THE LOWER PALAEOZOIC STRATIGRAPHY AND SEDIMENTOLOGY OF THE BRABANT MASSIF IN THE DYLE AND ORNEAU VALLEYS AND OF THE CONDROZ INLIER AT FOSSES: AN EXCURSION GUIDEBOOK

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(31 figures, 9 photo-plates)

ABSTRACT. The excursion guidebook describes in an introduction the Brabant Massif and the Condroz Inlier in its tectonic and palaeogeographic context during the Cambrian to Lower Devonian. The excursion describes the main Lower Palaeozoic lithostratigraphic units, their sedimentology, fossil content and structures present in the Brabant Massif, in a north-south transsect using two parallel valleys: the Dyle-Thyle valleys (Cambrian and Ordovician) and the Orneau valley (Ordovician and Silurian). Contrasting lithostratigraphic units, sedimentology, fossil content and structures are shown in the Condroz Inlier, in its western extremity, where two structural units are present: the central Condroz part and the Puagne Inlier. A new relative dating and a local biozonation, both with chitinozoans, are presented of (parts of) the Corroy, Fumal and Vichenet formations in the Brabant Massif and of the Vitrival-Bruyère, Fosses, Génicot formations and the base of the Criptia Group in the Condroz Inlier. Sedimentological and biostratigraphical arguments are proposed that may indicate that the Génicot Formation was deposited just before, during and just after the Hirnantian (latest Ordovician) glaciation. This means the first indication of the presence of the latter in Belgium.

KEYWORDS: lithostratigraphy, sedimentology, Lower Palaeozoic, Brabant Massif, excursion guidebook.

1. General introduction

The Lower Palaeozoic formations in Belgium crop out in six areas: the Brabant Massif in the north, north-west and central part of the country, the Condroz Inlier, in the south-centre of the country and four inliers in the Ardennes in the south and east of the country, traditionally called "massifs" in the regional literature: the two larger Stavelot and Rocroi inliers and the two smaller Givonne and Serpont inliers (Figs. 1 & 2). Four of these inliers extend into the surrounding countries: the Brabant Massif into the Netherlands and France, the Rocroi Inlier/Massif and the Givonne Inlier into France and the Stavelot Inlier/Massif into Germany where it is called Stavelot (-Venn) Inlier/Massif. All the massifs or inliers are unconformably covered by Devonian rocks. The unconformity has been called Caledonian in the literature and results from orogenic deformations caused by the Avalonia-Baltica-Laurentia collisions during the Late Ordovician to Early Devonian times. A summary of the new hypotheses is given in Verniers *et al.* (2002a)

The Brabant Massif (Dumont, 1847) is defined by the present day outcrop and (sub-Mesozoic) subcrop of Lower Palaeozoic rocks in central and western Belgium, northern France and southwestern Netherlands unconformably covered by Middle Devonian strata. The depositional basin is called the Brabant Basin, with in the Late Ordovician and Silurian a shelf area in the southwestern part, south of the line Ronse - Veurne, the southwestern Brabant Shelf, and a basinal area north of it, the central and north Brabant

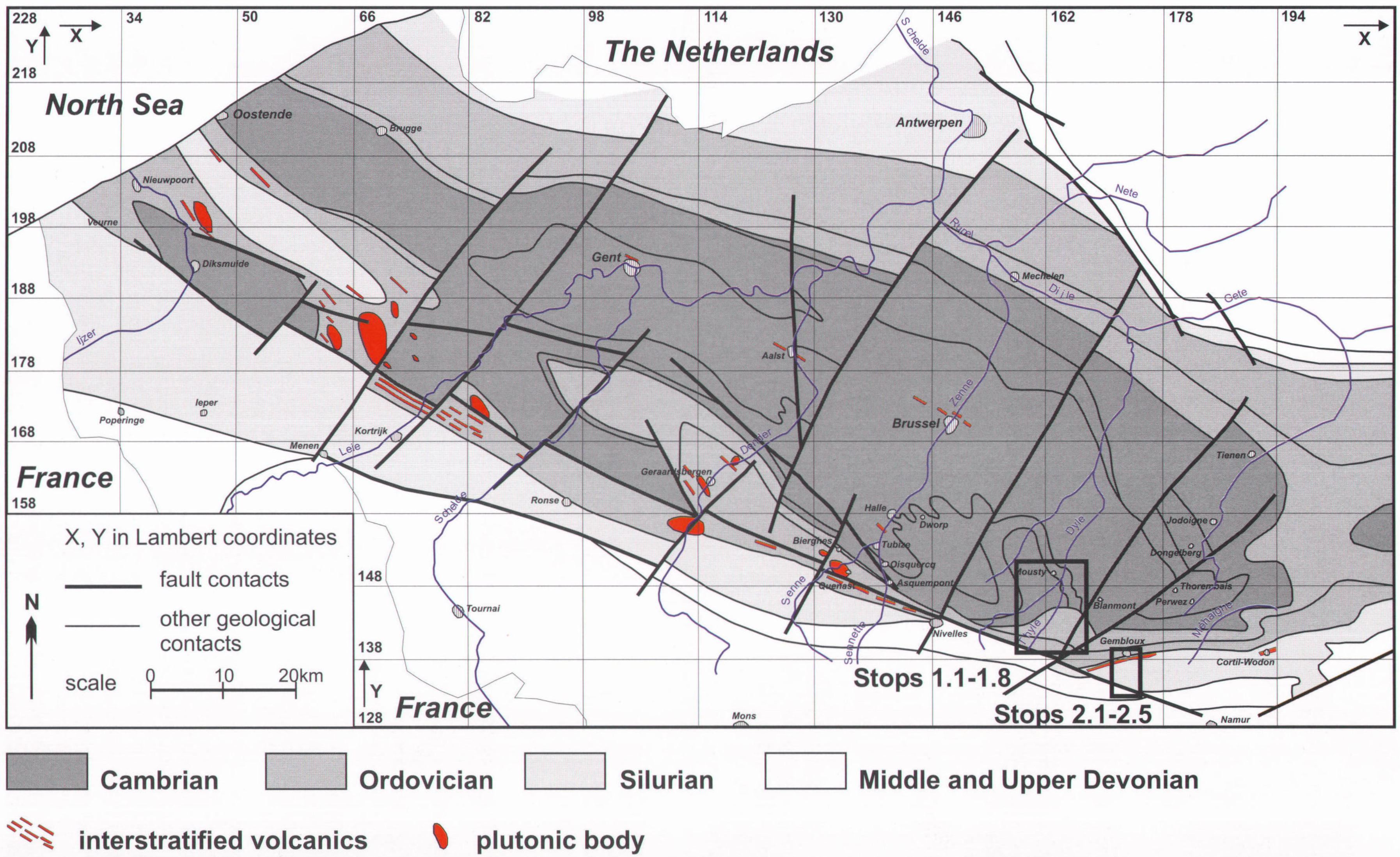

Figure 1. Geological subcrop map of the Lower Palaeozoic Brabant Massif with location of the stops (redrawn after De Vos *et al.*, 1993).

