# 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

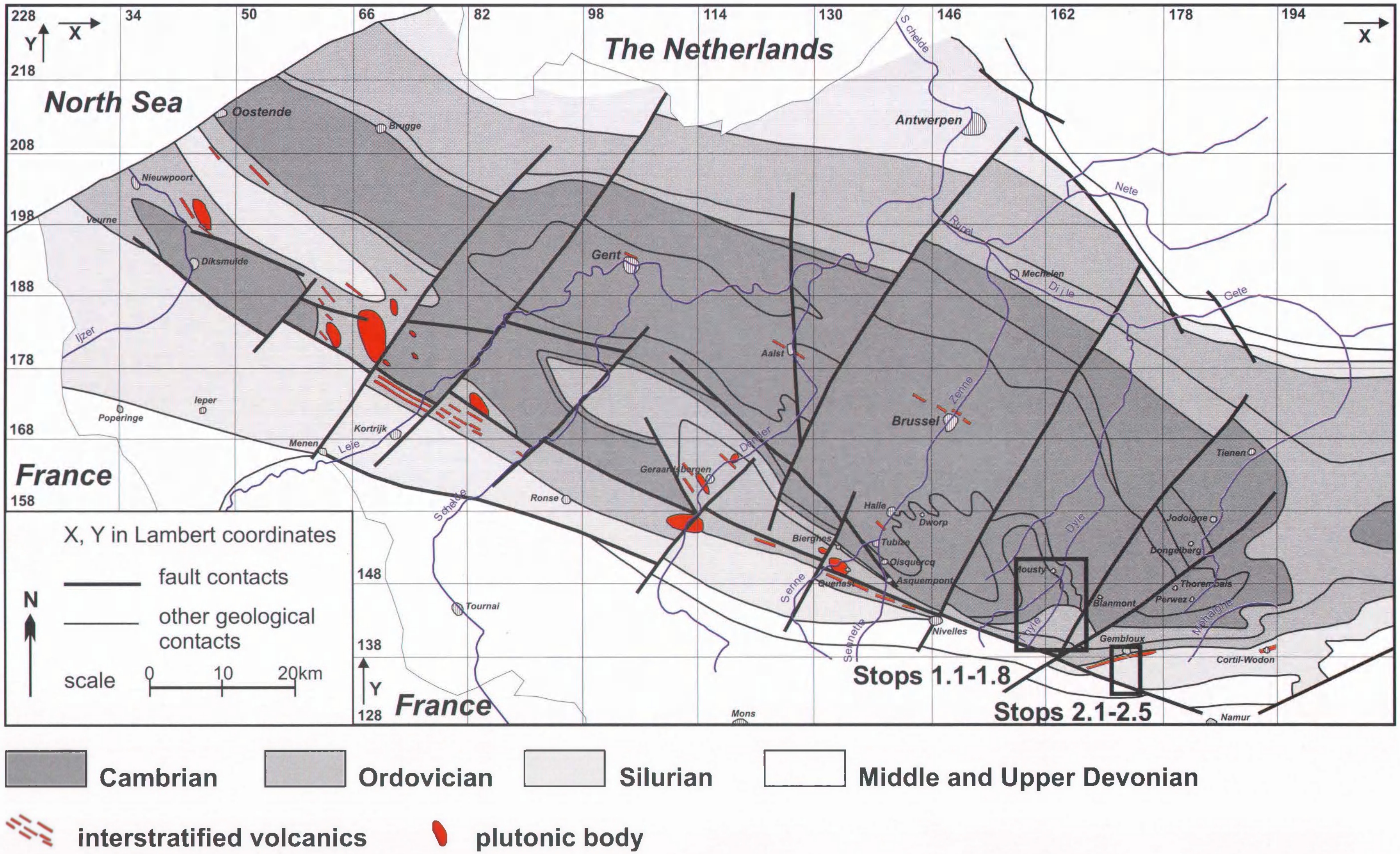


**Figure 1**. Geological subcrop map of the Lower Palaeozoic Brabant Massif with location of the stops (redrawn after De Vos *et al.*, 1993).

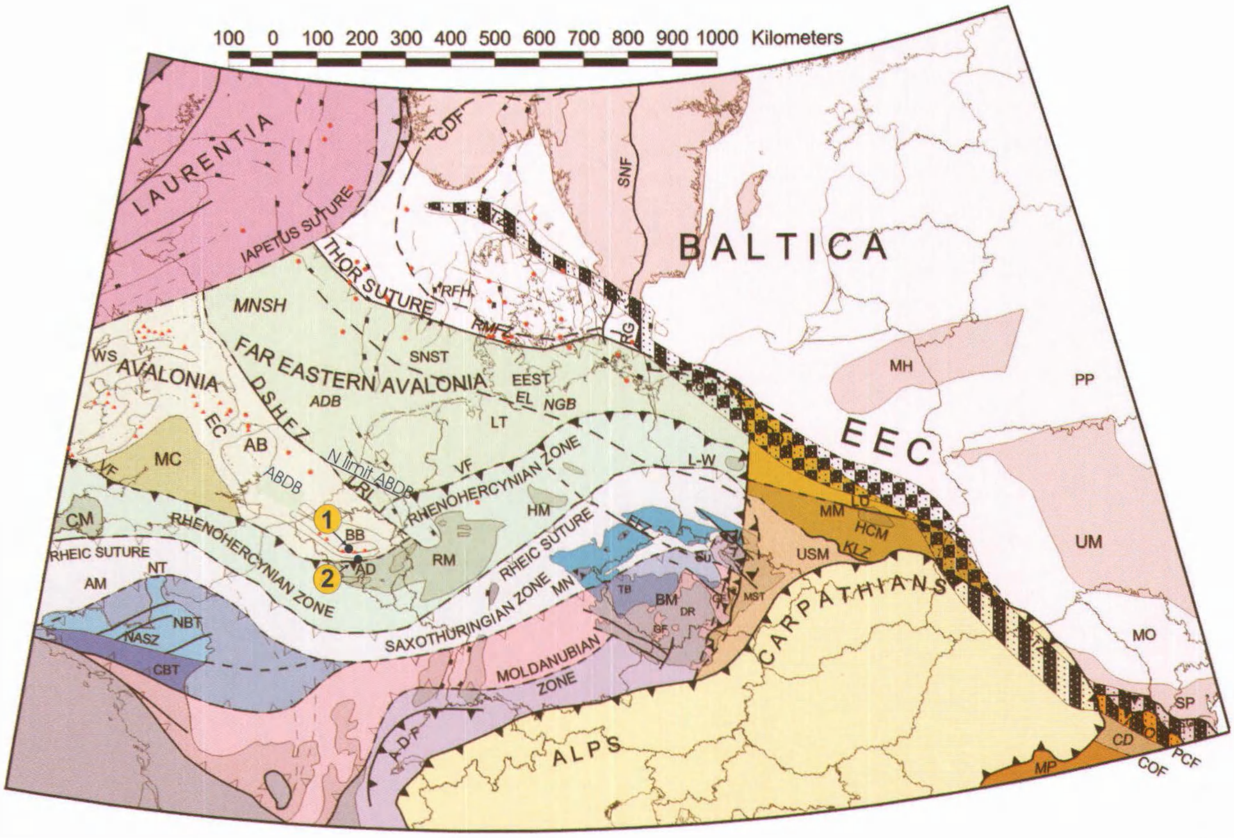


**Figure 2**. Basement tectonic sketch map of NW Europe after PACE TMR Network Team & Winchester (2002). Oceanic sutures, open ticks; orogenic frontal zones, filled ticks; key boreholes, solid dots; Ordovician arc volcanic rocks in Avalonia, triangles. Geographic locations mentioned in text: Liz, Lizard; 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

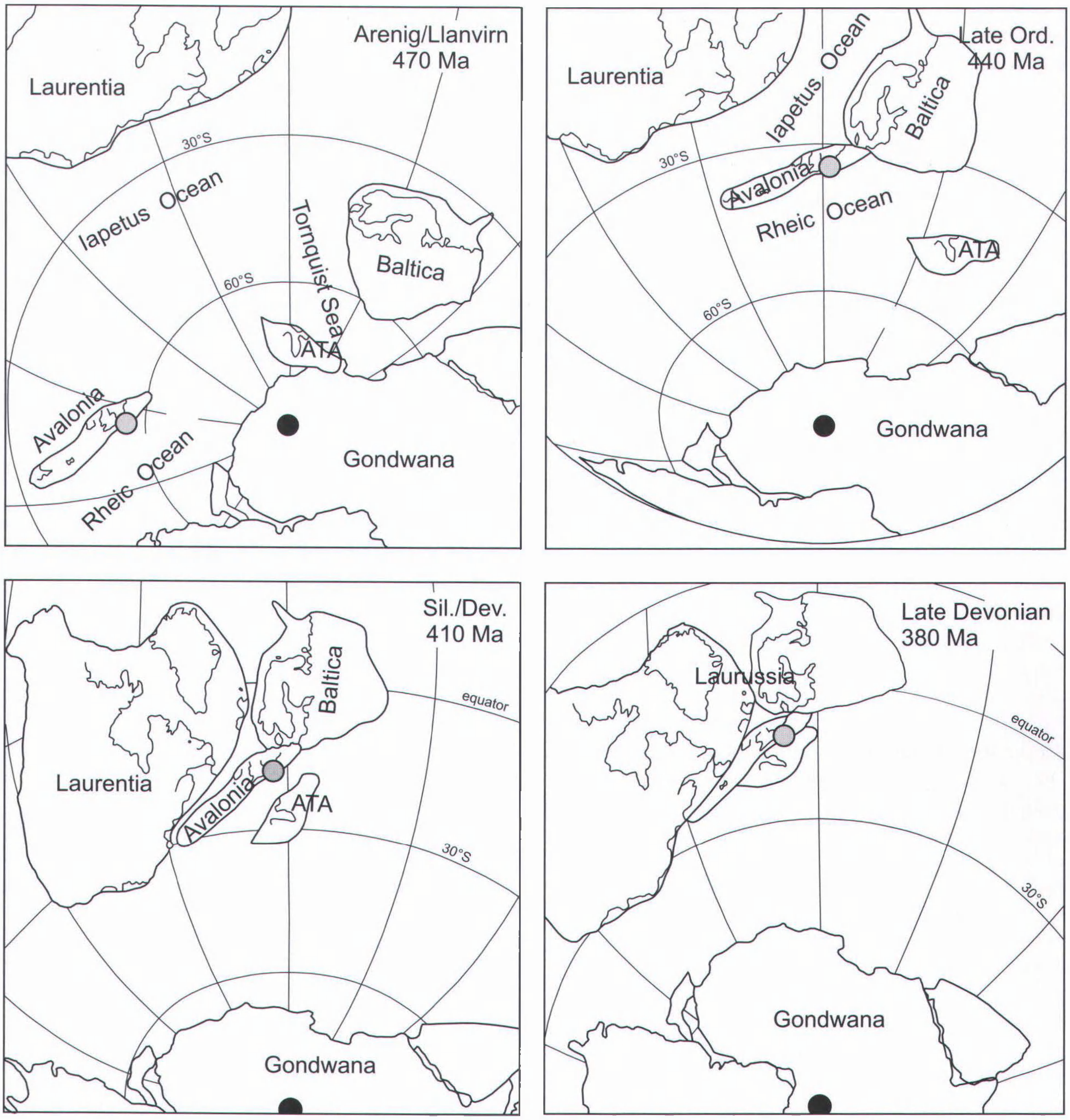


**Figure 3**. Palaeogeographical reconstruction for the Arenig/Llanvirn, Late Ordovician, Silurian/Devonian and Late Devonian, with position of the palaeocontinents and the palaeocontinents. ATA: Armorican Terrane Assembly. The black circle indicates the position of the south pole and grey circle the 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 microcontinent. 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.

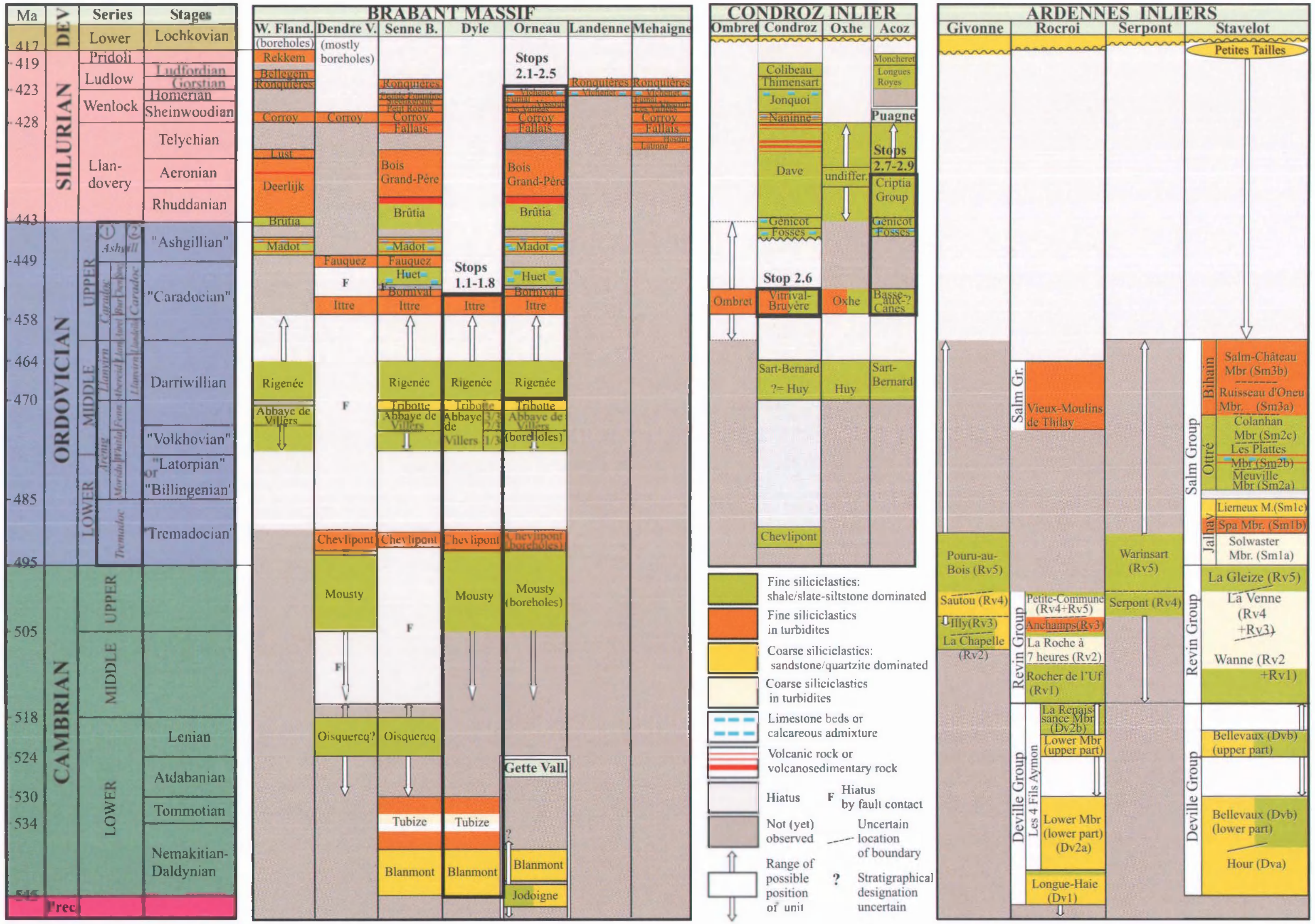


**Figure 4**. Chronostratigraphical position of the Lower Palaeozoic lithostratigraphic units of Belgium. The fat rectangle in the Series column of the chronostratigraphy shows under 1 the British Ordovician chronostratigraphy revised by Fortey *et al.* (1995) and under 2 the position of the Llanvirn 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).

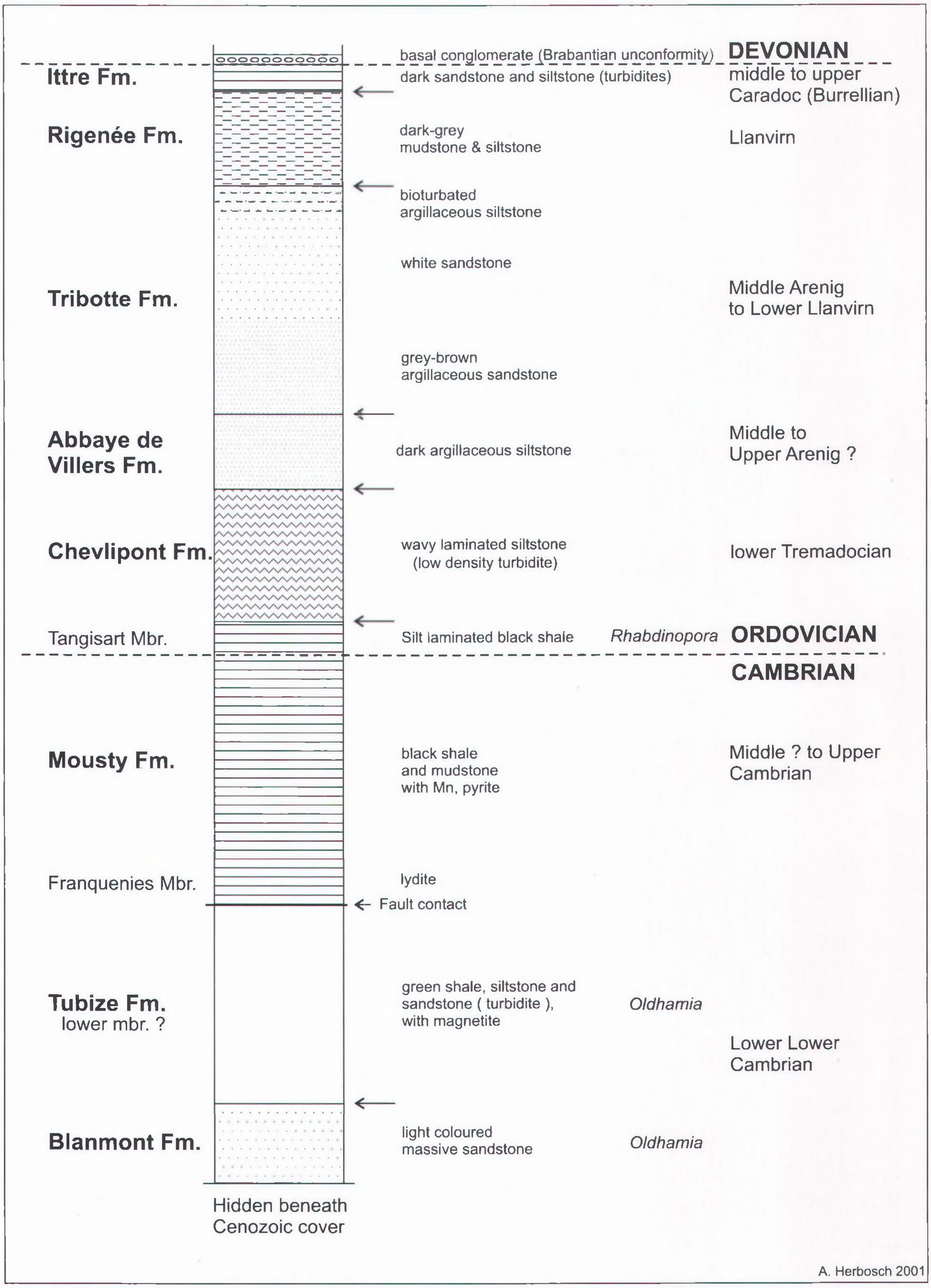


**Figure 5**. Stratigraphical log of the Dyle basin. Not to scale.

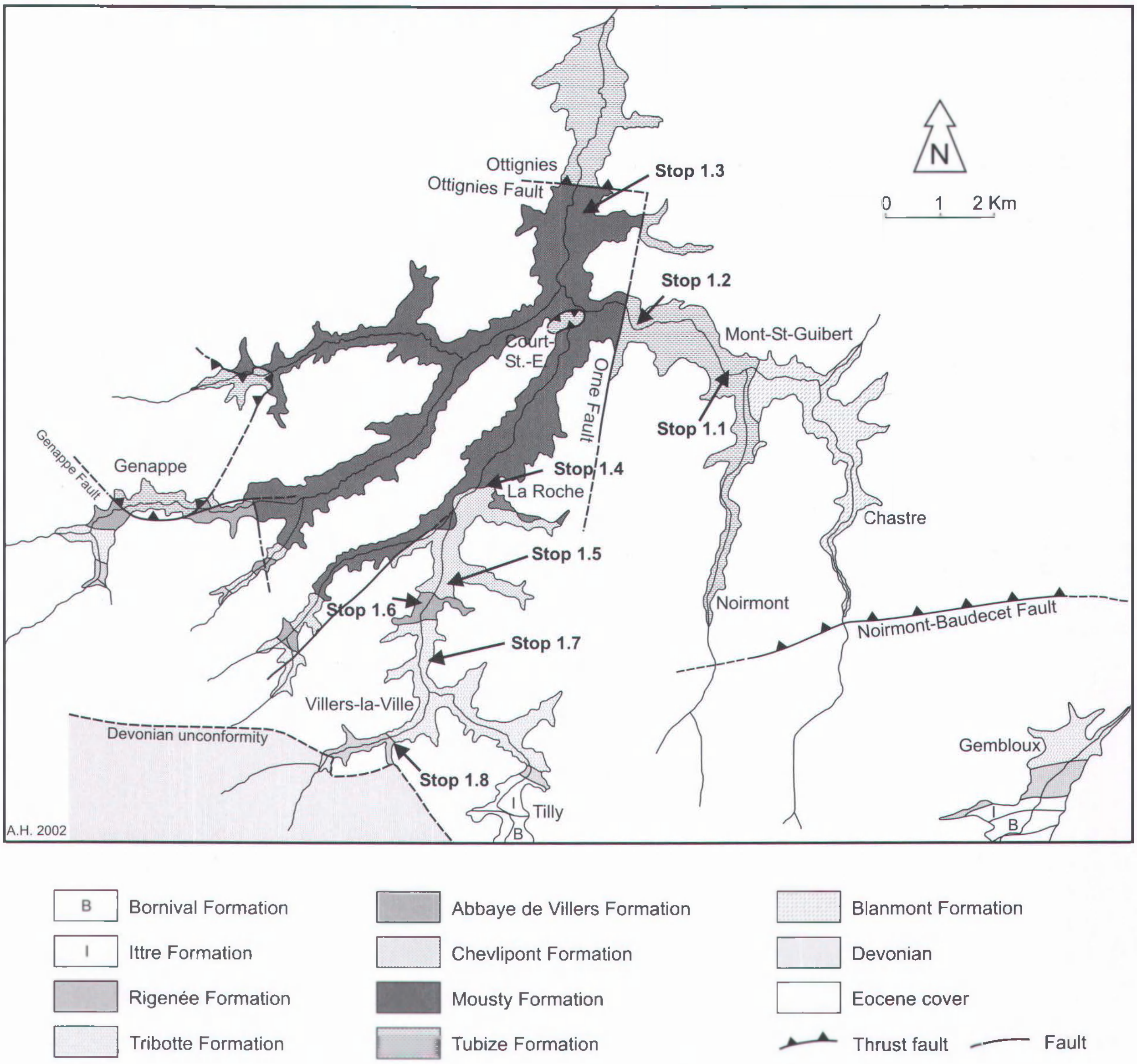


**Figure 6**. Geological map of the Lower Palaeozoic of the Dyle basin and northern part of the Orneau river (Gembloux).

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).

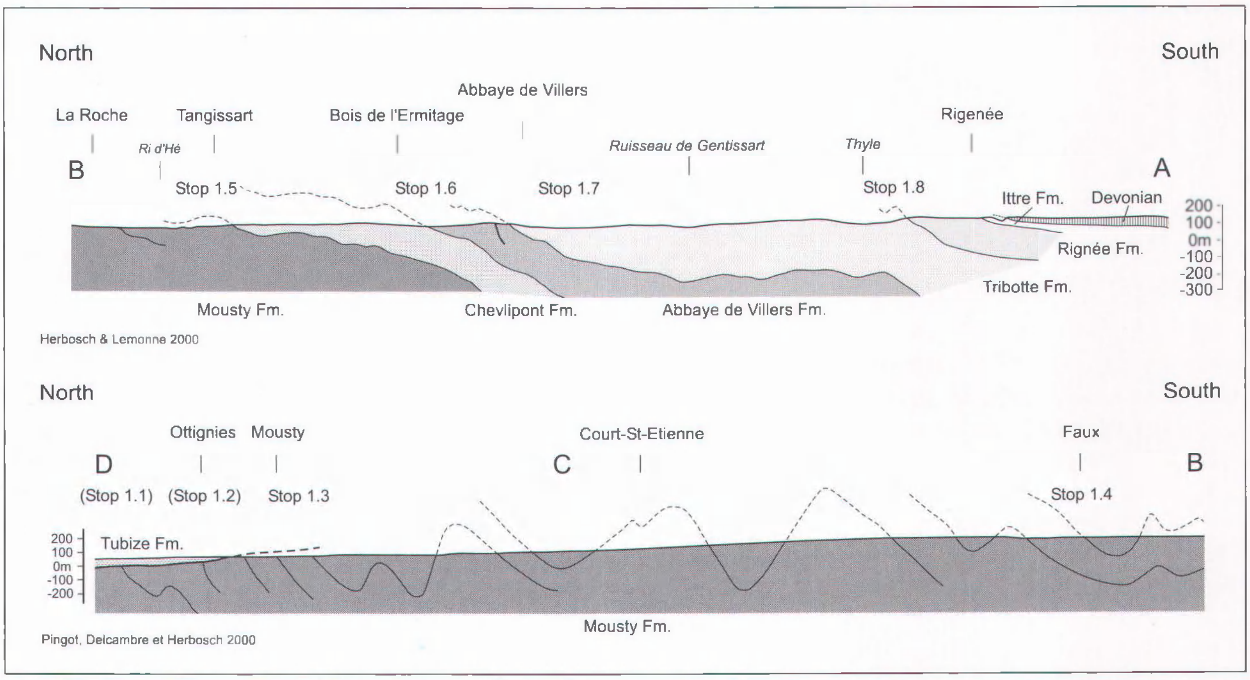


**Figure 7**. N-S cross-section in the central part of the geological map on Figure 6. The position of this ABCD section is shown on the structural map (Fig. 10). No vertical exaggeration. Stops between brackets are not situated in the cross section but projected

3.2.9. Ittre Formation (Beugnies, *in* Robaszynski & Dupuis, 1983; Servais, 1991a)

Description. Rhythmic alternation of light grey fine sandstone, grey siltstone and dark mudstone (slate) in decimetric beds. The sandstone and siltstone beds show parallel, oblique and convolute lamination, fining upward grading and load casts. All of these features are characteristic of Bouma-type sequences (Servais, 1991a, in the Sennette Valley). The base of this formation crops out with exactly the same facies in the Dyle basin (near Rigenée, Ri des Goutailles, Herbosch & Lemonne, 2000). The occurrence of a very large slump within the Ittre and the Bornival Formations has been suggested in the Sennette valley (Debacker, 2001; Debacker *et al.*, 2001). Thin volcanic beds are observed in the basal part of the formation in the Sennette valley (Corin, 1963; Debacker, 2001; Debacker *et al.*, 2001).

