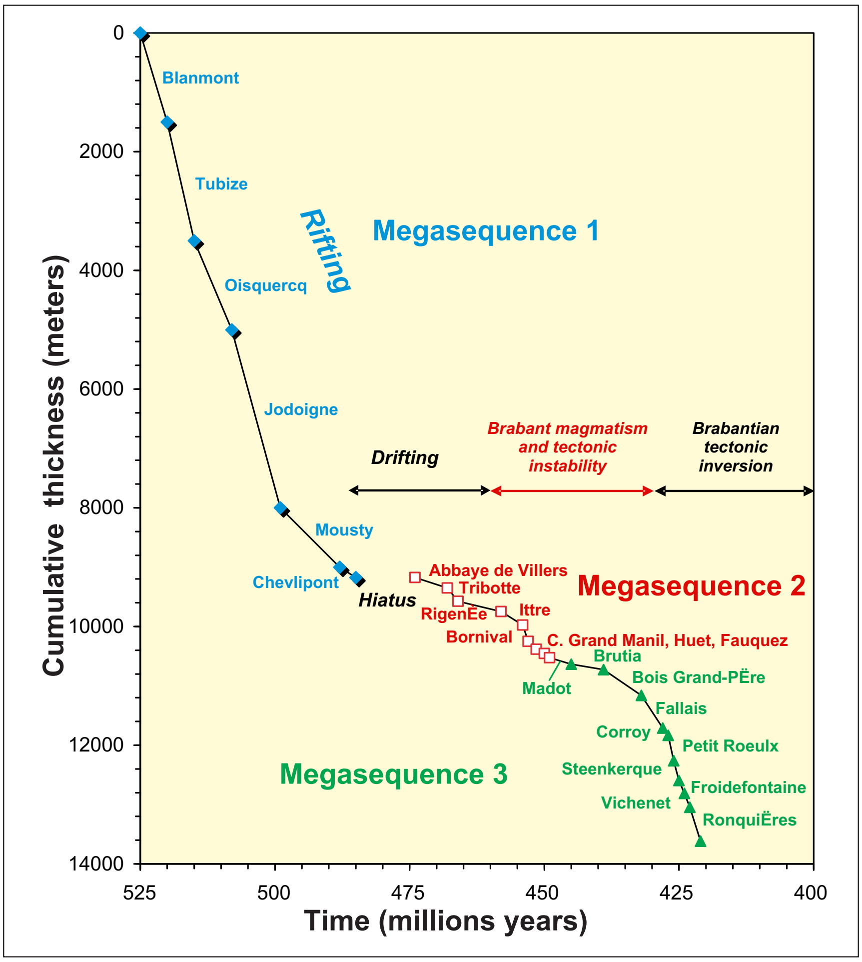


Figure 7. Cumulative thickness curve of the Lower Palaeozoic sediments of the BM plotted against the stratigraphic age (thickness from Herbosch & Verniers, 2013, 2014 and for the Silurian from Verniers et al., 2001).

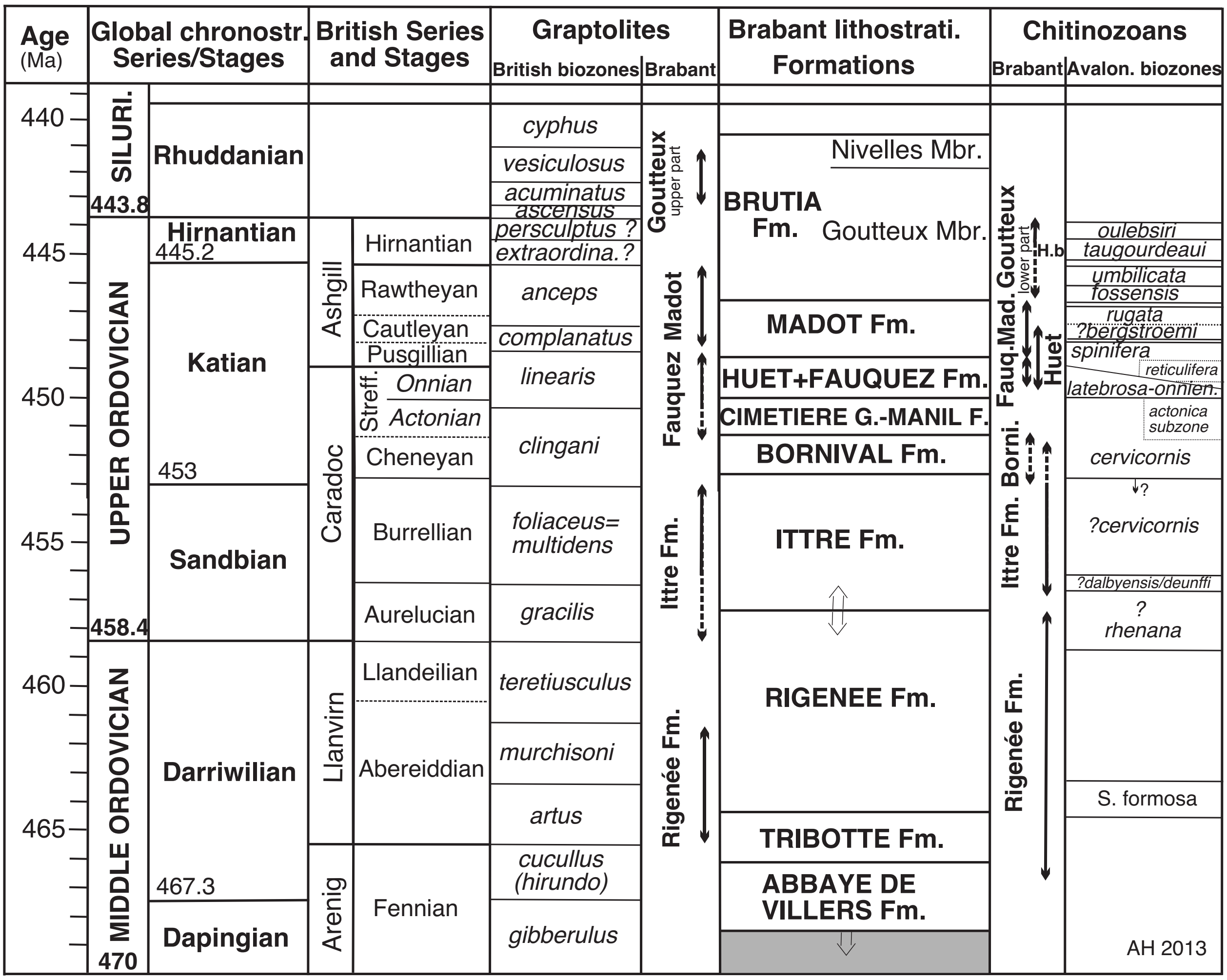


Figure 8. Comparison of the chronostratigraphic position obtained with graptolites and chitinozoan occurrences from the Middle Ordovician to the lowermost Silurian formations of the BM. Chronostratigraphy of the global Series/Stages, of the local British Series and Stages and of the British graptolite biozones after Cooper & Sadler (2012). Composite Avalonian chitinozoan biozonation after Vandenbroucke (2008), except S. Formosa Zone after Cooper & Sadler (2012). H.b: Harelbeke borehole. After Herbosch & Verniers (2014).

The palaeogeographic reconstruction of Cocks and Torsvik (op. cit., Fig. 6) is confirmed by palaeontological investigations done in the BM and its southern prolongation, the Condroz Inlier (Brabant-Condroz Basin). In the latter, assemblages of trilobites show typical northern Gondwana (Avalonian) faunas up to the Sandbian (Upper Ordovician), whereas in the Katian these faunal assemblages witness the increasing proximity of Baltica (Verniers et al., 2002; Owen & Servais, 2007). Chitinozoan assemblages seem even more sensitive: until the mid-Darriwilian, they have N-Gondwana affinities whereas a Baltoscandian signature exists from the lower Sandbian until the end of the Ordovician (Samuelsson & Verniers, 2000; Vanmeirhaeghe et al., 2005; Vanmeirhaeghe, 2006).

2.4. Tectonics and the "Brabantian orogeny"

Since Fourmarier (1920), deformation within the Brabant Massif, or the "Brabantian orogeny", is unanimously considered to have taken place between the end of Gorstian (*circa* 426 Ma) and the Givetian (*circa* 389 Ma), leaving a time hiatus of about 36 Ma. These ages correspond respectively to the youngest deformed basement rocks in outcrop (*Ronquières Fm.*) and the oldest cover rocks above the major Brabantian unconformity (*Bois de Bordeaux Fm.*), at the base of the Brabant Parautochthon. This hiatus was further narrowed by the discovery of Pridoli to lowermost Lochkovian (*circa* 417 Ma) deposits under the unconformity in the W Vlaanderen borehole (Verniers & Van Grootel, 1991; Van Grootel et al., 1997). However, the nature, timing and geodynamic significance of the "Brabantian orogeny" have only been established recently (Sintubin *et al.*, 2009 and references therein). By contrast to the old view of an anticlinal culmination, the architecture of the Brabant Massif is currently interpreted as a NW-SE trending compressional wedge consisting of a central steep belt composed of predominantly Cambrian metasediments, symetrically bordered by less deformed Ordovician and Silurian metasediments (Figs. 2A, 9) (Sintubin & Everaerts, 2002; Verniers *et al.*, 2002; Debacker *et al.*, 2005).

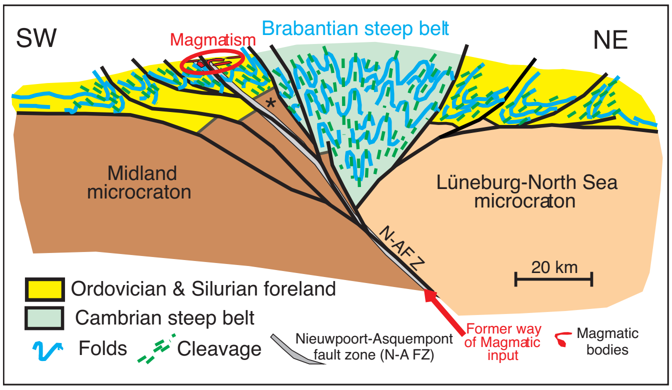


Figure 9. Schematic SW-NE cross-section of the Brabantian belt along the Dendre river (see Fig. 2A). Modified after Sintubin & Everaerts (2002) and Sintubin et al. (2009).

The sudden increase of subsidence and the initiation of deep marine turbiditic sedimentation in the upper Telychian (Fig. 7) mark the onset of a foreland-basin development in the Brabant rim domain. This suggests that a tectonic inversion began around 433 Ma in the Cambrian core of the BM. This inversion is called "Brabantian orogeny" (Debacker *et al.*, 2005) by reference to the pioneering work of Michot (1980). The Lower-Middle Devonian hiatus in the Brabant Parautochthon and the Silurian to lower Lochkovian reworked acritarchs (Steemans, 1989) found in the Lower Devonian just to the south in the Dinant basin (Fig. 3; Ardenne Allochthon) suggest that the central part of the Brabant Massif emerged since the Lochkovian. From the middle Silurian to early Devonian, the progressive deformation results in a continued uplift of the core and a gradual spreading of the deformation toward the rim of the massif. This Brabantian inversion cannot have lasted after *circa* 388 Ma, the age of the oldest undeformed rocks in the Brabant Parautochthon, the Bois de Bordeaux Fm. Considering that the Burnot conglomerates (late Emsian to early Eifelian, 395-392 Ma) in the northern rim of the Dinant basin resulted from the last increment of deformation of the Brabantian orogeny and using $^{40}Ar/^{39}Ar$ ages (426 to 393 Ma) on syn- to post-cleavage metamorphic muscovite/sericite grains, Debacker et al. (2005) placed the end of the Brabantian orogeny at the Emsien-Eifelien boundary (*circa* 393 Ma), thus very near the end of the Lower-Middle Devonian hiatus (Fig. 15).

Detailed field studies have shown that throughout the outcrop there is evidence for one single progressive deformation event only. The main features associated with this deformation are folds cogenetic with a well-developed axial plane cleavage: in the core, steeply plunging folds are associated with subvertical to steeply dipping cleavage while towards the south-eastern side, subhorizontal to gently plunging folds are associated with steep to moderately north dipping cleavage (Fig. 9) (Sintubin, 1999; Debacker, 2001, 2002; Verniers et al., 2002; Debacker et al., 2003, 2004a, 2005; Debacker, 2012). The uplift of the central Brabant core could also explain the development of pre-cleavage and pre-folding low angle detachments at the edges of the steep core like the "Asquempont Detachment System" described by Debacker (2001) and systematically observed during the geological mapping of the outcropping area (Fig. 2A) (Debacker et al., 2003, 2004a, 2004b, 2011; Debacker & Sintubin, 2008; Herbosch & Blockmans, 2012; Herbosch & Verniers, 2013, 2014).

There is a main structure running NW-SE along the southwestern flank of the Brabant Massif: the Nieuwpoort-Asquempont fault zone (Figs. 2A, 9), marked by a pronounced aeromagnetic and gravimetric lineament (cover figures, Debacker et al., 2004b and references therein). Although initially considered to be a strike-slip fault zone (André & Deutsch, 1985), recent observations indicate that this fault zone corresponds to subvertical movements along normal faults (Debacker et al., 2003; 2004b; 2011). These normal faults have been repeatedly reactivated, especially at *circa* 375 Ma (Frasnian; André and Deutsch, 1985), but also in historical times (1938 Grammont earthquake; Legrand, 1968). The rapidly decreasing intensity of deformation to the SW of the N-AFZ and the inferred presence at depth of the Midlands microcraton to the SW and the Lüneburg-North Sea microcraton to the NE (Fig. 9), led Sintubin & Everaerts (2002) to propose that the Brabantian orogeny was the result of the convergence of these two microcratons. This convergence would be the result of the anti-clockwise rotation of the former (Piper, 1997), during an intracontinental process within the Avalonia microcontinent (Sintubin & Everaerts, 2002; Sintubin et al., 2009). Another interpretation (Linnemann et al., 2012) could be a far-field effect of the collision Baltica-Avalonia with Laurentia which could be the direct cause of the invoked rotation.

2.5. Magmatism in the Brabant Massif

The main igneous activity in the BM extends from the late Ordovician up to the lower Silurian, with a peak in the late Katian during the deposition of the Madot Formation (Fig. 5). The magmatism is located in an arcuate belt along the SW margin of the BM (Figs. 2A, 9) (André et al., 1986, 1991; Verniers et al., 2002). The activity is mainly represented by interbedded pyroclastic rocks of dacitic and rhyolitic composition, but several hypabyssal microdioritic bodies intrude the Ordovician sediments of the SW margin of the Cambrian core (Figs. 2A, 9). Best documented are the Quenast plug and the Bierghes and Lessines sills which are extensively quarried. This activity has been interpreted as related to a continental active margin (André et al., 1986) or to a volcanic arc (Van Grootel et al., 1997) and the presence of an important negative gravimetric anomaly (cover illustration) suggests the presence of a granitic batholith underlying the magmatic arc (Everaerts et al., 1996;

Mansy et al., 1999). However this low-density body is also interpreted as an elevated cratonic basement block in the framework of a compresional wedge model (Fig. 9; Sintubin & Everaerts, 2002).

Recent geochemical investigations (Linnemann et al., 2012) show that the magmatic rocks have, despite their strong alteration, a high-K calc-alkaline to alkali-calcic character (shoshonitic, Fig. 10A & B), a negative restricted range of initial epsilon Nd (-3.7 to -4.9, Fig. 10) and a model age between 1.32 and 1.73 Ga wich are indicative of a predominantly crustal origin and of an intracontinental post-collisional period. Dating of the zircons by LA-ICP-MS (Linnemann et al., 2012; new unpublished data; Fig. 15) shows also that the magmatic rocks have been intruded during the 460-430 Ma time span, most of them between 450 and 440 Ma (Figs. 5, 11, 12, 15).