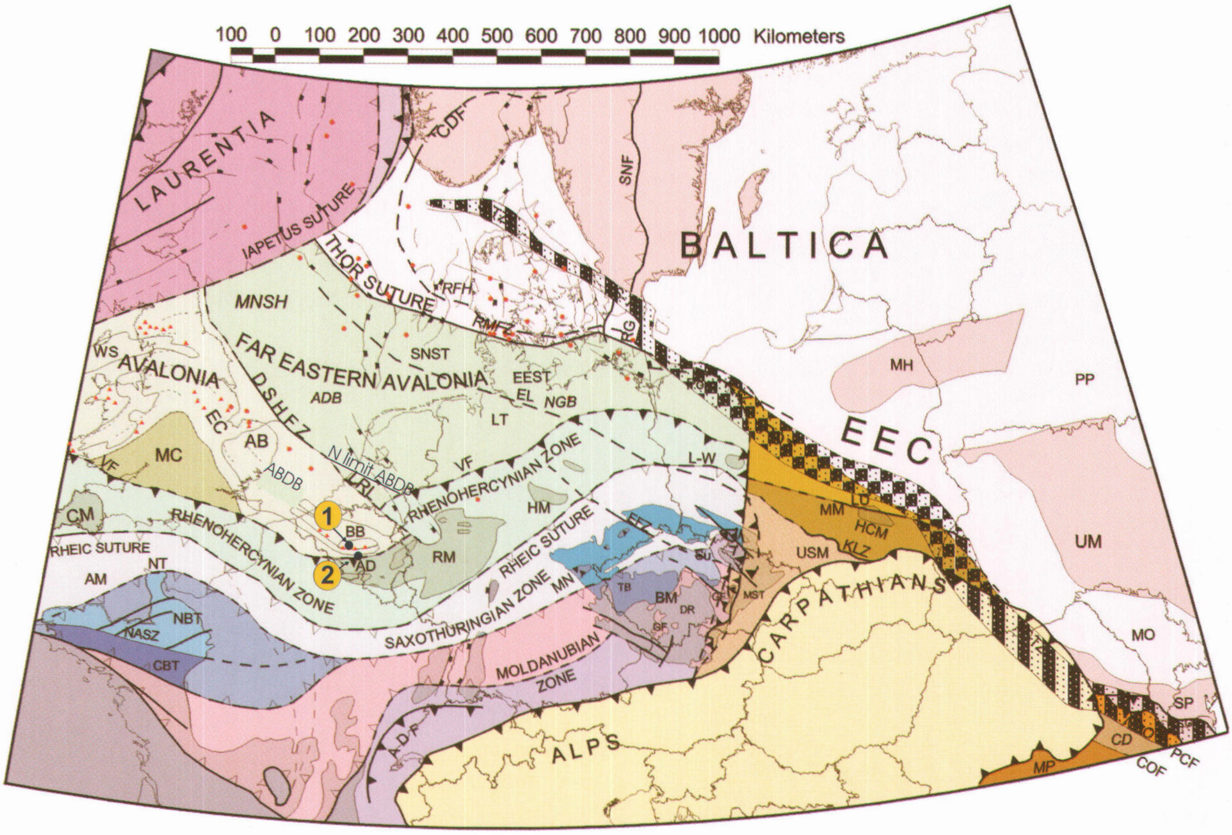

Figure 2. Basement tectonic sketch map of NW Europe after PACE TMR Network Team & Winchester (2002). Oceanic sutures, open ticks; orogenic frontal zones, filled ticks; key boreholes, solid dots; Ordovician arc volcanic rocks in Avalonia, triangles. Geographic locations mentioned in text: Liz, Lizard; Pom, Pomerania; SG, Sowie Góry. Key: **Post-Palaeozoic basins and platforms**: ADB, Anglo-Dutch Basin; CD, Central Dobrogea; MNSH, Mid-North Sea High; MP, Moesian Platform; NDO, North Dobrogea Orogen; NGB, North German Basin; POT, Polish Trough; RFH, Rynkøbing-Fyn High; RG, Rønne Graben; RMFZ, Rømø-Møn Fracture Zone; SP, Scythian Platform. **Postulated Palaeozoic terranes and possible terrane/sub-terrane boundaries**: DSHFZ, Dowsing-South Hewett Fault Zone; EEST, East Elbian Suspect Terranes; EL, Elbe Lineament; KLZ, Kraków-Lubliniec Zone; LRL, Lower Rhine Lineament; LT, Lüneburg Terrane; LU, Lysogory Unit (?Terrane); MM, Malopolska Massif (?Terrane); MST, Moravo-Silesian Terrane; NT, Norannian Terrane; PCF, Peceneaga-Camena Fault; SNST, southern North Sea Terrane; SGF, Sfantu Gheorghe Fault. **Proterozoic-Palaeozoic tectonic elements**: ABDB: Anglo-Brabant Deformation Belt; AB, Anglian Basin; AD, Ardennes Massifs; AM, Armorican Massif; BB, Brabant Massif; BM, Bohemian Massif; CBT, Central Brittany Terrane; CDF, front of Caledonian deformation; CM, Cornubian Massif; COF, Capidava-Ovidiu Fault; DR, Drosendorf Unit (of BM); EC, eastern English Caledonides; EEC, East European Craton; EFZ, Elbe Fault Zone; CF, Gföhl Unit (of BM); HM, Harz Mountains; HCM, Holy Cross Mountains; L-W, Leszno-Wolsztyn Basement High; MC, Midlands Microcraton; MH, Mazurska High; MN, Münchberg Nappe (of BM); MO, Moldavian Platform; NASZ, N Armorican Shear Zone; NBT, N Brittany Terrane; PP, Pripyat Trough; RM, Rhenish Massif; USM, Upper Silesian Massif (= MST); SNF, Sveconorwegian Front; SASZ, S Armorican Shear Zone; S-TZ, Sorgenfrei-Tornquist Zone; Su, Sudetes Mountains; TB, Tepla-Barrandian Basin (of BM); T-TZ, Teisseyre-Tornquist Zone; UM, Ukrainian Massif; VF, Variscan Front; WS, Windermere Supergroup.

Basin (Verniers *et al.* 2002a). Most of the Brabant Massif, except for the southwestern part, was folded, faulted and slightly metamorphosed during the Acadian orogeny and hence belongs to the Anglo-Brabant Deformation Belt (Van Grootel *et al.*, 1997, PACE TMR Network Team & Winchester, 2002).

The Brabant Massif contains a thick siliciclastic often turbiditic and rather complete Lower Palaeozoic sequence, from the lowest Cambrian to the uppermost Silurian. Its prolongation below the Devonian cover can be traced by boreholes and geophysical data to the north under the Campine Basin until the Ruhr Valley Graben, in the south at least under the northern half of the Namur Synclinorium. According to potential field images (Lee *et al.*, 1993; Pharaoh *et al.*, 1993) and a few boreholes the massif continues to the west under the North Sea into the concealed Caledonides of East-Anglia. The entire orogenic belt is now called the Anglo-Brabant Deformation Belt (PACE TMR Network Team & Winchester, 2002). It was first called the Caledonides of the Midlands and Brabant Massif (Verniers *et al.*, 1991), modified as the Anglo-Brabant Massif by Pharaoh *et al.* (1993) and the Anglo-Brabant fold belt by Van Grootel *et al.* (1997). It is the folded, faulted and weakly metamorphosed belt in the subcrop of East Anglia and of the Brabant Massif. Within the Brabant Massif three tectonic domains were recognised: (1) a southwestern undeformed domain, south of the Caledonian/Acadian deformation front (Sintubin, 1999), or structural area I in Debacker (2001), (2) a southern domain (Sintubin, 1997), called the Ordovician-Silurian domain in Sintubin (1999), and structural areas II and III in Debacker (2001) and (3) a northern domain (Sintubin, 1997b), also called Cambrian core domain (Sintubin, 1999), steep belt (Sintubin & Everaerts, submitted) or structural area IV in Debacker (*ibid.*). We will visit structural areas II and IV.