Sedimentology. This rhythmic sedimentation has been interpreted as Bouma-type distal turbidites (only the intervals c, d, and e are present, Servais, 1991a). Van Grootel *et al.* (1997) have recently shown that Unit II of the Lessines borehole also belonged to this formation. Herbosch *et al.* (1991) observed over 86 m rhythmic low density turbidite sequences (decimetric Piper sequences E1E2E3, Stow, 1986) which towards the top become high density distal turbidite sequences (c, d, and e intervals). The whole formation often shows slumps and microbreccias tens of centimetres in size. This is interpreted as a clastic deep marine environment which extends from the continental slope to the turbidite plains.

Thickness. More than about 180 m in the Sennette valley (Debacker, 2001; Debacker *et al.*, 2001) and 86 m in the Lessines borehole (Herbosch *et al.*, 1991).

Age. Graptolites described by Martin and Rickards (1979) and Degardin (*in* Herbosch *et al.*, 1991) indicate a Caradoc age. These same graptolites later re-examined by Maletz and Servais (1998) indicate the *Nemagraptus gracilis* or '*Diplograptus multidens*' Biozone (Aurelucian or Burrellian), with a preference for the latter biozone (Fig. 8). The chitinozoans, moderately well preserved and diverse with the presence of *Belonechitina* cf. *robusta*, indicate a Burrellian age. Both fossil groups together indicate a Burrellian age (Caradoc) (Verniers *et al.*, 1999; Samuelsson & Verniers, 2000) (Figs. 8 & 9).

3.2.10. Upper Ordovician formations

In the Dyle basin, the highest visible formation is the Ittre Formation. The overlying rocks of the Upper Ordovician outcrop in the Senne and Orneau basins (Fig. 3) where they have been recently studied by the research team of J. Verniers (Van Grootel *et al.*, 1997, 1998; Verniers *et al.*, 1999, 2001, 2002a; Debacker, 2001) and recently mapped during the Wallonia map-

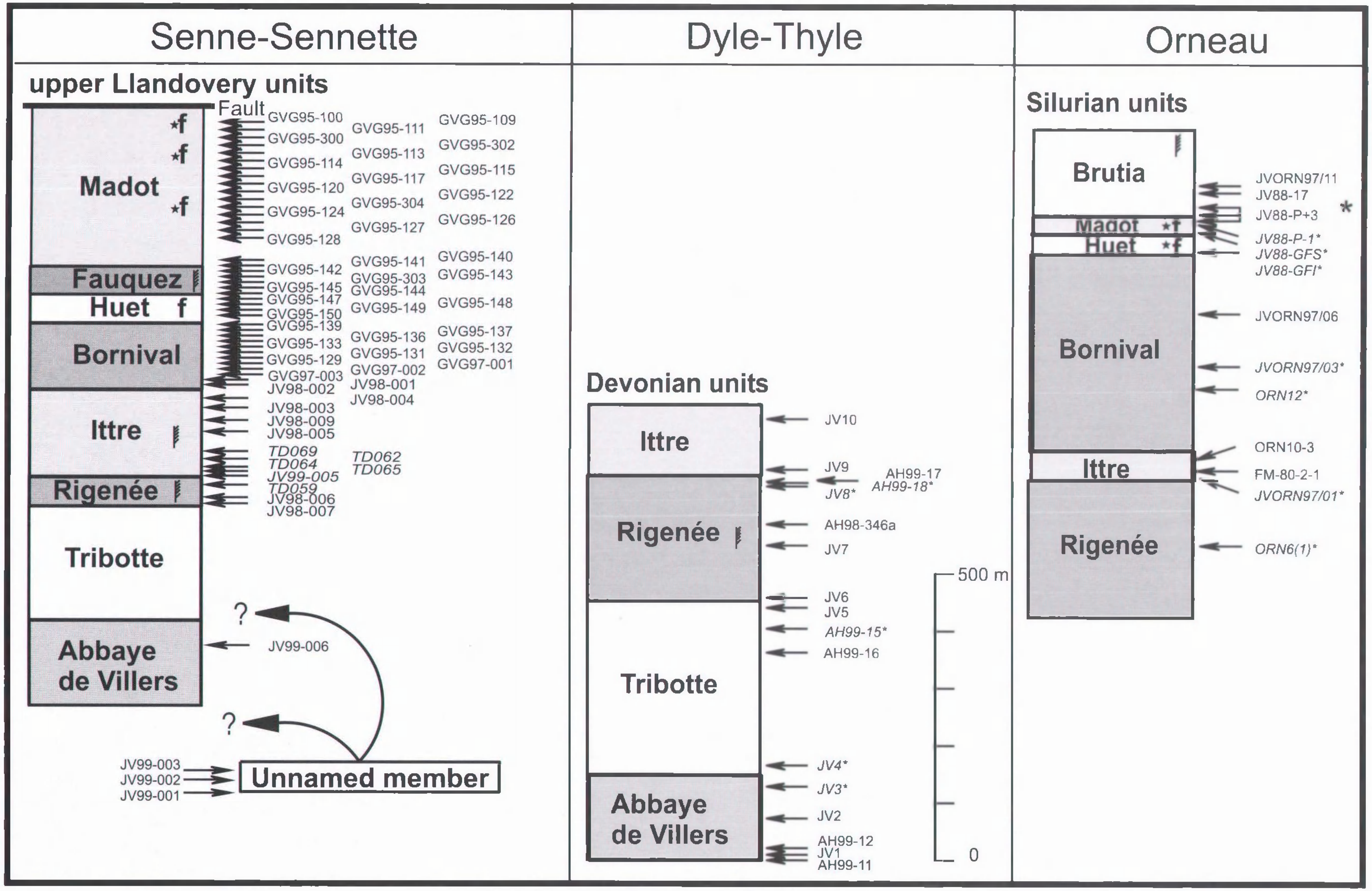


**Figure 8**. Schematic Ordovician lithostratigraphy in the three valleys of the outcrop area in the Brabant Massif with approximate thickness of the formations and location of the horizons sampled for chitinozoans in Samuelsson & Verniers (2000). Sample numbers in italics are barren. The star (*) indicates the location of the barren samples JV88-P+7 to JV88-P+15, and sample JV88-P+16 which yielded a single chitinozoan (see Fig. 9). The five-pointed stars indicate that macrofossils such as bryozoans, brachiopods, cystoids and trilobites are recovered in these units. The stylised graptolites denote levels with graptolites. Chitinozoans from the Bornival (upper members), Huet, Fauquez and Madot formations from the Sennette valley were studied by Van Grootel *et al.* (1997) and Van Grootel & Verniers (1998).

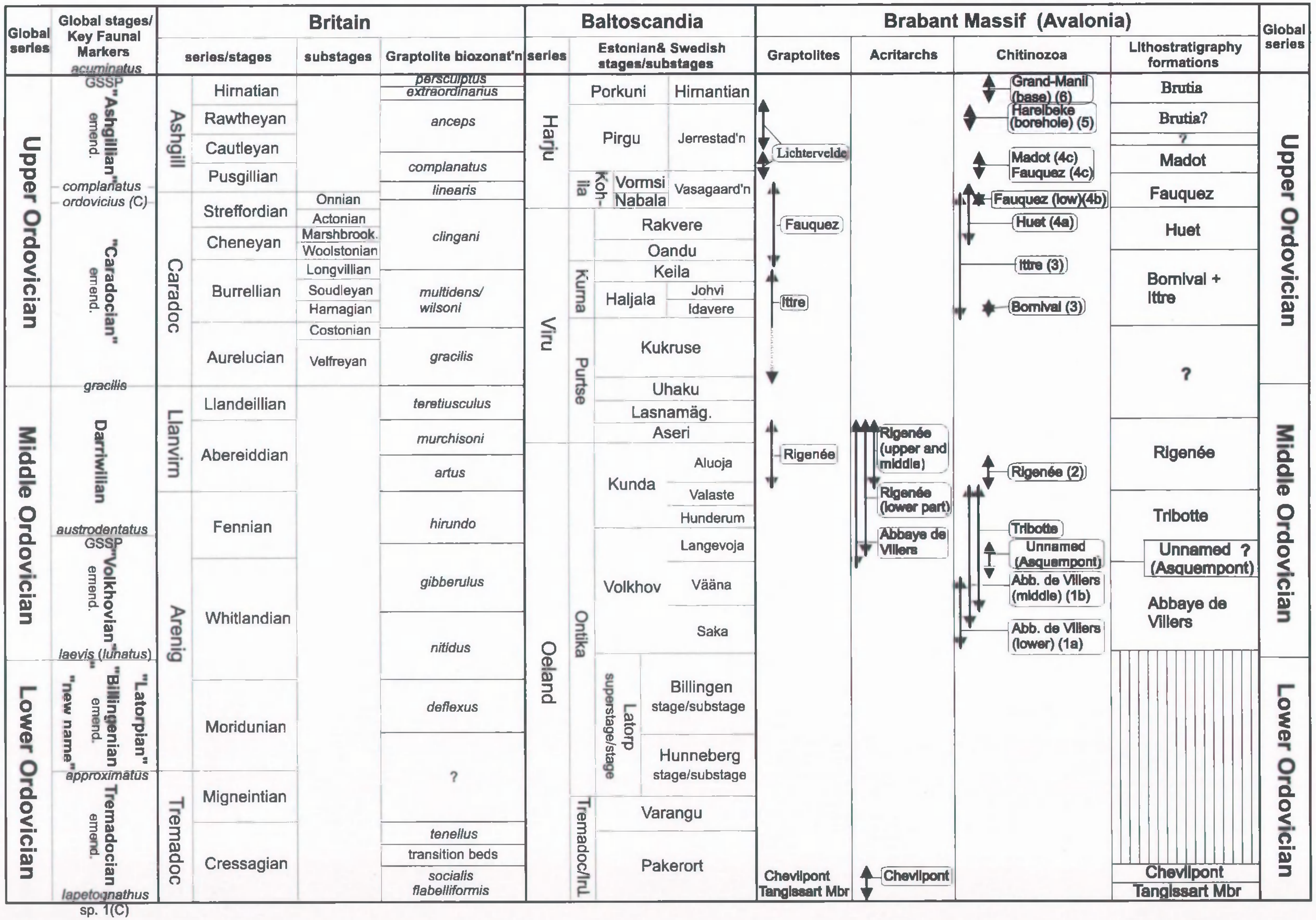


**Figure 9**. Schematic correlation of the Ordovician Brabant Massif formations with chitinozoan, acritarch and graptolite biozonations and the global, British and Baltoscandian chronostratigraphy and British graptolite biozonation (after Fortey *et al.*, 1995; Webby, 1998). Vertical hatching indicates hiatus, non-deposition, erosion or faulted contact. Arrows in biostratigraphy column indicate the range of the biozones versus the chronostratigraphy on the left. The number between brackets in the chitinozoan column is the number of the local biozonation. The Tangissart member is now considered the top part of the Mousty Formation (after Samuelsson & Verniers, 2000).

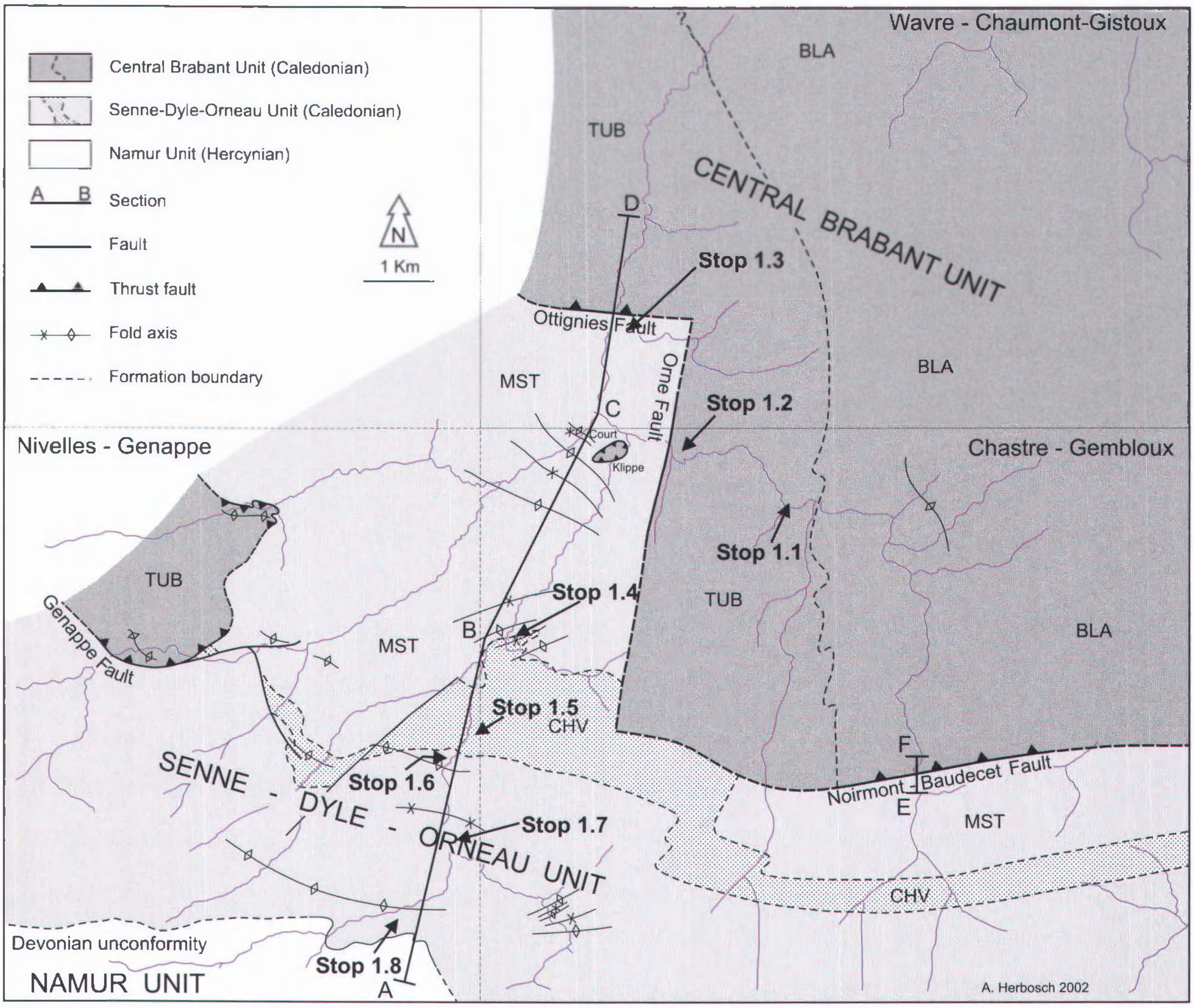


**Figure 10**. Structural map of the Dyle basin. BLA = Blanmont Fm.; TUB = Tubize Fm.; MST = Mousty Fm.; CHV = Chevlipont Fm. ABCD cross-section of Fig. 7 and EF cross-section of Fig. 11.

ping program (Chastre-Gembloux map, Delcambre *et al.*, 2002 and Braine-le-Comte-Feluy map, Hennebert & Eggermont, 2002). A brief description of these formations which we will not see is given below:

a) Bornival Formation: centimetric alternation of dark grey siltstone and dark grey to blackish mudstone with very fine sandstone beds (with plane or oblique laminations); same chitinozoan assemblage and age as the Ittre Formation.

b) Huet Formation: greenish to grey siltstone and fine sandstone, with characteristic brown to orange alveoli of decalcified fossil fragments; late Caradoc (Verniers *et al.*, 2001; Samuelsson & Verniers, 2000) (Figs. 8 & 9).

c) Fauquez Formation: a black shale facies which shows fine centimetric alternation of dark grey silty slate with black slate showing abundant pyrite and graptolites. The rhythmic sequences and grading point to a low density turbiditic environment (Lessines borehole, Herbosch *et al.*, 1991); latest Caradoc to earliest Ashgill (Verniers *et al.*, 2001; Samuelsson & Verniers, 2000) (Figs. 8 & 9).

d) Madot Formation: unit containing many volcanic and volcano-sedimentary rocks (André *in* André *et al.*, 1991; Van Grootel *et al.*, 1997) interbedded with dark shale and siltstone as well as siltstone rich in macrofossils (bryozoans, brachiopods, crinoids, trilobites, rugosa corals, etc.); early or middle Ashgill (Verniers *et al.*, 2001; Samuelsson & Verniers, 2000) (Figs. 8 & 9).

e) Brutia Formation: compact dark grey mudstone and slate with a very characteristic level of bioturbated mudstone. Upper part composed of about 40 m of volcanic rocks (eurite of Grand-Manil Member or Nivelles Member, Corin, 1965; Herbosch & Lemonne, 2000). Age: middle Ashgill to early Llandovery (Verniers *et al.*, 2001) (Figs. 8 & 9).

### *3.3. Cartography and general tectonics*

#### 3.3.1. New geological maps

The geological map given in Figure 6 is a schematic synthesis of three new geological maps (scale 1/25,000) recently surveyed for the Ministry of the Walloon Region. These new maps are: Nivelles-Genappe (Herbosch & Lemonne, 2000), Chastre-Gembloux (Delcambre *et al.*, 2002) and Wavre-Chaumont-Gistoux (Herbosch *et al.*, in prep. a). They cover all the Lower Palaeozoic and the Devonian outcrops of the Dyle basin and only the northern part (the Ordovician) of the Orneau valley.

As has been seen previously (see 3.2; Fig. 5) the stratigraphic column encompasses the lowest Lower Cambrian (Blanmont Formation) to the Upper Ordovician (Ittre Formation, Bornival Formation only visible at Gembloux) and the best sections occur in the Middle-Upper Cambrian (Mousty Formation) and in the Lower to Middle Ordovician (Chevlipont to Ittre formations). This new mapping definitively demonstrates the absence of the Oisquercq Formation in the Dyle basin (Figs. 5 & 4). Previously it had been confused (Anthoine & Anthoine, 1943) with a non-magnetite bearing facies of the Tubize Formation which is very similar. The Silurian is absent in the Dyle basin, hidden below a Middle Devonian cover to the south. However, it is present in large outcrops in the Orneau valley, just SE of the edge of the map (south of Gembloux, Fig. 6), a site which will be visited (see 1.3.2). Palaeozoic rocks are covered for the most part by Eocene formations (in white on the map, Fig. 6).

The cross-section of the Thyle Valley between Rigenée and La Roche (Fig. 7, cross-section AB) is the only section that is well constrained due to the relative quality and continuity of the outcrops (Michot, 1978). In this NNE-SSW cross-section we observe two large (km scale) low amplitude gentle anticlinal antiforms (Bois de l'Ermitage and Rigenée) separated by a large (km scale) low amplitude gentle synclinal synform (from Abbaye de Villers to Thyle river). The southern limbs of the two anticlines appear relatively short with inclinations of 30-35°, whereas their northern limbs are long, undulating and close to horizontal. The asymmetry of these folds is barely perceptible because of their large interlimb angles. The cross-section BCD, which continues to the north in the direction of Ottignies (Fig. 7), is much less well constrained because of the scarcity of outcrops and the almost complete absence of marker beds in the Mousty Formation. The only clear observation is that one moves from the large (km scale) gentle folds observed in the south (section AB) into a series of smaller (hm scale) open to close folds in the north, and that one progressively proceeds down series, as a lydite marker bed is encountered in the vicinity of Mousty (Franquenies Member). The Tubize Formation occurs in the extreme northern part of the cross-section in fault-contact (nowhere visible in outcrop) with the Mousty Formation (cf. 3.3.2).

#### 3.3.2. Structural Implications

The structural map can be found in Figure 10. It can be observed that the Caledonian basement (Brabant Massif) covers most of the mapped area except to the SW where the Devonian of the Namur Synclinorium unconformably overlies this basement. The Devonian unconformity which separates these two major units is marked by a basal conglomerate (Bois de Bordeaux Formation).

Two structural units can clearly be defined in the Caledonian basement (Fig. 10):
– the Central Brabant Unit, formed by the oldest rocks of the Blanmont and Tubize formations (and probably also the Jodoigne Formation present to the NE), crops out poorly and shows a general NNW-SSE direction with folds characterised by frequent steep plunges, variable trends and a north dipping cleavage (Lembeek type cleavage-fold relationship, Sintubin *et al.*, 1998). However, sub-horizontal to gently plunging folds have also been observed (Wavre-Chaumont-Gistoux map sheet, Herbosch *et al.*, in prep. a).
– the Senne-Dyle-Orneau Unit, formed by all the younger formations (from Mousty to the Silurian formations), shows E-W to NW-SE fold axes and open upright folds with a steeply N-dipping axial planar cleavage (Fauquez type cleavage-fold relationship, Sintubin, 1997a).

The tectonic break between these two domains has been described repeatedly since Fourmarier (1921) and is clearly visible on the aeromagnetic map (Chacksfield *et al.*, 1993; Sintubin, 1999; Everaerts, 2000).

The contact between these two units is always faulted (Genappe, Ottignies, Orne and Noirmont-Baudecet faults, Fig. 10). The Tubize Formation typically overlies the Mousty Formation but can also be found overlying the Abbaye de Villers or Chevlipont Formations (Fig. 6). At Court-St-Etienne (centre north, Figs. 6 & 10), a small zone with Tubize Formation is completely surrounded by the Mousty Formation suggesting a klippe (Anthoine & Anthoine, 1943). However, these faults have never been observed in the field due to the scarcity of outcrops, and thus only geophysical measurements (magnetism), trenches or boreholes have allowed us to deduce their presence and the position of their trace:
– at Baurieux, a magnetic ground survey allowed de

Magnée and Raynaud (1944) to prove the existence of the Orne Fault, as well as its trace under the Cenozoic cover (Fig. 14);
– at Noirmont (Figs. 10 & 11, section E-F), several boreholes and an old quarry along a N-S profile indicate the presence of an E-W trending fault, the Noirmont-Baudecet Fault. Other boreholes scattered to the east allow us to show its continuation until Baudecet (N of Gembloux). This fault could corresponds with an E-W magnetic lineament recognised by Sintubin (1997b);
– at Ottignies, the Ottignies Fault (Herbosch & Blockmans, in press a) is less well constrained but nevertheless its existence is proven along several hundred metres by trenching (Van Tassel, 1986);
– the trace of the Genappe Fault and its prolongation towards the NE (Cala valley) is poorly constrained in outcrop, but can be clearly seen on aeromagnetic maps (Herbosch & Lemonne, 2000).

As a consequence, we believe that in the Dyle-Thyle area all the Lower Cambrian of the Central Brabant Unit was thrusted onto the Cambrian-Silurian foreland. According to this hypothesis, the Tubize Formation outcrop at Court-St-Etienne is a klippe, the Ottignies and Noirmont-Baudecet faults are a part of a single unique thrust fault and the Orne Fault, more vertical could be interpreted as a tear fault (Herbosch & Blockmans, in press a).

3.3.3. Stratigraphic implications

The hypothesis that the Lower Cambrian of the Central Brabant Unit was thrusted onto its foreland also has implications for an old stratigraphic problem that has not yet received an entirely satisfactory resolution. The problem is the following: how can the absence of the Mousty Formation in the Senne basin be explained when it is present in the Dyle basin whereas, on the contrary, the Oisquercq Formation is present in the Senne basin but absent in the Dyle? The schematic drawing in Figure 12 illustrates this problem.

The most widely accepted explanation (Vanguestaine, 1978) was that there were lateral variations in the facies between the two regions, implying that these two formations had the same age (not well known at the time). This simple solution which did not satisfy everyone (Legrand, 1967, 1968), can no longer be considered since Vanguestaine (1991, 1992; Herbosch *et al.*, 1991) demonstrated, using micropalaeontology, that the Oisquercq Formation is upper Lower to Middle Cambrian in age and thus underlies the Mousty Formation which is Middle Cambrian to earliest Tremadocian in age (Fig. 4).

Now that almost all of the outcrop areas of the Brabant Massif have been mapped the following explanation can be proposed:
– let us begin with the Senne basin. Mapping of the Ittre-Rebecq region (Herbosch *et al.*, in prep. b) demonstrates that the Asquempont Fault, as Beugnies (*in* Waterlot *et al.*, 1973, Figs. 48 & 49) has clearly shown, runs from the Sennette valley through the Senne valley and then continues to the NNW. This important fault causes the disappearance of the upper part of the Oisquercq Formation, all of the Mousty Formation and

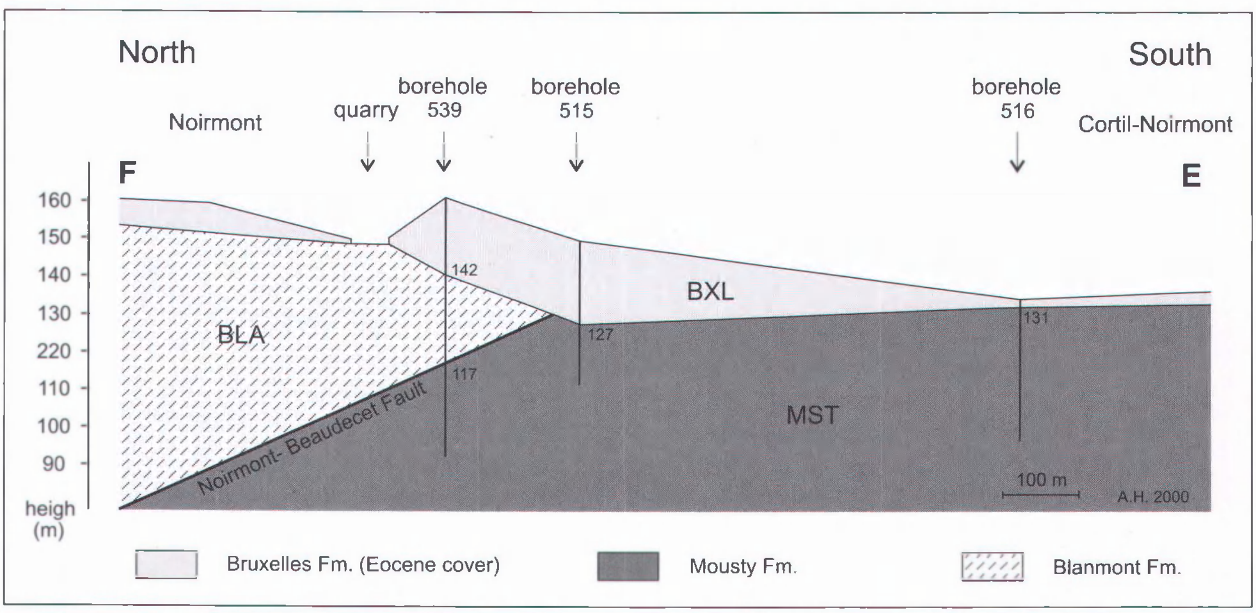


**Figure 11**. N-S cross-section of the Noirmont region. The position of this EF section is shown on the structural map (Fig. 10). Several boreholes and an old quarry allow the existence of the Noirmont-Baudecet Fault to be proved. A large range of dip to the north is possible. Vertical exaggeration x5.

almost all of the Chevlipont Formation. Indeed, the work done by Vanguestaine (1978, *in* André *et al.*, 1991) and by Lenoir (1987) shows that there remain several or at the most ten metres of the Chevlipont Formation in the four sections where the Asquempont Fault is observed (Fig. 12). This situation is identical to that of the Lessines borehole (Herbosch *et al.*, 1991).