The previous geodynamic interpretation of these intrusions (André, 1983; Pharaoh et al., 1993) in the framework of a SSW-dipping subduction zone beneath Avalonia preceeding the closure of the Tornquist Ocean (Fig. 6) is no longer tenable. The subduction zone is too far from the BM (more than about 100 km), and the chemical composition and crustal origin of these rocks are uncommon in subduction setting but frequent in post-collisional periods. It is the same for the more recent hypothesis of Verniers et al. (2002) wich is also based on a subduction hypothesis. In consequence, Linnemann et al. (2012) propose that the Avalonia-Baltica docking induced the reactivation of the lithospheric discontinuity between the Midlands and Lüneburg-North-Sea microcratons (Fig. 9), inducing movement of extension or transtension along a proto Nieuwpoort-Asquempont Fault Zone, generating the partial melting of the lithosphere. These events give rise to the BM magmatism between 460 and 430 Ma, a period characterized by tectonic instability (Fig. 15). At *circa* 430 Ma the magmatism ceased in response to the beginning of the Brabant core inversion that corresponds to the beginning of the Baltica/Avalonia with Laurentia collision.

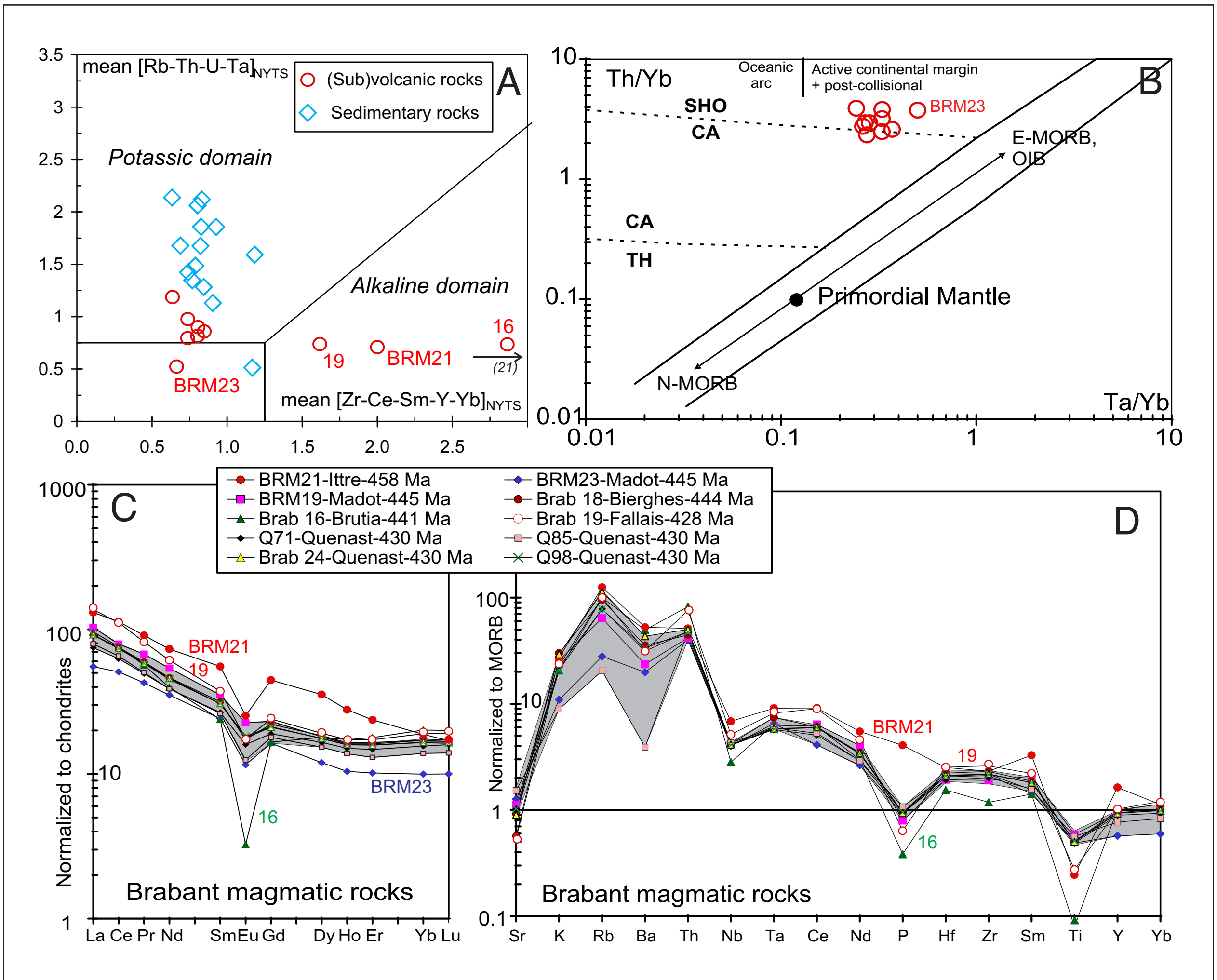


Figure 10. Geochemistry of Brabant magmatic rocks. A: sliding normalized values for the Yenchichi-Telabit reference series (Liègeois et al., 1998). B: Th/Yb vs. Ta/Yb diagram (Pearce, 1982). C: spectrum of Rare Earth Elements normalized to chondrites (Taylor & Mc Lennan, 1985). D: trace elements normalized to MORB (Sun, 1980; Pearce, 1982). Gray areas enhance the Brabant samples without outliers. From Linnemann et al. (2012).

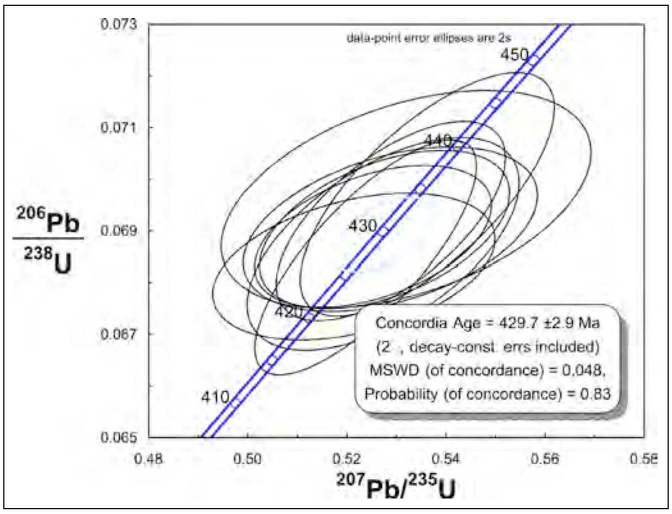


Figure 11. U-Pb age of the zircons of the microdiorite Quenast plug (from Linnemann et al., 2012).

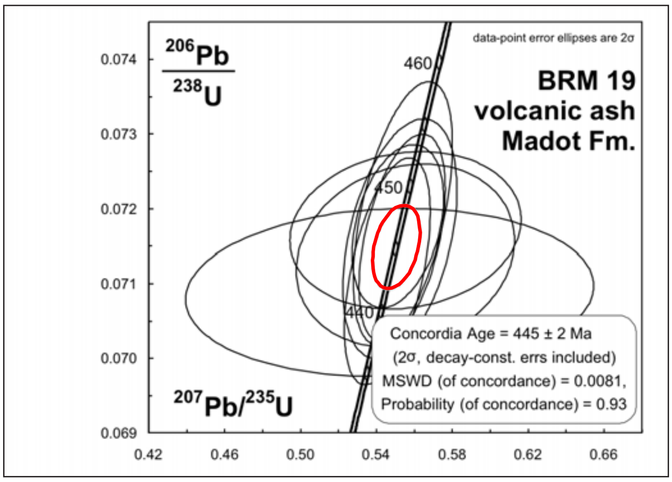


Figure 12. U-Pb age of the zircons of the tuffite of the Madot Formation (ibid.).

2.6. Contributions of detrital zircon ages and Nd isotopes

Detrital zircons from 7 sandstones (covering the whole BM succession) and from one Givetian conglomerate of the Brabantian unconformity were dated with LA-ICP-MS (Linnemann et al., 2012). Fig. 13 gives an example of the graphic result obtained for the Tubize Fm. (88 concordant ages over 120 spots). These diagrams show that about 53% of the zircons have a Neoproterozoic age and about 40% a Palaeoproterozoic age.

The variations of these age spectra through time are very interesting because they reveal the changes in the sources of the zircons, and thus the changes in palaeogeography. 17 argillaceous samples were also analysed for their eNd that give information about the nature of the continental crust that was eroded. The model age NdTDM, calculated from the eNd, corresponds to the mean ages of the initial sources of the sediments.

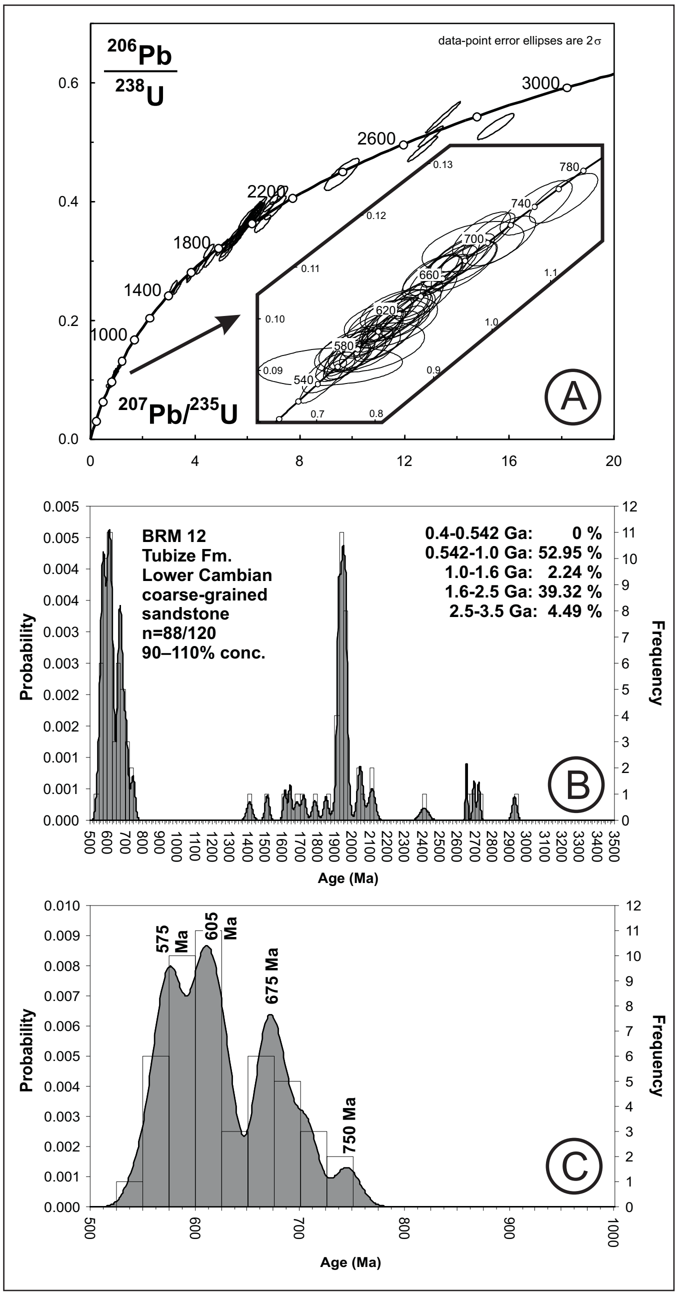


Figure 13. U-Pb ages of detrital zircon grains from greywackes of the Tubize Fm. Concordia diagram (A) and combined binned frequency and probability density distribution plots of ages in the range 500-3.500 Ma (B) and of 500 to 1000 (C). From Linnemann et al. (2012).

The main results of these investigations are synthesized in Fig. 14: (1) from the lower Cambrian until the mid-Silurian, Neoproterozoic rocks were the most important zircon sources (35-75%). The Pan-African domain, from which the debris are most probably derived, could have been the main Gondwana (W Africa) basement, the Avalonia basement itself, or both; (2) the second dominant source is represented by a continental basement composed of Palaeoproterozoic rocks. Such units are documented in all major potential source domains: Amazonia, Baltica, W Africa. (3) A Mesoproterozoic contribution exists in some Cambrian and Silurian sedimentary rocks. Such basement occurs in Baltica and Amazonia. Interestingly, the Mesoproterozoic zircon supply vanished during the Ordovician rift-drift stage and reappeared during the Silurian foreland basin stage. (4) Ordovician and Silurian igneous rocks contributed to the Palaeozoic debris and should thus have been exposed during the time of sedimentation.

The eNd at the sedimentation age of the rocks vary from -5.5 to -9.3 (Fig. 14) and the model ages vary from 1.44 to 1.70 Ga, which indicates globally the presence of an old component in the BM sedimentary rocks. It could be either from a Palaeoproterozoic basement or from the mixing of a Neoproterozoic and Palaeoproterozoic basements, the two possibilities not being exclusive.

These new approaches give additional constraints to decipher the wandering history of Avalonia that are complementary to palaeomagnetism and palaeontology with a much better time resolution thanks to the precise stratigraphic control available for the BM sedimentary record.

2.7. New palaeogeographic reconstruction of Avalonia for the Cambrian

Important and numerous stratigraphical, sedimentological and geochemical similitudes between the BM and the Harlech Dome in Wales ("eastern" Avalonia) on the one hand, and in the Meguma terrane in Nova Scotia (Canada, peri-Gondwana) on the other hand (Waldron et al., 2009, 2011; Fig. 16), allow the reconstruction of a new lower Cambrian palaeogeographic map of Avalonia (Fig. 17; Linnemann et al., 2012). The main arguments are the huge thickness of the Cambrian successions (especially in Meguma >11 km and in BM >9 km), the similarity in the stratigraphical, sedimentological and geochemical successions. This amazing similarity strongly suggests that these terranes were deposited in the same deep rift basin since the lower Cambrian. More arguments too long to be developed in this guide can be found in the latter cited paper.

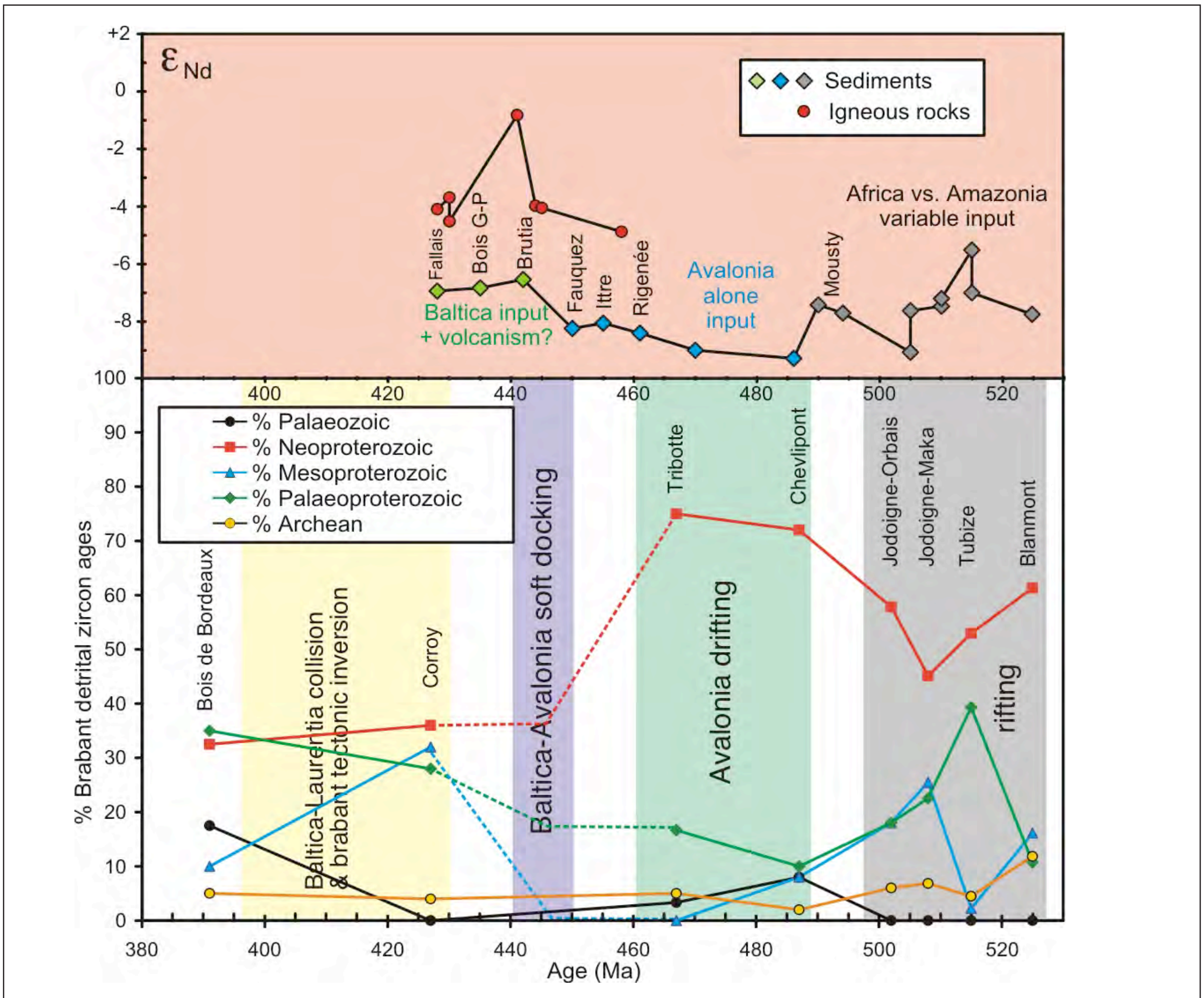


Figure 14. Below: variation through time of the proportion of BM detrital zircon ages grouped in five classes. Above: ƐNd variation through time of the signatures of the BM sediments and magmatic rocks and the deduced signature of the main concerned terranes (from Linnemann et al., 2012).