The Condroz Inlier (Dumont, 1847), previously called *Bande Sambre-et-Meuse* (Omalius d'Halloy, 1842), Sambre and Meuse strip, *Bande condruzienne*, Condroz ridge (Verniers & Van Grootel, 1991), is a long and narrow Ordovician-Silurian inlier, south of the Sambre and Meuse rivers and north of the Condroz. It contains Ordovician and Silurian siliciclastic sediments generally not turbiditic, except at Ombret, and mostly of a deeper shelf facies, the Condroz Shelf. Except for the Ombret area, they are present in tectonic wedges in a major central part and two smaller southern parts, the Oxhe Inlier in the east and the Puagne area in the southwest. They may have had different tectonic histories and are brought together by the Variscan thrust faults in the present day Condroz Inlier. We will visit the outcrops around the city of Fosses, in the western extremity of the Condroz Inlier, the central part and the Puagne area.

The four Ardennes inliers with Caledonian deformed Lower Palaeozoic strata are similar in composition: they contain a thick siliciclastic, often turbiditic sequence from the lowest Cambrian to the top of the Middle Ordovician. They are all situated in the Variscan Ardennes Allochton, previously called Dinant Nappe, thrusted from the south and south-east about ten to one hundred kilometres into their present position.

2. General palaeogeography

As shown by Cocks and Fortey (1982) benthic fauna and flora endemic to a continent can be used to distinguish palaeocontinents and their limits. They proved that in the Cambrian to Early Ordovician southern Britain, Ireland and eastern Newfoundland, together called Avalonia, were situated at high latitudes attached to Gondwana, whereas Scotland and western Newfoundland were at equatorial latitude attached to the North American continent, with in between the Iapetus Ocean. At temperate latitudes the Baltic continent was situated in between both other continents. They proved with fossils that in the Ordovician Avalonia moved from Gondwana in the direction of Baltica with the Tornquist Sea in between and in its wake the opening Rheic Ocean. Certainly from the late Caradoc, faunas from southern Britain were very similar to those from the Baltic continent. Llanvirn trilobites and middle Caradoc trilobites both at Oxhe in the Condroz Inlier (Dean, 1991) also indicate its affinity with Avalonia (Cocks *et al.*, 1997). Benthic faunal evidence of the first contact of our part of Avalonia with Baltica appears in the Ashgill of the Fosses Formation, where brachiopods, trilobites and cystoids show a North European affinity closest to Scandinavia and the Baltic area and a limited faunal relationship with Bohemia and the Armorican Massif (Regnell, 1951; Lespérance & Sheehan; 1987; Sheehan, 1987). The Fosses bryozoans are closest to Wales with some similarity with Baltica. Rugose and tabulate corals show affinities with the Baltic area and less obvious with Wales and northern England. Algae and corals indicate a tropical position (Tourneur *et al.*, 1993).

Trench and Torsvik (1991) and Torsvik *et al.* (1993) were the first to prove with palaeomagnetism that from the Cambrian till the Tremadocian Avalonia was attached to or close to northern Gondwana, at a high southern latitude. With sedimentological, palaeoecological and palaeontological arguments a position near the northern coast of South America is suggested by McKerrow *et al.* (1992). By the Llanvirn Avalonia was moving away from Gondwana, leading to the opening of the Rheic Ocean in between Avalonia and Gondwana (Prigmore *et al.*, 1997). By the Late Ordovician a position of Avalonia close to Baltica is deduced with an inferred closure of the Tornquist Sea. In the Silurian Avalonia is very close to Baltica and later also to Laurentia forming by Middle Devonian the Laurussia palaeocontinent (see Tait *et al.*, 1997 and Fig. 3).

Hence, the Lower Palaeozoic history of Belgium has to be viewed in three very different geodynamic periods (Verniers *et al.*, 2002a). (1) In a first period from earliest Cambrian to late Tremadocian Belgium was attached to the large continent of Gondwana. First on a shelf and soon in deep basins thick siliciclastic sediments are deposited in extensional basins. The mid- or end Tremadocian to middle Arenig hiatus, observed all over the country, corresponds with the rifting event of the microcontinent Avalonia away from Gondwana. (2) The second period corresponds to the drift of Avalonia on its own towards Baltica, from middle Arenig till

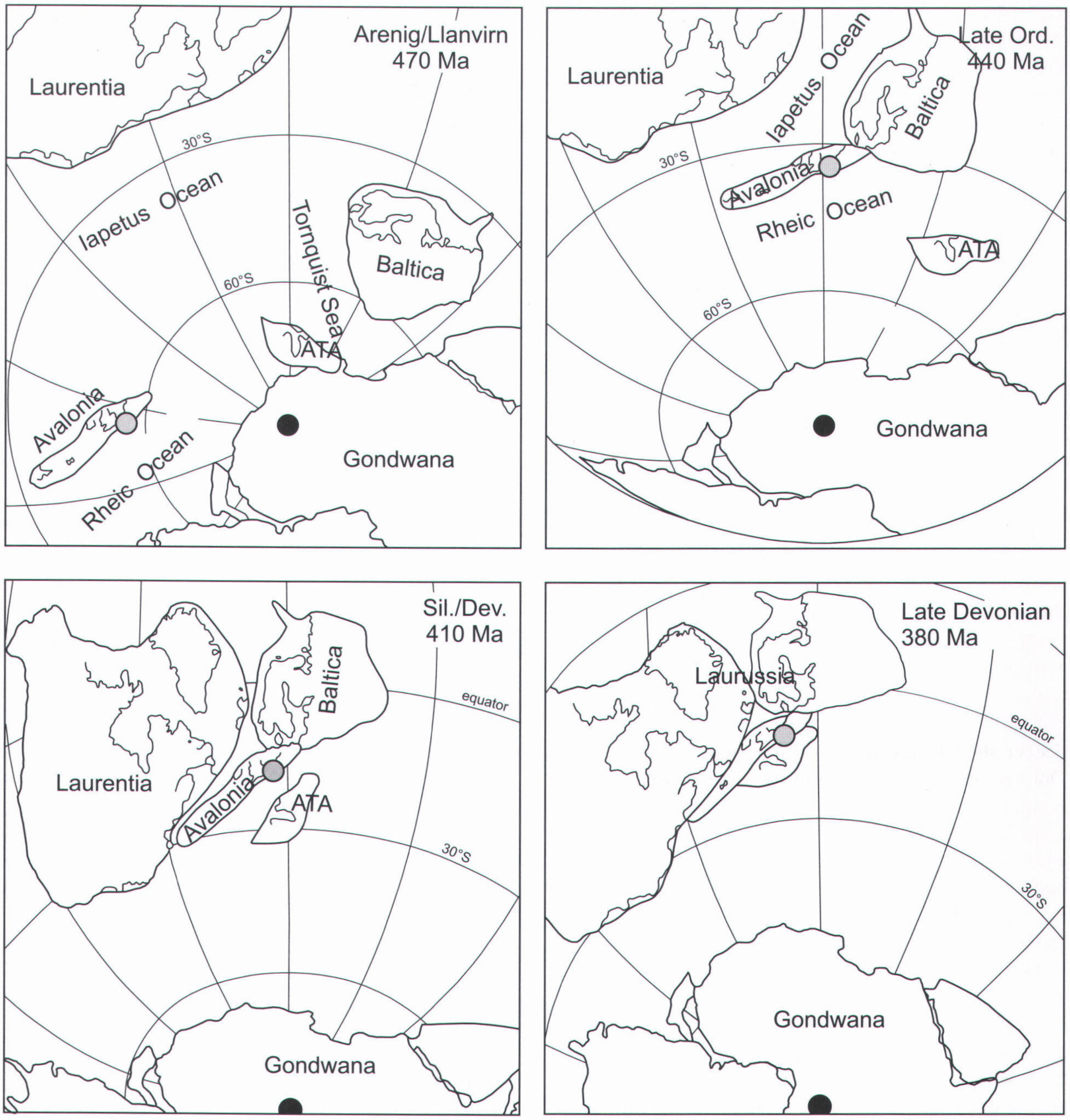

Figure 3. Palaeogeographical reconstruction for the Arenig/Llanvirn, Late Ordovician, Silurian/Devonian and Late Devonian, with position of the palaeocontinents and the palaeocontinents. ATA: Armorican Terrane Assembly. The black circle indicates the position of the south pole and grey circle the location of Belgium (modified after Tait *et al.*, 1997).