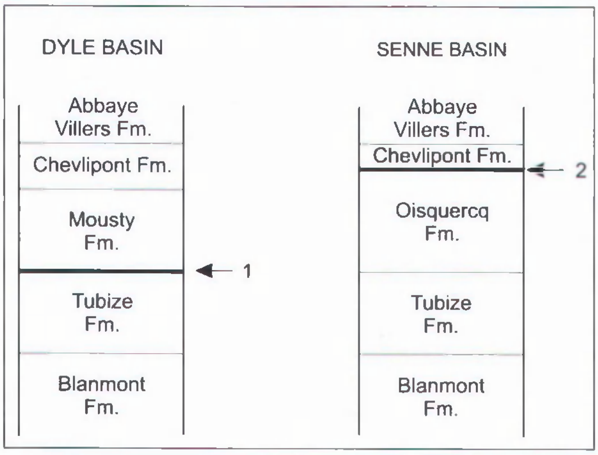


**Figure 12**. Stratigraphical succession in the Senne and Dyle basins (not to scale). 1=thrust fault contact 2=Asquempont Fault.

The interpretation by these authors of the contact between the Oisquercq Formation (Unit IV of the borehole) and the Chevlipont Formation (Unit III of the borehole) must now be revised. What was previously interpreted as a basal conglomerate is more likely to be a fault breccia (Debacker, 2001 and Debacker & Herbosch, unpublished data);
– in the Dyle basin, if we accept the hypothesis of a thrust fault, which is suggested by the complex and lobed aspect of the thrust front as well as the presence of a klippe, one could suppose that the Oisquercq Formation as well as a basal part of the Mousty Formation have been hidden underneath the thrust sheet.

## 4. Description of excursion stops DAY 1

### *Stop 1.1. Church of Mont-Saint-Guibert*

Location. Escarpment at the southern base of the church.

General structure. The bedding attitude is approximately N-S (strike N10°-20°E, dip 80°W) while the cleavage attitude in the argillaceous beds is: strike N45°E, dip 82°W. This implies a steeply WNW-plunging cleavage-bedding intersection. This peculiar geometry has already been described in the Senne basin by Sintubin *et al.* (1998). It is often observed in the structural unit forming the central part of the Brabant Massif (Central Brabant Unit of Fig. 10, also see introduction 3.3.2).

Lithostratigraphy. Lower part of the Tubize Formation; the Blanmont Formation crops out about 400 m to the east (Figs. 5 & 6).

Lithology. Rhythmic sequences of sandstone, siltstone and mudstone (slate). Beds show graded bedding, plane and convolute lamination. Here in the lower part of the formation the sandstone has a high mica content but does not contain any rock debris or feldspars, as opposed to the Rogissart Member (Senne basin, middle part of the Tubize Formation).

Sedimentology. A detailed study of this 5 meter-long section (Fig. 13) shows sedimentological features supporting a turbiditic origin of this rhythmic sequence. This sequence can be described in terms of the classical Bouma division. Incomplete Tab and Tae sequences, with a mean thickness ranging from 10 cm to 1 m, are most frequent. While load casts and graded bedding are frequently observed, flute casts are scarce. Only one complete Tabcde sequence, with rip-up clasts, has been recognised. All bedding polarity criteria indicate a younging of the sequence towards the WNW.

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

### *Stop 1.2. Beaurieux, Orne Valley*

Location. Western side of a small valley, northern tributary of the Orne, situated 150 m west of the motorway bridge which cuts across the Orne Valley (Fig. 14).

General structure. A series of small outcrops each several meters in size, dispersed across the wooded hillside. The bedding which is often poorly visible is subvertical (>80°W, more rarely E) and cuts across the hillside following an approximately N-S direction (strike N350° to 020°E). The cleavage, relatively rough, is vertical and runs in an E-W direction (N090° to 110°E, 65°-75°N), which is the common cleavage direction of the Dyle basin.

Lithostratigraphy. Lower part of the Tubize Formation (Fig. 5).

Lithology. The facies is homogeneous (no visible stratification or lamination) and is composed of grey-green siltstone (slate) with numerous millimetric crystals of magnetite.

Sedimentology. In view of the presence of sandy episodes which have been interpreted as turbidites, we consider these silty and argillaceous sequences to be deep-sea pelagic to hemipelagic deposits (for more details see 3.2.2).

Age. Cf. stop 1.1.

Magnetic survey and the Orne Fault. A magnetic survey of the Beaurieux area has been published by de Magnée & Raynaud (1944) with the intention to verify the

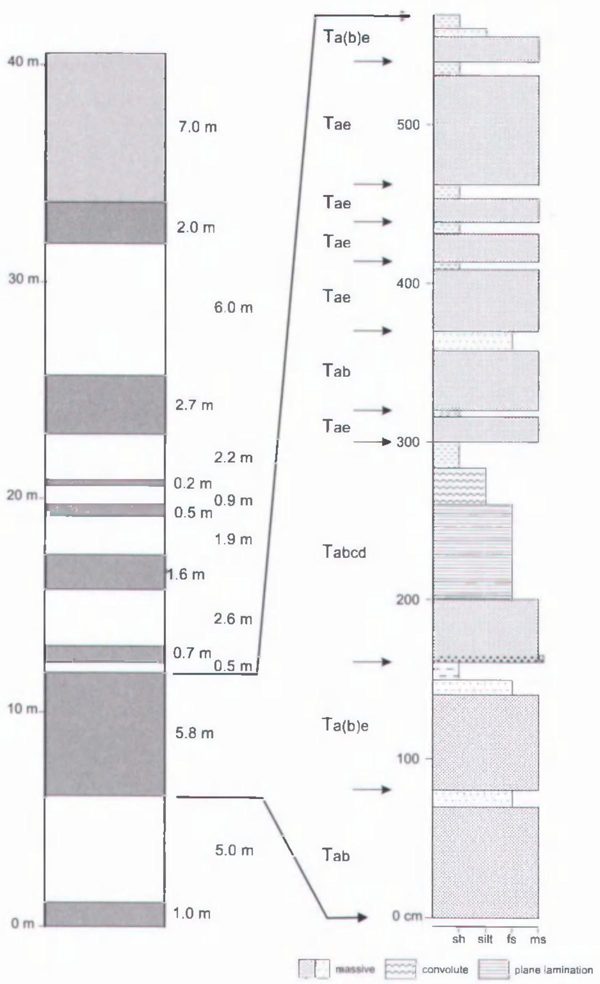


**Figure 13**. Left: general stratigraphical section of all the outcrop present below and south of the church of Mont-Saint-Guibert (lower part of Tubize Formation, lower Lower Cambrian); grey: in outcrop, white: present, but not outcropping; right: detailed lithological log of part of the section. Tab = type of Bouma turbidite sequence; Sh = shale; fs = fine sandstone; ms = medium sandstone.

hypothesis that the Orne Fault is a thrust fault (Anthoine & Anthoine, 1943). This work (Fig. 14) shows that there are several magnetic horizons (stripes in Fig. 14) oriented in a NNE to NNW direction which are abruptly interrupted towards the west where a negative anomaly occurs corresponding to the Mousty Formation. Thus the Orne Fault must surely dip to the east and must run close to and east of the interruption of the magnetic horizons. The Orne Fault has been traced essentially on the basis of this magnetic survey; its prolongation to the NW is only based on a few outcrops (Figs. 6 & 10).

***Stop 1.3. Former Franquenies quarry***

Location. Private property situated at the Franquenies locality, SE of Céroux-Mousty, on the northern slope of the Angon brook, about 200 m west of the Brussels-Namur railway line.

General structure. The wall of this old quarry, 25 m long and 5-6 m high, cuts into the wooded flank of the valley. Bedding is clearly visible (strike N040°-050°W, dip 50°-60°E) as well as cleavage parallel to the wall; the strike is approximately E-W and the dip is sub-vertical (N or S). The unconformity with the Cenozoic cover (Brussels Formation, Eocene) is marked by a thin basal conglomerate which can be seen several dozen meters to the north.

Lithostratigraphy. Franquenies Member, lower visible part of the Mousty Formation.

Lithology. Black shale (slate) massive or laminated, grey to grey-black in colour, locally grey-brown (in the yard of the house). The organic material has been transformed into graphite by metamorphism and when touched the rocks leave traces on the fingers. The pyrite is not visible any more due to the alteration in this old quarry, but it was noted in older descriptions. Silicified zones can also be seen, in particular in the centre and towards the top of the wall (inaccessible overhanging zone) which forms extremely black and hard concretionary masses with botryoidal structures. Several blocks that have fallen to the foot of the wall can be observed. These are lydite essentially formed of quartz which has been completely obscured by a finely dispersed black graphitic pigment and sometimes a bit of pyrite. Certain samples, in thin section, show numerous transparent objects formed by microcrystalline silica without any graphitic pigment which are embedded in a completely opaque matrix (finely dispersed graphite). These objects are about 150 to 200 microns in size, are oblong and elongated along the bedding (compaction). These objects strongly suggest radiolarian phantoms and in certain exceptional cases we have been able to observe very poorly preserved radioles. Unfortunately,

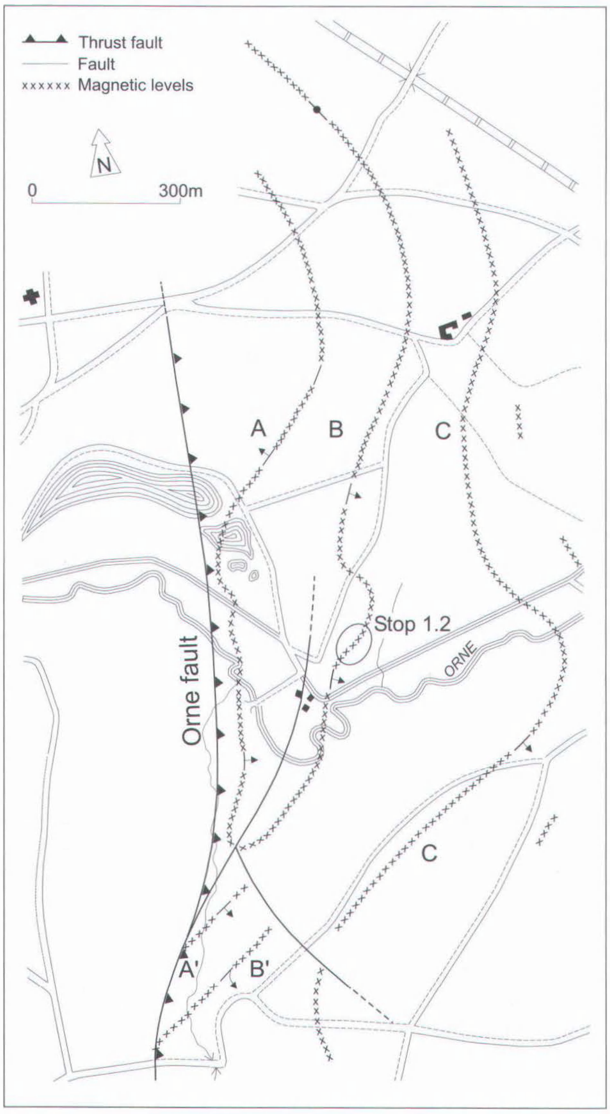


**Figure 14**. Geological interpretation of the field magnetic survey done by de Magnée and Raynaud (1944). Location of stop 1.2. (A. Herbosch modified after de Magnée & Raynaud, 1944).

M. Caridroit (University of Lille I) has not been able to confirm our hypothesis through acid extractions.

Sedimentology. Pelagic or hemipelagic shales deposited in a deep-sea anoxic environment as substantiated by the abundance of organic material, pyrite (only seen in boreholes) and manganese (in the form of garnet, see 3.2.4). The presence of lydites (radiolarians?) is also in agreement with this type of environment.

Biostratigraphy. Despite numerous attempts, no acritarchs have yet been obtained on samples coming from outcrops in the Dyle basin (Vanguestaine, pers. comm.). This is probably due to the strong surface alteration affecting these pyritic shales. The same facies in boreholes contains acritarchs giving ages between the Middle and Late Cambrian (Vanguestaine, 1992).

Radon anomaly. Tondeur *et al.* (1994) have shown that the towns of Villers-la-Ville and Court-St-Etienne are areas where there is a risk of high levels of radon in the houses. These risk areas coincide perfectly with the outcrop zones of the Mousty Formation (Tondeur *et al.*, 2001). Black shales are well known in the geochemical literature for their enrichment in uranium (average shale U content = 2 to 4 ppm; average black shale U content = 8 ppm but the variability is very high, up to 189 ppm in Swedish alum shale).

### *Stop 1.4. Charleroi-Ottignies railway section: La Roche - Faux cut*

Location. Railway section, km 36.2-36.1, 500 m to the north of the La Roche station.
General structure. Very discontinuous outcrops on the eastern side of the track. This cut has been described in detail by Van Tassel (1986) at a time when the outcrop was in much better condition (Fig. 15). The bedding, of variable direction, is cut by numerous almost vertical faults that run approximately E-W.
Lithostratigraphy. Tangissart Member, uppermost part of the Mousty Formation. This member is a maximum of 100 m thick and marks the transition between the Mousty and the Chevlipont formations.

Lithology. Black shales alternating with dark shales with lighter siltstone laminae. The number and the size of these laminae vary considerably from sample to sample. The thickest are rarely greater than 0.5 cm. Horizontal bioturbation is also observed (*Planolites*).

Sedimentology. The silty laminae which appear episodically, and higher up more and more frequently within the black shales are interpreted as very distal low density turbidites (mud turbidites, Fig. 16). This interpretation is only possible within the extended framework of the sedimentological study of the Chevlipont Formation (cf. stop 1.5 and see sedimentological introduction 3.2.5) and the gradual transition from the Mousty to Chevlipont formations. As a result, the depositional environment is interpreted as an anoxic deep marine environment. However, instead of having a purely pelagic sedimentation a terrigenous component, typical of hemipelagic sedimentation, appears progressively. This is interpreted as the beginning of a regression which continued into the overlying formation.

Biostratigraphy. It is within this section and in the talus of the Faux-La Roche road (now hidden by a wall) that Lecompte (1948) described *Rhabdinopora flabelliforme* which proves the early Tremadocian age of the

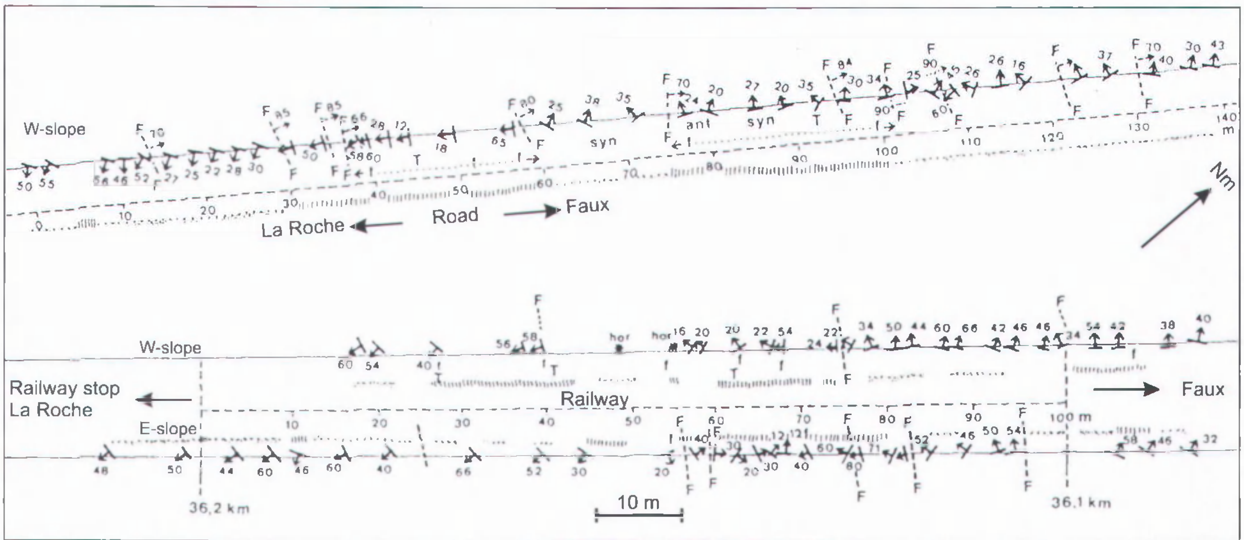


**Figure 15**. Location map by Van Tassel (1986) of the road and Charleroi-Ottignies railway sections N of La Roche station (stop 1.4). Tangissart Member, uppermost part of Mousty Fm. Stippling = silty laminated shales. Hatching = black shales. f = graptolite. T = trilobite. F = fault.

member (see also Van Tassel, 1986). However, Vanguestaine (*in* André *et al.*, 1991) has only found rare and poorly preserved acritarchs.

### *Stop 1.5. Charleroi-Ottignies railway section at Chevlipont*

Location. Railway crossing to the NE of Chevlipont (old water mill along the Thyle river). NE side, km 37.93-38.08.

General structure. This quite continuous 160 m long outcrop shows regular bedding gently dipping to the N (strike N30°-40°W, dip 20°N). The north dipping beds correspond to the N limb of a low amplitude (hm scale) gentle antiform ("Bois de l'Hermitage dôme", Michot, 1978). This gentle antiform is part of the long undulating N limb of the larger (km scale) Bois de l'Hermitage anticline, which extends from La Roche to the Bois de l'Hermitage (Fig. 7).

Lithostratigraphy. Middle part of Chevlipont Formation (Fig. 5).

Lithology. Grey siltstone with characteristic wavy bedding consisting of rhythmic alternations (mm to cm) of light grey siltstone laminae and dark grey clayey siltstone to 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-size horizontal bioturbation (*Planolites*) is frequently observed in the clayey laminae at the tops of the sequences. This facies is regularly interrupted by cm to dm fine sandstone beds which are either massive or present plane parallel bedding or convolute structures.

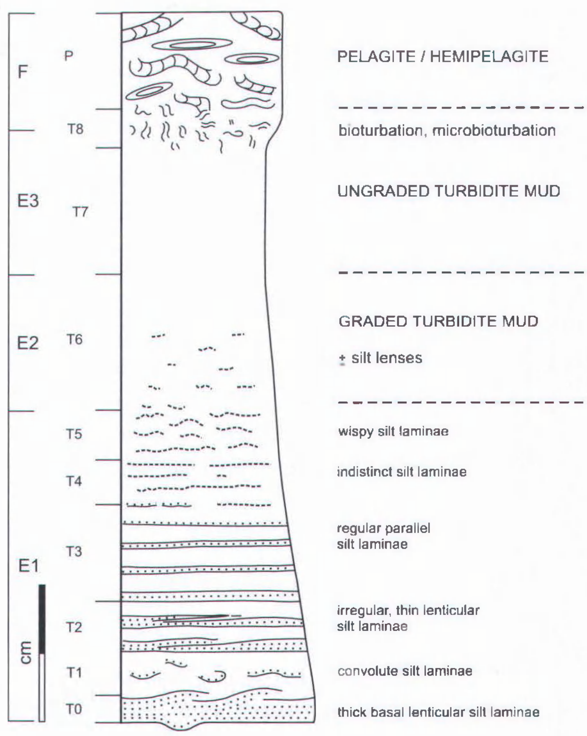


**Figure 16**. Low density turbidite facies model. Structural division E1 to F redrawn after Piper (1978), T0 to P redrawn after Stow & Shanmugam (1980).

Sedimentology. Study of the outcrops and boreholes allows the interpretation of this facies as low density turbidite sedimentation typical of "mud turbidites"

(Stow & Shanmugam, 1980; Stow, 1986). This laminar siltstone is the result of the repetition of incomplete Stow's sequence 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). These top-cut-out type of sequences are transitional to silt turbidite facies (Stow & Piper, 1984). Repetition of these truncated sequences with minor variations led to the development of a distinctive facies, thin parallel to wavy silt-laminated mud, in which individual turbidite units may not be readily distinguished.

Several other arguments support this interpretation: frequent presence in boreholes of slumps and intraformational breccias; occurrence of thin sandy beds interpreted as high density distal turbiditic episodes; lateral variation to high density turbidites; sedimentary continuity with the pelagic shales of the Mousty Formation. The depositional environment is always deep, relatively reducing (pyrite and organic matter), with a slightly increased slope (slumps), certainly closer to the continental slope than the depositional environment of the Mousty, although we can not be any more precise.

Biostratigraphy. Tremadocian, probably Lower Tremadocian. This is based on the presence of abundant and well preserved acritarchs (Martin, 1976; Vanguestaine *in* André *et al.*, 1991; Vanguestaine, 1992; Herbosch *et al.*, 1991), as well as the presence of rare graptolites (Lecompte, 1949).

***Stop 1.6. La Roche - Abbaye de Villers road, km 30.75***

Location. West slope of the Thyle Valley, on the road from the Villers Abbey to La Roche at about km 30.75 (outcrop just above a stone wall).

General structure. An almost continuous outcrop over 60-70 m. A succession of three antiforms (meter to decameter scale) with step fold geometry can be observed. Further north the beds remain sub-horizontal to gently S dipping with an approximately E-W strike. On the AB cross-section of Figure 7, this outcrop is situated in the fold hinge zone of the northern km scale antiform (Bois de l'Hermitage). Although the folds are considered to be tectonic they may well have formed along a previous soft deformation structure (Debacker, 2001). Michot (1978) has previously described slump structures not far from this outcrop in the same formation.

Lithostratigraphy. Lower third of the Abbaye de Villers Formation (Fig. 5).

Lithology. Grey to dark grey argillaceous to sandy siltstone, roughly laminar to bedded. This laminar structure is marked either by set of numerous mm thick silty laminations or by lighter coloured silty/sandy beds which interrupt the argillaceous sedimentation. The abundance of mm sized micas (illite-chlorite stacks in thin section) is also very characteristic. The stratification is frequently disturbed by a strong horizontal or more rarely vertical bioturbation (plurimillimetric to centimetric). Unfortunately, this outcrop does not show the oblique laminated bedsets that are frequently observed elsewhere in this formation (for example in the Abbaye de Villers railway trench).

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 continental shelves. The dominantly argillaceous character demonstrates that the depositional environment is certainly below the fair weather wave base. This depth which corresponds to the middle shelf 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).

A significant sedimentological gap is observed between the Chevlipont and Abbaye de Villers Formations as we move directly from a deep marine environment probably situated close to turbidite plains to a relatively shallow continental shelf environment.

Biostratigraphy. No macrofossils are found. Chitinozoans in the lower third of the formation in the Dyle basin contain *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 acritarch 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 thus present between the Chevlipont and the Abbaye de Villers formations (Figs. 4 & 8).

### *Stop 1.7. Bois Pinchet pond, 500 m S of Abbaye de Villers*

Location. East slope of the Thyle Valley to the NE of the Bois Pinchet pond. Corresponds with km 40.5 of the Charleroi-Ottignies railway line.

General structure. Numerous small outcrops dispersed over about a hundred metres in a general N-S direction

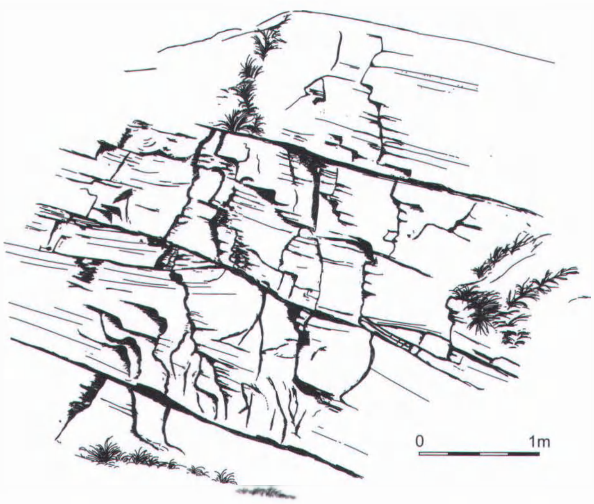


**Figure 17.** Set of oblique stratifications (sand ridge?) in the Abbaye de Villers Formation; km 38.842 of the railway cut just east of the ruin of the Abbaye de Villers (drawing F. Boulvain).

on the Bois Pinchet slope. The beds have a WNW-ESE strike and are slightly dipping to the south (strike N60°W, dip 15°-20°S). On the AB cross-section (Fig. 7) this outcrop is situated in the northern descending limb of the large km scale syncline which extends from Abbaye de Villers to the Thyle river.

Lithostratigraphy. Lower third of the Tribotte Formation.

Lithology. This lower part contains brownish grey, clayey sandstone and siltstone, with coarse laminations and strong bioturbation (mainly horizontal). The transition is progressive and is marked by an increase in grain size (average 70 to 90 microns), an increase in the quartz content and the disappearance of the dark colouring. Thin section examination shows abundant potassic feldspar (up to 18%) and plagioclase in the sandstone from the base of the formation (Jodard, 1986; Herbosch *in* André *et al.*, 1991).