In the new palaeogeographic reconstruction BM and Harlech Dome constitute embayments of the main rift within the European part of Avalonia ("eastern Avalonia") along weakness zones formed by the eastern and western boundaries of the Midlands microcraton. Meguma is in the western continuation of the main rift, currently located in the American part of Avalonia ("western Avalonia") (Fig. 17). In consequence, **the Meguma terrane**, which is up to now of uncertain palaeogeographic provenance**, should belong to the Avalonia microplate** that rifted from Gondwana later in the Tremadocian.

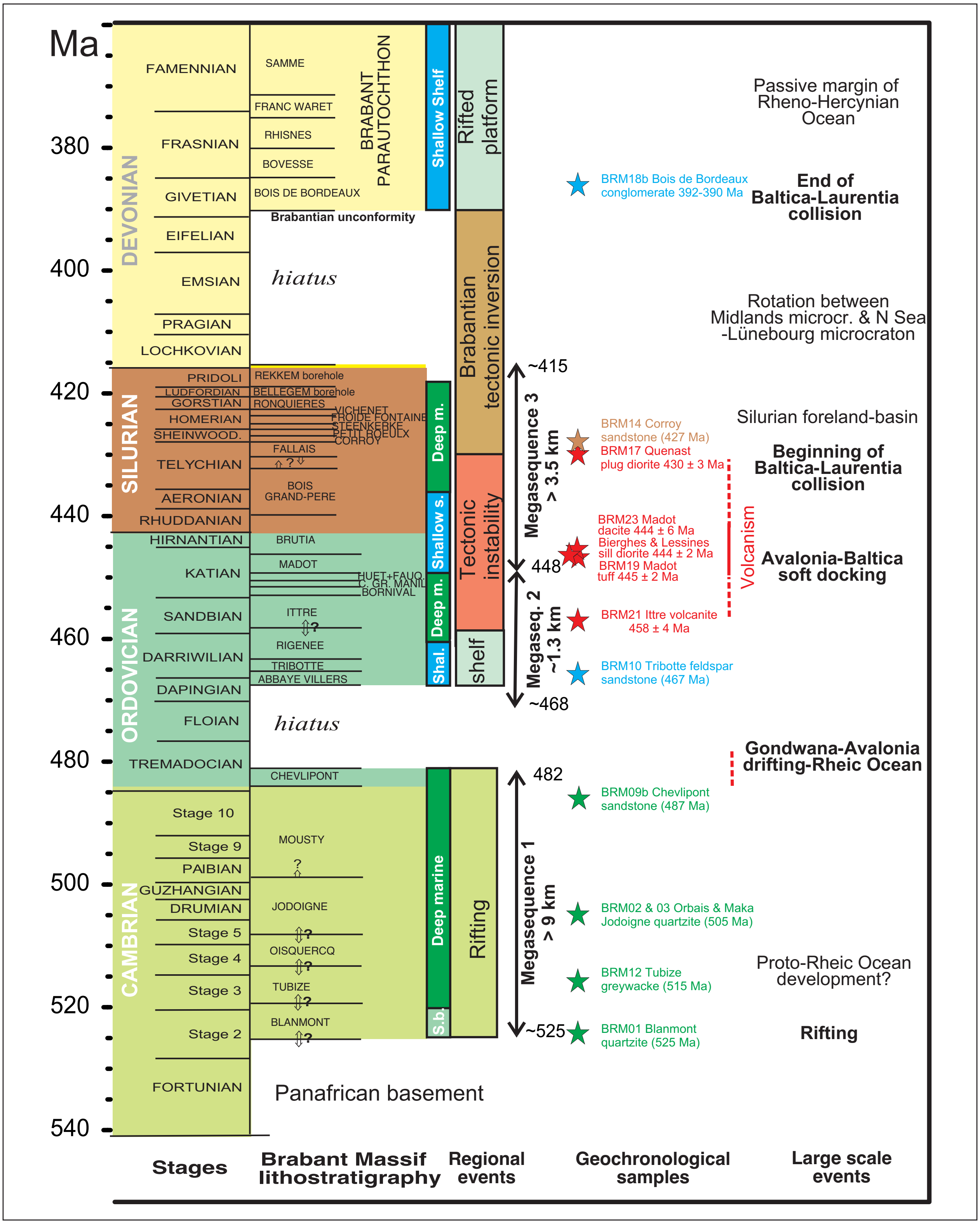


Figure 15. Synthesis of the Cambrian to Devonian odyssey of the Brabant Massif from Gondwana to Laurentia within the wandering Avalonia microcontinent history (modified after Linnemann et al., 2012).

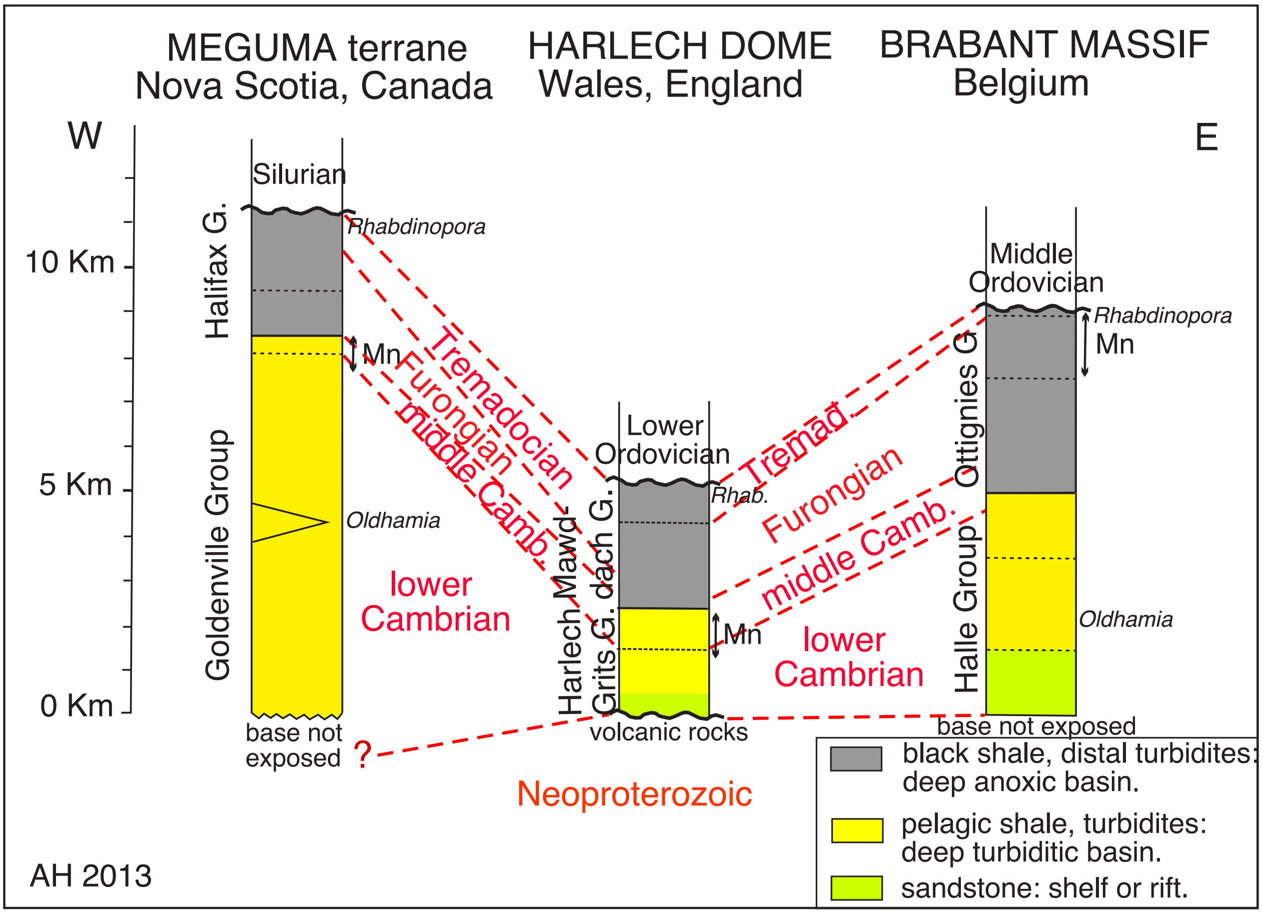


Figure 16. Comparison of the sedimentary registration between the BM, the Harlech Dome (Wales) that belong to Avalonia, and Meguma (Nova Scotia) that, up to now, is assumed to belong to uncertain N-Gondwana terrane. Herbosch (unpublished) based on Waldron et al. (2011) for Harlech and Meguma stratigraphy and sedimentology.

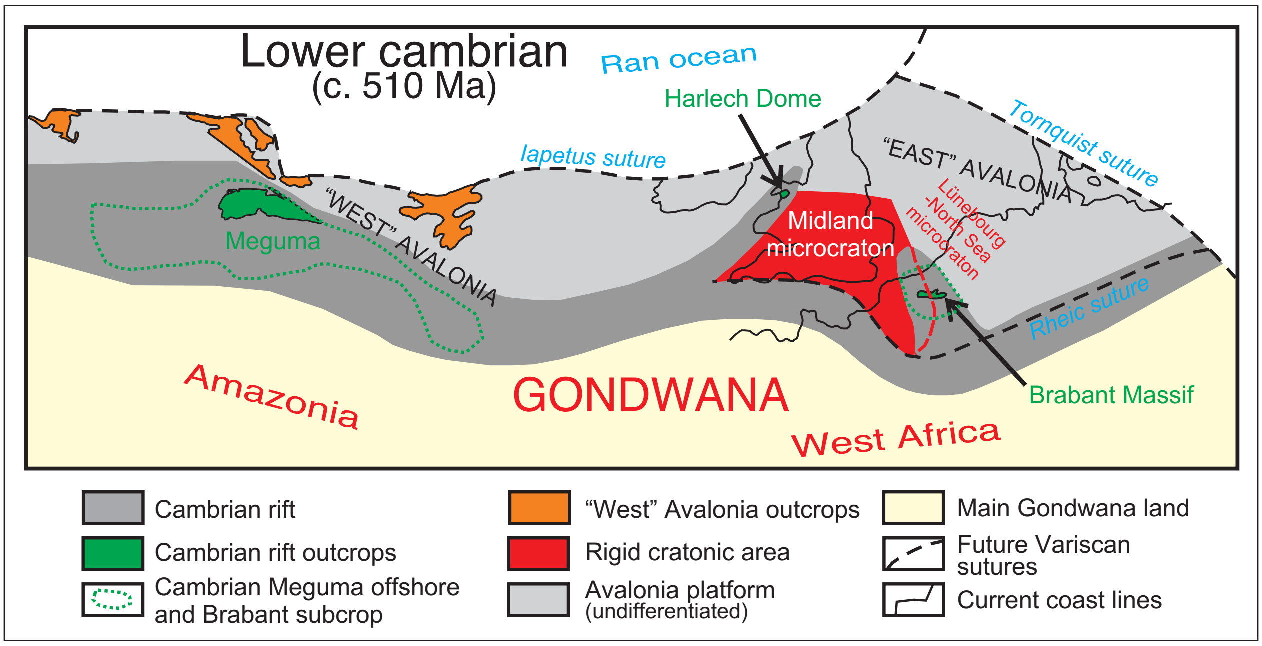


Figure 17. Upper lower Cambrian palaeogeographic reconstruction of the large rift separating Gondwana from Avalonia that had led to the formation of the Rheic Ocean. General palaeogeography after Cocks & Torsvik (2005); contour of the SE prolongation under the BM of the Midlands microcraton after Sintubin et al. (2009); offshore extension of the Meguma terrane from Pe-Piper & Jansa (1999) and White (2010). After Linnemann et al. (2012).

3. Description of the Dyle Basin stops (first day)

STOP 1: Blanmont Fm. in the "Al Vau" old quarry, along the Orne river.

Location: New geological map 1/25,000 Chastre-Gembloux (Delcambre et al., 2002).

5°37'47.56" N 4°38'01.77" E

Stratigraphy: Blanmont Formation, position in the formation unknown. Old occurrences of the ichnofossil *Oldhamia radiata* in slate intercalations allow provisional allocation of the formation to the upper part of the Cambrian Stage 2 (uppermost Terreneuvian) (Fig. 5). The thickness is estimated to be more than 1.5 km in the Grande Gette valley (Jodoigne area) where two large flooded quarries occur (Debacker & Herbosch, 2011).

Observations: Massive whitish quartzite and greenish argillaceous sandstone are visible in meter-size beds that are difficult to distinguish. The absence of slate intercalations should be noted. Nevertheless slate intercalations were described in the region when the quarries were in full activity at the end of the 19th century. The stratification is subvertical.

Interpretation: The Blanmont Fm. is today considered to be the oldest unit in the outcrop or subcrop area of the BM (Herbosch et al., 2008a). It is poorly known even if it is observed in numerous small outcrops appearing in the river valleys, in old flooded quarries or in palaeoreliefs from the central part of the BM (Fig. 19). The eastern outcrops from Offus to Dongelberg (Fig. 19) belong to the northern flank of the Brabant anticline and Offus seems to mark the periclinal end of the Cambrian core.

The huge thickness of sandstone with very scarce slate intercalation, the presence of sedimentary structures characteristic of a rapid sedimentation in a shallow environment (cross laminations, escape structures, pseudo nodules) and old descriptions of conglomeratic intercalations testify to a shallow but rapidly subsiding rift environment.

Subvertical stratification is usual in the Cambrian core (see discussion at Stop 4).