middle Ashgill. Sedimentation is on a shelf and not very active, reflecting the limited source area of the micro-continent. By the end of the second period the Ardenne inliers are folded, cleaved and weakly metamorphosed during the Ardenne deformation phase, situated in the sedimentation gap between the early Caradoc and latest Silurian. (3) The third period marks the collision with Baltica and the closure of the Tornquist Sea with sediments passing the suture by middle Ashgill (Samuelsson *et al.*, 2001) and later with Laurentia where sediments start to pass the suture of the Iapetus Ocean in the middle Wenlock (Kneller *et al.* 1993). This third period starts in the late Ashgill with sedimentation continuing till the Pridoli. During the third period a deep shelf is present all the time in the central and southern parts on the Condroz Inlier. At the end of the Llandovery in the central and north Brabant Basin of the Brabant Massif, a foreland basin develops, with very thick Silurian turbiditic formations. Most of the Brabant Massif was deformed during the long-lived

Brabantian deformation phase (part of the larger Acadian Phase) from Silurian in places, continuing till the (early) Eifelian (Debacker 2001; Verniers *et al.*, 2002a).

Subduction-related magmatism is observed from middle Caradoc till early Llandovery (see André, 1991) in the Brabant Massif, which is linked with a rather short-lived subduction of oceanic crust in the first Avalonia-Baltica collision, at the end of the second period and the beginning of the third period. The magmatism ended in the Brabant Massif before the foreland basin started to develop (Verniers *et al.*, 2002a).

3. Cambrian-Ordovician of the Dyle Basin (DAY 1) and Orneau Valley (DAY 2)

3.1. Introduction

The lithostratigraphy of the Lower Palaeozoic in the Brabant Massif, the Condroz and Ardennes inliers has recently been revised by the Belgian Lower Palaeozoic Stratigraphical Subcommission (Verniers *et al.,* 2001) (see Fig. 4).

We will be specifically concerned in this excursion with the only outcropping zone located in the southern rim of the Brabant Massif and in particular the stratigraphy of the Senne, Dyle and Orneau valleys (Figs. 4 & 5). This has been the subject of active investigation by several research teams over the past ten years. This has given rise to: (1) the adoption of an unified lithostratigraphy per massif or inlier, whereas until recently the formation names varied from valley to valley (Legrand, 1968; André *et al.*, 1991; Servais *et al.*, 1993; Verniers *et al.,* 2001); (2) the lithostratigraphic column can be considered as rather complete and comprises five Cambrian and nine Ordovician formations; (3) most of these formations are well dated at the present time. The most recent progress in dating these rocks has been through the use of chitinozoans for the Ordovician (Van Grootel *et al.*, 1997; Verniers *et al.*, 1999; Samuelsson & Verniers, 2000) as well as the revision of graptolites (Maletz & Servais, 1998; Maletz, 1999); (4) the only large hiatus observed is between the Chevlipont and Abbaye de Villers formations and probably extends from the middle part of the Tremadocian (Vanguestaine, pers. comm.) to the upper part of the middle Arenig (Samuelsson & Verniers, 2000, Figs. 4 & 5).

The sedimentology was studied in a somewhat piecemeal fashion by the research team of A. Herbosch (Vander Auwera & André, 1985; Jodard, 1986; Lenoir, 1987; Herbosch, 1991; Herbosch *in* André *et al.,* 1991; Herbosch, 1996; Herbosch & Lemonne, 2000). The following introduction (see 3.2) integrates this work and gives a much more complete sedimentological synthesis. It attempts to interpret the depositional environments from the Lower Cambrian (Tubize Formation) to the base of the Upper Ordovician (Ittre Formation).

This basic research has both permitted and stimulated the mapping of the southern limit of the Brabant Massif on a scale of 1/25,000, mapping which has almost been completed (seven 1/25,000 scale maps covering the Senne, Dyle and Orneau basins). A synthesis for the Dyle basin is given in Chapter 3.3 and Figures 6, 7 and 10.

3.2. Stratigraphy and sedimentology

Along the Dyle river and its tributaries such as the Thyle and Orne rivers (Fig. 6), between Wavre to the north and Sart-Dames-Avelines/Tilly to the south, outcrops can be seen with ages from earliest Early Cambrian (Blanmont Formation, Figs. 5 & 6) to early Late Ordovician (Ittre Formation, Figs. 5 & 6). These outcrops will be visited on this field trip. However the oldest beds (Blanmont Formation) will not be seen as there are no longer any good outcrops.

It is also important to emphasise here that the Oisquercq Formation (uppermost Lower Cambrian to Middle Cambrian, Fig. 4) is not observed in the Dyle basin whereas it outcrops extensively in the Senne basin. This is an old problem that has not yet been explained in a thoroughly satisfactory manner (see discussion in 3.3.3).

3.2.1. Blanmont Formation (Malaise, 1873)

This formation is poorly studied. Its outcrops are dispersed, small in size, with large distances between them and occurring principally in numerous abandoned quarries now under water. It occurs always stratigraphically below the Tubize Formation and towards the centre of the Brabant Massif.

Description. Mostly light coloured, whitish, bluish, or greenish, massive, fine to coarse-grained quartzite. Stratification is not well marked, except where thin intercalated beds of grey or green siltstone and slate occur. Earlier descriptions mention oblique stratification in coarse-grained arkose and also fine-grained conglomerate. The upper boundary with the Tubize Formation and the lower boundary with the Jodoigne Formation is nowhere observed.

Sedimentology. No recent study.

Thickness. Estimation not possible.