Sedimentology. At the base of this formation there is a progressive evolution from a shelf environment situated under the wave base, the depositional environment of the Abbaye de Villers Formation, towards a more agitated (sand) and oxygenated (lighter coloured rocks) environment, probably situated close to the wave base (clays are still present). The arkosic episode at the base of this zone has not yet been explained sedimentologically or geodynamically.

Biostratigraphy. Neither macrofossils nor acritarchs have been observed but numerous ichnofossils: 'Fucoids' (Malaise, 1911; de la Vallée Poussin, 1930), bilobites (Legrand, 1967). A poor chitinozoan assemblage however, 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).

### *Stop 1.8. "Chemin creux de Rigenée", NW of Rigenée, Thyle Valley*

Location. Along a dirt road going to the village of Rigenée, east of the bridge over the Thyle, which leads to the old castle Le Chatelet.

General structure. Discontinuous outcrops in the road and on its south side. They begin about 60 m from the bridge and continue until the road curves to the S and becomes sunken. The bedding attitude is WNW-ESE with a moderate dip (strike N70°-80°W, dip 55°-60°S). This moderate dip corresponds to the southern limb of a small scale (hm) anticline which is situated in the fold hinge zone of the southernmost large scale anticline (Rigenée, Figs. 6 & 7). The cleavage can be clearly seen in the argillaceous rocks and is always in an E-W direction and sub-vertical (dipping to the N, occasionally to the S).

Lithostratigraphy. Top of the Tribotte Formation and base of the Rigenée Formation (Fig. 5).

Lithology. Dark grey to bluish grey laminated siltstone and mudstone. The siltstone has often experienced bioturbation. The lower boundary of the unit marks a rapid upward change (over about 10 metres, Servais, 1991b; Herbosch & Lemonne, 2000) from light clayey siltstone (upper part of Tribotte Formation) to dark finely laminated argillaceous siltstone to mudstone.

Sedimentology. The sedimentology has not yet been systematically studied. The extremely rapid passage from the light coloured silto-argillaceous rocks of the Tribotte Formation into the dark laminar siltstone of the lower part of 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 middle and upper part 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 (probable depositional environment of the Ittre Formation).

Biostratigraphy. According to Martin (1969a), acritarchs indicate an (early) Llanvirn age. Acritarchs studied by Servais (1993) indicate that for the base of the formation (this outcrop) 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 late 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).

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) (Fig. 8).

## 5. Silurian of the Orneau Valley (DAY 2)

### *5.1. Introduction*

The Lower Palaeozoic sequence of the Brabant Massif can be well observed in the Dyle Valley for the Cambrian-Ordovician part. The small Orneau Valley passing thorough the city of Gembloux offers a good continuation upward of the sequence. Between Gembloux and the hamlet of Les Mautiennes, near Mazy (Fig. 18) this valley cuts through the Ordovician and Silurian sediments of the Brabant Massif, 20 to 45 m below the base of the Upper Eocene Bruxelles Formation, present in the upper parts of the valley (Fig. 19). The valley is oriented at right angles to the fold structures. The generally south dipping strata with some folds in the southern part allows the observation of a rather continuous succession with much of the Middle Ordovician to Silurian formations (Figs. 1 & 20). The series of long outcrops along the railway, roads and small quarries, known since long (Dumont, 1848), was chosen a century ago as the type area for the three-fold division of the Silurian in the Brabant Massif, with the "Assise de Grand-Manil" for the Llandovery, the "Assise de Corroy" for the Wenlock and the "Assise of Vichenet" for the Ludlow (Malaise 1900, 1910). Later stratigraphical studies corrected this division in other outcrop areas of the Brabant Massif as in Ronquières (Legrand, 1967) and in the Mehaigne area (Verniers, 1983a). It allowed to subdivide these too large and heterogeneous units in more than eight formations (Fig. 4). The units were well dated with graptolites, acritarchs and chitinozoans (see overview in Verniers & Van Grootel, 1991 and Verniers *et al.*, 2001, 2002b). In the stratigraphical revision of the Lower Palaeozoic formations of Belgium, four formations have their type locality in the Orneau Valley: the Brûtia, Bois Grand-Père, Corroy and Vichenet formations (Fig. 4) (Verniers *et al.*, 2001).

In an unpublished M.Sc. thesis by De Schepper (2000) the four type localities in the Orneau Valley were studied in some detail. Samples from all Silurian formations were taken for chitinozoan studies, but only from five formations the organic microfossils could be extracted. As usual in most of the Brabant Massif, the chitinozoans are dark to opaque and only moderately preserved. They have a concentration between 0.04 and 7.0 chitinozoans per gram of rock and a diversity of 1 to 12 species per sample (Fig. 22 & 23). The indicative and well preserved species are shown on Plates 1 to 5. Illite crystallinity studies in this area indicated that the metamorphism reached the low anchizone (Van Grootel *et al.*, 1997), which could have destroyed part of the microfossils. The assemblages of chitinozoans corroborated the age of the Corroy Formation, as deduced earlier by graptolites and they could date for the first time the Vichenet Formation (not Ludlow as assumed earlier but late middle to early late Wenlock). The lower limit of the latter formation in the type locality could be located (stop 2.4, see below).

The following Silurian formations are present in the Orneau Valley and are summarised from the stratigraphical revision by Verniers *et al.* (2001).

### *5.2. Brûtia Formation*

Defined during the geological mapping by Delcambre *et al.* (2002) the formation has its stratotype in Grand-Manil, between the localities Try-al-Vigne and Brûtia. It contains the lower part of the previous "Assise de Grand-Manil".

Description. The lower unit contains (greenish) medium to dark grey mudstone and slate, compactly bedded. At one third up in the unit, a mottled grey mudstone member occurs, a few meters thick, consisting of dark grey lenses (about 1-2 mm wide and 2-5 mm long) in a medi-

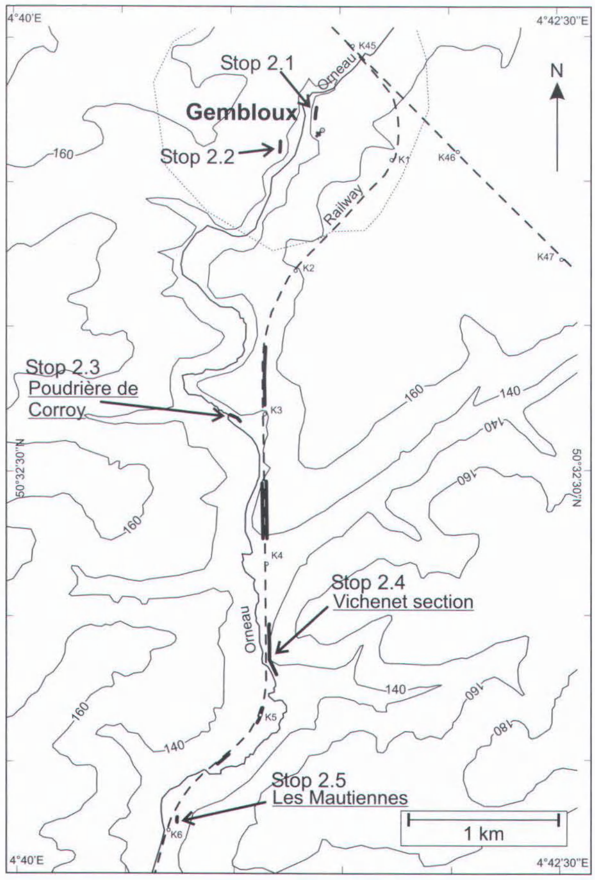


**Figure 18**. Simplified topographic map of the Orneau valley south of Gembloux, showing the position of the three visited outcrops. Also the position of other outcrops is shown (thick black lines).

um grey mudstone, interpreted as an ichnofossil (*Chondrites* sp.). The topmost part of the formation is a fine-grained quartzitic tuff, very hard and white, light pink or yellow in colour and is called the "Eurite of Grand-Manil Member". By weathering it was transformed into kaolinite and was locally exploited for the ceramic industry.

Sedimentology. The fine-grained sediments with occasional benthic fauna and bioturbation could indicate a deep shelf.

Thickness. In the Orneau Valley 40 m for the "Eurite" de Grand-Manil Member and between 80 and 100 m for the whole formation (Delcambre *et al.*, 2002).

Age. Trilobites in the lower/middle part of the formation were described by Malaise (1903) without using them for dating. The chitinozoans from the mottled mudstone member are dominated by *Belonechitina* cf. *gamachiana*, indicating a middle or late Ashgill, possibly Hirnantian age (Samuelsson & Verniers, 1999, 2000). Graptolites described from the top of the formation in slates below the "Eurite of Grand-Manil member" in the Orneau Valley, belong to the *Coronograptus cyphus* Biozone (Elles *in* Maillieux, 1930a). Graptolite collections in the same levels mentioned in Gerlache (1956) and determined by Bulman (1950) as *Climacograptus scalaris* indicate the *Parakidograptus acuminatus* Biozone (the basal graptolite biozone of the Rhuddanian) or slightly above or below. A middle or late Ashgill to Rhuddanian age is proposed for the formation.

### *5.3. Bois Grand-Père Formation*

The formation was also created by Delcambre *et al.* (2002) and contains the middle part of the previous "Assise de Grand-Manil". Its stratotype is on private grounds in two small abandoned quarries and outcrops, a few hundred meters south of the Tri à la Vigne and Brûtia.

Description. Mostly medium to dark grey shale or slate unit, dark greenish grey at the base; at certain levels interbedding of many medium grey laminated siltstone beds and light coloured quartzitic very fine sandstone, a few to 20 cm thick. Some beds can be calcareous. The lower boundary is taken just above the "Eurite of the Grand-Manil member", with the upper boundary not observed.

Sedimentology. The sandstone beds show lamination or fine oblique stratification with undulating bedding planes; at least partly deposited as turbidites.

Thickness. About 200 m (Delcambre *et al.*, 2002); estimated between 375 and 500 m (Verniers *et al.*, 2001).

Age. Graptolites have been described from the *Coronograptus cyphus*, *Monograptus convolutus* and *Monograptus crispus* Biozones (Maillieux, 1930a, 1930b, 1933; Michot, 1930). A middle Rhuddanian to middle Telychian age is tentatively proposed (Verniers *et al.*, 2001). One of the three Chitinozoa samples contains only a few specimens, but amongst them *Conochitina* spp. cf. *Vitreachitina* sp. 2 Nestor, 1994 indicates a Telychian age (Nestor, 1994).

### *5.4. Fallais Formation*

The formation informally defined by Verniers (1976, 1983a) in the Mehaigne Valley, around the village of Fallais is poorly outcropping in the Orneau Valley.

Description. Light green, olive-greenish grey, or light grey chloritic mudslate and mudstone, with rare siltstone and fine sandstone beds from distal turbiditic ori-

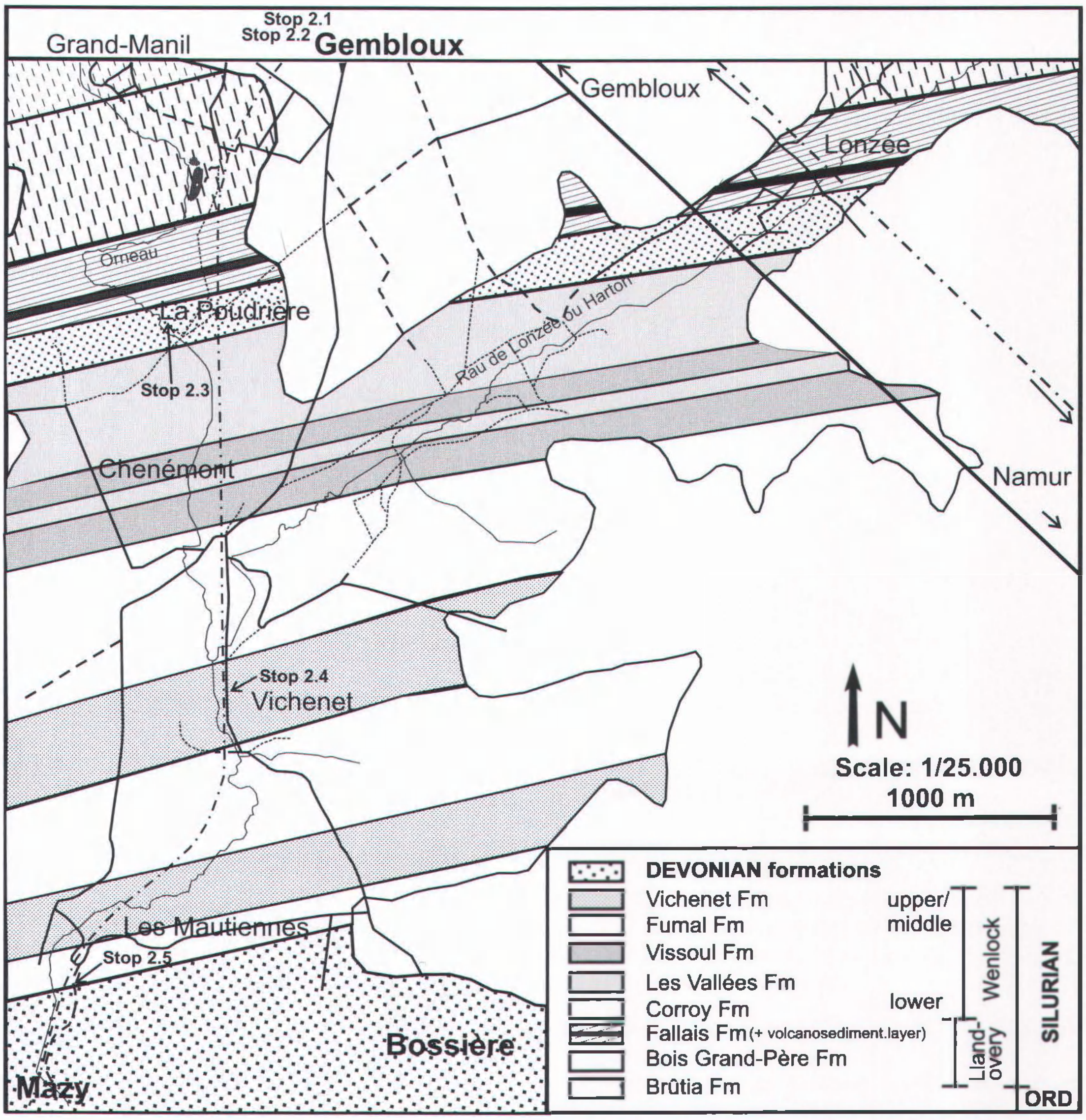


**Figure 19**. Geological map of the Orneau Valley after De Schepper (2000).

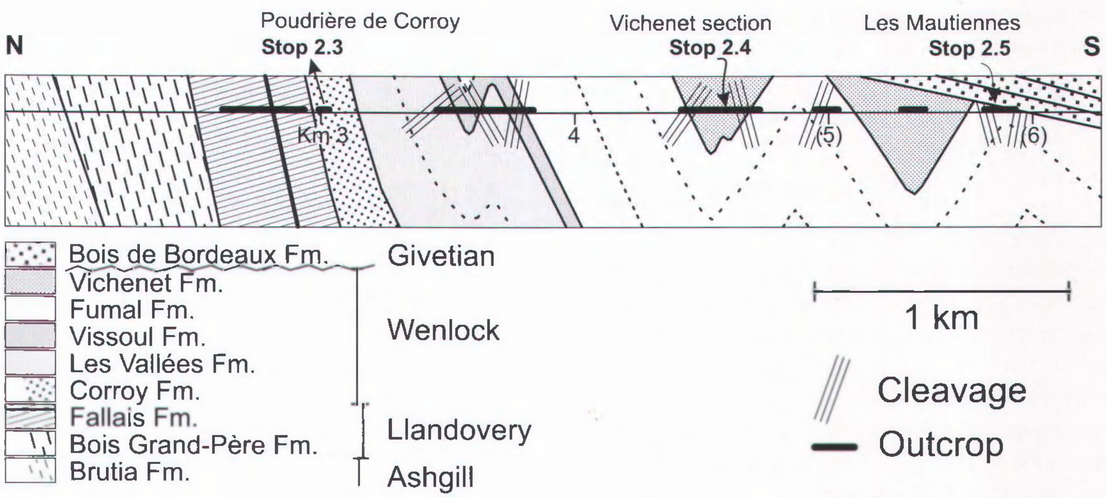


**Figure 20**. N-S cross-section of the Orneau valley south of Gembloux, after De Schepper (2000), Belmans (2000) and Debacker (2001) (vertical exageration not to scale).

gin, mostly without laminated hemipelagite. In the Mehaigne area six members occur with a 26 m thick volcanic layer near the top, the volcano-sedimentary layer of Pitet. In the Orneau Valley along the railway cut at km 2.809 a similar volcano-sedimentary rock is present, only a few meters thick and located at about 20 m below the top of the formation.

Sedimentology. The unit contains typical Bouma-type turbidites, with mostly distal Tde and more rarely Tcde divisions and no laminated hemipelagite present. A turbiditic basin is hence deduced. The unit is relatively thick in the Mehaigne area (estimated at 600 m) for a relatively short period of time.

Thickness. Difficult to measure in the Orneau Valley but estimated at 300-400 m (Delcambre *et al.*, 2002).

Age. Undated in the Orneau Valley. Telychian, late Llandovery, based on acritarchs in the Sennette Valley (Martin, 1969a). The chitinozoans from the Mehaigne area indicate for the lower half of the formation the *Angochitina longicollis* global Biozone which is calibrated versus the graptolite biozonation from the post-*Monoclimacis griestoniensis* to the pre-*Cyrtograptus insectus* Biozones. The *Margachitina margaritana* global Biozone is observed in the upper half. According to the calibration by Mullins (1998) the upper half corresponds to the *Cyrtograptus insectus* graptolite Biozone, the uppermost biozone of the Telychian, late Llandovery (Verniers *et al.*, 2001).

### *5.5. Corroy Formation*

The formation as defined by Malaise (1900) and later by Legrand (1961) is restricted by Verniers (1982), Verniers & Van Grootel (1991) and Verniers *et al.* (2001) to its lower third part. The type locality is situated in the abandoned quarry *La Poudrière de Corroy*, near the Orneau river, 150 m west of km 3.050 of the railway (stop 2.3).

Description. Mudslate, mudstone, siltstone and fine sandstone, alternating in thin beds, decimetric to a few cm thick. The sandstone is light-coloured, obliquely stratified, sometimes convolute bedding, with often undulating base and bounce and other current marks. Interbedded are dark grey laminated hemipelagite layers containing most of the graptolite levels. Some of the beds are slightly calcareous. The mudslate is greenish grey in the lower part of the formation, indicating its softer, chloritic composition. The colour changes to dark grey in the upper part of the formation, indicating a more quartzic or illitic composition. The lower boundary occurs via a gradual transition from the Fallais Formation and is marked by the lowest presence of at least three, decimetric, sandstone beds per meter of sediment. The upper boundary is marked by the highest sandstone bed of more than 5 cm thickness.

Sedimentology. Typical Bouma-type turbidites, with mostly Tde sequences, frequent to very frequent T(b)cde sequences, generally twice as thick as the average Tde sequence, with sandstone Tc-divisions in thickness between 2 and 12 cm. The turbidites alternate with laminated hemipelagites indicating an autochthonous sedimentation in the basin. Throughout the outcrop area of the Brabant Massif there is an overall tendency in this formation to thinner turbidite sequences from east to west (Verniers *et al.*, 2001). The sudden increase of energetic turbidites in the formation in relation to the under- and overlying formation indicates a more shallow period which is also observed on the Baltica platform as e.g. in Gotland (Laufeld, 1979). No eustatic sea-level low is observed at this time (see Johnson *et al.*, 1985; Kaljo *et al.* 1995), and hence the shallowing event is probably caused by a tectonic event affecting

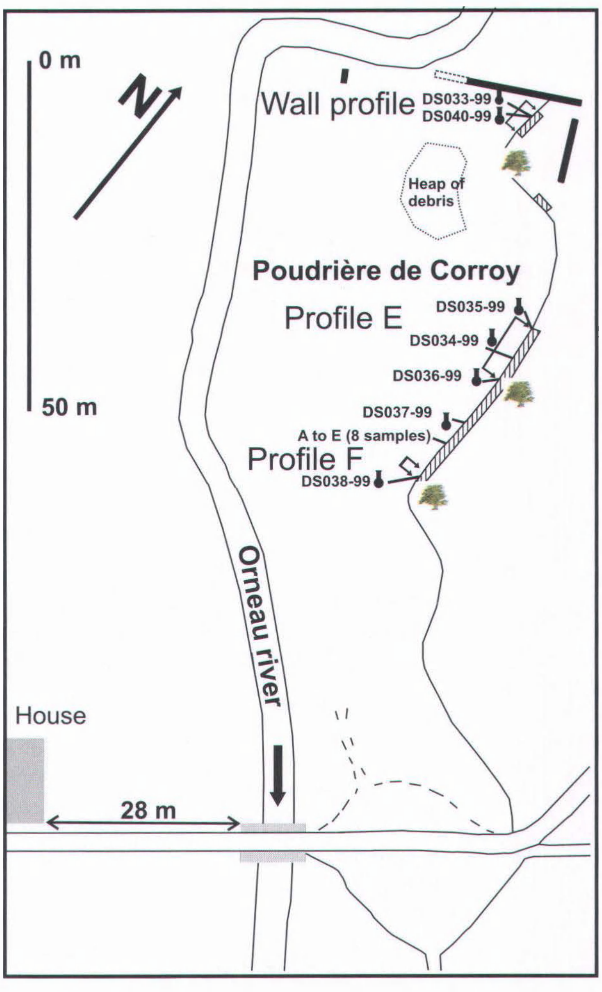


**Figure 21**. Location map of the samples for chitinozoan studies in the Poudrière de Corroy, type locality of the Corroy Formation, after De Schepper (2000).

**Figure 22.** Composition of chitinozoan assemblages and concentrations in samples of the Silurian of the Orneau valley after De Schepper (2000).

| Formations | Orneau | *Ancyrochitina ancyrea* | *Ancyrochitina* spp. | *Angochitina longicollis* | *Bursachitina* sp. A in Sutherland (1994) | *Cingulochitina burdinalensis* | *Cingulochitina cingulata* | *Cingulochitina dreyensis* | *Cingulochitina ithoniensis* | *Cingulochitina pitetensis* | *Cingulochitina* spp. | *Conochitina acuminata* | *Conochitina claviformis* | *Conochitina fortis* | *Conochitina proboscifera* | *Conochitina subcyatha* | *Conochitina truncata* | *Conochitina tuba* | *Conochitina tuba?* in Paris (1981) | *Conochitina visbyensis* | *Conochitina* spp. | *Desmochitina opaca* | *Desmochitina* spp. | *Eisenackitina anulifera* | *Eisenackitina causiata* | *Eisenackitina* spp. | *Margachitina margaritana* | *Salopochitina bella* | *Sphaerochitina jaegeri* | *Sphaerochitina serpaglii* | *Sphaerochitina* sp. B in Verniers (1982) | *Sphaerochitina* spp. | *Conochitina* spp. cf. *Vitreachitina* sp. 2 | Chitinozoa spp. | Total number of chitinozoans | N° of specimens determined to genus level | N° specimens determined to species level | Weight of dissolved rock (g) | Number of chitinozoans per gram of rock | Acritarchs |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Vichenet | Sect. B: DS-027-99 | 3? | | | 3 | | 9 | | | | 1 | | 3 | 1? | | 1 | | 5 | | | 6 | | 1 | | | 2 | | | 2 | 4 | | | | 4 | 41 | 10 | 27 | 66.31 | 0.56 | >400 |
| | Sect B: DS-026-99 | | | | | | | | | | | | 1 | 3 | | | | 2 | | | 10 | | | | | | | | | | | | | 1 | 17 | 10 | 6 | 27.15 | 0.59 | >100 |
| | Sect. A: DS-025-99 | | | | | | | | | | | | | 1 | | | | | | | | | | | | | | | | | | | | 1 | 2 | 0 | 1 | 25.91 | 0.04 | >7 |
| | Sect. A: L9 base | | | | | | 1 | | | | | | | 1? | | | | | | | 2 | | | | | 2 | | | | | | | | 1 | 6 | 6 | 2 | 31.04 | 0.23 | x |
| | Sect. A: LHP 8/9 | | | | | | | | | | | | | 2 | | 1 | | 2 | | | | | | | | | | | | | | | | 0 | 5 | 3 | 5 | 29.9 | 0.17 | x |
| | Sect. A: L8 top | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 29.92 | 0.00 | x |
| | Sect. A: DS-024-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 35.27 | 0.00 | 0 |
| | Sect. C: DS-029-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 5 | 5 | 0 | 0 | 13.19 | 0.00 | ? |
| | Sect. C: DS-028-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 1 | 1 | 0 | 0 | 14.25 | 0.00 | >5 |
| | Sect. D: DS-031-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 12.95 | 0.00 | 0 |
| | Sect. D: DS-032-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 14.21 | 0.00 | >100 |
| Fumal | DS-052-99 | | | | | | | | | | | | | | | | | | | | 1 | | | | | | | | | | | | | | 1 | 1 | 0 | 12.83 | 0.08 | >100 |
| | DS-054-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 12.54 | 0.00 | 10 |
| Vissoul | DS-051-99 | | 1 | | | | 1 | | | | | | | 2 | | | | | | | 1 | | | | | | | | | | | | | 3 | 8 | 2 | 3 | 12.16 | 0.41 | >15 |
| | DS-055-99 | | | | | | 13 | | | | | | | 3 | | | | | | | | | | | | | | | | | | | | 3 | 19 | 0 | 16 | 12.97 | 1.23 | >100 |
| Les Vallées | DS-050-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 12.22 | 0.00 | >250 |
| Corroy | DS-038-99 | | | | | 9 | | | | | | | | | 1 | | 1 | | | | 1 | | | | 2 | | | 1 | | | | 1 | | 4 | 20 | 2 | 14 | 11.78 | 1.36 | x |
| | E Te (01-0814) | | | | | 3 | | | | | | 2 | | | | | | | | | 2 | | | | 1 | | | | 3 | | | | | 4 | 15 | 9 | 11 | 12.42 | 1.29 | x |
| | E Tc (01-0815) | 1 | | | | | | | | | | | | | | | | | | | 3 | | | 1 | | | | | | | | | | 2 | 7 | 2 | 5 | 12.41 | 0.56 | x |
| | D Td (01-0813) | 2 | | | | 4 | | | | | | | | | | | | | | | 2 | | | | 2 | | | | | | | | | 5 | 15 | 7 | 10 | 11.72 | 1.37 | x |
| | C Te (01-0816) | | | 2 | | 4 | | 2 | | | | 5 | | | | | 2 | | | | 10 | | | 6 | 9 | | | 1 | 2 | | | 1 | | 7 | 51 | 34 | 44 | 19.06 | 2.83 | x |
| | C Tc (01-0817) | 1 | | 3 | | 1 | | | 1 | | | | | | 1 | | 11 | | | | 26 | 7 | | 5 | 9 | | | | | | | | | 4 | 69 | 39 | 65 | 19.91 | 7.03 | x |
| | B Te (01-0812) | | | | | 5 | | | | | | 10 | | | | | | | | | 8 | | | 2 | 2 | | 1 | | | | | | | 2 | 30 | 20 | 28 | 22.6 | 1.50 | x |
| | B LHP (01-0806) | | 4 | | | 2 | | | | | | 2 | | | | | | | | | | | | 3 | | | | | | | | | | 0 | 11 | 7 | 11 | 23.43 | 0.30 | x |
| | A Te (01-0805) | | | | | 2 | | | | | | | | | | | | | | | 3 | | | | | | | | | | | | | 1 | 6 | 2 | 5 | 22 | 0.27 | x |
| | DS-037-99 | 1? | | 8 | | 2 | | 4 | | 1 | | 6 | | | 3 | | 2 | | 1 | | 23 | | | 3 | 3 | | | | 2 | | | 2 | | 5 | 65 | 25 | 35 | 52.88 | 1.13 | >200 |
| | DS-036-99 | | | | | | | 2 | | | | 2 | | | | | | | | | 3 | 1? | 1 | 1 | 3 | | 2 | | | | | | | 3 | 17 | 4 | 10 | 12.35 | 1.13 | >500 |
| | DS-034-99 | | | | | | | 1 | | | | | | | | | | | | | 3 | | | | | | | | | | | | | 0 | 4 | 3 | 1 | 12.13 | 0.33 | 16 |
| | DS-035-99 | | | 2 | | | | 2 | | | | 5 | | | 2 | | | | | | 5 | | | | | 1 | | | | | | 1 | | 4 | 22 | 7 | 11 | 12.41 | 1.45 | 3 |
| | DS-033-99 | | 2 | 8 | | 6 | | 19 | | 4 | | 27 | | | 17 | | 8 | 17 | | 2 | 11 | | | 2 | 8 | | | | | | | 1 | | 51 | 183 | 14 | 118 | 52.69 | 2.51 | >100 |
| | DS-040-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 0 | 12.19 | 0.00 | 0 |
| Fallais | DS-039-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 11.94 | 0.00 | 0 |
| Bois Grand-Père | DS-046-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 12.68 | 0.00 | 0 |
| | DS-042-99 | | | | | | | | | | | | | | | | | | | | 1 | | | | | 1 | | | | | 1 | 1 | 2 | | 6 | 3 | 3 | 54.36 | 0.11 | >350 |
| | DS-041-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 13.2 | 0.00 | 0 |
| | DS-049-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 0 | 12.81 | 0.00 | 0 |
| Brutia | DS-048-99 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 1 | 1 | 0 | 0 | 12.58 | 0.00 | 0 |
| | Total | 4 | 7 | 23 | 3 | 38 | 24 | 30 | 1 | 5 | 1 | 59 | 4 | 11 | 24 | 2 | 24 | 26 | 1 | 2 | 121 | 7 | 2 | 23 | 39 | 6 | 3 | 2 | 9 | 4 | 1 | 7 | 2 | 112 | 627 | 210 | 431 | 790.37 | 0.79 | |

both areas and linked with the Scandian orogeny. During the excursion M.P. Dabard (Rennes) and A. Loi (Cagliari) criticised the turbiditic nature of this formation and indicated features suggesting a tempestite sedimentation. A detailed sedimentological study of the outcrop is hence needed.