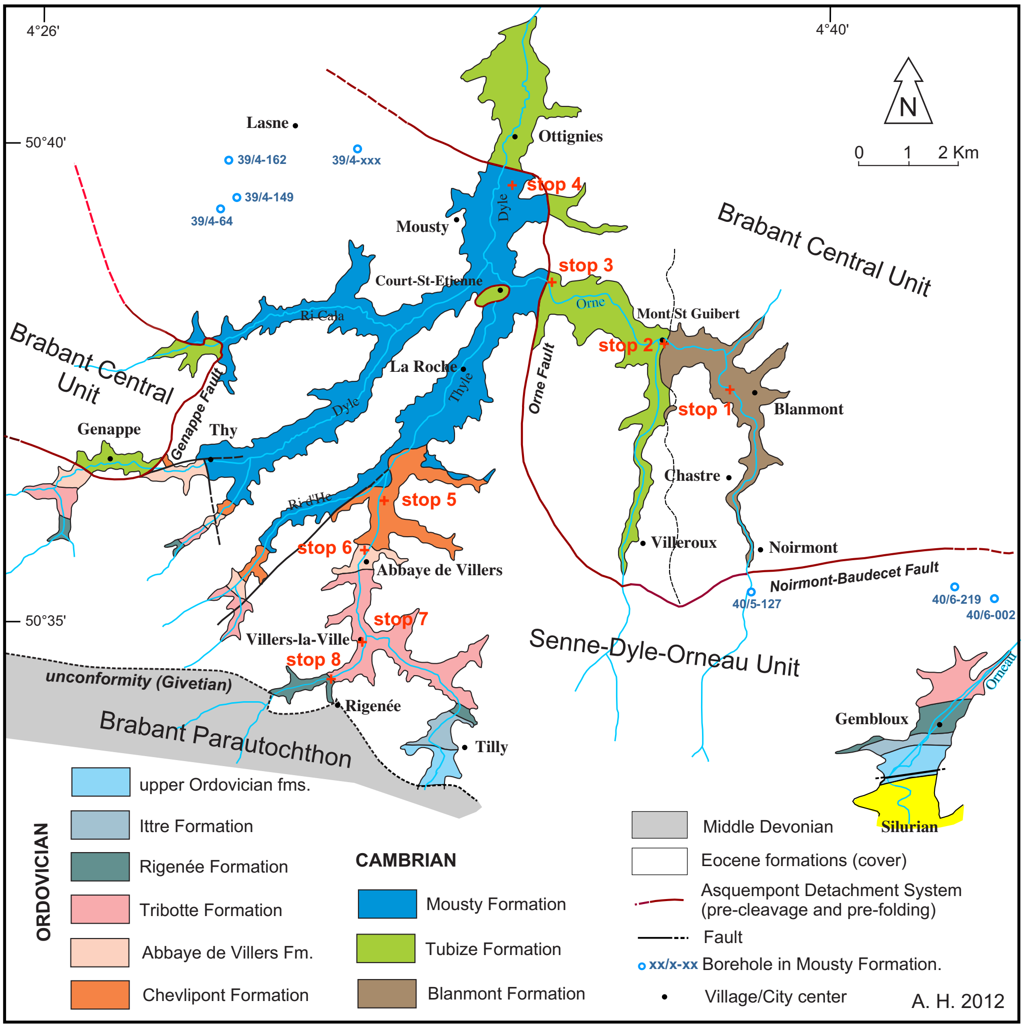


Figure 18. Geological map of the Lower Palaeozoic of the Dyle and Orneau (pro parte) basins (see location in Fig. 2A) (modified after Herbosch & Lemonne, 2000; Delcambre et al., 2002; Debacker et al., 2004a; Herbosch & Blockmans, 2013). Brabant Central Unit and Senne-Dyle-Orneau Unit are two tectonic sub-units separated by the Asquempont Detachment System.

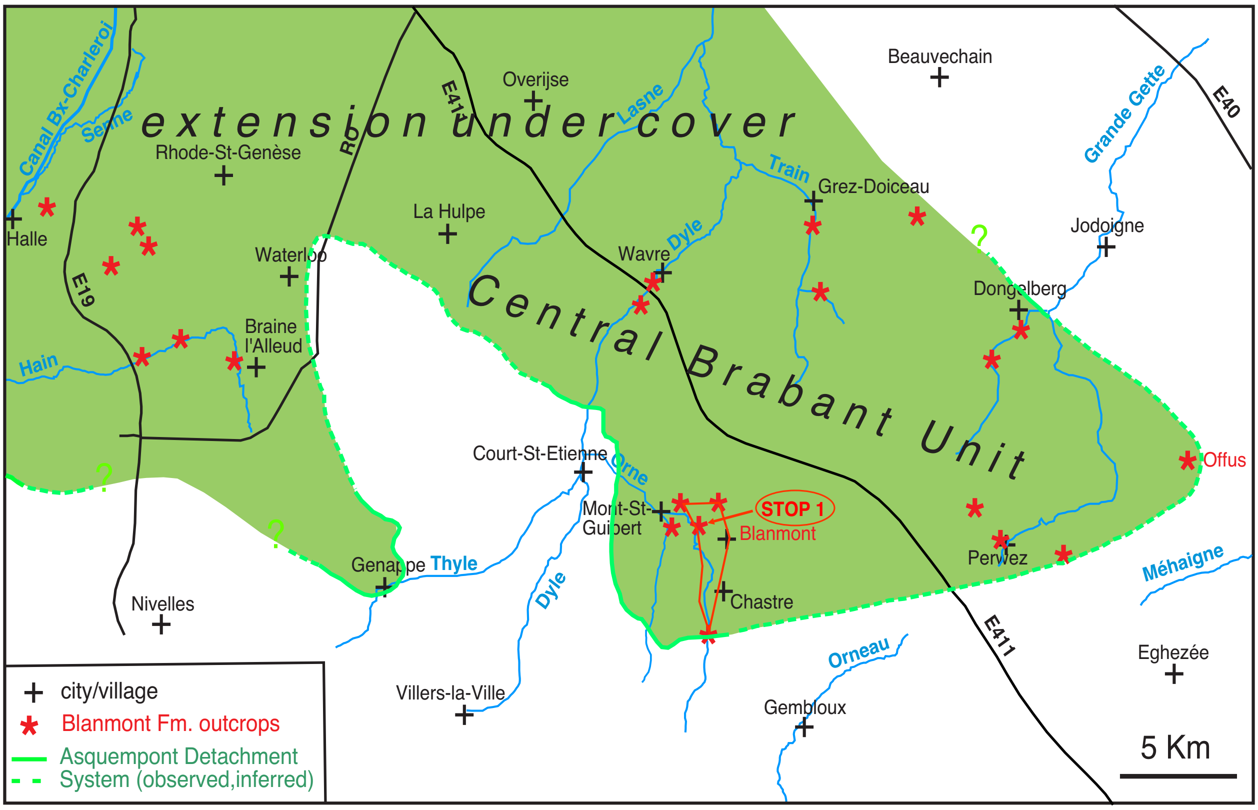


Figure 19. Geographic situation of all the outcrops of the Blanmont Fm. (red stars). In green the trace of the Asquempont Detachment System that corresponds to a low-angle unconformity forming the limit between the Cambrian core (Central Brabant Unit in light green) and the Ordovician-Silurian rim (in white) of the BM. The extension of the Central Brabant Unit to the north is from borehole evidence.

STOP 2: Tubize Fm. at Mont-St-Guibert, under the church

Location: New geological map 1/25,000 Chastre-Gembloux (Delcambre et al., 2002)

50°38'00.19" N 4°36'38.45" E

Stratigraphy: Mont-St-Guibert Mbr. (lower) of the Tubize Fm. Stage 3 (Fig. 5; lower Cambrian) based on the ichnofossil *Oldhamia* observed in all the three members. Very huge thickness estimated at >2.5 km with about 800 m for the middle Rogissart Mbr. only.

Observations: rhythmic alternations of dm-size beds of sandstone, greenish siltstone and greenish slate. Very nice loadcasts can be observed at the base of the beds and flute marks have also been observed in a temporary outcrop not far from this outcrop. Stratification and cleavage are sub-vertical which means that the folds are of steeply plunging type B (classification of Debacker, 2001; Figs. 25, 26). See discussion in the Stop 4.

Interpretation: sequences of high-density turbidites of the Bouma model (Fig. 20). A quite complete Bouma sequence Tabcde can be observed in the center of the outcrop. We will observe a comparable facies at Stop 9 in the middle member of the Tubize Fm. (day 2).

STOP 3: Tubize Fm. at Beaurieux near the highway bridge, Orne valley.

Location: New geological map 1/25,000 Chastre-Gembloux (Delcambre et al., 2002)

50°38'28.17" N 4°35'11.47" E

Stratigraphy: Unknown member of the Tubize Fm. Age and thickness same as Stop 2.

Observations: Massive green slate (siltstone) with numerous tiny magnetite crystals. The stratification is difficult to observe but it shows a NNE-SSW direction with a 70-80° dip that corresponds to the direction of the hill (Fig. 21). These hard slates are nearly perpendicular to the Orne river that shows a deviation of its trajectory and a narrow valley ("cluse").

Interpretation: The slates are interpreted as hemipelagic to pelagic sediments. Magnetite is frequently observed in the 3 members of the Tubize Formation mainly in the siltstone facies. Magnetite was formed by late diagenesis and/or metamorphism which reaches the upper anchizone to greenschist levels in most of the BM. Ground magnetic survey shows that the magnetite is more abundant in some layers, forming magnetic ridges (Fig. 21).

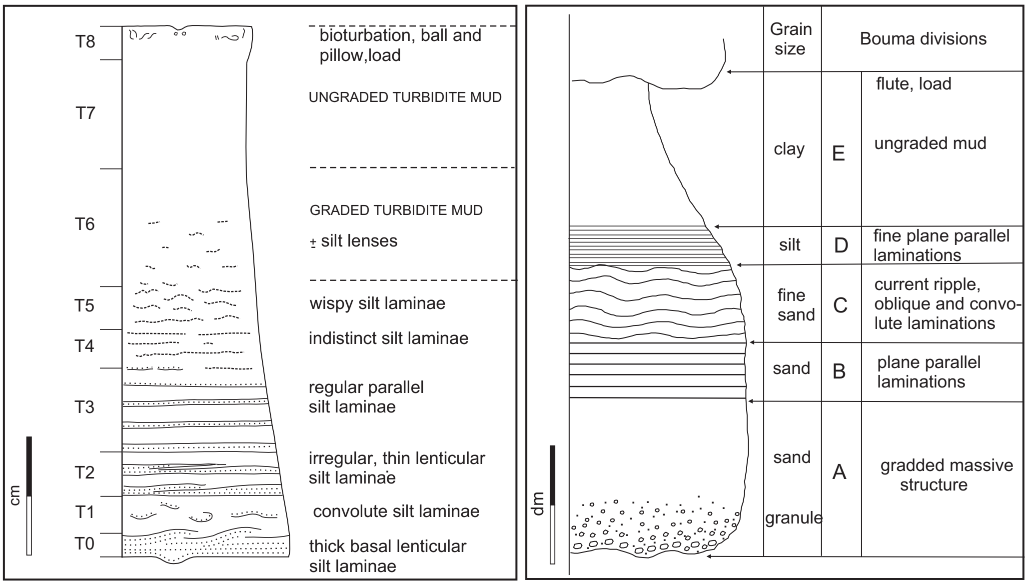
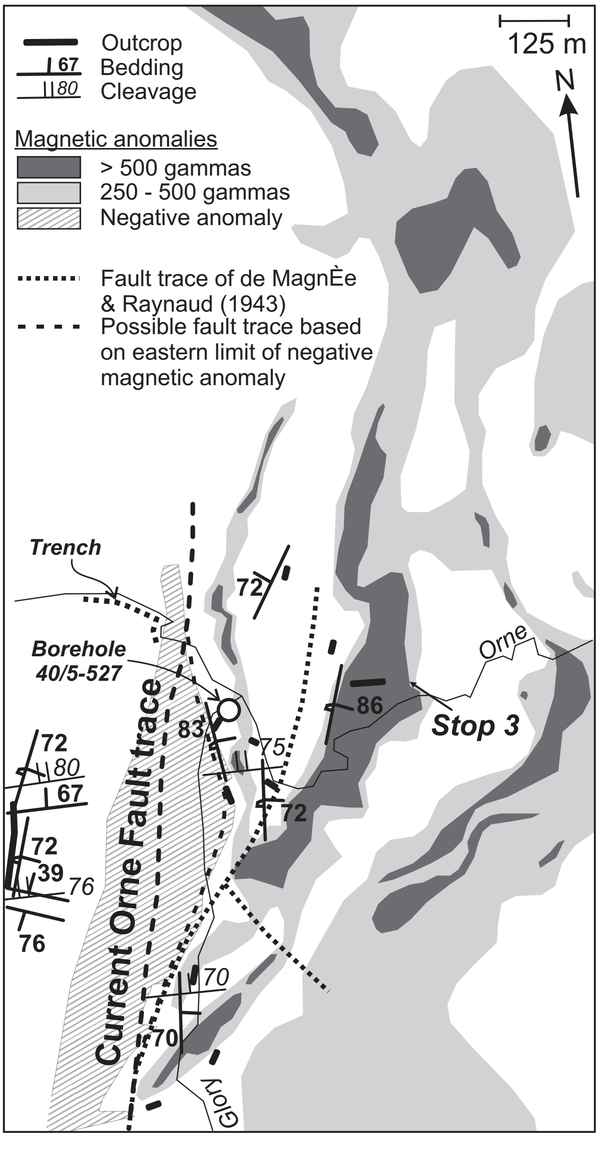


Figure 20. High-density turbidite model (Bouma model, at right) compared with low-density turbidite model (Stow model, at left): note the difference in scale between the two models.



Discussion: **Asquempont Detachment System (ADS)**. Stop 3 is the place where de Magnée & Raynaud (1943) demonstrated, by a ground magnetic survey, that the contact between the Tubize Fm. and the Mousty Fm. was faulted and that the "Orne Fault" should be a thrust fault (Figs. 18, 21). At the beginning of the new mapping of the BM (Herbosch & Lemonne, 2000; Delcambre et al., 2002; Herbosch & Blockmans, 2012) we interpreted this fault and its continuation to the NW and SE (Fig. 18) as a gently N-dipping large displacement thrust. However it is now interpreted as a pre-cleavage and pre-folding low-angle extensional detachment that belongs to the Asquempont Detachment System (see Figs. 18, 19, 2A) (Debacker, 2001, 2012; Debacker et al., 2004b; Piessens et al., 2006; Debacker & Herbosch, 2011). The ADS is now observed in all the outcropping zones and boreholes from the southern border of the BM (Fig. 22) and this new model gives a better explanation of all the field constraints. The early formation of this fault system is responsible for its present-day curved contours on the map (see Figs. 2A, 18, 19, 31) and of the large stratigraphical gaps that are systematically observed between the Cambrian core and the Ordovician rim (Fig. 22). It is also responsible for the astonishing difference in the local stratigraphy between the three major outcrop zones (Senne, Dyle and Gette basins) (compare Fig. 18 with Fig. 31). It was the reason for numerous discussions and misinterpretation since the pioneering work of Fourmarier (1920).

We can observe the Asquempont Fault in the Senne Basin on Stop 10 (Day 2).

Figure 21. Re-interpretation of the Orne Fault position after the magnetic survey of de Magnée & Raynaud (1943) with new data. The Orne Fault trace is constrained only by the truncation of the magnetic ridge to the S and less accurately by the eastern limit of the negative magnetic anomaly that corresponds to the Mousty Fm. After Debacker et al. (2004b).