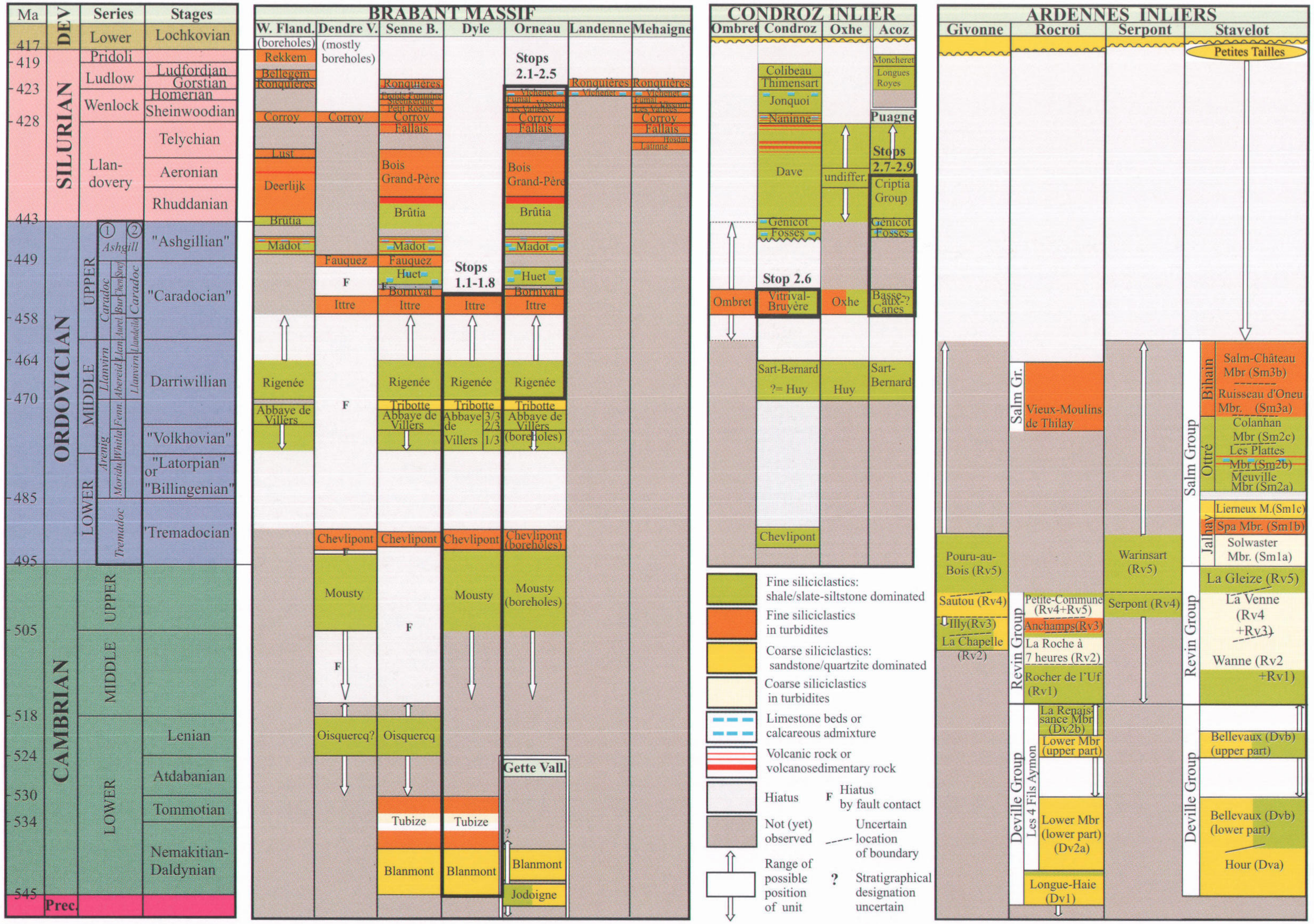

Figure 4. Chronostratigraphical position of the Lower Palaeozoic lithostratigraphic units of Belgium. The fat rectangle in the Series column of the chronostratigraphy shows under 1 the British Ordovician chronostratigraphy revised by Fortey *et al.* (1995) and under 2 the position of the Llanvirn and Llandeilo stages, before the latter was abolished in 1995. Abbreviations: Prec.: Precambrian; Moridu.: Moridunian; Whitla.: Whitlandian; Fenn.: Fennian; Abereid.: Abereiddian; Llan.: Llandeilian; Aurel.: Aurelucian; Bur: Burrellian; Chen: Cheneyan; Stref: Streffordian; Dev: Devonian. Visited units are marked in black squares (after Verniers *et al.*, 2001).

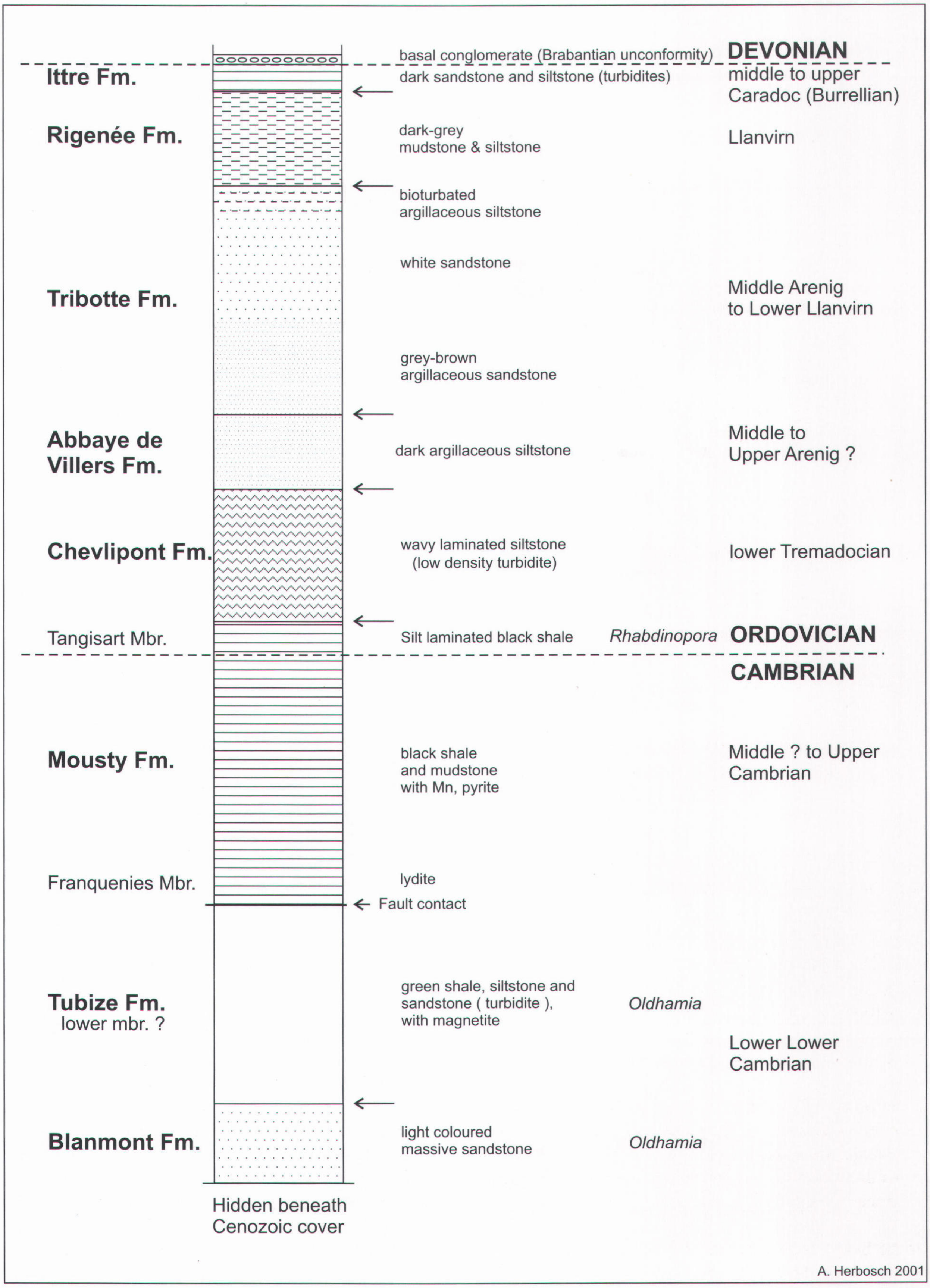

Figure 5. Stratigraphical log of the Dyle basin. Not to scale.