Thickness. About 100 m (Verniers *et al.*, 2001), with at least 53 m present in the quarry *Poudrière de Corroy* representing the upper part of the formation but probably not its top (De Schepper, 2000).

Age. Graptolites from the *Poudrière de Corroy* belong to the *Cyrtograptus murchisoni* and *Monograptus riccartonensis* Biozones (Malaise, 1900; Legrand, 1961); graptolites in the Mehaigne area indicate the presence of the *Cyrtograptus centrifugus* and *Cyrtograptus murchisoni* Biozones in the lower part of the formation, and the upper part of the *Monograptus riccartonensis* Biozone and the middle Wenlock *Monograptus flexilis* Biozones (now *Monograptus dubius* Biozone) in the upper part of the formation: lower to middle Sheinwoodian (lower and basal middle Wenlock) (Verniers & Rickards, 1979).

The results of the recent chitinozoan study in the *Poudrière de Corroy* by De Schepper (2000) on fourteen samples are shown on Fig. 23 &24. The chitinozoans for this part of the formation were described as belonging to subzone C3 (Verniers, 1981, 1982), to the *Margachitina margaritana* global Biozone of Verniers *et al.* (1995) and to the *Cingulochitina burdinalensis* local biozone of Verniers (1999). The *Margachitina margaritana* global Biozone starts in the uppermost Telychian in the *Cyrtograptus insectus* graptolite Biozone (Mullins, 1998). The former chitinozoan biozone spans the Wenlock-Llandovery boundary and hence this boundary cannot be located exactly with chitinozoans. The presence of Wenlock graptolites close to the base of the formation however suggests that the base of the Wenlock coincides or lies close to the base of the Corroy Formation.

### 5.6. *Les Vallées Formation*

The formation is defined in the Mehaigne and Burdinale valleys by Verniers (1976), Verniers & Van Grootel (1991) and Verniers *et al.* (2001) where mostly the lower part is visible, while the upper part of the formation is visible in the northern part of the railway section of the Orneau Valley, 350-500 m east and northeast of the "Ferme de Chenémont" (parastratotype section). The unit was considered earlier as the middle part of the Corroy Formation (Legrand, 1961; Verniers, 1983b). It is not visited during the excursion.

Description. Grey mudslate, mudstone, siltstone and fine sandstone with non-calcareous quartzic pelite in the Te-divisions in 10 to 40 cm thick Bouma-sequences.

Sedimentology. More energetic but still distal turbidites, alternating with thin-bedded laminated hemipelagites.

Thickness. Estimated at more than 225 m.

Age. middle Wenlock, based on the relative position of the formation: above the formation which at its top contains graptolites of a middle Wenlock post-*Monograptus riccartonensis* Biozone and chitinozoans of the middle Wenlock and below a formation with middle Wenlock graptolites (Verniers, 1982, 1999).

### 5.7. *Vissoul Formation*

The unit was considered as the upper part of the Corroy Formation by Malaise (1910) and Legrand (1961). The very different sedimentology in comparison with the Corroy Formation in its type locality, the *Poudrière de Corroy*, was the reason why this new unit was defined by Verniers (1976, 1983a, b) (see Verniers & Van Grootel, 1991; Verniers *et al.*, 2001).

Description. Grey mudslate, mudstone, siltstone and fine sandstone with non-calcareous quartzic pelite in the Te-division.

Sedimentology. Distal to slightly proximal turbidites alternating with thin-bedded laminated hemipelagites; the Tde sequences are medium thick and the T(b)cde sequences are frequent (about 15% of all sequences) and normally about 50% thicker than the average Tde sequences; the Tc-divisions are thin (2-8 cm). This formation forms a clearly different sedimentology from the adjacent units Les Vallées and Fumal formations.

Thickness. Estimated at more than 30 m.

Age. Graptolites from the Burdinale Valley are not characteristic for one particular biozone but the assemblage indicates a range from the *Pristiograptus dubius* to the *Cyrtograptus lundgreni* biozones, middle Sheinwoodian to early Homerian (middle to late Wenlock) (Verniers & Rickards, 1979; Zalasiewicz *et al.*, 1998); the chitinozoans belong to subzone D1 (Verniers, 1982) and to the *Cingulochitina cingulata* global Biozone or possibly the *Conochitina pachycephala* global Biozone of Verniers *et al.* (1995), spanning the same interval as indicated by the graptolites.

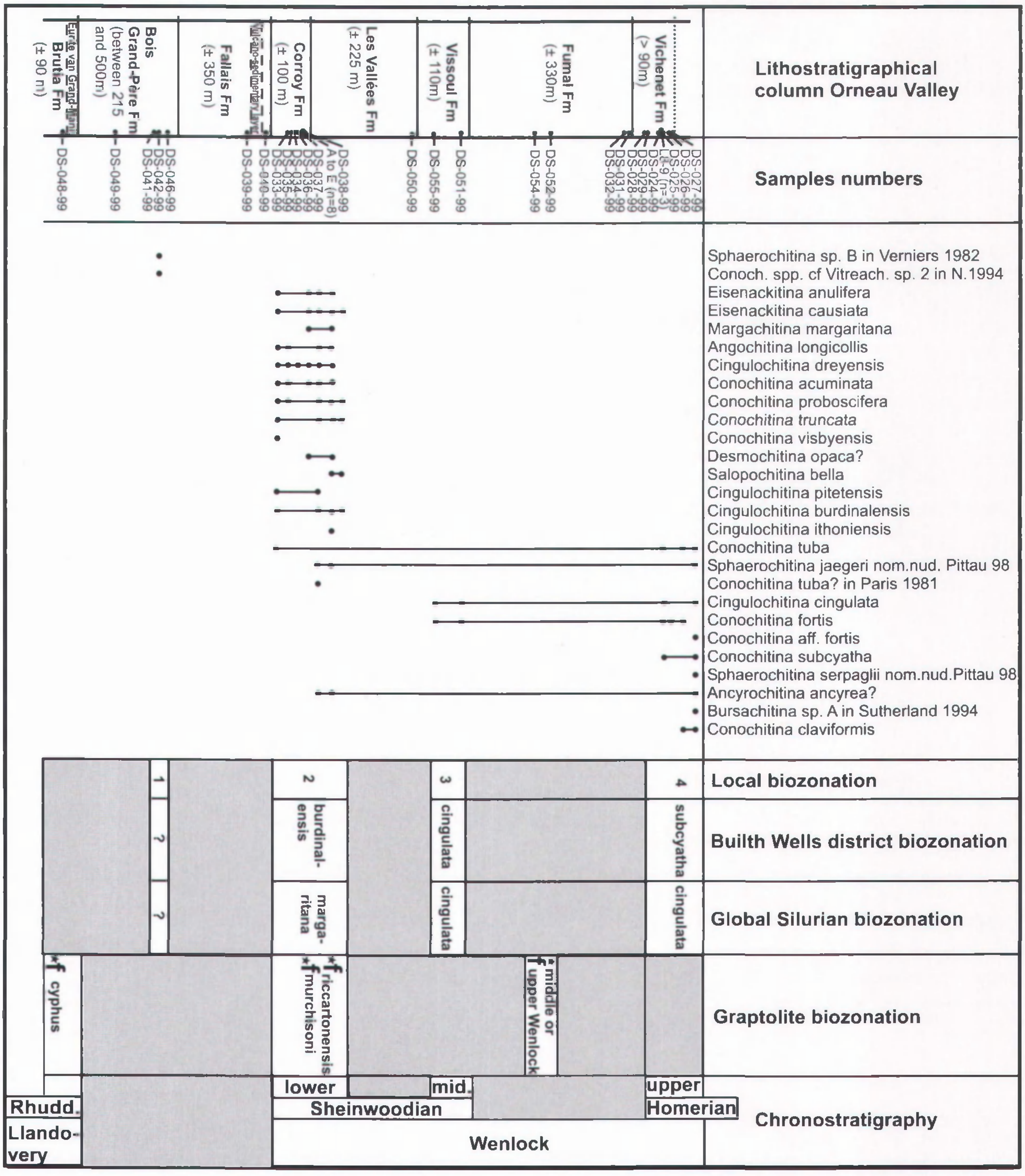


**Figure 23**. Ranges of selected chitinozoan species in the Silurian of the Orneau valley after De Schepper (2000), with local biozonation, correlation with the Builth Wells district (Verniers, 1999), with the global Silurian biozonation (Verniers *et al.* 1995), with the graptolite biozonation (see text) and with the deduced chronostratigraphy.

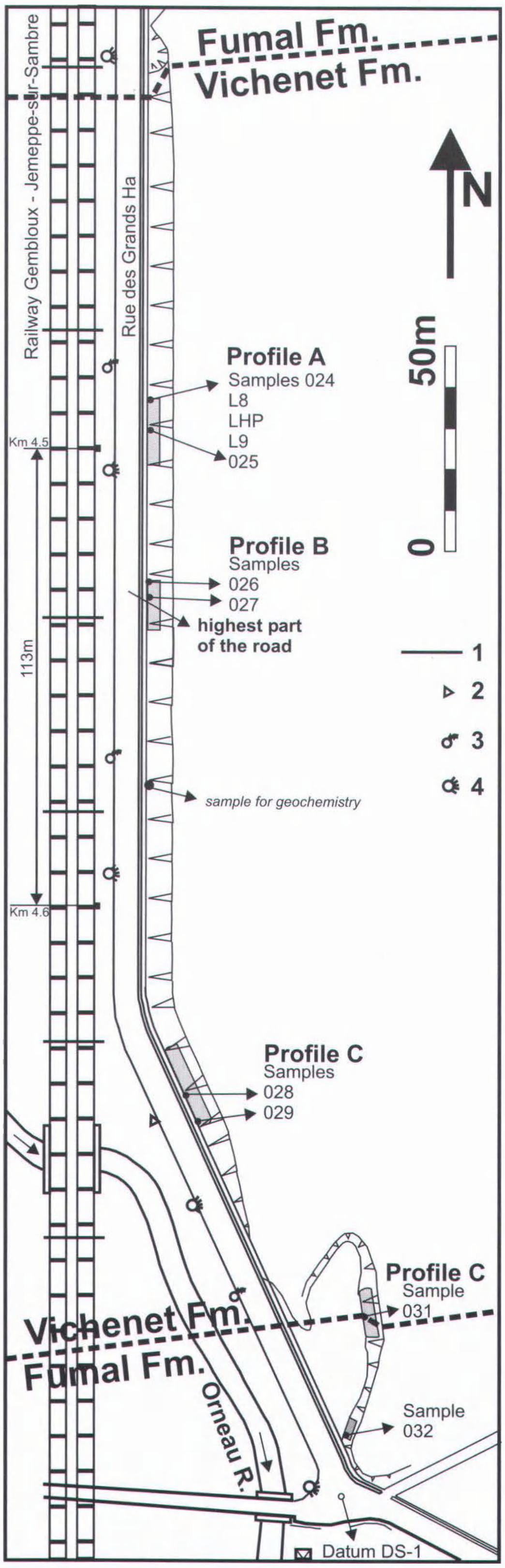


**Figure 24.** Location map of the Vichenet section, type locality of the Vichenet Formation and of the samples for chitinozoan studies after De Schepper (2000); compare with section in Fig. 25; 1: road sign, 2: electricity pole, 3: pole for street lighting.

### *5.8. Fumal Formation*

The formation resembles much the Les Vallées Formation and can only be distinguished by its relation towards the neighbouring formations. The name was informally proposed by Michot (1957), Verniers (1976, 1983a) and Verniers & Van Grootel (1991) and formalised in Verniers *et al.* (2001).

Description. Grey mudslate, mudstone, siltstone and fine sandstone with quartzic pelite in the Te-divisions.

Sedimentology. Three types of sequential patterns occur: (a) thick Tde sequences (average between 24 and 28 cm) with absent or very rare T(b)cde sequences; (b) medium thick to thick Tde sequences (average between 17 and 25 cm) with rarely present to very frequent T(b)cde sequences (6 to 30% of all sequences) of about the same thickness as the Tde sequences; the c-divisions are either thicker or thinner than 10 cm but there is a higher frequency (80%) of c-divisions thinner than 10 cm; (c) medium thick to thick Tde sequences (average between 14 and 18 cm) and no T(b)cde sequences. Lower and upper boundaries are not observed in the Mehaigne area: the upper boundary is situated in an observation gap of ± 20 m between this formation and the Vichenet Formation.

Stratotype. Stratotype sections south-east of the church of Fumal; parastratotype in the outcrops just south to south-west of the church of Fumal in the Mehaigne Valley.

Thickness. Estimated in the Mehaigne area: 330 m (Verniers, 1983a).

Age. Middle Wenlock to early Homerian. The chitinozoans, the only fossils found in this formation in the Mehaigne area, belong to the *Cingulochitina cingulata* global Biozone and possibly the *Conochitina pachycephala* global Biozone of Verniers *et al.* (1995), middle Sheinwoodian to lower Homerian (middle to late Wenlock). In the Mehaigne area most of the upper two thirds of the formation can be assigned to the Welsh *Conochitina subcyatha* local biozone, lower Homerian (late Wenlock), and the uppermost part of the formation to the lower or middle part of the Homerian (late Wenlock) (Verniers, 1999).

### *5.9. Vichenet Formation*

Unit created by Malaise (1910) with type locality in a 350 m long road section north of the abandoned railway station of Vichenet-Bossière, east of the castle of Vichenet in the Orneau Valley. It was meant to represent the Ludlow in the Brabant Massif. A graptolite locality described by him, could not be located, nor could the fossils of his collection. Michot (1954) proposed the name Ronquières Formation for the Ludlow and abandoned the term "Assise de Vichenet". The unit was informally reinstalled by Verniers (1976, 1983a), Verniers & Van Grootel (1991) and formally by Verniers *et al.* (2001). The section exposes only the lower part of the formation while the higher parts seem to be missing in this valley.

Description. Grey mudslate, mudstone, siltstone and rare fine sandstone with often calcareous quartzic-chloritic pelite in the thick Te-divisions. Characteristic are the very thick Te or Tde turbidite sequences and very rare or absent Tcde sequences, the grey to greenish colour, its sometimes calcareous content up to 5%, alternating with laminated hemipelagite beds. De Schepper (2000) placed the lower limit of the formation in the type section above the highest sandstone bed of a Tcde sequence.

Sedimentology. Alternating distal thick-bedded turbidites with thin-bedded laminated hemipelagites.

Thickness. Measured in the type locality at about 90 m for the basal part of the formation by De Schepper (2000); in the Mehaigne area estimated at more than 210 m (Verniers, 1983a) and in the Landenne outcrop area at more than 350 m (De Winter, 1998).

Age. Until recently no fossils were found in the Orneau Valley and the age was estimated middle to upper Wenlock, based on its relative position in relation to a fossil locality below the unit (Rickards, pers comm. *in* Verniers, 1983b). Chitinozoans studied by De Schepper (2000) presented here (Figs. 23 & 24) indicate a lower to middle part of the Homerian (late Wenlock). In the Mehaigne area only chitinozoans are found and limited to the middle part of the formation. The assemblage belongs to the *Sphaerochitina lycoperdoides* global Biozone of Verniers *et al.* (1995), lower to middle part of the Homerian (upper Wenlock) (Verniers, 1982, 1999). In the Landenne area the chitinozoans belong to the same biozone for most of the formation and for the lower part of the formation to the *Cingulochitina cingulata* local biozone, middle of the Sheinwoodian (middle Wenlock) to the middle Homerian, late Wenlock (De Winter, 1998).

### *5.10. Structural geology of the Silurian in the Orneau Valley*

Like the southern Sennette Valley (cf. Legrand, 1967), also the Silurian in the southern Orneau Valley shows folds with a pronounced convergent cleavage fanning (Fourmarier, 1921; Kaisin, 1933; Mortelmans, 1954). Mortelmans (1954; cf. Kaisin, 1933) interpreted these fans as the result of two deformations: a first-phase Caledonian deformation, responsible for folding and cleavage development, followed by a Variscan overprint, causing a post-cleavage fold amplification accompanied by a passive rotation of the cleavage in the fold limbs, hence resulting in convergent cleavage fans. However, following the construction of the inclined ship lift at Ronquières, Legrand (1967) demonstrated that, at least in the Sennette Valley, this hypothesis is highly questionable, since the Silurian-Devonian unconformity can be seen truncating the convergent cleavage fans. Instead, Legrand put forward a two-phase Caledonian origin for the convergent cleavage fans. More recently, partly in accordance with Legrand (1967), the convergent cleavage fans in the Ronquières area were explained by means of a single-phase progressive deformation, consisting of a continued fold amplification after cleavage development (Debacker *et al.*, 1999). However, although Legrand (1967) questioned the influence of the Variscan deformation in the southern rim of the Brabant Massif, a Variscan origin for the convergent cleavage fans and kink bands was still advocated in the Orneau Valley by Vandenven (1967).

The origin of the convergent cleavage fans in the southern Orneau Valley, was studied structurally in an unpublished M.Sc. thesis by Belmans (2000).

#### 5.10.1. Les Mautiennes

The outcrop shows a steeply inclined, gently W-plunging antiform with a slightly N-verging asymmetry and a well-developed convergent cleavage fan, truncated by the Middle Devonian angular unconformity (Fig. 20). The angular unconformity is sharp and straight. Detailed observations show that, in contrast to the opinion of Kaisin (1933), there is no evidence of slip along the unconformity. In this respect, the observations of Belmans (2000) in the Orneau Valley point to a similar situation as in Ronquières (cf. Legrand, 1967) and, contradicting the models of Kaisin (1933), Mortelmans (1954) and Vandenven (1967), they point to a pre-Givetian origin of the convergent cleavage fans.

#### 5.10.2. Vichenet

This outcrop is the most accessible and one of the longest outcrops of the Orneau Valley. It contains a large open upright, gently W-plunging synform, comprising several smaller folds, which are cross-cut by small faults. All the folds show a convergent cleavage fanning. However the cleavage fan angle does not seem to be related to the fold interlimb angle: tighter folds may have a less developed convergent cleavage fan than more open folds (Fig. 25; cf. Fig. 20).

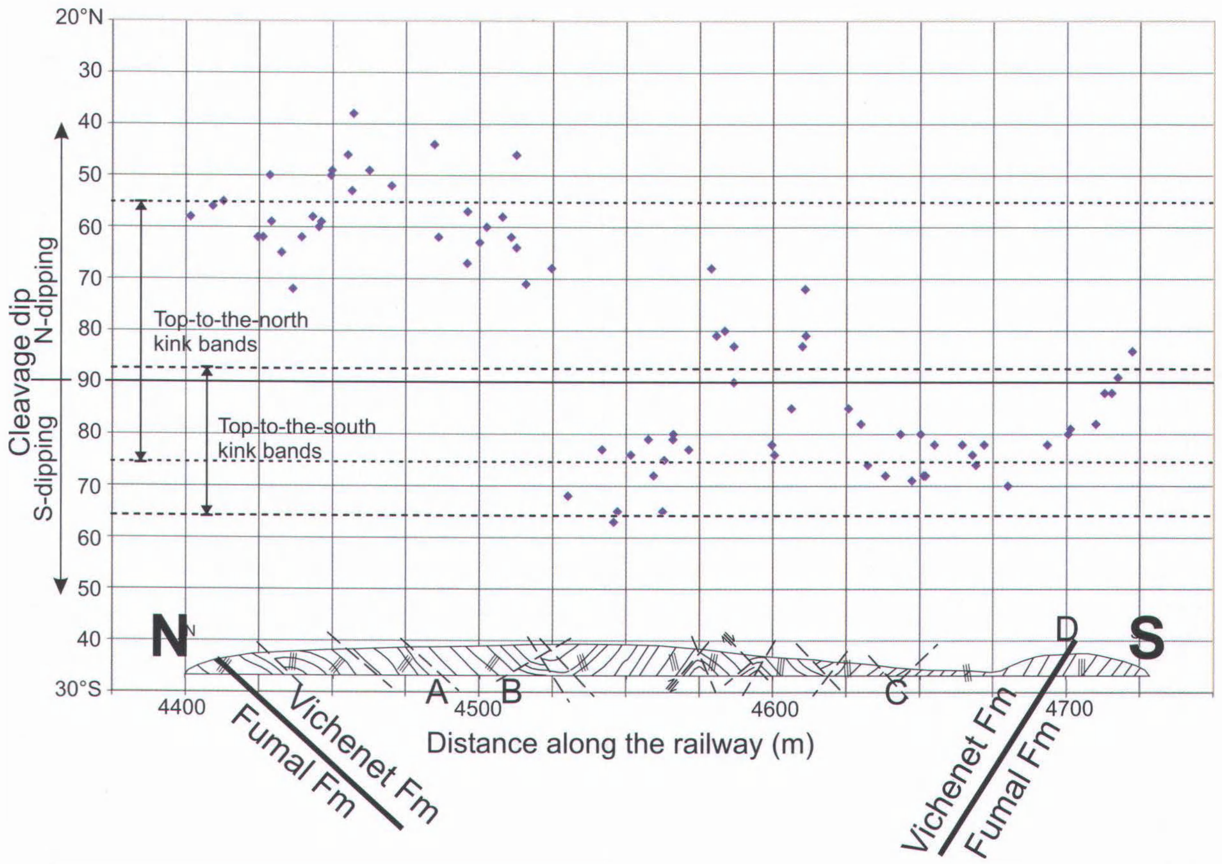


**Figure 25**. Graph showing the change in cleavage dip along the Vichenet section. Below a cross-section along the Vichenet section, oriented parallel to the railway, is added for better orientation. Cleavage data and the cross-section are taken from Belmans (2000). Note the change in cleavage dip across the folds, reflecting a pronounced convergent cleavage fanning. Also shown are the cleavage dip intervals in which top-to-the-north and top-to-the-south kink bands are encountered. Conjugate sets are expected, and occasionally encountered, in the zone of overlap, between cleavage dips 87°N and 64°S. A, B and C: visited places in the outcrops (figure after Debacker, 2001).