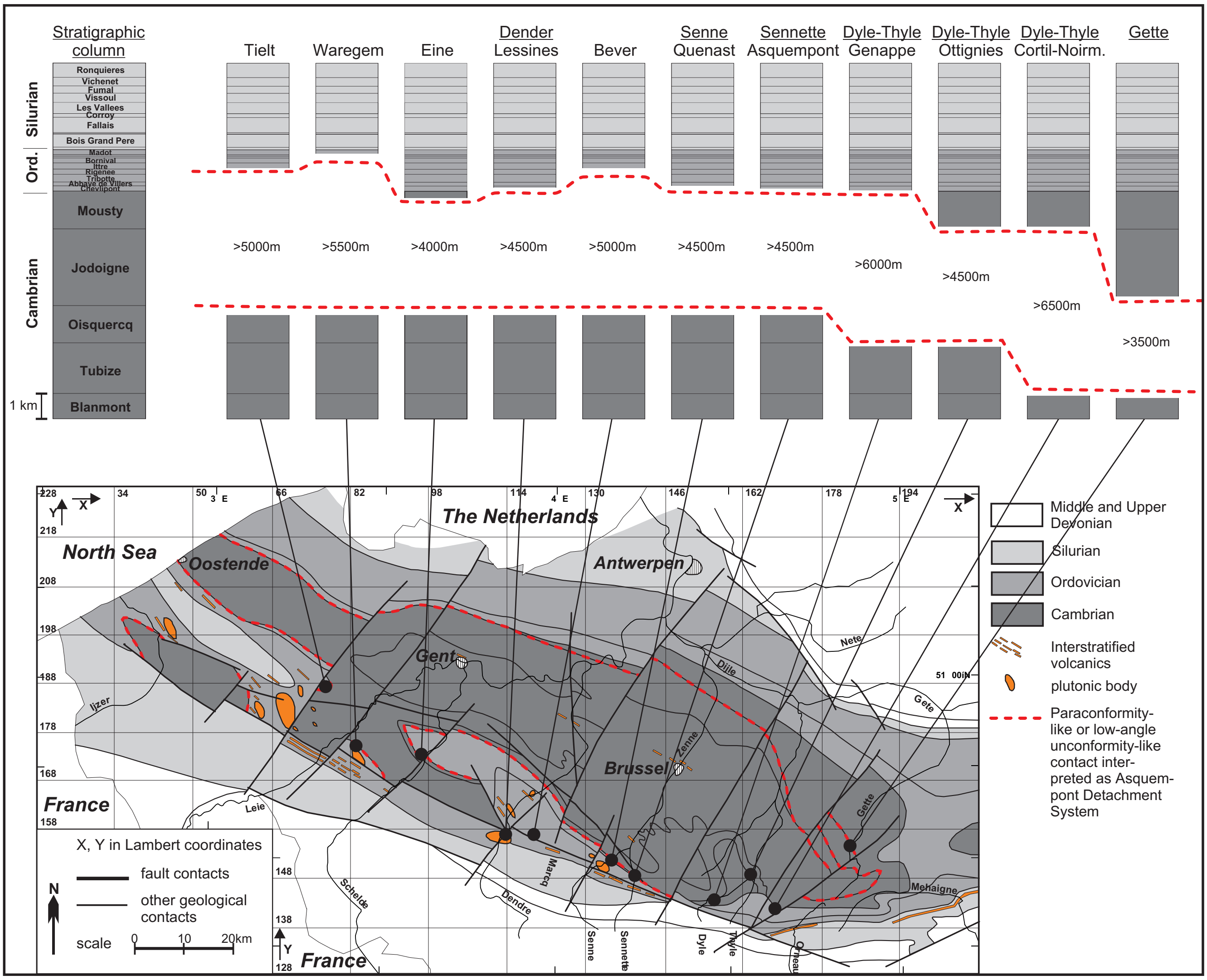


Figure 22. Geological map of the BM showing the effect of the Asquempont detachment (ADS) on the stratigraphy. The same composite stratigraphic column is used for each point of observation and the part removed by the ADS is shown in white together with the minimum thickness estimate (from Debacker et al., 2008).

STOP 4: Mousty Fm. in the old quarry of Franquenies, Ry Angon valley.

Location: New geological map 1/25,000 Wavre – Chaumont-Gistoux (Herbosch & Blockmans, 2012). Private property.

50°39'31.84" N 4°34'27.04" E

Stratigraphy: This outcrop is situated in the Franquenies Mbr., lower part of the Mousty Fm., that extends from the Furongian (upper Cambrian) to the lowermost Tremadocian. The Cambrian part was dated by acritarchs (Vanguestaine, 1992) and the uppermost part by the dendroid graptolite *Rhabdinopora flabelliformis* (Lecompte, 1948). The formation crops out poorly, its lower contact is always faulted (ADS). It is subdivided in three members: the Fanquenies Mbr., an unnamed thick middle mbr. and the upper Tangissart Mbr. The thickness is roughly estimated at > 1500 m.

Observations: Monotonous graphitic and pyritic slate in cm- to dm-size beds. The frequent enrichments in Mn generated spessartine (Spessartine 68 Almandine 26 Grossular 6), Mn ilmenite and Mn chlorite during the Brabantian metamorphism. The Franquenies Mbr. is characterized by dm-size lenses of chert ("phtanite") interstratified in the slate. In thin section, these cherts show numerous transparent objects of 100 to 150 microns formed by cryptocrystalline quartz having the shape and size of radiolarians embedded in a very fine opaque matrix (Fig. 24). The identity of these objects cannot yet be confirmed by specialists due to metamorphic overprint.

The stratification is 60° E with a steep dip and the cleavage is parallel to the rock wall. Old and new observations in the surroundings of the Franquenies quarry (Fig. 23; Debacker et al., 2004a) show a steeply plunging cleavage/bedding intersection lineation and the folds are steeply plunging (type B folds).

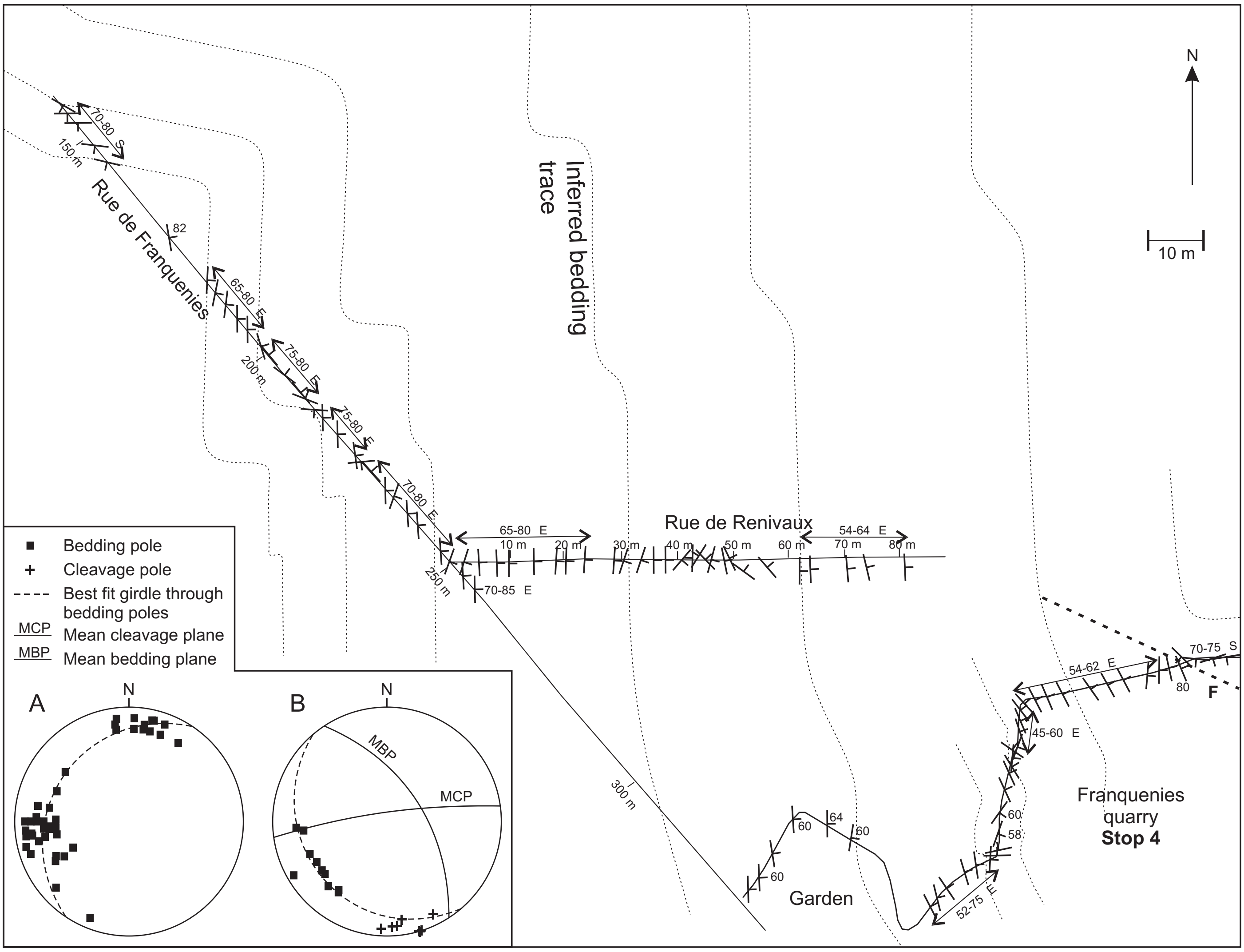


Figure 23. Old (Van Tassel, 1986) and recent observations in the Mousty Fm. in the surroundings of the Franquenies quarry. Bedding and cleavage data indicate the presence of steeply plunging folds of type B. From Debacker et al. (2004b).

The Eocene sand of the cover rests, with only some quartzite and perforated black slate pebbles, in unconformity on the Mousty Fm.

Discussion: **Type of folds in the BM**. Debacker (2001, see synthesis 2012) has shown that there is evidence for one single progressive deformation event in the BM. The main features associated with this deformation are folds cogenetic with a well developed axial plane cleavage: in the core (Brabant Central Unit), steeply plunging folds are associated with subvertical to steeply dipping cleavage while towards the southern side (Senne-Dyle-Orneau Unit) subhorizontal to gently plunging folds are associated with steep to moderately north dipping cleavage (Figs. 9, 23, 25). Type B folds are quite exclusively observed in the steep belt that forms the core of the BM (Fig. 9).

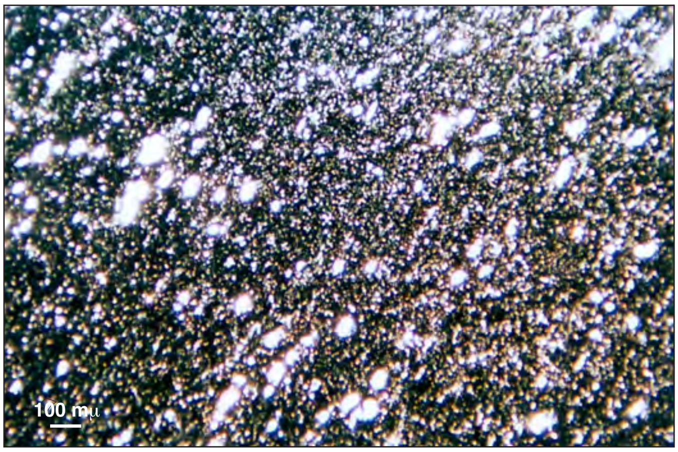


Figure 24. Franquenies Mbr. of the Mousty Fm., Bois des Rêves, Mousty. Thin section in a chert level ("phtanite"), the black matrix is mainly formed by graphite and the white elongated holes (filled with microquartz) are interpreted as phantoms of radiolarians. (Normal light).

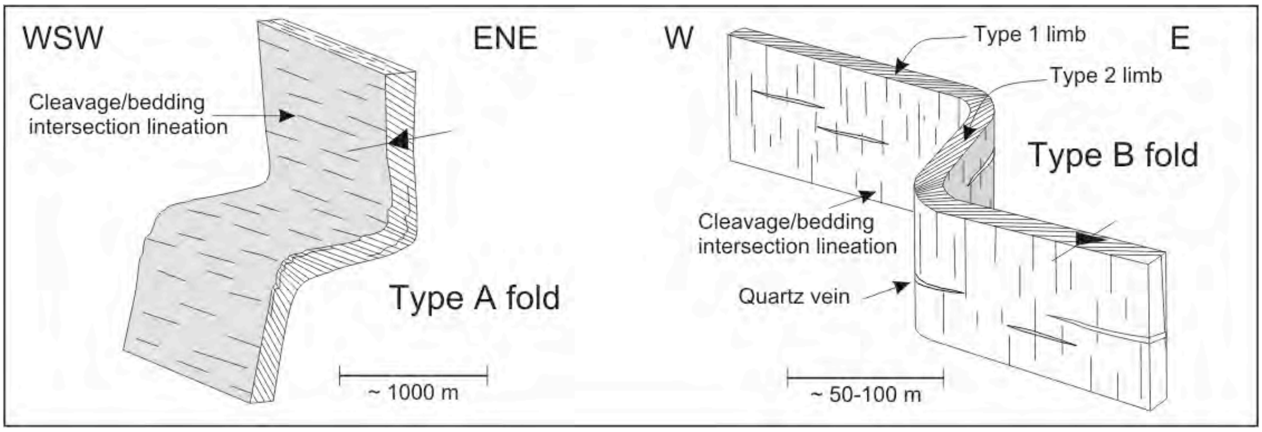


Figure 25. The two main fold types in the BM. The sense of the stratigraphic younging is given by the heavy black arrows (from Herbosch et al., 2008b).

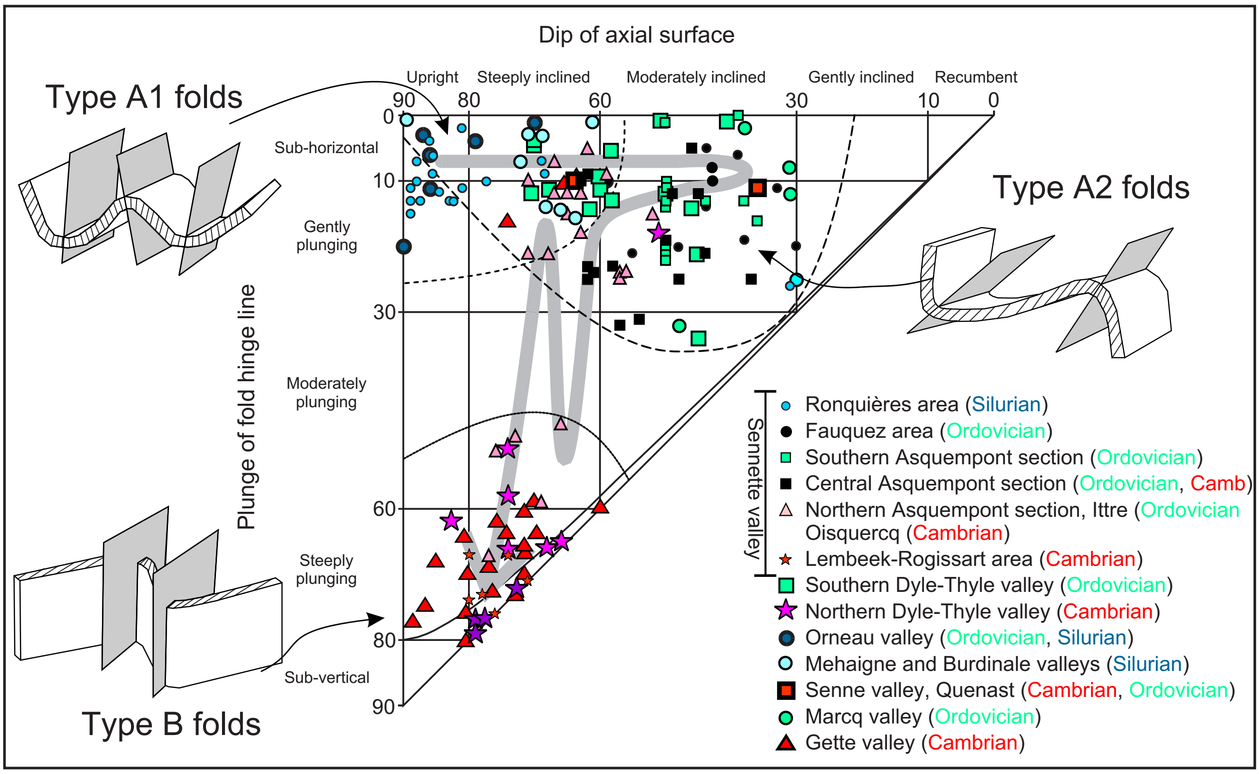


Figure 26. Fleuty diagram with data of the three different types of folds within the main outcrop areas of the BM (modified after Debacker, 2012).

STOP 5: Chevlipont Fm., railway cut near the old mill of Chevlipont, Thyle valley.

Location: New geological map 1/25,000 Nivelles-Genappe (Herbosch & Lemonne, 2000). East slope of the railway cut km 38,2.

50°35’55.29” N 4°32’09.64” E

Stratigraphy: Chevlipont Fm. was dated by the graptolite *Rhabdinopora flabelliformis* (Lecompte, 1948) and by acritarchs (Vanguestaine *in* André et al., 1991) from the lower part of the Tremadocian, lowermost Ordovician (Figs. 5, 28). Its thickness is about 150 m.