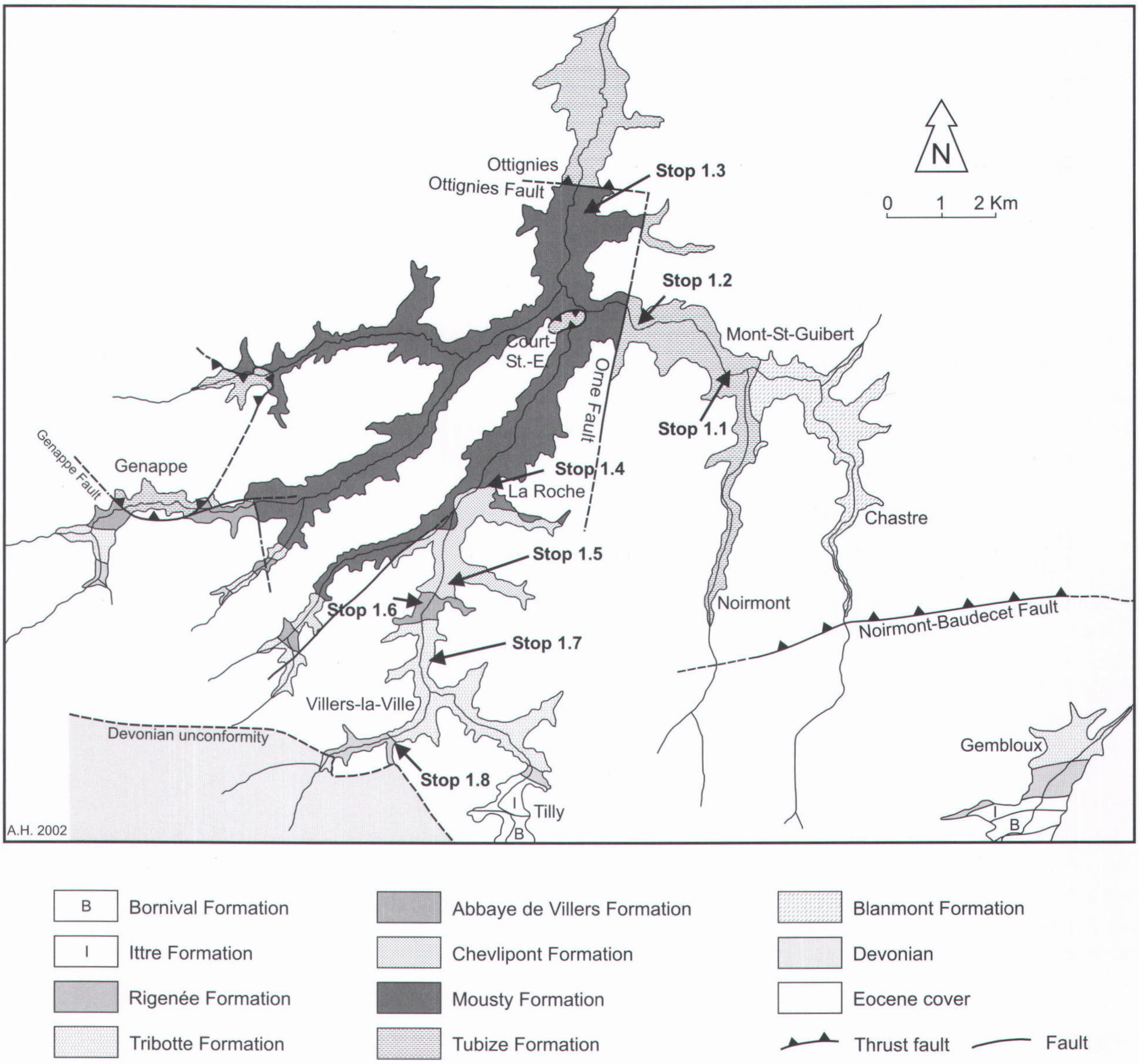

Figure 6. Geological map of the Lower Palaeozoic of the Dyle basin and northern part of the Orneau river (Gembloux).

Age. The presence of the ichnofossil *Oldhamia* suggests an early Early Cambrian age (cf. Tubize Formation).

3.2.2. Tubize Formation (Malaise, 1873)

Description. This formation is essentially made up of mudstone and siltstone, but sandstone, arkose as well as greywacke are also important. It is easily recognisable by its dominant grey-green colour and by the frequent presence of magnetite.

In the Senne basin, the middle and upper parts of the formation can be clearly observed and in particular the middle part of the Rogissart Member (Vander Auwera & André, 1985), which is very characteristic. It contains light coloured often greenish sandstone, feldspathic sandstone, arkose, greywacke in decimetric to metric beds, alternating with green siltstone and slate forming upward fining sequences. Magnetite is often present, mostly in the siltstone and slate. The coarser beds are often graded and the sandstone and siltstone show plane, oblique and convolute laminations. All these structures are characteristic of Bouma sequences. The thick and coarse-grained sequences (1-2 m and 3-4 mm) at the base of the Rogissart Member seem to evolve upward into progressively thinner and finer-grained sequences. The upper part, the Les Forges Member (new name in Herbosch *et al.,* in prep. b, Ittre-Rebecq map sheet) is mostly formed of homogeneous to zoned mudstone and siltstone, but often with millimetre to centimetre thick laminations, grey-green in colour or also dark grey to grey-blue (a distinctive characteristic colour of this member). Mostly complete decimetric Bouma sequences in series or isolated within shale (slate) or siltstone continue to be observed. Magnetite remains frequent in this member.

In the Dyle basin, whether towards Genappe to the west or towards Mont-St-Guibert and Ottignies to the east, the same general characteristics are found, green colour and the occurrence of magnetite, but overall the rocks are finer-grained and more argillaceous. Neither the coarse-grained arkose and greywacke of the thick Rogissart Member nor the sometimes grey-blue finer-grained rocks of the Les Forges Member are to be found here. It is possible that we are stratigraphically in a lower member (the Blanmont Formation occurs near by, Fig. 6), composed principally of green shale (slate) and siltstone that are either massive or show varied grey-green laminations with frequent magnetite. Zones of sandstone, siltstone and mudstone forming decimetric Bouma sequences similar to those seen at Mont-Saint-Guibert (stop 1.1) can be observed locally.

Sedimentology. The sedimentation of this thick formation is generally of the clastic deep sea-type environment (Reading, 1978), oscillating between pelagic (?) to hemipelagic sedimentation and high density turbidites.

The lower member is composed essentially of pelagic to hemipelagic mudstone and siltstone intercalated with several high density turbidite sequences. Some of the mudstone or claystone can be interpreted as true pelagic deposits.

The rhythmic sedimentation of the Rogissart Member is interpreted as Bouma type high density turbidite sequences (Vander Auwera & André, 1985; Herbosch, unpublished data). Most of the determining criteria are present: rhythmic sedimentation, grading of the intervals and of the sequences, thickness of the sequences (metric), sedimentary structures associated with each of the intervals, basal structure, lateral continuity and total thickness of the member (about 1000 m).

A large scale vertical evolution can be observed: the transition is abrupt from the shale and siltstone (slate) of the lower member (which does not outcrop in the Senne basin) to the overlying thick (>1 m) and coarse grained (3-4 mm) sequences that form the base of the Rogissart Member. These sequences are incomplete (often only showing the a interval), frequently amalgamated and with very little siltstone and shale. The higher sequences become thinner (tens of cms to 1 m), more complete, with occasional recurrences of coarse-grained layers. Then hemipelagic siltstone and shale episodes begin to appear, intercalated between the turbidite-like episodes, becoming more and more frequent.

When we enter the upper Les Forges Member, the hemipelagic episodes show numerous types of lamination, some of which demonstrate typical characteristics of high density distal turbidites (cm thick sequences with plane and or convoluted laminations) while others show typical characteristics of low density turbidites (numerous mm laminations with silty bases in shale, Fig. 16).

Thickness. Difficult to estimate but certainly more than 2 to 3 km. The Rogissart Member on its own is estimated at about 1000 m (Herbosch *et al.*, in prep. b, Ittre-Rebecq map sheet).

Age. An Early Cambrian age is proposed as the only fossil present is the trace fossil *Oldhamia* (Malaise, 1883), a genus which occurs in the Lower to Middle Cambrian, but not below the Cambrian-Precambrian boundary (Crimes, 1992; Verniers & De Vos, 1995). According to new observations by A. Seilacher (pers. comm., 1998) the *Oldhamia* sp. is more restricted in time and only present in the Tommotian or Nemakit-Daldynian (early Early Cambrian), an age proposed for the formation.