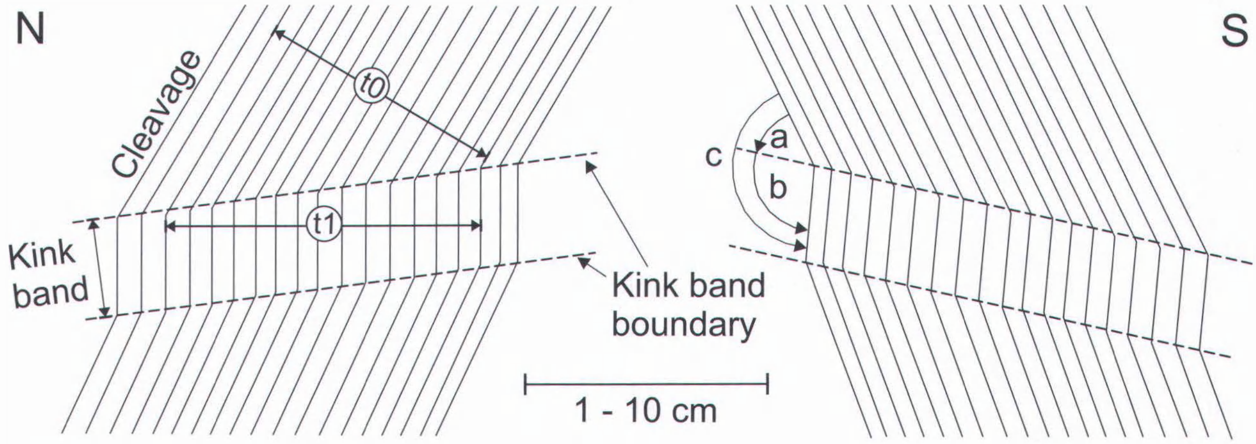


**Figure 26**. Top-to-the-south and top-to-the-north contractional kink bands from the Vichenet section. The internal geometry of most of the kink bands points to a volume increase inside the kink band (compare thickness t0 with thickness t1), which can be calculated by means of the formula (sin b/sin a) – 1. Figure after Debacker (2001).

The fold train is cross-cut by numerous faults, post-dating the cleavage. The faults have their highest concentration in this outcrop. Due to the scarcity of fault striations and other kinematic indicators and the rather homogeneous lithology and the resulting lack of marker horizons, the sense and amount of fault displacement is difficult to determine. Nevertheless, it appears that both S- and N-dipping, normal and reverse E-W-trending faults occur, some apparently forming conjugate sets. The faults usually occur in the hinge zones of the smaller folds, possibly indicating a relationship between fold geometry and faulting (Fig. 25). A zone of turbidite sequences with cross-bedded c-divisions (De Schepper, 2000) was used by Belmans (2000) to correlate both limbs of the large-scale synform in the section. From this analysis, it follows that the net result of the normal and reverse faults is hardly reflected across the synform. Hence, there is no evidence for important faults.

Like for the faults, the highest concentration of kink bands is also found in this outcrop. The kink bands have sub-horizontal to gently, occasionally moderately dipping kink band boundaries, and sub-horizontal to gently plunging, E-W-trending kink axes. As previously noted by Vandenven (1967), kink bands with a top-to-the-south-geometry as well as kink bands with a top-to-the-north-geometry, occur. Vandenven (1967) argued that the former type of kink bands only occur in zones with a S-dipping cleavage, whereas the latter are restricted to zones with a N-dipping cleavage. He concluded that the kink bands were formed during fanning of the cleavage. Inspired by Mortelmans (1954), he attributed both the fanning and the kink bands to a second deformation phase of Variscan origin. However, although most kink bands are compatible with the generalisation of Vandenven (1967), Belmans (2000) also observed top-to-the-south kink bands in zones with sub-vertical to steeply N-dipping cleavage and top-to-the-north kink bands in zones with a sub-vertical to steeply S-dipping cleavage (respectively termed top-to-the-south, B and top-to-the-north, B by Debacker, 2001). As such, based on kink band geometry and position, Belmans (2000) was able to distinguish four different types of kink bands, two with a top-to-the-south geometry (called top-to-the-south, A and top-to-the-south, B by Debacker, 2001) and two with a top-to-the-north geometry (called top-to-the-north, A and top-to-the-north, B by Debacker, 2001). Locally, conjugate sets may be observed (e.g. top-to-the-south, A kink band and top-to-the-north, B kink band in zone of steeply S-dipping cleavage). All the kink bands are of a contractional kind and reflect development under the influence of a steeply plunging shortening with a N-S-directed extension. Because of the seemingly preferred orientation of the top-to-the-south and top-to-the-north kink bands with respect to cleavage dip, these structures formed after development of the convergent cleavage fans (Fig. 25). Conjugate kink bands develop where the shortening is parallel to the anisotropy (e.g. Price & Cosgrove, 1990). As shown on Figure 25, top-to-the-south and top-to-the-north kink bands (and occasional conjugate set), both occur in zones of sub-vertical to steeply S-dipping cleavage. Hence it is suggested that the kink bands in the southern Orneau Valley formed after development of the convergent cleavage fans, under the influence of a sub-vertical to steeply S-plunging shortening. The preferential orientation of the different types of kink bands may partly be attributed to the orientation of the cleavage within the convergent cleavage fans and partly to a change in local strain or stress orientation under the influence of bedding orientation and fault tip zones.

### 5.10.3. Discussion and conclusion

The truncation of the folds and the well-developed convergent cleavage fans by the angular unconformity questions a Variscan origin for the convergent cleavage fans. Instead, as previously put forward by Legrand (1967) in the southern Sennette Valley, the convergent cleavage fans may be attributed to a pre-Givetian, "Caledonian" deformation. However, as in the Ronquières area, also the rocks in the southern Orneau Valley only show evidence for one single deformation phase (Belmans, 2000). Hence, following the hypothesis of Debacker *et al.* (1999), Belmans (2000) explained the convergent cleavage fans in the Silurian of the Orneau Valley as the result of continued fold amplification after cleavage development.

The apparent preferred position of the faults in the fold hinge zones, in combination with the small displacements, the parallelism of the fault trend with the cleavage trend and the fold hinge lines, and the frequently occurring apparent conjugate sets of reverse faults suggest a genetic relationship between folding and fault development. Possibly the small reverse faults formed as accommodation structures during folding. Several of the normal faults initially may have developed in the fold hinge zones or may reflect reactivated reverse faults (Belmans, 2000; Debacker, 2001; cf. Legrand, 1967).

As argued by Vandenven (1967), the thick overburden necessary for kink band development cannot be attributed to the Middle Devonian to Carboniferous deposits considered to have been previously overlying the Brabant Massif (cf. Patijn, 1963; Van den haute & Vercoutere, 1990). Instead, Vandenven (1967) suggested that the convergent cleavage fans were formed by an accentuation of the pre-existing "Caledonian" folds during Variscan thrusting and that this was accompanied by simultaneous kink band development due to the weight of the hypothetical thrust sheet. However, several objections can be made. Firstly, from a kinematic point of view, the shortening necessary for forming the convergent cleavage fans (sub-horizontal, N-S-directed) is at high angles to the shortening responsible for kink band development (sub-vertical). As such, a contemporaneous development of both the kink bands and the convergent cleavage fans seems unlikely (Belmans, 2000). Secondly, there is no evidence for a post-Givetian deformation in the southern Orneau Valley (Belmans, 2000) and the cleavage fanning can entirely be explained by means of a Caledonian single-phase progressive deformation (Belmans, 2000; cf. Debacker *et al.*, 1999). Thirdly, in order to explain the well-developed cleavage and the anchizonal degree of metamorphism in the Silurian deposits of the Orneau Valley (cf. Geerkens & Laduron, 1996), a significant overburden can be inferred, which was eroded prior to deposition of the Givetian continental sediments, which do not have a cleavage and only reflect diagenetic burial conditions. The load, necessary for kink band development, may have been caused by the now eroded post-Wenlock to pre-Givetian deposits previously overlying the Silurian deposits in the southern Orneau Valley.

In our opinion (Belmans, 2000; Debacker, 2001), the deformation of the southern Orneau Valley can be depicted as follows. Roughly N-S-directed shortening led to folding and cleavage development which eventually resulted in the presence of convergent cleavage fans. During shortening, small reverse faults likely developed in the fold hinge zones as fold accommodation structures. As shortening progressed a gradual change in stress pattern occurred. The initial, approximately sub-horizontal, N-S-directed maximum compressive stress gradually changed orientation and became steeply plunging during the final stages of shortening. The steep maximum compressive stress was responsible for kink band development and normal faulting. The orientation and geometry of the kink bands may partly be attributed to the orientation of the cleavage and partly to a change in local strain or stress orientation under the influence of bedding orientation and nearby fault tip zones. During the last stages of shortening and after shortening, the overburden was removed by erosion. The last erosion products are represented by the Givetian conglomerates.

## 6. Description of excursion stops DAY 2 (morning)

### *Stop 2.1. Outcrop below the Saracens Tower, Gembloux*

Location. Outcrops along the Rue de Moulin, between houses N°44 and 42, below the walls of the *Faculté universitaire des Sciences Agronomiques de Gembloux*, close to the *Tour d' enceinte des Sarrasins* (dated AD 1156).

General structure. Bedding difficult to observe; steeply bedded strata (S0: N090/65-75°S: S1: N090/70-80°S). Some folds and faults are visible.

Lithostratigraphy. Rigenée Formation.

Lithology. Fine silty dark grey shale with dispersed pyrite and very rare centimetric sandstone or siltstone beds, with weak oblique bedding.

Sedimentology. See stop 1.8.

Biostratigraphy. No graptolites were found, nor chitinozoans in this outcrop (Samuelsson & Verniers, 2000); acritarchs were studied by Martin (*in* Verniers, 1983b) and by Servais (1991). The latter discovered here a rich assemblage characterised by *Arkonia virgata*, *Frankea sartbernardensis, F. hamata and F. hamulata*. The latter species indicates a late Llanvirn or younger age. With this assemblage this outcrop can be correlated with the middle to upper part of the Rigenée Formation in the Thyle Valley at stop 1.8.

### *Stop 2.2. Gembloux city centre, Place de l'Orneau and Grand-Manil, Try à la Vigne (flying stops)*

Location: Below the restaurant (in 1980) of the school St-Guibert on the Place the l'Orneau and until recently in the school yard one could observe clearly the turbiditic Ittre Formation, dated elsewhere with graptolites and chitinozoans as Burrellian (low-middle Caradoc).

Further to the south between the city centre and the neighbourhood Grand-Manil several temporary outcrops indicated the monoclinal Upper Ordovician succession of the Bornival, Huet and Madot formations. The Fauquez Formation normally situated in between the latter two formations is not observed here. The Huet and the Madot formations contain rather rich macrofossil bearing levels with brachiopods, bryozoans, crinoids, trilobites, pelmatozoans, etc. studied by Lespérance and Sheehan (1987). The fauna of both levels was referred to in the literature as the "*grauwacke* *fossilifère de Grand-Manil*". It is now estimated that the lower level is late Caradoc and the second early Ashgill. Outcrops in both beds at present are very small or temporary. We will visit a similar but slightly younger facies in stops 2.8 and 2.9.

### *Stop 2.3. "La Poudrière de Corroy", abandoned quarry near the Orneau river, Gembloux*

Location. Old abandoned quarry, used in the 18th century for the construction of a powder factory within the quarry. It was destroyed in the 19th century. It is located on the left bank of the Orneau river, 200 m west of km 3 of the railway line (map see Fig. 21).

General structure. The outcrop is now about 55 m long and 3-6 m high; strike of the bedding N074°/66-86°S, cleavage nearly parallel to the bedding.

Lithostratigraphy. Type locality of the Corroy Formation in its old and new definition.

Lithology. Centi- to decimetric shale/siltstone/sandstone alternations, with obliquely bedded, undulating sandstone beds with laminated siltstone and shale beds and interlayered centimetric laminated hemipelagites, in which graptolites can be found.

Sedimentology. Laminated hemipelagites are deposited in a deep-sea anoxic environment and alternate with the Bouma-type turbidites.

Biostratigraphy. The outcrops in the quarry are dated with graptolites and chitinozoans as early Wenlock (see Chapter 5.5).

Remarks. This turbidite facies shows that the foreland basin, started in the underlying Fallais Formation, is fully developing in the Brabant Massif. This is a clearly different facies than in the Condroz Inlier where a deep shelf sedimentation occurs in the same time interval.

### *Stop 2.4. Vichenet section, Gembloux*

Location. A 300 m long outcrop along the road "Chemin du Grand Ha", close to the old railway station Vichenet-Bossière (see detailed map on Fig. 24 and section on Fig. 25).

General structure. A general synclinal or synclinorial structure with a series of folds with an E-W fold axes (see Chapter 5.10).

Lithostratigraphy. Type locality of the Vichenet Formation. De Schepper (2000) suggested that only the lower part of the formation is present in the type locality. The lower contact with the underlying Fumal Formation can be seen in the north and the south of the section (see Fig. 25).

Lithology and sedimentology. Thick-bedded silty shale in beds of 10 to 100 cm thick, interpreted as distal turbidites, alternating with dark grey centimetric laminated hemipelagites.

Biostratigraphy. Chitinozoans (this study) are the first fossils discovered in this formation in its type locality and indicate that the limit between the Sheinwoodian-Homerian (middle-late Wenlock) boundary runs in this long outcrop somewhere close to section A (see fig. 23).

Remarks. This stop, together with stops 2.3 and 2.5, illustrate the variation in the turbiditic sedimentation during the foreland basin development.

### *Stop 2.5. Les Mautiennes, Acadian ("Late Caledonian") unconformity, Gembloux*

Location. Outcrop, along a dirt road, east of the railway and 57 m south of the middle of the railway bridge, where the "rue de Mautienne" crosses the railway line.

General structure. The outcrop shows the Fumal Formation with 10-20 cm thick distal Bouma-type turbidites in a medium grey silty shale, siltstone and some fine sandstone. A closed fold is clearly visible showing a cleavage fanning. This is cut by the Middle Devonian (Givetian) conglomerate and sandstone beds, dipping 10-14°S. It clearly shows the Caledonian folding and cleavage formation prior to the Givetian.

Lithostratigraphy. Fumal Formation and Les Mautiennes Member (Bois de Bordeaux Formation).

Biostratigraphy. No chitinozoans were recovered from the Fumal Formation in this outcrop, but elsewhere the unit is dated middle Wenlock (Sheinwoodian). The Devonian is dated by the presence of the brachiopod *Stringocephalus burtini*, an index fossil of the Givetian, in the basal conglomerate of a borehole in the vicinity (Bultynck *et al.*, 1991).

## 7. The Lower Palaeozoic of the Condroz Inlier in the Fosses area

### *7.1. Introduction*

The Fosses area (Fig. 27), is the western termination of the Condroz Inlier around and west of the city of Fosses-la-Ville and consists structurally of two tectonic units: the central part of the Condroz Inlier in the north, and the Puagne Inlier in the west and south. Both contain Ordovician and Silurian strata, unconformably covered by Lochkovian (Lower Devonian) strata. Both are Variscan tectonic wedges situated in the Variscan Deformation Front. The northern tectonic unit underwent no or little metamorphism (diagenetic zone), while the Puagne Inlier underwent anchizonal metamorphism as deduced from the degree of carbonisation of the organic material and from the illite crystallinity (Steemans, 1994). It is interpreted that both sedimentation areas were originally more distant from one another (a few to a few tens of kilometre) and were thrusted northwards with different throws and became adjacent to each other during the Variscan shortening and thrusting in the Late Carboniferous. The Puagne area is more metamorphosed because it is interpreted to have derived from a more southern and deeper structural level than the central part of the Condroz Inlier.

In the Condroz Inlier Ordovician and Silurian sediments are deposited in a different environment than in the Brabant Massif, except for the Lower and Middle Ordovician, but which will not be visited nor discussed. The Upper Ordovician and Silurian sediments are mostly fine siliciclastics and interpreted as deposited on a deep shelf. As a rule units are thinner than in the Brabant Massif and turbidites are not found, except in the Ombret area, in the north-east of the Condroz Inlier. The Fosses area is the type area for several units, such as the Vitrival-Bruyère, Basse-aux-Canes, Fosses, Génicot and Thimensart formations, the Bois de Presles Member and the Cocriamont conglomerate bed (Figs. 4 & 28).

Recently Billiaert (2000) studied in some detail two of the type localities: Vitrival-Bruyère and Basse-aux-Canes. Two other units, the Fosses and Génicot formations have the Fosses area as their type area, but no type section has yet been defined. B. Delcambre and J.-L. Pingot (pers. comm., 1999) suggested after their extensive mapping in the area, that the *Parc de Sart-Eustache* section and the area around the *Étang du Diable* might be a good candidate for a type section, being the most complete section in the area. The results of the study of this section are presented here. Michot (1931) was the first to postulate in the Puagne area (Fig. 27& 29) the presence of the Cocriamont conglomerate at the base of the Fosses Formation, overlying an unconformity, in the locality Gazelle. In the *Parc de Sart-Eustache* section, 2.5 km west of Gazelle, one cannot observe the base of the formation. The contact with the underlying units is located below the pond of the *Étang du Diable*, where probably a fault contact or an unconformity occurs. In the long *Parc de Sart-Eustache* section, however, it became evident that the Cocriamont conglomerate is not situated at the base of the formation, but much higher and according to the new definition of the Fosses Formation (Delcambre & Pingot, in press; Verniers *et al.*, 2001) not even in that formation but in the overlying Génicot Formation. In an 11 m thick interval, a lower 2.2 m interval with two thick sandstone beds occurs with the possible presence of small siltstone

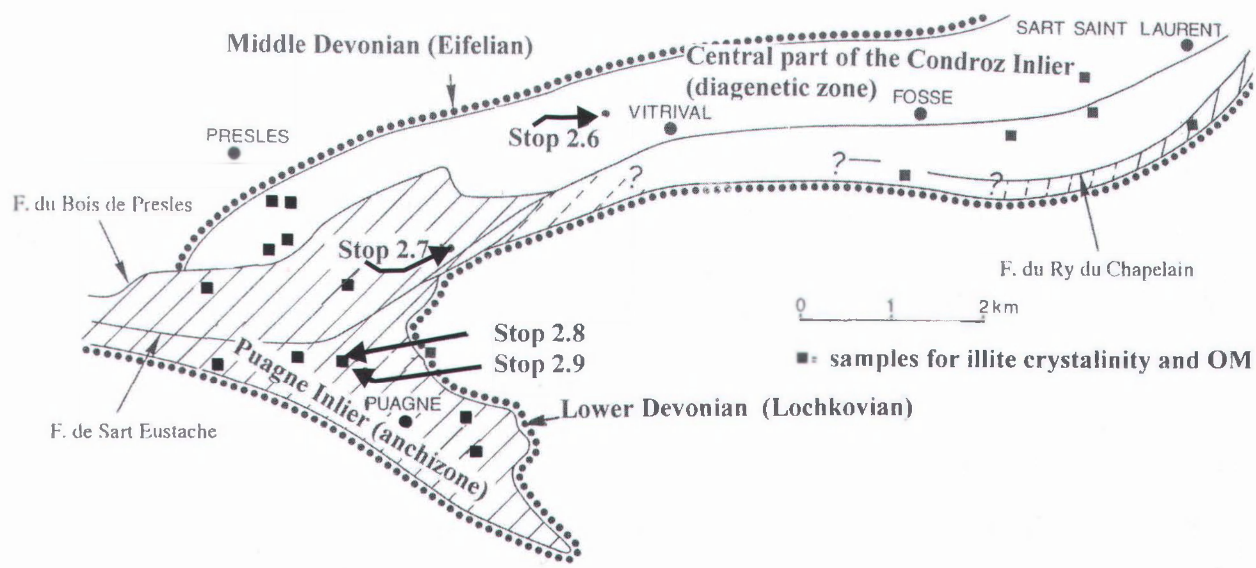


**Figure 27**. General sketch map of the western extremity of the Condroz Inlier, around Fosses (Fosse-la-Ville), after Steemans (1994). The hatched area corresponds to the area where a cleavage is observed after Michot (1934). Metamorphism data of Steemans (1994) are added and indicate that the area with cleavage corresponds to a higher metamorphic zone. The affinity of the area between the Bois de Presles Fault and the Sart-Eustache Fault to either the central part of the Condroz Inlier or to the Puagne Inlier has still to be established.

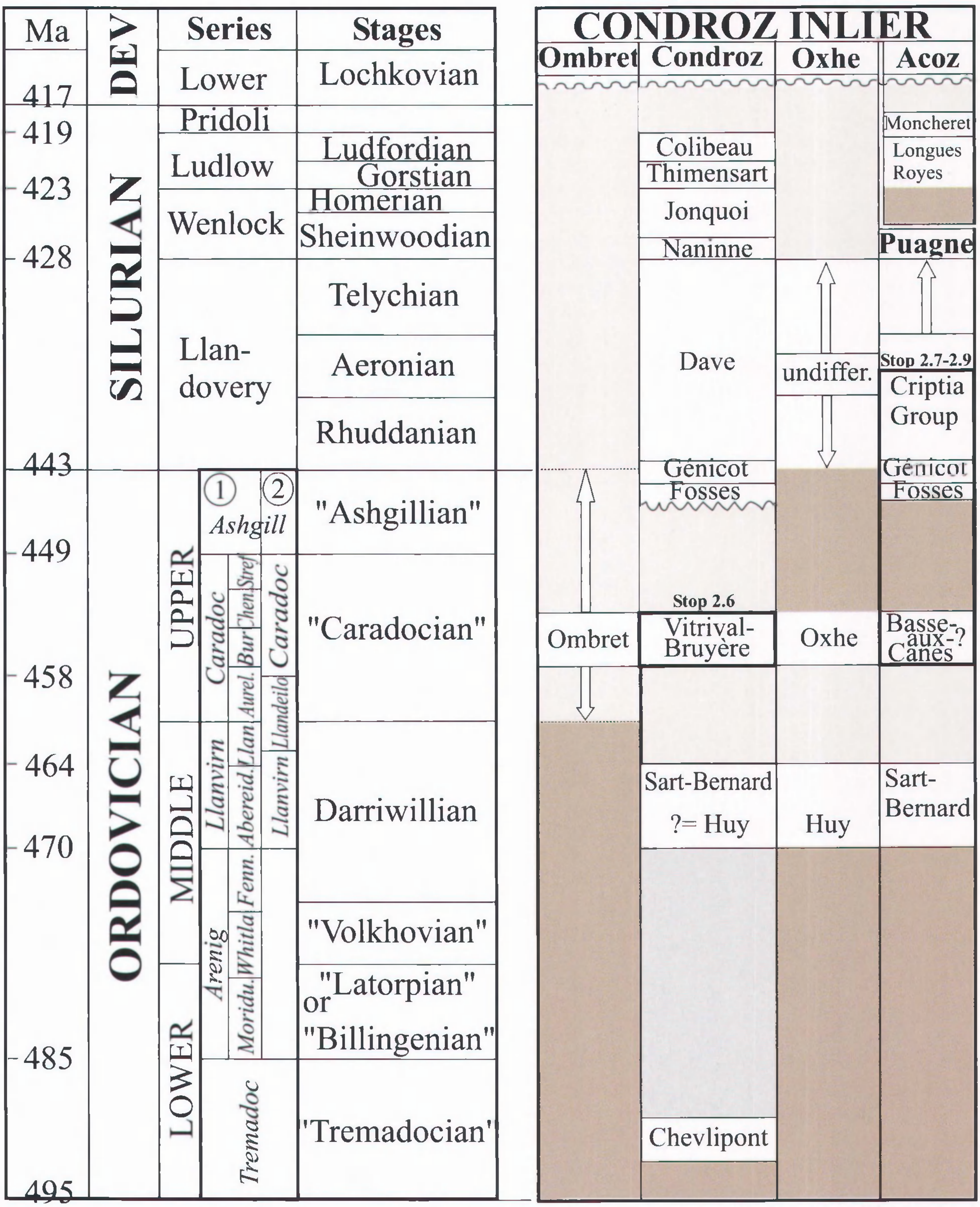


**Figure 28**. Chronostratigraphical position of the Lower Palaeozoic lithostratigraphic units in the Condroz inlier (detail of fig. 4) (after Verniers *et al.* 2001).

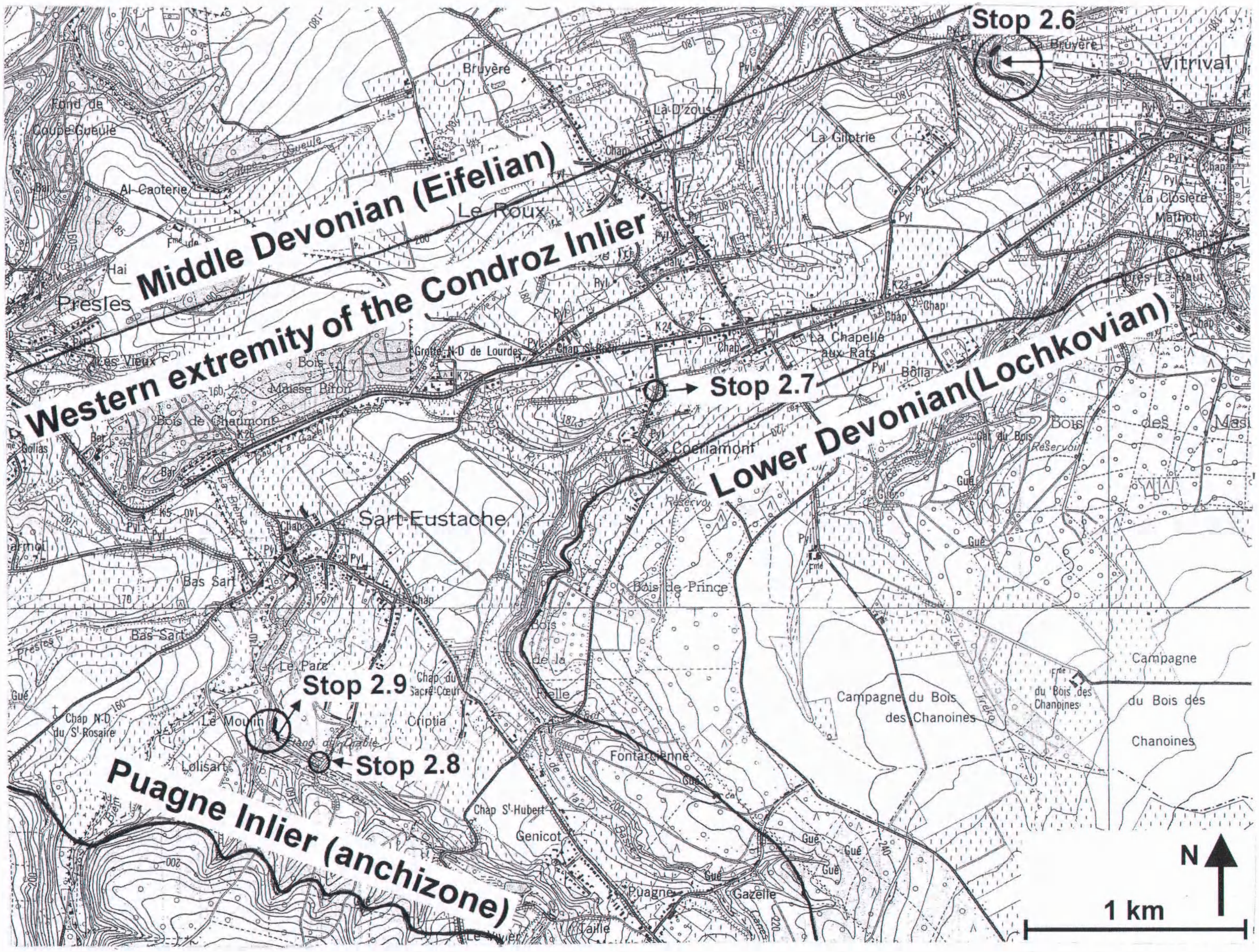


**Figure 29**. Detailed topographic map (1/25,000) of the Fosses area. The line indicates the limit of the Devonian cover. The location of the visited outcrops is indicated.

pebbles; this is covered by a 7 m thick mostly shaly, silty and sandy interval, followed by a 2 m thick conglomeratic level, which resembles the Cocriamont conglomerate in its type section in the rue Cocriamont (stop 2.7). This conglomeratic level contains besides pebbles of siltstone and sandstone also well-rolled large cobbles (up to 55 cm in length) of limestone, with up to 57% of $CaCO_3$, rich in macrofauna, indicating an eroded carbonate platform nearby. Forty samples were taken for chitinozoan studies and in five formations the organic microfossils could be extracted. As usual in most of the Condroz Inlier, the chitinozoans are dark to opaque and only moderately preserved. They have a concentration between 0.1 and 14.5 chitinozoans per gram of rock and a diversity of 1 to 10 species per sample (Fig. 30). The indicative species are illustrated on the Plates 6-9.