Observations: Gray siltstone (called “quartzophyllade” in older literature) with characteristic wavy bedding consisting of mm- to cm-size rhythmic alternations of light gray siltstone and mudstone (Fig. 27). Silty laminae occur in small lenses a few cm long and a few mm thick with oblique laminations (ripples). Sparse interstratified cm- to dm-size beds with massive, plane- parallel or convolute structures can also be observed.

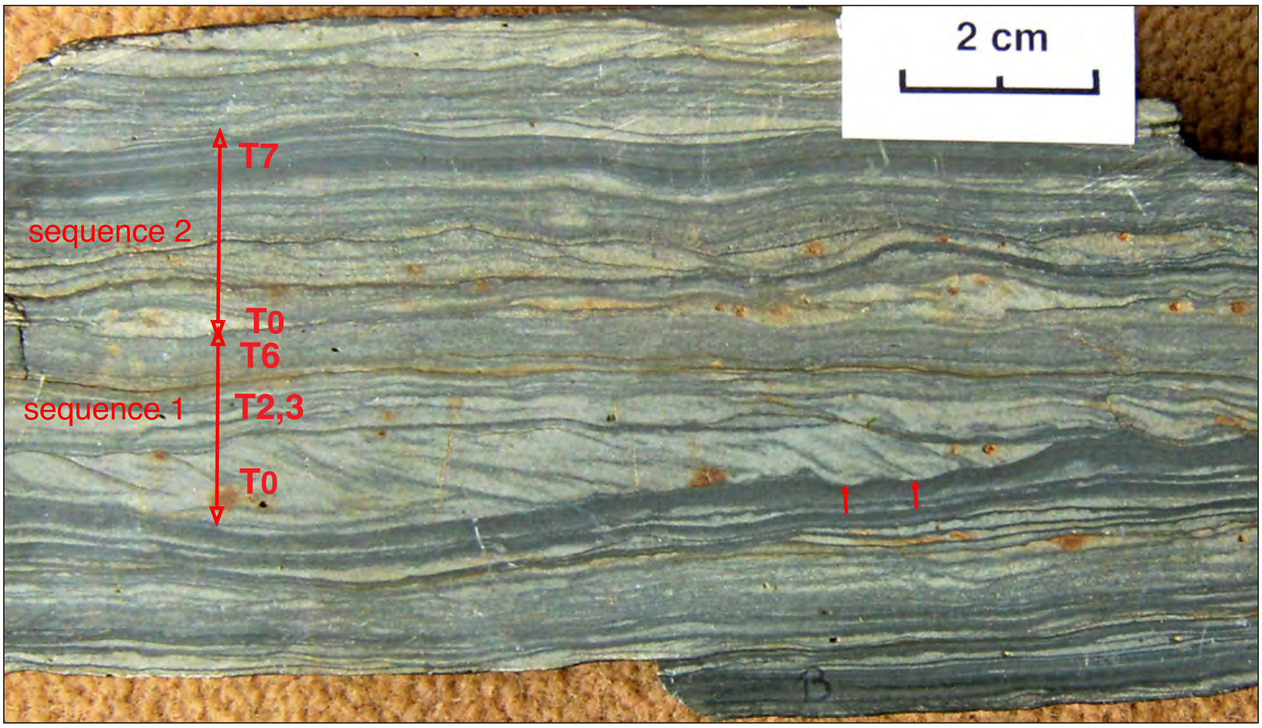


Figure 27. Macrophoto of the Chevlipont Fm. facies. Two top-cut sequences of silt-turbidites described with the Stow model nomenclature of the Fig. 20. Red arrows show load structures at the base of the rippled and cross-laminated T0 division.

Along the same section but about 1 km to the S a decametric slump was described (Beckers, 2004) with a classic turbidite facies.

Interpretation: The wavy bedded siltstones are interpreted as silt turbidites, more precisely as top-cut sequences of low-density turbidites of the Stow model (Stow & Piper, 1984; Fig. 20). The rippled silty and lenticular laminae at the base of each sequence are identified as the term To of the Stow model, the other divisions are more difficult to identify because the sequences are not complete. The very continuous beds with massive, plane-parallel or convolute structures are interpreted as very distal high-density turbidite (Bouma model). The presence of slump is classic in such a turbiditic environment with slope.

Discussion: Mousty-Chevlipont formation transition. It is the only formation boundary visible in the Cambrian and the transition is sedimentologically gradational. Unfortunately we cannot visit this section along the railway. The black shale of Mousty facies shows more and more silt laminae towards the top: packages of black shale alternate with packages with silt laminae, the limit was placed after the last black shale package. This transition could be explained by a regressive trend marking the drifting of Avalonia from Gondwana and the birth of the Rheic Ocean (Fig. 6). The transition zone forms the Tangissart Mbr. and shows *Rhabdinopora flabelliformis,* so the uppermost part of Mousty Fm. is in the lowermost Tremadocian (Figs. 5, 28).

STOP 6: Abbaye de Villers Fm.,
W side of the Thyle valley N of the Abbaye

Location: New geological map 1/25,000 Nivelles-Genappe (Herbosch & Lemonne, 2000). Western side of the road Villers-la-Ville - Laroche.

50°35’39.54” N 4°31’48.57” E

Stratigraphy: Upper Dapingian to lowermost Darriwilian (upper Arenigian) based essentially on acritarchs (Fig. 28; Vanguestaine & Wauthoz, 2011). The thickness is about 150 to 200 m.

Observations: Gray to dark gray fine-grained sandstone to mudstone with regular to irregular, lenticular to wavy cm-scale laminations. Locally, some pyrite and also siderite are observed. Characteristically, the fine-grained sandstone laminae have rather diffuse limits. Frequent and abundant bioturbations, mostly along bedding planes, disturb or even rub out the laminar structure.

Large-scale bedforms and quite frequent slumps are described in other sections, the latter with a S-dipping palaeoslope (Beckers & Debacker, 2006; Debacker & De Meester, 2009).

Interpretation: mid to outer shelf environment of deposition, suboxic (pyrite and siderite).

Discussion: Stratigraphic hiatus between Chevlipont and Abbaye de Villers formations. The contact between the Chevlipont and Abbaye de Villers formations is nowhere visible in the BM. But biostratigraphic work has proved a large hiatus between the two formations (Verniers et al., 2002) and recent investigations (Vanguestaine & Wauthoz, 2011) show that the hiatus is more important than previously thought. It extends from the lower Tremadocian to the latest Dapingian or early Darriwilian, i. e. about 14 Ma (Fig. 28).

A similar observation has also been made in the Wépion borehole (Condroz Inlier, Fig. 3) where a thin conglomerate overlies in accordance the top of the hiatus (Graulich, 1961) that extends from the lower Tremadocian to the middle Darriwilian (Servais & Maletz, 1992; Vanmeirhaeghe, 2006). **An unconformity at the top of the Chevlipont Fm.** is thus demonstrated. This unconformity marks the end of Megasequence 1 (Fig. 5) and is interpreted as resulting from the rifting of Avalonia from Gondwana that occurs during the mid Tremadocian (Cocks & Torsvik, 2002, 2005; Linnemann et al., 2012).

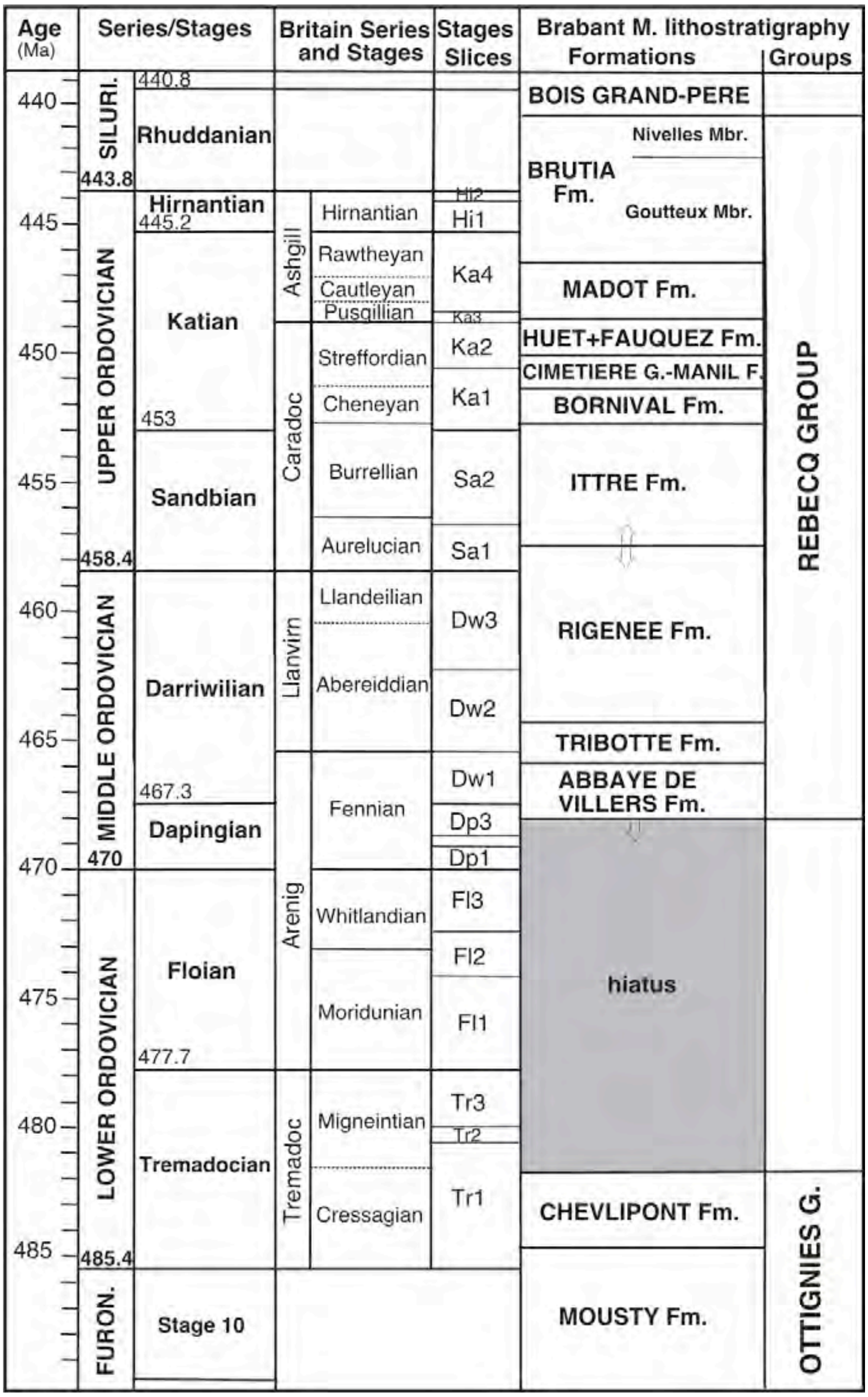


Figure 28. Chronostratigraphic position of the Ordovician and lowest Silurian lithostratigraphic units in the BM. Chronostratigraphy after Cooper & Sadler (2012). From Herbosch & Verniers (2014).

STOP 7: Tribotte Fm.,
wall of the Villers-la-Ville church.

Location: New geological map 1/25,000 Nivelles-Genappe (Herbosch & Lemonne, 2000). No good outcrop except along a railway. The church was constructed with rocks from the upper part of the Tribotte Fm.

Stratigraphy: The Tribotte Fm. was dated by chitinozoans (Samuelsson & Verniers, 2000) and acritarchs (Vanguestaine & Wauthoz, 2011) from the lower Darriwilian (Fig. 28) (late Arenigian to earliest Llanvirnian). Its thickness is estimated at 200 to 250 m.

Observations: Yellowish gray to dark gray siltstone with flaser bedding. Bioturbation is strong with predominance of oblique and vertical burrows (Fig. 29). This facies forms the upper third part of the formation.

Figure 29. Macrophoto of the upper part of the Tribotte Fm. facies. The bioturbations are mostly vertical with spreiten. Intertidal facies. The yellow scale bar is 6 cm long.

Interpretation: The lithology, the flaser bedding and the vertical bioturbations are characteristic of an intertidal environment.

STOP 8: Old sunken road to Rigenée
near the Le Chatelet castle, Thyle valley.

Location: New geological map 1/25,000 Chastre-Gembloux (Delcambre et al., 2002).
50°34'17.01" N 4°31'16.34" E

Stratigraphy: Lower boundary of the formation. The Rigenée Fm. is dated from the upper two thirds of the Darriwilian (Llanvirn) based on graptolites (Maletz & Servais, 1998), acritarchs (Servais, 1991, 1993; Vanguestaine & Wauthoz, 2011) and chitinozoans (Samuelsson & Verniers, 2000) (Figs. 8, 26). Its thickness is about 200 m.

Observations: dark gray to bluish slate (siltstone to mudstone), vaguely or coarsely laminated or without well marked stratification, locally bearing pyrite.

Interpretation: The lower boundary marks a rapid transition over 5 to 10 m and a sharp change from the light coloured clayey siltstone of the upper part of the Tribotte Fm. to the dark slate of the Rigenée Fm. This abrupt change of lithofacies records a regional transgression well known in north Gondwana and Baltica as the « Formosa flooding Event » (Paris et al., 2007). Indeed, the *Siphonochitina formosa* Biozone is exactly placed at the Tribotte-Rigenée transition (Fig. 8) in the lower Llanvirn (Cooper & Sadler, 2012).

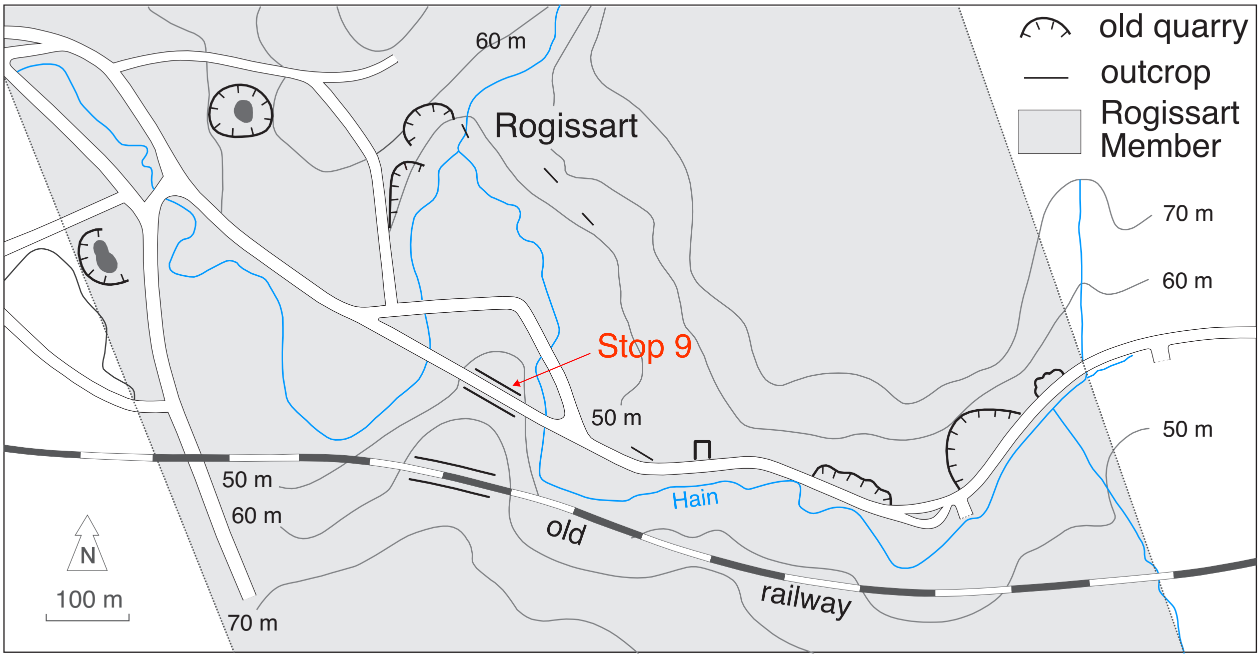


Figure 30. Geological situation of the Rogissart Mbr. of the Tubize Fm. in the Hain valley. Mean directions: S0 N334°E dip 76°NE; S1 N308°E dip 71°NE.