3.2.3. Oisquercq Formation (Malaise, 1873)

Its presence in the Dyle basin is disputed but the recent survey of the Chastre-Gembloux map (Delcambre *et al.*, 2002) definitively demonstrates its absence. Hence only a brief description will be given of this formation, which we will not see.

Description. The lower Ripain Member is made up of extremely homogeneous and fine-grained blue-grey slate (claystone). Stratification is rarely visible. The Asquempont Member forms the upper part of the formation, and is made up of greyish or greenish slate (claystone to mudstone) and siltstone in the upper part. The transition between the two members is gradual over about ten meters and consists only of a change in colour from blue to green.

Age. Based on acritarchs in the upper member from the Lessines, Oudenaarde, Eine and Kruishoutem boreholes: latest Early Cambrian to Middle Cambrian (Vanguestaine, 1992).

3.2.4. Mousty Formation (Malaise, 1883, 1900)

This formation outcrops widely in the Dyle basin (Fig. 6) but is completely absent in the Senne basin (see discussion 3.3.3).

Description. Shale or slate, sometimes mudstone, grey-blue to grey-black in colour, graphitic and pyritic. Often massive (without visible stratification) or more rarely finely laminated as in typical black shale, where rhythmic variations in clay and organic matter content occur. Stratification is also frequently marked by light or greenish coloured more silty beds or laminae, or by banded colour variations. Sometimes grey, more or less clayey siltstone with pyrite occurs, and occasionally centimetric to decimetric fining upward sandstone or siltstone layers, interpreted as distal turbidites. The shale is frequently manganiferous (de Magnée & Anciaux, 1945), which is marked in thin section by the presence of Mn-garnet, Mn-ilmenite and Mn-chlorite porphyroblasts (Jodard, 1986; André *et al.*, 1991).

The very thick and monotonous shaly formation shows only two characteristic levels: the Franquenies Member, in the lower part of the formation, shows siliceous beds or lenses of lydite within typical black shale; the Tangissart Member, at the top of the formation, is characterised by increasingly recurrent intervals of dark shale with abundant millimetric light coloured silty laminae. The disappearance of the last black shale intervals marks the boundary with the overlying Chevlipont Formation.

The lower boundary is not observed, only faulted contacts with the underlying Tubize Formation are present.

Sedimentology. The very argillaceous character of the sedimentation, its monotony and its considerable thickness, the sporadic presence of fine grained distal turbidite sequences demonstrate that we are dealing here with pelagic to hemipelagic shales deposited in a deep marine environment far from direct terrigenous sources. The abundance of organic material and of pyrite, the conservation of delicate bed structures or of zoned variations in colour, the high levels of manganese are all characteristic of an anoxic and calm environment. The lydites observed in the Franquenies Member, whose microscopic characteristics indicate that they are probably radiolarites, also indicate a pelagic environment. The presence of episodic distal turbidites (cm to dm in thickness), of more silty pyritic shale or even siltstone with pyrite, as in the Court-St-Etienne borehole (Herbosch, Delcambre & Pingot, unpublished data), demonstrates, however, that we are never very far from the continent.

The Tangissart Member marks a progressive transition to the Chevlipont Formation. The recurrent intervals of shales with silty millimetre thick laminations can be interpreted as low density, very distal turbidites (Stow & Shanmugam, 1980; Stow & Piper, 1984; Stow, 1986) (Fig. 16) deposited in a shallower environment or one closer to terrigenous sources, as for example turbidite plains. The member marks the beginning of a shallowing which continues at an accelerated rate in the overlying Chevlipont Formation.

Thickness. Difficult to estimate; more than 500 m is tentatively suggested.

Age. Middle Cambrian to earliest Ordovician. Acritarchs from the upper part of the Mousty Formation, below the Tangissart Member, studied by Martin (1976), Vanguestaine *et al.* (1989) and Vanguestaine (1992 and unpublished data) indicate a Middle or Late Cambrian age. Graptolites (*Rhabdinopora* sp.) and acritarchs in the Tangissart Member prove the early Tremadocian age of that member (Lecompte, 1948, 1949).

3.2.5. Chevlipont Formation (Anthoine & Anthoine, 1943)

Description. Grey siltstone (called "quartzophyllade" in older literature), with characteristic wavy bedding consisting of rhythmic alternations (mm to cm) of light grey siltstone laminae and dark grey clayey siltstone and mudstone laminae. Each of these centimetric rhythmic sequences is graded. Silty laminae occur frequently in small lenses, a few cm long and a few mm thick, with oblique lamination and load structures. These wavy silty laminae occur at the base of the most complete sequences. Millimetre-sized horizontal bioturbation (*Planolites*) is observed in the clayey laminae at the top of the sequence. This facies is regularly interrupted by centimetric to decimetric fine sandstone beds, with massive, plane parallel or convolute structures. These beds are quite continuous and are also present in the Lessines borehole (Herbosch *et al.*, 1991). In this borehole frequent slumping and decimetric intraformational breccias are observed.
The dominant facies, can sometimes be replaced by typical Bouma sequences, tens of cms in thickness (Dyle basin, Herbosch & Lemonne, 2000; Marcq area, Debacker, 1999). The contact with the overlying Abbaye de Villers Formation is nowhere observed in the Brabant Massif.

Sedimentology. Study of the outcrops and boreholes allows the interpretation of this facies as a low density turbidite sedimentation typical of "mud turbidites" (Stow & Shanmugam, 1980; Stow & Piper, 1984; Stow, 1986). The laminar siltstone is the result of the repetition of incomplete Stow sequences-model (Fig. 16) whose lower intervals (T0 to T4), and in particular the silty basal lamination with ripples (T0), are well developed in comparison to the upper argillaceous intervals (T7 is almost always absent). Several other arguments support this interpretation: the frequent presence in boreholes of slumps and intraformational breccias (palaeoslope); the systematic occurrence (in boreholes and in outcrops) of thin sandy beds interpreted as high density distal turbidite episodes; the observation of occasional high density turbidite episodes; the sedimentary continuity with the pelagic shales of the Mousty Formation (appearance of shales with silty laminations in the Tangissart Member); the absence of benthic fauna and the presence of planktonic fauna (acritarchs, graptolites). The depositional environment is always deep, relatively reducing (pyrite), with a slightly increased slope (slumps), probably closer to the continental slope although one can not be any more precise.

Sedimentological comparison with Stavelot. It is important here to note that exactly the same sedimentological succession is observed in the Stavelot Massif (Fig. 4). The black shales of the Rv5, which mark the top of the Cambrian, are overlain by high density turbidites of the Sm1a followed by low density turbidites of the Sm1b (Lamens, 1985). The overlying Sm1c which represents the upper part of the Tremadocian (Vanguestaine, 1992) and which is not present in the Brabant Massif, shows a shelf facies. For Lamens (*ibid.*), this sequence is thought to represent the filling of an epicontinental basin by a northward prograding clastic wedge characterised by a high sedimentation rate.

Thickness. On the order of 150-200 m in the Dyle valley; at least 92 m in the Lessines borehole (Herbosch *et al.*, 1991).

Age. Early Tremadocian, for the lower part of the formation. This is based on the presence of the dendroid graptolites *Rhabdinopora flabelliformis* ssp. *socialis* and *typica* (Lecompte, 1948, 1949), as well as acritarchs (Martin, 1969a, 1969b, 1976; Vanguestaine *in* André *et al.*, 1991). In contrast to the Stavelot Massif where acritarchs of the upper Tremadocian are present (Vanguestaine, 1992), only acritarchs of the lower Tremadocian seem to be present in the Brabant Massif (Vanguestaine, pers. comm.).