The biozonation with chitinozoans could corroborate the age given with graptolites for the Vitrival-Bruyère Formation and the base of the Criptia Group. For the other three formations a detailed age can now be proposed for the first time (see below). The Fosses Formation was dated earlier with trilobites and brachiopods as pre-Hirnantian Ashgill by Sheehan (1987) and Lespérance & Sheehan (1987), mainly because no Hirnantian brachiopods were encountered. Hence an broad age of early to middle Ashgill age was inaccurately adopted, but a pre-Hirnantian late Ashgill age was also possible according to these macrofossils.

The most important result from the chitinozoans is that the Fosses and Génicot formations around the *Etang du Diable* and in the *Parc de Sart-Eustache* section are middle and late Ashgill in age and for the uppermost levels possibly Rhuddanian (early Llandovery). The samples in which the chitinozoans are frequent and diversified can be correlated with several biozones of Baltoscandia (Nolvak & Grahn, 1993; Nolvak, 1999), although not all of the Baltoscandian biozones are recognised here. The chitinozoans in the basal sample of the Fosses Formation can be correlated with the local biozone 10 (Fig. 9) in the Brabant Massif, present in the Harelbeke borehole (Samuelsson & Verniers, 2000), correlatable with the Baltoscandian *Tanuchitina bergstroemi* Biozone, dated as late Vormsi or early Pirgu, middle Ashgill. This age for the basal part of the Fosses Formation is the same age or younger than the top of the Madot Formation, the unit with a similar facies in the Brabant Massif (see stop 2.2). The chitinozoan assemblage from a higher level of the formation in the top of the quarry near the *Étang du Diable* can be correlated with the *Conochitina rugata* Biozone of Baltoscandia (Nolvak & Grahn, 1993; Nolvak, 1999), corresponding to the middle Pirgu (also middle Ashgill). Hence a broad middle part of the Ashgill is proposed for the whole Fosses Formation in its new definition. Only a few but well preserved chitinozoans are present in the limestone cobbles of the Cocriamont conglomerate and indicate a late Ashgill age. The assemblage is very comparable with the assemblage in a shaly level about 4 m below that conglomerate. Samples from above the upper conglomerate contain a different chitinozoan assemblage possibly correlatable with the *Conochitina scabra* Biozone of Baltoscandia dated as late Porkuni (Nolvak & Grahn, 1993; Nolvak, 1999). The early Porkuni is considered to be the time of the Hirnantian glaciation, which on Gondwana is often expressed sedimentologically as two (very) coarse intervals. It is tempting to interpret the two coarse levels as the sedimentological expression of a sea level drop, resulting from the Hirnantian glaciation. It would be difficult to interpret otherwise the sudden presence of two coarse levels in between the dominantly fine clastic sedimentation in the Condroz Inlier. Well preserved and rich chitinozoans in the base of the Criptia Group corroborate the late Rhuddanian age proposed with graptolites in this section. The transition between the Génicot Formation and the Criptia Group is however not observed here, due to an observation gap over 17 m in the section.

In the part of the Condroz Inlier east of the city of Fosses-la-Ville, the area that we visit, the following formations are present and their descriptions are summarised from the recent revision by Verniers *et al.* (2001). The results of the study presented here however change already slightly the descriptions and ages of the formations in that revision. It appears that the Fosses Formation as defined now, e.g. on the new geological map of the area by Delcambre & Pingot (in press) corresponds only with what was previously included in the Bois de Presles Member, and that the recently defined Génicot Formation corresponds with what was called the Faulx-les-Tombes Member (Lassine *in* Martin, 1969a). The Cocriamont conglomerate would be situated in the middle part of the Génicot Formation. It can be mentioned here that after the excursion new samples of the Fosses and the Génicot Formations have been studied for chitinozoans in seven sections of the Puagne area (Vanmeirhaeghe & Verniers, submitted).

The new stratigraphical results presented above, can offer an explanation for the apparent contradiction in sedimentological interpretation for the deposition of the Fosses Formation (and the Génicot Formation) by Tourneur *et al.* (1993). Either a deposition as bioclastic turbidites with interbedded shales near a platform-ramp margin was suggested or a regressive event on a shelf followed by a transgressive sequence. Due to lack of a type section the exact position of samples in the lithostratigraphy and macrofossils was largely unknown. A

| Formations | Samples | Conochitina dolosa | Belenochitina capitata | Cyathochitina kuckersiana | Conochitina minnesotensis | Laufeldochitina stentor | Desmochitina spp. | Cyathochitina calix | Cyathochitina campanulaeformis | Desmochitina cocca | Desmochitina? mayensis | Desmochitina minor var. A | Desmochitina ovulum | Desmochitina typica | Conochitina spp. | Belonochitina spp. | Cyathochitina spp. | Rhabdochitina spp. | Belenochitina robusta | Belenochitina wesenbergensis | Tanuchitina bergstroemi | Spinachitina spp. | Spinachitina bulmani | Conochitina rugata | Cyathochitina jenkinsi? | Cyathochitina vaurealensis | Armoricochitina nigerica | Sphaerochitina spp. | Ancyrochitina spp. | Angochitina spp. | Eisenackitina aff. toddingensis | Conochitina scabra | Hercochitina cf. gamachiana | Eisenackitina spp. | Conochitina electa | Conochitina iklaensis | Densichitina densa? | Belenochitina postrobusta | Chitinozoa spp. | Total number of chitinozoans | Weight of dissolved rock (g) | Concentration (chitinozoans per gram of rock) | Local biozones | Baltoscandian biozones (Nolvak & Grahn, 1993 and Nolvak, 1999) | Global Silurian biozones (Verniers et al., 1995) |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Criptia Group | BB40,10 | | | | | | | | | | | | | | 3 | 1 | | | | | | | | | | | | | | | | | | | | | | 8 | 0 | 12 | 22,25 | 0,54 | 12 | ? | postrobusta |
| | BB40,11 | | | | | | | | | | | | | | | | 12 | | | | | | | | | | | | | | | | | | 155 | 45 | 13 | 1 | 41 | 267 | 20,37 | 13,11 | | | |
| Génicot | Aisemont | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 37,21 | 0,00 | | | |
| | BB342000J | | | | | | | | | | | | | | x | x | | | | | | | | | | | | x | | | | | x | | | | | | x | 6 | 21,42 | 0,28 | 11 | | |
| | BB40,12 | | | | | | | | | | | | | | 3 | 1 | | | | | | | | | | | | | | | | | | | | | | | 0 | 4 | 20,06 | 0,20 | | | |
| | BBextra1 | | | | | | | | | | | | | | 7 | | 1 | | | | | | | | | | 1 | 1 | | | | | 1 | 5 | | | | | 0 | 16 | 37,21 | 0,43 | | | |
| | BB342000I | | | | | | | | | | | | | | | | x | | | | | | | | | | | | | | | x | | | | | | | x | 6 | 22,22 | 0,27 | 10 | scabra | |
| | BB342000H | | | | | | | | | | | | | | x | | x | | | | | x | | | | | x | | x | | | x | | | | | | | x | 30 | 22,85 | 1,31 | | | |
| | BB342000E | | | | | | | | | | | | | | x | x | | | | | | | | | | | x | | | | | x | | | | | | | x | 25 | 20,05 | 1,25 | | | |
| | BB342000A | | | | | | | | | | | | | | x | | | | | | | x | | | | | | | | | | x | | | | | | | x | 15 | 21,95 | 0,68 | | | |
| | BB33,20 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 1 | 12,01 | 0,08 | | | |
| | BB342000C (pebble) | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | x | | | | | | | | x | 1 | 13,84 | 0,07 | 9 | | |
| | BB33,18 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 1 | 1 | 17,58 | 0,06 | | | |
| | BB342000D (pebble) | | | | | | | | | | | | | | x | | | | | | | | | | | | | x | x | | x | | | | | | | | x | 9 | 40,71 | 0,22 | | | |
| | BB33,17 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 23,6 | 0,00 | | | |
| | BB33,16 | | | | | | | | | | | | | | | | | | | | | | | | | | | 2 | | 1 | 19 | | | | | | | | 1 | 23 | 15,5 | 1,48 | | | |
| | BB33,11(pebble) | | | | | | | | | | | | | | x | | | | | | | | | | | | | | x | | | | | | | | | | x | 34 | 25,23 | 1,35 | 8 | | |
| | BB342000B | | | | | | | | | | | | | | x | | x | | x | | | | x | | | | x | x | | | | | | | | | | | x | 28 | 13,04 | 2,15 | 7 | | |
| | BB33,12 | | | | | | | | | | | | | x | x | | x | | x | | | | x | x | | | x | x | | | | | | | | | | | x | 64 | 13,12 | 4,88 | | | |
| | BB33,15 | | | | | | | | | | | | | | 3 | 2 | | | | | | | | | | | | | | | | | | | | | | | 0 | 5 | 12,09 | 0,41 | | | |
| | BB15,6 | | | | | | 3 | | 2 | | | | | 3 | 5 | | | | 4 | | | | | | | 2 | 7 | | | | | | | | | | | | 20 | 46 | 27,44 | 1,68 | | | |
| | BB33,14 | | | | | | | | | | | | | | 4 | | | | 2 | | | | | | | | | | | | | | | | | | | | 9 | 15 | 14,49 | 1,04 | 6 | | |
| | BB33,10 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 12,12 | 0,00 | | | |
| Fosses | BB33,9 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 12,9 | 0,00 | | | |
| | BB33,8 | | | | | | | | | | | | | | | | | | | | | | 18 | | | | | 1 | | | | | | | | | | | 12 | 31 | 42,68 | 0,73 | | | |
| | BB33,6 | | | | | | | | | | | | | | x | | x | | | | | | x | x | x | | | | | | | | | | | | | | x | 33 | 13,52 | 2,44 | 5 | rugata | |
| | BB33,5 | | | 1 | | | | | | | | | | | | | | | | | | | | 59 | 9 | | | | | | | | | | | | | | 48 | 117 | 42,68 | 2,74 | | | |
| | BB33,4 | | | | | | | | | | | | | | 3 | | | | | | | | 72 | | | | | | | | | | | | | | | | 0 | 75 | 41,46 | 1,81 | 4 | | |
| | BB33,3 | | | | 4 | | | | | | | | | | | 1 | 2 | | | | | | 94 | | | | | | | | | | | | | | | | 123 | 224 | 42,44 | 5,28 | | | |
| | BB40,13 | | | | 5 | | | | | | | | | | | | | | | 69 | 42 | | | | | | | | | | | | | | | | | | 0 | 116 | 45,25 | 2,56 | 3 | bergstroemi | |
| Vitrival-Bruyère | BB40,8 | | | | 1 | | | | | | | | | | 1 | | | | | | | | | | | | | | | | | | | | | | | | 2 | 4 | 20,08 | 0,20 | 2 | | |
| | BB40,9 | | 15 | 2 | 18 | | 53 | | 5 | 9 | 4 | 3 | 1 | 7 | | | | | | | | | | | | | | | | | | | | | | | | | 13 | 130 | 20,08 | 6,47 | | | |
| | BB40,7 | | | | | | 1 | | | | | | | 1 | 4 | | | | | | | | | | | | | | | | | | | | | | | | 10 | 16 | 21,83 | 0,73 | | | |
| | BB40,6 | | | | | | | | | | | | | | 7 | 1 | | | | | | | | | | | | | | | | | | | | | | | 0 | 8 | 21,82 | 0,37 | | | |
| | BB40,5 | | | | 3 | | 2 | 54 | 3 | 2 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 17 | 81 | 20,58 | 3,94 | | | |
| | BB40,4 | | | | | | | | | | | | | 1 | 2 | | | | | | | | | | | | | | | | | | | | | | | | 0 | 3 | 23,75 | 0,13 | | | |
| | BB40,3 | | | | | 10 | 3 | | | 2 | | | | | 3 | | | | | | | | | | | | | | | | | | | | | | | | 0 | 18 | 26,3 | 0,68 | | | |
| | BB40,2 | | | | 1 | | | | | | | | | | | 1 | | | | | | | | | | | | | | | | | | | | | | | 8 | 10 | 22,8 | 0,44 | | ? | |
| | BB40,1 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 0 | 20,35 | 0,00 | | | |
| Basse-aux-Cannes | BB33,2 | 14 | 48 | 2 | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | | 0 | 64 | 34,1 | 1,88 | 1 | | |
| | Total | 14 | 63 | 5 | 32 | 10 | 62 | 54 | 10 | 13 | 4 | 3 | 1 | 12 | 45 | 7 | 15 | 0 | 6 | 69 | 42 | 0 | 184 | 59 | 9 | 2 | 8 | 4 | 0 | 1 | 19 | 0 | 1 | 5 | 155 | 45 | 13 | 9 | 305 | | | | | | |

**Figure 30**. Composition of chitinozoan assemblages and concentrations in samples of the Ordovician-Silurian of the Condroz Inlier around Fosses after Billiaert (2000) and more recently studied samples, with the local biozonation and correlation with the Baltoscandian biozonation of Nolvak & Grahn (1993) and Nolvak (1999) and with the global Silurian biozonation of Verniers *et al.* (1995).

revision of the studied sedimentological samples and macrofossils should relocate these in the section. Because of the drastic environmental changes involved due to the glaciation-induced sea-level drops both sedimentological interpretations in Tourneur *et al.* (1993) are now possible, and not mutually exclusive any more.

### *7.2. Vitrival-Bruyère Formation*

Description. Blackish, silky and fine micaceous shale with quartzite intercalations and graptolites (Maillieux, 1926); micaceous pelitic unit (Michot, 1928); the higher part contains clayey quartzitic sandstone in thick beds, covering black or blue shale with intercalations of black or dark coloured mudstone, siltstone and fine sandstone. The sedimentology has not been studied yet. The stratotype (stop 2.6) lies in the outcrops in the river bed of or along the Ruisseau Le Treko, west of the hamlet La Bruyère, 500 to 1050 m east of the church of the village Vitrival. Possibly the Oxhe Formation is a lateral equivalent or synonym for the unit in the Oxhe Inlier, situated 52 km to the NE.

Thickness: Difficult to estimate, except for the upper part with quartzite beds: 20-30 m.

Age. Graptolites in a locality at Vitrival-Bruyère in the upper part of the unit indicate the *Climacograptus peltifer* Biozone (Maillieux, 1933). A restudy of the fauna by Bulman (1950) confirmed the presence of the *Climacograptus peltifer* Biozone (now equivalent to the *Diplograptus foliaceus* Biozone), which is situated in the uppermost Costonian to lower Longvillian substages (uppermost Aurelucian and most of the Burrellian stages, Caradoc) (Fortey *et al.,* 1995). The chitinozoans in the present study indicate the same assemblage, and the sampled levels belong to the local biozone 2 (Fig. 30) with as age indicators *Cyathochitina calix* and *Laufeldochitina stentor*. Comparing the range of these species with Baltoscandia (Nolvak, 1999), a broad late Llanvirn to early Caradoc age is attributed. By the presence of the first species it is more or less correlatable with the local biozone 4 of the Brabant Massif present in the Rigenée Formation (Samuelsson & Verniers, 2000).

### *7.3. Basse-aux-Canes Formation*

Description. Greenish grey or dark grey silty shale, often sandy siltstone, with irregular jointing. White mica often present. The siltstone weathers with dark or rusty patches of iron and manganese oxides, described as a characteristic shining blue colour on freshly broken surface. No sandstone beds observed. The formation overlays the Sart-Bernard Formation. The unit resembles slightly the Vitrival-Bruyère Formation, but has no sandstone beds; it is possibly a lateral facies change of (a part of) that formation. The stratotype area lies in the southern part of the Puagne area, south-east of Sart-Eustache but a section is not defined yet (Michot, 1934, 1954; Martin, 1969a; Delcambre & Pingot, in press; Verniers *et al.*, 2001).

Thickness. Estimated 100 to 150 m.

Age. Without fossils it could only be estimated as Llanvirn or early Caradoc (Middle or Late Ordovician). Two samples with chitinozoans described in the present study form the local biozone 1 (Fig. 30). The range of *Belonechitina capitata* and *Conochitina dolosa* compared to Baltoscandia (Nolvak, 1999) narrow the age down from the late Llanvirn to middle Caradoc.

### *7.4. Fosses Formation*

Description. From top to bottom green and dark green fine sandy shale, with trilobites and brachiopods; greywacke, calcareous shale with locally some limestone beds rich in brachiopods (Michot, 1934). Martin (1969a after Michot, 1927, 1934) divided the formation in two members: a lower Bois de Presles Member with clayey limestone, calcareous shale, fossiliferous with brachiopods, cystoids and trilobites, in an alternation of thin-bedded muddy limestone and calcareous shale, in and in the upper part of the formation calcareous shale with thin limestone layers. The brachiopods and trilobites were described by Sheehan (1987) and Lespérance & Sheehan (1987), the crinoids, bryozoans, echinoderm debris, cystoids, molluscs and algae were described by Tourneur *et al.* (1993) and the corals by Servais *et al.* (1997). The base is not yet observed. Several sections were informally designated as stratotype: the northern bank of the Fuette and Rosière rivers one km east of the city of Fosses (Malaise, 1900) and the area between the city of Fosses and the village of Sart-Eustache. In the present study we propose the outcrops and the abandoned quarry north of the *Etang du Diable* and the section in the *Parc de Sart-Eustache* as the stratotype section (Fig. 31).

Thickness. About 50 m (Tourneur *et al.*, 1993); more than 38 m in the *Parc de Saint*-Eustache section; estimated at more than 75 m.

Age. Based on trilobites and brachiopods: pre-Hirnantian Ashgill (Sheehan, 1987; Lespérance & Sheehan, 1987); based on chitinozoans: local biozones 3 to part of 6 (Fig. 30); Baltoscandian biozones of Nolvak & Grahn (1993) and Nolvak (1999): at the base of the formation *Tanuchitina bergstroemi* Biozone (upper Vormsi to lower Pirgu) and in the middle of the formation the *Conochitina rugata* Biozone (middle Pirgu), middle to late Ashgill.

### *7.5. Génicot Formation*

Description. Black shale, dark grey laminated siltstone and often thin beds of medium grey sandstone. In the middle part two sandstone beds occur, separated by a sandy-shaly unit from an overlying conglomeratic bed with large limestone cobbles. The latter bed was sometimes described as the Cocriamont conglomerate, a coarse crinoidal limestone above a basal conglomerate, which was considered by earlier authors (Michot, 1931) as the base of the Fosses Formation. High in the formation a few meters thick bed occurs with dark coloured microbioturbations, a few mm long (*Chondrites* sp.?) dispersed in the shale or siltstone. They were called "*schistes mouchetés*" by Lassine (unpublished *in* Martin, 1969a) or described as green sandy shale with on breaking surfaces blackish elliptic or fusiform spots and a rare macrofauna level (Martin, 1969a after Michot, 1927, 1934). The base of the formation is located above the highest rich macrofossil bearing horizon of the Fosses Formation. The top of the formation is formed by the highest siltstone/sandstone bed below the Criptia Group. It was considered by Martin (1969a) as the Faulx-les-Tombes Member, the upper member of the Fosses Formation. The *Parc de Sart-Eustache* section is proposed herein as the stratotype section.

Thickness. At least 85 m in the stratotype section (this study), but probably between 150 and 200 m (Delcambre & Pingot, in press).

Age. The chitinozoan assemblages allow the distinction between local biozones 6 (part) to 11 (Fig. 30). All of them contain the rare but distinct *Armoricochitina nigerica*, indicative for the late Ashgill. Above the conglomerate of Cocriamont level, the *Conochitina scabra* Biozone occurs, dated as late Porkuni by Nolvak & Grahn (1993) and Nolvak (1999), the part of the latest Ashgill after the Hirnantian glaciation. *Hercochitina* cf. *gamachiana* in this assemblage resembles the similar or same species in the local chitinozoan biozone 11 of the Brabant Massif in Samuelsson & Verniers (2000). If this can be confirmed by more studies it would place this biozone 11 also in the post-Hirnantian glaciation latest Ashgill. No chitinozoans were discovered yet in the top of the formation.

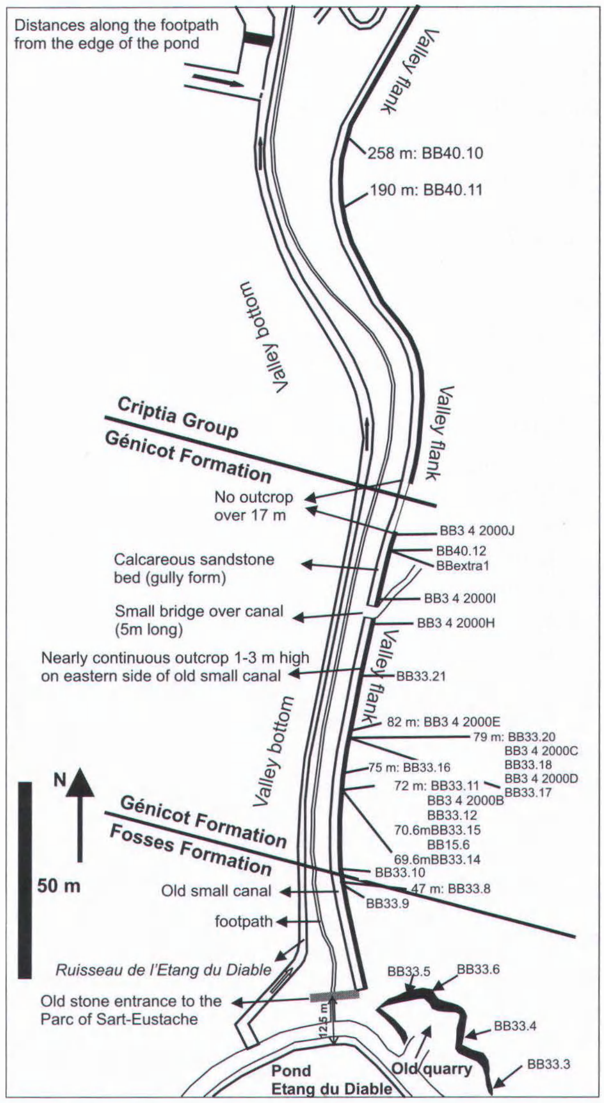


**Figure 31**. Detailed location map of the samples for chitinozoan studies in the quarry near the *Etang du Diable* and in the *Parc de Sart-Eustache*, the proposed new type locality and section of the Fosses and the Génicot formations, changed after Billiaert (2000).

### *7.6. Criptia Group*

Description. A group described as "Schistes de Criptia" by Michot (1928) and re-used by Delcambre & Pingot (in press) for a thick unit of shale and silty shale, homogeneous with various colours (greenish, grey, ochre) and a stratification difficult to distinguish. In the group a green to greenish grey soft shale unit can be differentiated. The fine-grained and compact sedimentation points to a deep shelf environment. The Génicot Formation is supposed to be the underlying unit. The covering unit is a more silty and darker coloured unnamed shale unit. The stratotype area is around the village of Sart-Eustache, but not defined yet.

Thickness. Unknown, tentatively estimated at several hundred meters.

Age. Only at the base well preserved and frequent chitinozoans are described in this study belonging to the *Belonechitina postrobusta* global Biozone of Verniers *et al.* (1995) of the Rhuddanian (early Llandovery).

Graptolites are found in higher levels but are still unstudied; an early Llandovery or possibly Wenlock to early Ludlow age is tentatively postulated (Delcambre & Pingot, in press; Verniers *et al.*, 2001).

## 8. Description of excursion stops DAY 2 afternoon

***Stop 2.6. Vitrival-Bruyère section***

Location. Outcrops along a track and in the river bed of the Ruisseau Le Treko, west of the hamlet La Bruyère, 500 to 1050 m east of the church of the village Vitrival.

General structure. Subvertical bedding with stratification only visible in the decimetre to metre thick sandstone beds.

Lithostratigraphy. Vitrival-Bruyère Formation.

Lithology. Micaceous pelitic unit, with dark grey, black or blue, silty and fine micaceous shale with some thick quartzite intercalations and other beds of black or dark coloured mudstone, siltstone and fine sandstone. The sedimentology is unstudied yet.

Biostratigraphy. See above in chapter 7.2.

Remarks. The outcrop shows the difficulty to describe complete lithostratigraphical sections in the Condroz Inlier. The complexity of the Variscan deformation is enhanced by the suspected Quaternary mass movements along the steep hill side. The thickness of the formation cannot be determined. The lithology is however considered typical for the formation and chitinozoans allowed a corroboration of the earlier dating (see above in 7.2).

***Stop 2.7. Cocriamont conglomerate type section***

Location. Rue de Cocriamont, Le Roux, Fosses-la-Ville. low escarpment in the west side of the road.

General structure. Steeply dipping beds, normal to overturned.

Lithostratigraphy. Cocriamont conglomerate level.

Lithology & sedimentology. See in 7.5.

Biostratigraphy. See in 7.4 and 7.5. One chitinozoan sample at 1.0 to 1.2 m above the conglomerate contains some specimens of species of the *Conochitina scabra* Biozone, suggesting tentatively the latest Ashgill.