4. Description of the Senne Basin stops (second day)

STOP 9: Rogissart Mbr. of the Tubize Fm. at Hameau du 45, Hain valley.

Location: New geological map 1/25,000 Ittre-Rebecq (Herbosch et al., 2013).

50°40’46.65” N 4°14’04.46” E

Stratigraphy: The trace fossil *Oldhamia radiata* was found in this outcrop (Malaise, 1883). An age could be assigned to the Rogissart Mbr. by first comparison with the upper part of the Deville Group in the Ardenne Inliers where *Oldhamia* was also discovered. These inliers were dated by acritarchs from the Zone 0 of Vanguestaine (1992). This biozone corresponds in the global stratigraphy (Peng et al., 2012) to the interval from the middle of the Cambrian Stage 3 (Series 2) to about 3/4 of the way up to Stage 5 (base of the *P.* gibbus Zone). As the Tubize Fm. is geometrically younger than the Blanmont Fm. that shows also *Oldhamia*, we could estimate that its age covers roughly the interval from the middle of Stage 3 to middle of Stage 4 (unnamed Series 2) (Fig. 5). The thickness of the Rogissart Mbr. is about 800 m and well constrained by outcrops (Fig. 30).

Observations: Fine to coarse-grained pale quartzitic sandstone, feldspathic sandstone, arkose and greywacke in dm- to m-size beds alternating with more or less clayey siltstone and green slate (claystone), together forming fining-upward sequences (Vander Auwera & André, 1985). The sandstone beds show massive, plane-parallel, oblique and sometimes convolute laminations. The stratification and the cleavage are subvertical which indicates type B folds.

Interpretations: The rhythmic sedimentation is interpreted as high-density turbidites of the Bouma type (Fig. 20). Some thick turbidites are amalgamated and probably proximal.

Type B folds are extremely frequent in the Central Brabant Unit (Fig. 26).

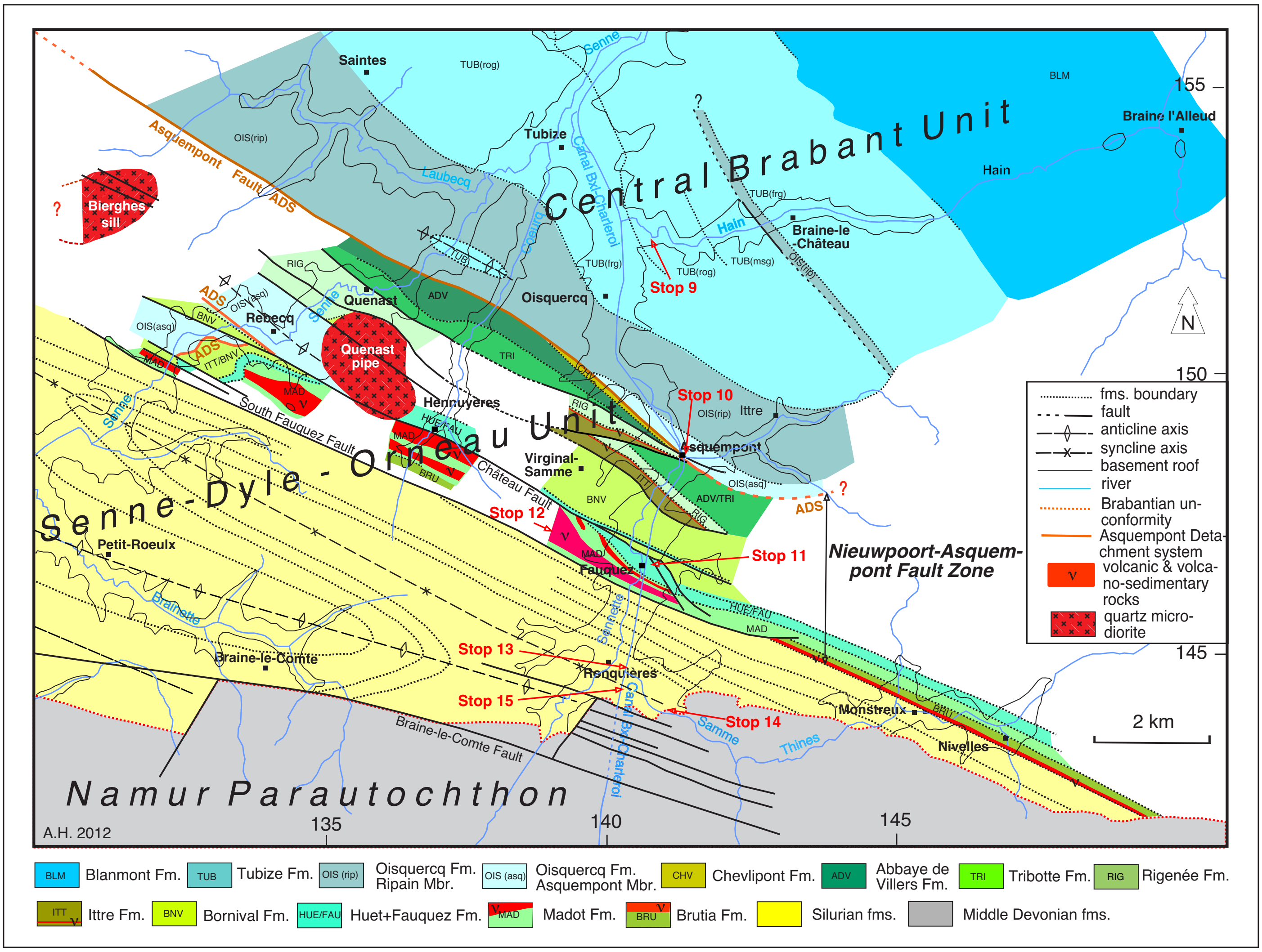


Figure 31. Subcrop geological map of the Senne Basin with location of the second day stops. The Asquempont Detachment System (ADS) separates the Central Brabant Unit (Cambrian) from the Senne-Dyle-Orneau Unit (Ordovician and Silurian). The Ordovician formations (in green) are cut by numerous faults that belong to the Nieuwpoort-Asquempont Fault Zone (Legrand, 1968). This zone strongly contrasts with the southern Silurian formations, gently folded and not affected by faults. Unpublished map based on personal observations, extended and modified after the new geological maps of Wallonia.

STOP 10: Bruxelles-Charleroi canal trench at Asquempont bridge, Sennette valley.

A. Asquempont Mbr. of the Oisquercq Fm. (Figs. 29, 31)

Location: New geological map 1/25,000 Braine-le-Comte – Feluy (Hennebert & Eggermont, 2002). East bank of the canal km 41.1-40.1
50°38’37.24 N 4°14’23.00 E

Stratigraphy: The formation was dated by acritarchs (Vanguestaine, 1991, 1992) from near the lower-middle Cambrian boundary, i.e. the interval between the middle of Stage 4 and the lower part of Stage 5 (Fig. 5). The thickness is about 1500 m for the entire formation.

Observations: Massive greenish gray to green very fine slate without any visible stratification, even in thin section. Good cleavage. The Ripain lower Mbr. of the same formation, not seen on the fieldtrip, presents exactly the same lithology but with a purplish colour.

Interpretation: The monotony, extremely fine grain and thickness favour deposition in the formation as a deep pelagic basin environment.

B. Asquempont Fault (Figs. 29, 32, 33)

Location: New geological map 1/25,000 Braine-le-Comte – Feluy (Hennebert & Eggermont, 2002). East bank of the canal km 40.12.
50°38’25.61 N 4°14’22.31 E

Observations: Fig. 32 shows a sketch of the faulted contact between the upper member of the Oisquercq Fm. (in green) and the Chevlipont Fm. (in yellow; dated by acritarchs). The fault F8 pre-dates the cleavage development and is affected by the post-cleavage overturning, a small normal fault and F7. This is the “true” Asquempont Fault. F7 truncates F8, thus only forming an apparent limit between the Cambrian and Ordovician. The fault is underlined by lenses of quartz and carbonates. It testifies to several episodes of movement, the early episodes being accompanied by folding, the later ones by brittle deformation.

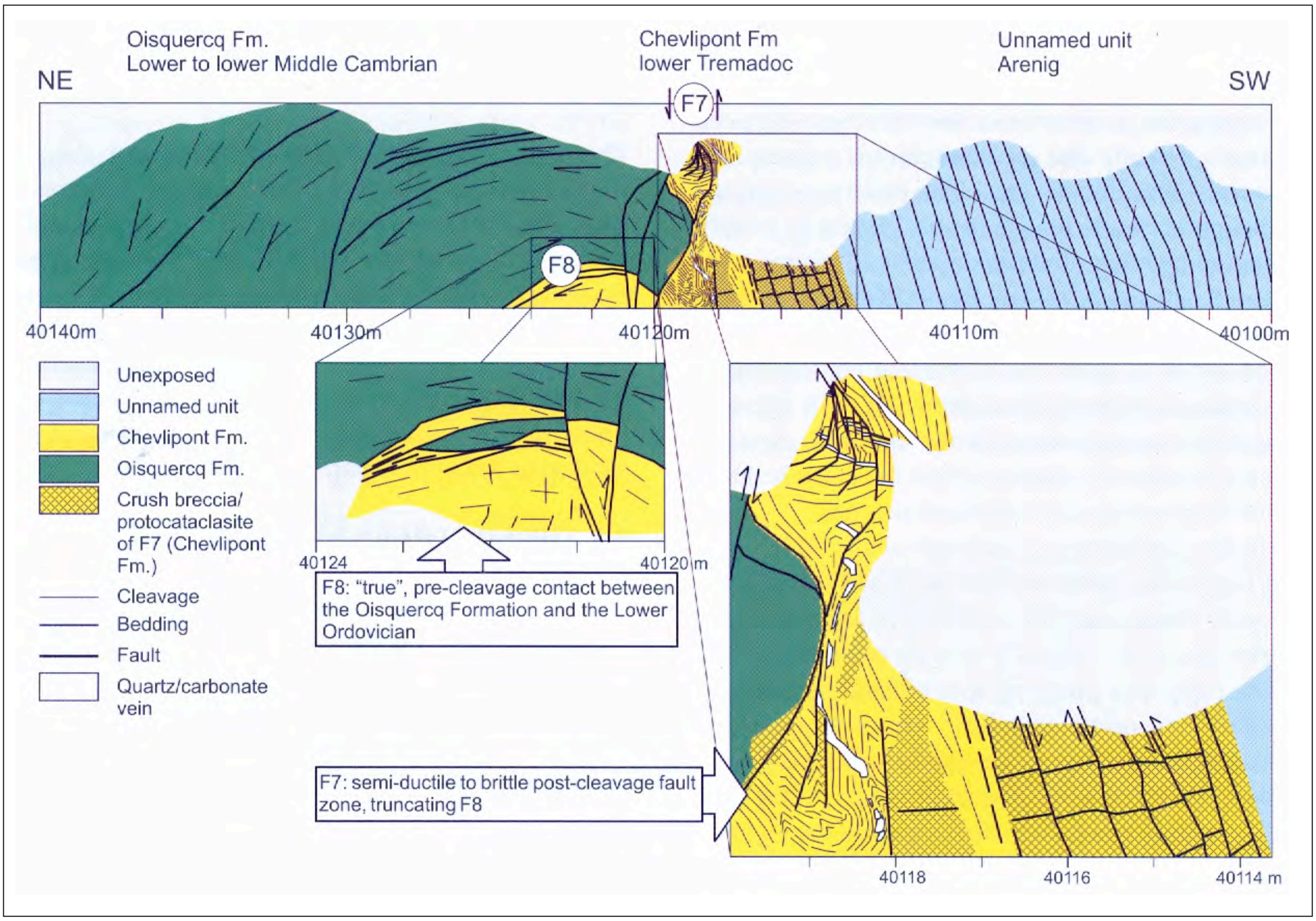


Figure 32. Cross-section along the Bruxelles-Charleroi canal, W side showing the Asquempont Fault (from Debacker et al., 2003). F8 is the Asquempont Fault.

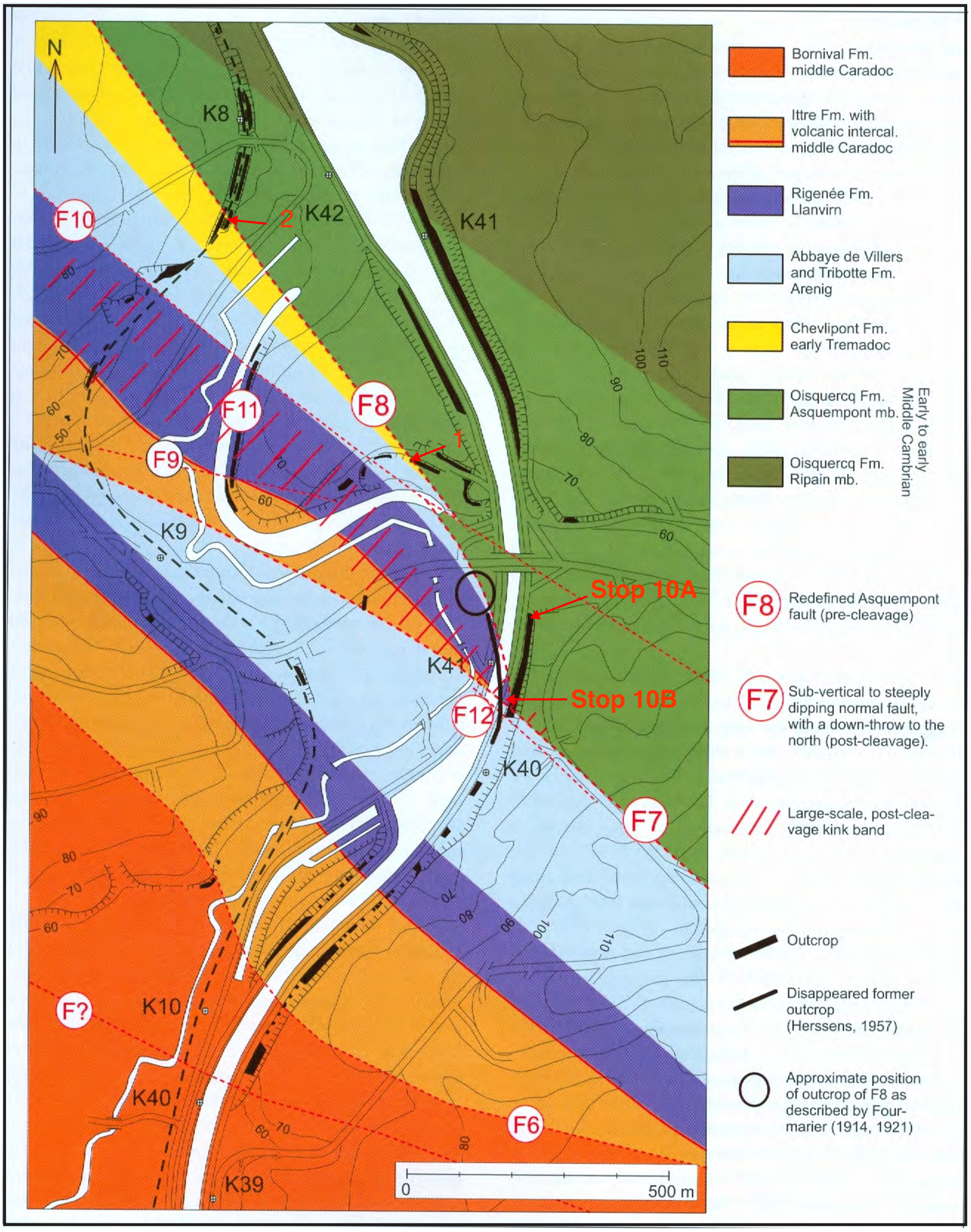


Figure 33. Detailed geological map of the Asquempont region, Sennette valley (from Debacker et al., 2003) with location of stop 10A, 10B and outcrops 1 and 2 (see text).