3.2.6. Abbaye de Villers Formation (Anthoine & Anthoine, 1943)

Description. Grey to dark grey argillaceous siltstone to mudstone, with distinct laminar to lenticular bedding. This laminar structure is marked either by sets of numerous mm thick silty laminations or by lighter coloured silty-sandy beds which interrupt the argillaceous sedimentation. The silty beds or laminae, with frequent load structures, are sometimes continuous, but are more often lenticular, representing low amplitude ripples. Some beds, particularly argillaceous and rich in organic material, contain pyrite. The abundance of mm size micas (illite-chlorite stacks in thin section) is also very characteristic.

These rocks have experienced bioturbation of variable intensity which disturbs or even effaces the laminar structure. The bioturbation is stronger horizontally than vertically, the horizontal burrows typically form flattened light-coloured nodules mms to cms in size. Vertical burrows, cms in size and with *spreiten*, are more rarely observed. Oblique stratification on the order of meters occurs quite frequently (Fig. 17).

The boundary with the overlying formation is progressive and is essentially marked by an increase in sand (increase in grain size and reduction of clays), as well as an increase in bioturbation and a change from dark colours to lighter colours (grey-brown to beige). These changes take place principally at the base of the Tribotte Formation.

Sedimentology. The alternation of silty-sandstone beds and clay beds implies significant periodic variations in the energy of the depositional environment, conditions that are characteristic of a continental shelf. The dominantly argillaceous character demonstrates that the depositional environment is certainly below the fair weather wave base. This bathymetry corresponds to an internal shelf and is compatible with the observed strong bioturbation in an essentially horizontal direction, with the strong tendency towards dysoxic conditions (dark colour, pyrite) and finally with the presence of submarine dunes with oblique stratification (Fig. 17). There is a significant sedimentological break between the Chevlipont and Abbaye de Villers Formations as is indicated by the abrupt change in depositional environment from a deep marine environment probably situated close to turbidite plains or continental slope (Chevlipont Formation) to a continental shelf (Abbaye de Villers Formation).

Thickness. Between 100 and 150 m in the Dyle area.

Age. No macrofossils are found. Chitinozoans in the lower third of the formation in the Dyle basin include *Eremochitina brevis* (Samuelsson & Verniers, 2000). The same assemblage occurs in the Grès Armoricain Formation in Brittany, and indicates a middle Arenig, Whitlandian (*pro parte*) or possibly late Arenig age (Paris, 1981). Acritarchs in the middle and upper part of the formation indicate a late Arenig or post-Arenig age according to Martin (1976) and Vanguestaine (*in* André *et al.*, 1991), which is corroborated by the chitinozoans (Samuelsson & Verniers, 2000) (Figs. 8 & 9). One of the genera present, *Frankea*, does not appear below the uppermost Whitlandian, top middle Arenig, in levels of the upper part of the *Isograptus gibberulus* graptolite Biozone, according to Servais (1993) and Brocke *et al.* (1995) or of the *Expansograptus hirundo* graptolite Biozones, upper Arenig.

An important time gap is hence present between the Chevlipont (early Tremadocian) and the Abbaye de Villers (middle-late Arenig) Formations (Figs. 4 & 8).

3.2.7. Tribotte Formation (Anthoine & Anthoine, 1943)

Description. In the Dyle basin, this formation shows three main lithofacies easily recognisable in the field. The lower third contains brownish grey, clayey sandstone and siltstone with coarse laminations and strong bioturbation (mainly horizontal). Some beds show oblique stratification on the scale of meters as in the Abbaye de Villers Formation. Thin section examination shows abundant potassic feldspar and plagioclase in the sandstone at the base of the formation (Jodart, 1986; Herbosch *in* André *et al.,* 1991).

The upper two thirds show yellowish grey to greenish grey sandstone and siltstone, clearly more clayey than the lower part of the formation. Bioturbation is relatively strong with principally oblique to vertical burrows (*Fucoides* in older literature). Burrows that are centimetric in length and millimetric in diameter are more abundant that the large burrows (centimetric in diameter) with *spreiten*. Clayey siltstone with flaser structures is frequently observed along with discontinuous clayey laminations and ripple mud drapes, the whole being affected by strong bioturbation. A relatively mature, bioturbated, yellowish grey, sandstone can be observed locally ("Strichon sandstone and psammite", Anthoine & Anthoine, 1943). The transition to the overlying formation is very rapid and abrupt.

Sedimentology. At the base of the formation there is a progressive evolution away from a shelf environment situated beneath the wave base, the depositional environment of the Abbaye de Villers Formation, towards a more oxygenated (lighter coloured rocks), more agitated (sand) environment, probably situated close to the fair weather wave base (there are still clays). The arkosic episode at the base of this zone still has not been explained sedimentologically or geodynamically. The Strichon sandstone probably corresponds to the shoreface, and the upper part of the formation indicates an intertidal environment characterised by silto-argillaceous rocks that have experienced strong vertical bioturbation.

Thickness. 200 to 300 m in the Dyle basin.

Age. Neither macrofossils nor acritarchs are observed. A poor chitinozoan assemblage in the uppermost part of the formation containing *Euconochitina vulgaris* indicates a middle Arenig to early Llanvirn age (Verniers *et al.*, 1999; Samuelsson & Verniers, 2000) (Figs. 8 & 9).

3.2.8. Rigenée Formation (Malaise, 1909a, b)

Description. Dark grey to bluish grey monotonous slate (mudstone), laminated or massive, locally bearing pyrite. The basal siltstone is more laminar and has experienced bioturbation. The lower boundary of the unit marks a rapid and sharp upward change (over about 10-20 metres; Servais, 1993; Herbosch & Lemonne, 2000) from light clayey and bioturbated siltstone to dark laminated siltstone and mudstone.

Sedimentology. The sedimentology has not yet been systematically studied. The rapid passage from the light coloured silto-argillaceous rocks of the Tribotte Formation into the dark laminar siltstone of the Rigenée Formation can only be explained by a rapid increase in water depth. Indeed, we move from an intertidal environment to an environment situated at least beneath the wave action zone (if not deeper). The overlying Ittre Formation can clearly be interpreted as a turbidite environment. Thus it can be deduced that the monotonous mudstone of the Rigenée Formation probably resulted from a relatively rapid event marking the transition from a shallow shelf environment to a deeper environment close to the continental slope.

Thickness. 150 to 200 m in the Dyle and Orneau valleys.

Age. Probably Llanvirn (Figs. 8 & 9). Graptolites in the lower or middle part of the formation in the Sennette valley belong to the lower Llanvirn *Didymograptus artus* Biozone according to Martin and Rickards (1979). Additionally, following a new examination of the fauna by Maletz and Servais (1998) these graptolites have been attributed to the *Didymograptus artus* and the *Didymograptus murchisoni* Biozones, corresponding to the entire Abereiddian (lower Llanvirn). Acritarchs from the Dyle basin indicate an (early) Llanvirn age (Martin, 1969a). Acritarchs studied by Servais (1993) indicate that for the base of the formation a late Arenig or younger age cannot be excluded. Higher levels of the formation seem to be (at least) late Llanvirn in age if not younger as indicated by the presence of *Frankea hamulata*, a species not found in rocks older than upper Llanvirn. A poor assemblage of chitinozoans with *Lagenochitina obeligis* and *Cyathochitina calix* indicates the same extended age bracket (Verniers *et al.,* 1999; Samuelsson & Verniers, 2000) (Figs. 8 & 9).