Interpretation: The Asquempont Fault is observed not only here but also on the other side of the canal along the old canal (Fig. 33 red arrow 1) and along the trench of the old railway (Fig. 33 red arrow 2). It is also observed to the NW in the Senne valley to the N of the Quenast railway station (see Fig. 31). Analysis of these 4 sections (Lenoir, 1987; Debacker et al., 2003) shows that the real contact between the Cambrian and Ordovician is formed by a pre-cleavage normal fault that can be followed over 6 km to the NW and more if we use the borehole information (Herbosch et al., 2013). The Asquempont Fault is a pre-cleavage fault and also older than the Brabantian tectonic inversion. Debacker (2001) has suggested that the Asquempont Fault is a pre-cleavage and pre-folding low-angle extensional detachment. This model was confirmed by the progress of the BM cartography: the Asquempont Detachment System was observed in the Dyle Basin (Debacker et al., 2004b), then in the Grande Gette Basin (Herbosch et al., 2008a, in press). See also Stop 3 discussion.

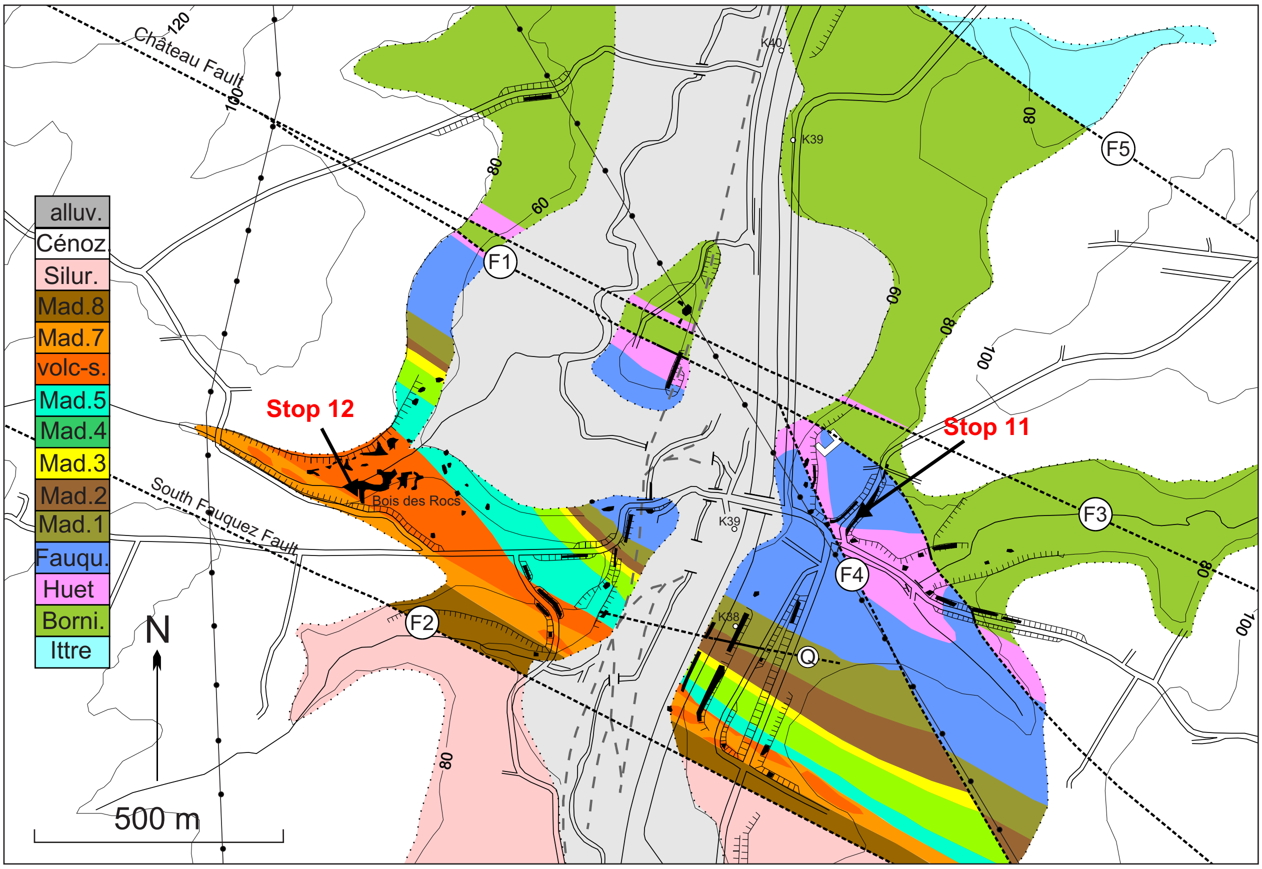


Figure 34. Detailed geological map of the Fauquez region with location of stops 11 and 12 (from Verniers et al., 2005).

STOP 11: Fauquez Fm. at the sunken road of Fauquez, Sennette valley.

Location: New geological map 1/25,000 Braine-le-Comte – Feluy (Hennebert & Eggermont, 2002). Small valley East bank of the canal.

50°37'28.91" N 4°13'57.84" E

Stratigraphy: According to the thesis of Vanmeirhaeghe (2006), this formation belongs to the *A. reticulifera* subzone of the *F. spinifera* Biozone that has probably a late Onnian to lowermost Pusgillian age in the British zonation (Fig. 8; Vandenbroecke, 2008). This confirms globally the former dating by graptolites: *P. linearis* to upper part of *D. clingani* zones (Maletz & Servais, 1998). The Fauquez Fm. is placed in a very short time interval in the late Onnian to lowermost Pusgillian from the middle Katian Stage (*circa* 448 Ma, Cooper & Sadler, 2012).

Observations: The formation is formed by graptolitic black slate, more precisely by a rhythmic and mm- to cm-size alternation of pyritic siltstone lamina and dark-gray to black slates (claystones) laminae (Fig. 35). The base of the sequences is always marked by mm-size euhedral pyrites that are completely dissolved at outcrop surface. From about 4 m above the base, abundant graptolites are found in the darker claystone laminae. The thickness is >35 m in the outcrops of the Senne Basin and > 58 m in the Lessines borehole.

Interpretation: The cm-scale rhythmicity and the structure of the basal pyritic laminae favour interpretation of these black slates as mud turbidites and permit comparison with the Stow model (Figs. 20, 35). The pyrite is formed during early diagenesis inside the more porous silt laminae (T0 to T4) by the microbial degradation of the organic matter. A deep marine anoxic environment of deposition, probably the slope, is postulated.

STOP 12: Madot Fm., Bois des Rocs in the Fauquez brook valley.

Location: New geological map 1/25,000 Braine-le-Comte – Feluy (Hennebert & Eggermont, 2002). Fauquez brook to the West of Fauquez.

50°37'32.40" N 4°13'10.51" E

Stratigraphy: Upper part of the Madot Fm. (Member 6 of Verniers et al., 2005). Based on chitinozoans (Vanmeirhaeghe, 2006), the formation can be ascribed to the interval lower Pusgillian to mid Rawtheian, mid Katian (Figs. 8, 28; *circa* 448-446 Ma, Cooper & Sadler, 2012). The thickness of the formation is variable from 200 to >300 m due to the variable amount of volcanic and volcano-sedimentary rocks.

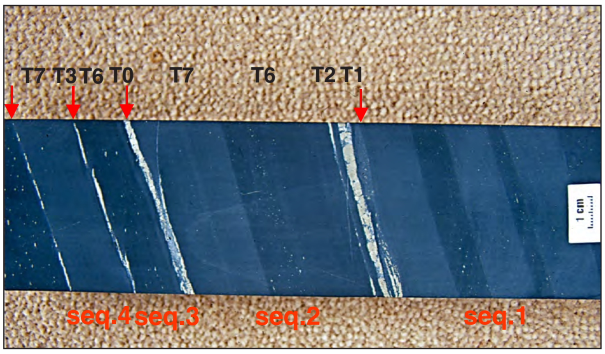


Figure 35. Polished surface of a core showing sequences of mud turbidite (red arrows) and the divisions of the Stow model (in black). Lessines borehole 149 m (Herbosch et al., 2008). Pyrite (in yellow) underlines the base of each sequence. Compare with the model of Fig. 20.

Observations: The towers of dacite lava that appear scattered in the valley are formed by erosion of the surrounding slate of the Madot Fm. The dacitic lavas present in many places a brecciated structure. They are also frequently foliated, and contain large slate xenoliths.

Interpretation: The dacites are interpreted as underwater lava flows (André, 1991). In thin section they present phenocrysts of plagioclase, pyroxene altered to chlorite and quartz, incorporated in a quartz-feldspatic mesostasis. Dating of the zircons of the dacite by LA-ICP-MS gives a concordant age of 444 ± 6 Ma (Linnemann et al., 2012) which is in good agreement with the biostratigraphic age.

The presence of abundant warm-water fossils (crinoids, bryozoans, brachiopods, trilobites, corals) in some members of the formation (and also in the Huet Fm.) records the rapid shift of Avalonia to lower latitudes (Fig. 6). Fortey & Cocks (2005) have also shown that a late Ordovician global warming episode, the Boda Event, has affected the low-latitude region of Gondwana. The movement of benthic faunas such as trilobites and brachiopods to progressively lower latitudes and the abrupt appearance of limestone formations during the mid-Katian are observed not only in Gondwana but also in Baltica (Boda limestones of Sweden) and Avalonia (N Wales). The Madot Fm. corresponds exactly to this stratigraphic interval (see also the introduction §2.4).

Discussion 1: Transition Fauquez-Madot formations and end of Megasequence 2: The transition from Fauquez to Madot fms. marks an abrupt change of bathymetry. We have seen in stop 11 that the Fauquez Fm. is formed by mud turbidites interpreted as a deep marine probably slope environment. The Madot Formation shows numerous levels of shelly facies and volcano-sedimentary facies formed in a shallow shelf environment. This change of bathymetry cannot be explained only by a sea-level change: it is more likely linked to the soft docking of the Avalonia microplate with the Baltica continent that occurs at the same age (Cocks & Torsvik, 2002; 2005; Cocks & Fortey, 2009; Linnemann et al., 2012). This event marks the end of Megasequence 2 (Herbosch & Verniers, 2014).

Discussion 2: The magmatism in the BM: The BM magmatism and a new hypothesis about its genesis have been described in the geological introduction (§ 2.6).

We do not have the opportunity to visit the pipe of Quenast which is dated at 430 ± 3 Ma (Fig. 11). Recent unpublished U-Pb S.H.R.I.M.P. dating of zircons from the sill of Bierghes, the sill of Lessines (see map of Fig. 31) and volcanic rocks from Marcq at 445 ± 2 Ma, show that the climax of the magmatic activity can be constrained to the interval 448-440 Ma that corresponds to the Madot and Brutia formations (uppermost Ordovician to lowermost Silurian). This period corresponds to the soft docking of Avalonia and Baltica (*circa* 445 Ma ; Cocks & Torsvik, 2002, 2005) and gives an explanation to this pre-tectonic magmatic event (Linnemann et al., 2012).

STOP 13: Ronquières Fm. at Ronquières, W side of the canal.

Locations: New geological map 1/25,000 Braine-le-Comte – Feluy (Hennebert & Eggermont, 2002). West side of the canal Bruxelles-Charleroi.

50°36'27.19" N 4°13'30.39" E

Stratigraphy: On the basis of graptolites and chitinozoans (Louwye et al., 1992) the formation is dated from the Gorstian Stage (lower Ludlow). Its minimum thickness is 530 m in the Sennette valley.

Observations: Dark to light gray dm and rhythmic alternance of silty slate and slate (mudstone). Some very fine sandstone layers with parallel or oblique laminations and darker finely laminated mudstone could be observed. Fine cm volcanic tuff levels could be also observed.

Interpretation: These rhythmic deposits are interpreted as distal turbidites of the Bouma type. The darker finely laminated mudstones are interpreted as laminated hemipelagites that represent the inter-turbidite sedimentation. They correspond to climatic variations (« varves »; Verniers & Van Grootel, 1991).

Discussion: Thickness of the Silurian and the development of a foreland basin.

As is visible in Fig. 7 the subsidence increases progressively from Brutia and Bois Grand-Père formations upwards. Then, from Corroy to Ronquières formations, the subsidence is extremely rapid (about 1.800 m in 7 Ma) and the sedimentation is of turbiditic type. The total thickness of Megasequence 3 is >3.5 km. This acceleration corresponds to the development of a foreland basin on both rims of the BM. This reflects the tectonic inversion of the core of the BM (Brabantian orogeny) following the Avalonia-Baltica docking with Laurentia (Fig. 15).

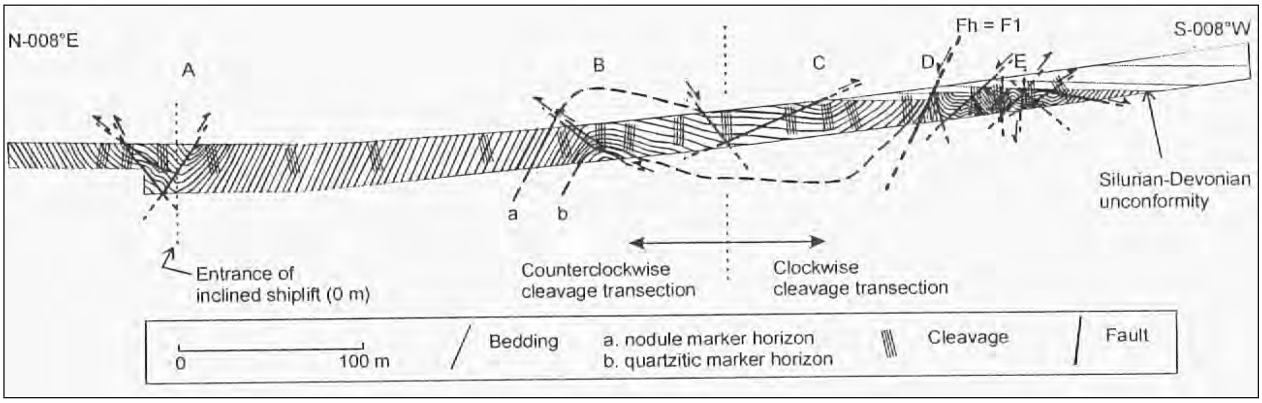


Figure 36. Section along the E-side of the inclined shiplift of Ronquières. A, B, C, D and E are respectively the Porte Avale synform, the central antiform, the Belvédère synform, the faulted antiform and the synform complex. The unconformity is at the southern extremity of the section. From Debacker et al. (1999).

STOP 14: Bois de Bordeaux Formation (Mbr. des Mautiennes), vallée de la Samme.

Location: New geological map 1/25,000 Braine-le-Comte – Feluy (Hennebert & Eggermont, 2002). North slope (very steep) of the Samme valley, nature reserve.

50°36’08.28” N 4°14’27.97” E

Stratigraphy: the formation is dated from the Givetian on the basis of plant remains and conodonts.

Observations: red conglomerates with dm (30-40 cm) sandstone boulders from the base of the member.

Interpretation: the conglomerate lies in angular unconformity upon the Silurian of the BM (next stop). The conglomerates are interpreted as continental deposits and the outcrop is interpreted as a river bed deposit.

STOP 15 (optional): View on the Brabantian unconformity, inclined shiplift trench.

Location: New geological map 1/25,000 Braine-le-Comte – Feluy (Hennebert & Eggermont, 2002). View from the W side of the inclined shiplift trench.

Stratigraphy: Below the unconformity, the Brabant basement is represented by the Ronquières Fm. of lower Ludlow age and above the unconformity, by the red conglomerate and shale from the Bois de Bordeaux Fm. of Givetian age. The stratigraphic hiatus is thus more than 30 Ma.

Observation: The red conglomerate is gently inclined to the S and contrasts with the Ronquieres Fm. rocks that are folded (figs. 36, 37).

Interpretation: The tectonic inversion of the BM is responsible for the Lower-partially Middle Devonian hiatus.

Figure 37. Photo of the unconformity, seen from the Stop 15 (from Sintubin et al., 2008).

Acknowledgements

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