***Stop 2.8. Outcrop east of the "Etang du Diable", Sart-Eustache***

Location. Escarpment in an outcrop at the corner of the road, east of the "*Etang du Diable*".

General structure. Steeply dipping beds and northwards younging.

Lithostratigraphy. Fosses Formation (lower part).

Lithology. Typical (slightly) calcareous mudstone with macrofossils dispersed or in centimetre thick layers.

Sedimentology. Not studied yet here.

Biostratigraphy. The chitinozoans are rather rich and belong to the *Tanuchitina bergstroemi* Biozone, middle Ashgill (see in 7.5).

***Stop 2.9. Abandoned quarry north the "Etang du Diable" and section in the "Parc of Sart-Eustache", Sart-Eustache***

Location. A continuous section can be followed from an abandoned quarry, along a public path, north of the "Etang du Diable" and in the private property of the Park of Sart-Eustache in a 750 m long nearly continuous section along a small disused canal (written permission should be asked to enter the property).

General structure. Steeply bedded normal to overturned beds with a main strike of N110-290. The sequence youngs to the north, evidenced by sedimentological structures.

Lithostratigraphy. In the first 250 m successively the Fosses and Génicot formations, with the Cocriamont conglomerate and the Criptia Group is outcropping (Fig. 31).

Lithology & sedimentology. See above in 7.4, 7.5 and 7.6.

Biostratigraphy. See above; the chitinozoan succession indicates the presence of middle and late Ashgill to Rhuddanian.

Remarks. In this section the presence of the Hirnantian glaciation could be postulated for the first time in Belgium.

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# Plate 1

a. *Bursachitina* sp. A Sutherland, 1994:
L = 130 µm; D = 70 µm; d = 43 µm;
DS-027-99/514 Ch 17

b. *Bursachitina* sp. A Sutherland, 1994:
L = 80 µm; D = 72 µm; d = 40 µm;
DS-027-99/514 Ch 17

c. *Desmochitina opaca*? Laufeld, 1974:
L = 210 µm; D = 120 µm; DS-036-99 Ch 6

d. *Cingulochitina burdinalensis* Verniers, 1999:
L = 160 µm; D = 150 µm;
DS-038-99/C Ch 2

e. *Cingulochitina burdinalensis* Verniers, 1999:
L = 108 µm; D = 102 µm;
DS-038-99/A Ch 3

f. *Cingulochitina burdinalensis* Verniers, 1999:
L = 137 µm; D = 138 µm;
DS-038-99/A Ch 12

g. *Cingulochitina burdinalensis* Verniers, 1999:
L = 140 µm; D = 128 µm;
DS-038-99/B Ch 1

h. *Cingulochitina cingulata* Eisenack, 1937:
L = 78 µm; B = 50 µm; d = 30 µm;
DS-027-99/607 Ch 1

i. *Cingulochitina cingulata* Eisenack, 1937:
L = 92 µm; B = 60 µm; d = 40 µm;
DS-027-99/607 Ch 21

j. *Cingulochitina cingulata* Eisenack, 1937:
L = 87 µm; B = 55 µm; d = 30 µm;
DS-027-99/514 Ch 1

k. *Cingulochitina cingulata* Eisenack, 1937:
L = 90 µm; B = 58 µm; d = 45 µm;
DS-055-99 Ch 14

l. *Cingulochitina cingulata* Eisenack, 1937: example of chain of two specimens;
L = 78 µm; B = 55 µm;
DS-055-99 Ch 19

PLATE 1

# Plate 2

a. *Cingulochitina dreyensis* Verniers, 1999:
L = 146 µm; D = 68 µm; DS-036-99 Ch 4

b. *Cingulochitina dreyensis* Verniers, 1999:
L = 164 µm; D = 65 µm; DS-036-99 Ch 3

c. *Cingulochitina dreyensis* Verniers, 1999:
L = 146 µm; D = 68 µm; DS-034-99 Ch 1

d. *Cingulochitina dreyensis* Verniers, 1999:
L = 146 µm; D = 68 µm; DS-033-99/608 Ch 39

e. *Cingulochitina pitetensis* Verniers, 1999:
L = 162 µm; D = 51 µm; DS-037-99/610 Ch 18

f. *Margachitina margaritana* Eisenack, 1937:
L = 100 µm; D = 92 µm; DS-036-99 Ch 1

g. *Margachitina margaritana* Eisenack, 1937:
L = 100 µm; D = 72 µm; DS-036-99 Ch 22

h. *Eisenackitina anulifera* Verniers, 1999:
L = 146 µm; D = 112 µm; DS-037-99/608 Ch 34

i. *Eisenackitina anulifera* Verniers, 1999:
L = 160 µm; D = 132 µm; d = 60 µm;
DS-037-99/469 Ch 17

j. *Eisenackitina causiata* Verniers, 1999:
L = 117 µm; D = 113 µm; DS-037-99 Ch 23

k. *Eisenackitina anulifera* Verniers, 1999:
L = 170 µm; D = 107 µm; d = 58 µm;
DS-037-99/469 Ch 14

l. *Salopochitina bella* Swire, 1990:
L = 160 µm; D = 132 µm; DS-038-99/B

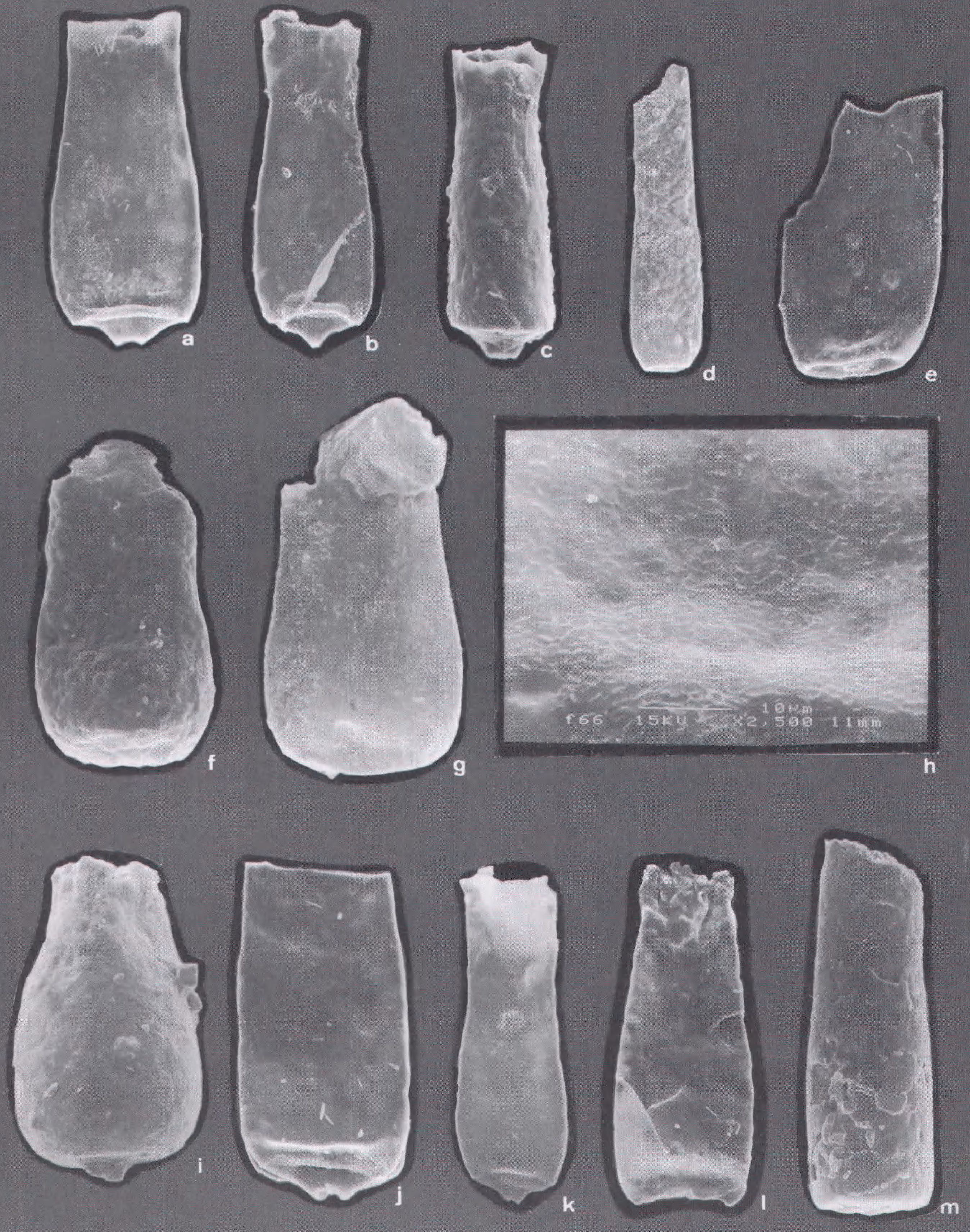
m. *Salopochitina bella* Swire, 1990: detail of the appendix; DS-038-99/B

PLATE 2

# Plate 3

a. *Conochitina acuminata* Eisenack, 1959:
L = 167 µm; D = 74 µm; d = 56 µm;
DS-037-99/610 Ch 31
b. *Conochitina acuminata* Eisenack, 1959:
L = 170 µm; D = 60 µm; d = 50 µm;
DS-035-99/A Ch 6
c. *Conochitina acuminata* Eisenack, 1959:
L = 160 µm; D = 50 µm; d = 45 µm;
DS-037-99/B Ch 21
d. *Conochitina claviformis* Eisenack, 1931:
L = 340 µm; D = 72 µm; DS-026-99 Ch 15
e. *Conochitina claviformis* Eisenack, 1931:
L = 130 µm; D = 77 µm; DS-027-99:607 Ch 15
f. *Conochitina fortis* Nestor, 1994:
L = 165 µm; D = 83 µm; d = 69 µm;
DS-026-99 Ch 2
g. *Conochitina fortis* Nestor, 1994:
L = 122 µm; D = 67 µm; d = 53 µm;
DS-055-99 Ch 10
h. *Conochitina fortis* Nestor, 1994: detail of the ornamentation on the tegument;
DS-025-99 Ch 2
i. *Conochitina* aff. *fortis* Nestor, 1994:
L = 83 µm; D = 50 µm; DS-027-99/607 Ch 12
j. *Conochitina proboscifera* Eisenack, 1937:
L = 170 µm; D = 65 µm; $L_{mucron}$ = 15 µm;
DS-037-99/610 Ch 4
k. *Conochitina proboscifera* Eisenack, 1937:
L = 216 µm; D = 77 µm; $L_{mucron}$ = 10 µm;
DS-037-99/610 Ch 27
l. *Conochitina proboscifera* Eisenack, 1937:
L = 170 µm; D = 68 µm; DS-035-99/B Ch 5
m. *Conochitina subcyatha* Nestor, 1982:
L = 185 µm; D = 60 µm;
DS-027-99/514 Ch 16æ.

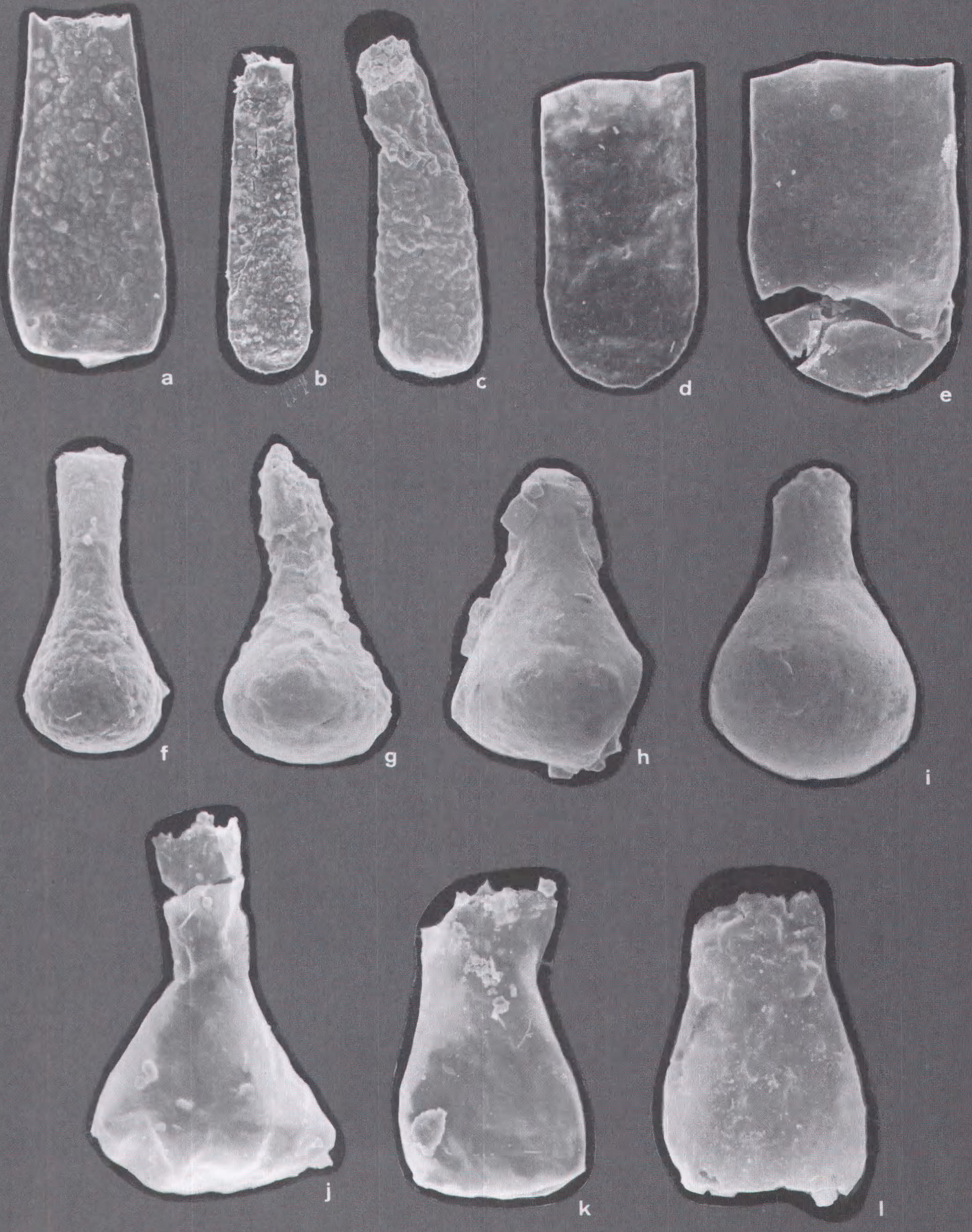
PLATE 3



## Plate 4

a. *Conochitina truncata*:
L = 305 µm; D = 115 µm; DS-038-99/C Ch 1
b. *Conochitina truncata*:
L = 340 µm; D = 90 µm; DS-033-99/608 Ch 27
c. *Conochitina tuba* Eisenack, 1932:
L = 220 µm; D = 80 µm; DS-026-99 Ch 1
d. *Conochitina tuba* Eisenack, 1932:
L = 197 µm; D = 102 µm; DS-033-99/608 Ch 15
e. *Conochitina* aff. *fortis* Eisenack, 1932 in Paris, 1981:
L = 170 µm; D = 105 µm; DS-037-99/469 Ch 8
f. *Sphaerochitina jaegeri* nomen nudum in Pittau, 1998:
L = 155 µm; D = 68 µm; d = 30 µm; DS-037-99/610 Ch 6
g. *Sphaerochitina jaegeri* nomen nudum in Pittau, 1998:
L = 107 µm; D = 56 µm; d = 20 µm; DS-027-99/514 Ch 4
h. *Sphaerochitina serpaglii* nomen nudum in Pittau, 1998:
L = 92 µm; D = 65 µm; d = 25 µm; DS-027-99/607 Ch 2
i. *Sphaerochitina serpaglii* nomen nudum in Pittau, 1998:
L = 85 µm; D = 54 µm; d = 20 µm; DS-027-99/607 Ch 5
j. *Sphaerochitina* sp. B in Verniers, 1982:
L = 93 µm; D = 62 µm; d = 20 µm; DS-042-99/609 Ch 2
k. *Vitreachitina* sp. 2 in Nestor, 1994:
L = 97 µm; D = 68 µm; d = 46 µm; DS-042-99/609 Ch 4
l. *Vitreachitina* sp. 2 in Nestor, 1994:
L = 100 µm; D = 61 µm; d = 40 µm; DS-042-99/609 Ch 5

PLATE 4



# Plate 5

a. *Angochitina longicollis* Eisenack, 1959:
L = 150 µm; D = 68 µm; d = 38 µm;
DS-037-99/469 Ch 2

b. *Angochitina longicollis* Eisenack, 1959: detail
stekels (max. 5 µm); DS-037-99/469 Ch 30

c. *Angochitina longicollis* Eisenack, 1959:
L = 173 µm; D = 80 µm; d = 40 µm;
DS-033-99/608 Ch 8

d. *Angochitina longicollis* Eisenack, 1959:
L = 200 µm; D = 70 µm; d = 37 µm;
DS-033-99/608 Ch 10

e. *Ancyrochitina ancyrea* ? Eisenack, 1931:
L = 120 µm; D = 79 µm; d = 40 µm;
DS-037-99/610 Ch 29

f. *Ancyrochitina ancyrea* ? Eisenack, 1931:
L = 75 µm; D = 65 µm; d = 30 µm;
DS-027-99/514 Ch 20

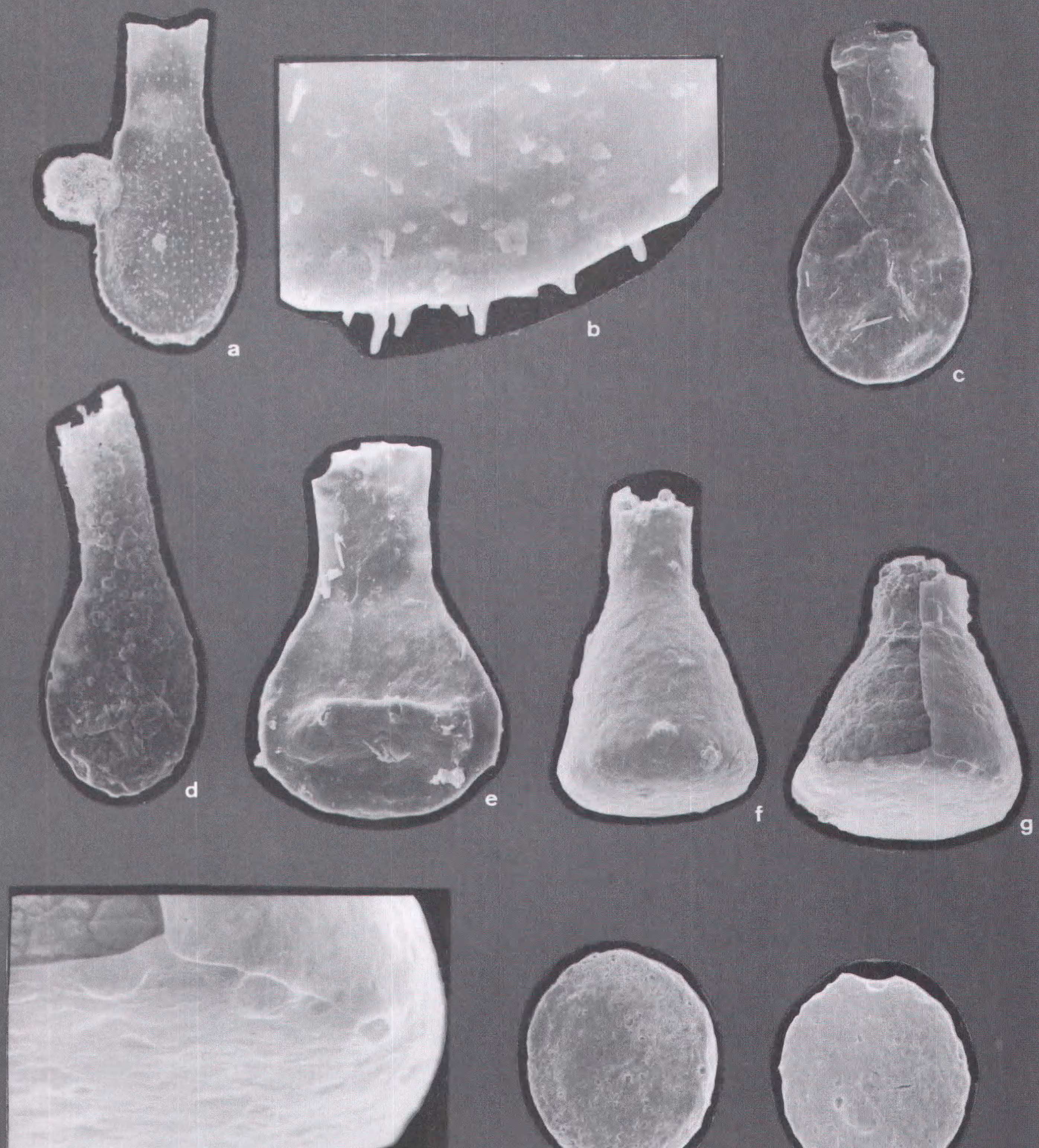
g. *Ancyrochitina ancyrea* ? Eisenack, 1931:
L = 50 µm; D = 55 µm; d = 24 µm;
DS-027-99/514 Ch 11

h. *Ancyrochitina ancyrea* ? Eisenack, 1931: detail of
broken appendix;
DS-037-99/610 Ch 29

i. Leiosphere:
L = 158 µm; D = 132 µm; DS-027-99/514 Ch 22

j. Leiosphere:
L = 160 µm; D = 130 µm; DS-025-99 Ch 5

PLATE 5



# Plate 6

Basse-aux-Canes and Vitrival-Bruyère formations

a. *Belenochitina capitata*, bb33.2, 99/0539A, 2, 0598, 400x
b. *Conochitina dolosa*, bb33.2, 99/0539A, 2, 0575, 110x
c. *Laufeldochitina stentor*, bb40.3, 99/0493, 2, 0372, 140x
d. *Desmochitina typica*?, bb40.7, 99/0501, 2, 0400, 600x
e. *Cyanthochitina campanulaeformis*, bb40.5, 99/0508, 2, 0324, 290x
f. *Cyanthochitina calix*, bb40.5, 99/0508-2, 2, 160x
g. *Laufeldochitina stentor*, bb40.3, 99/0493, 2, 0369, 100x
h. Chain of two *Desmochitina cocca*, bb40.5, 99/0508-2, 2, 350x
i. Chain of five *Desmochitina cocca*, bb40.9, 99/0495A, 1, 0222, 290x
j. *Belenochitina robusta*, bb40.7, 99/0501, 2, 0403, 290x
k. detail of spines of *Belonechitina robusta*, (6.10), 1500x

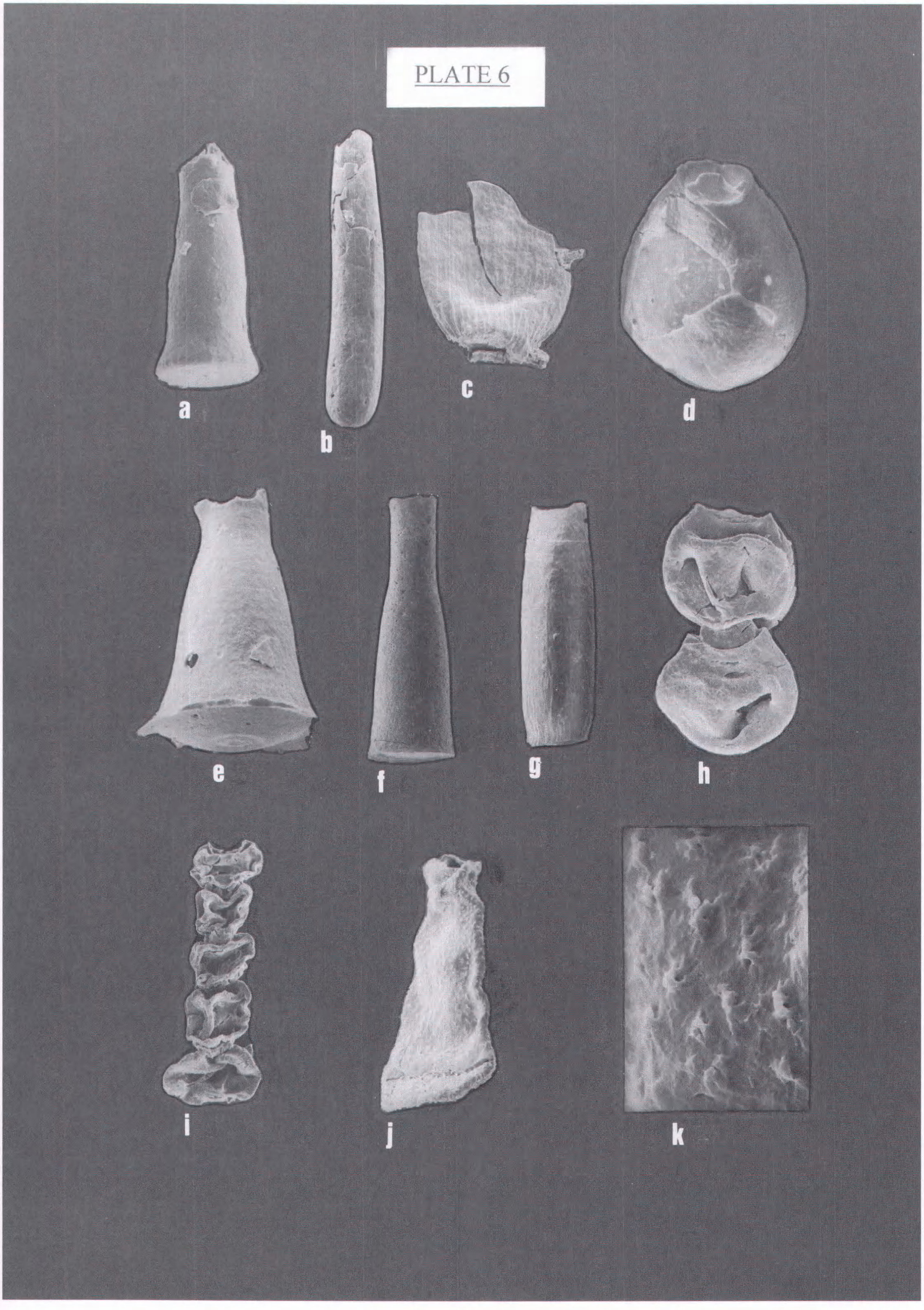


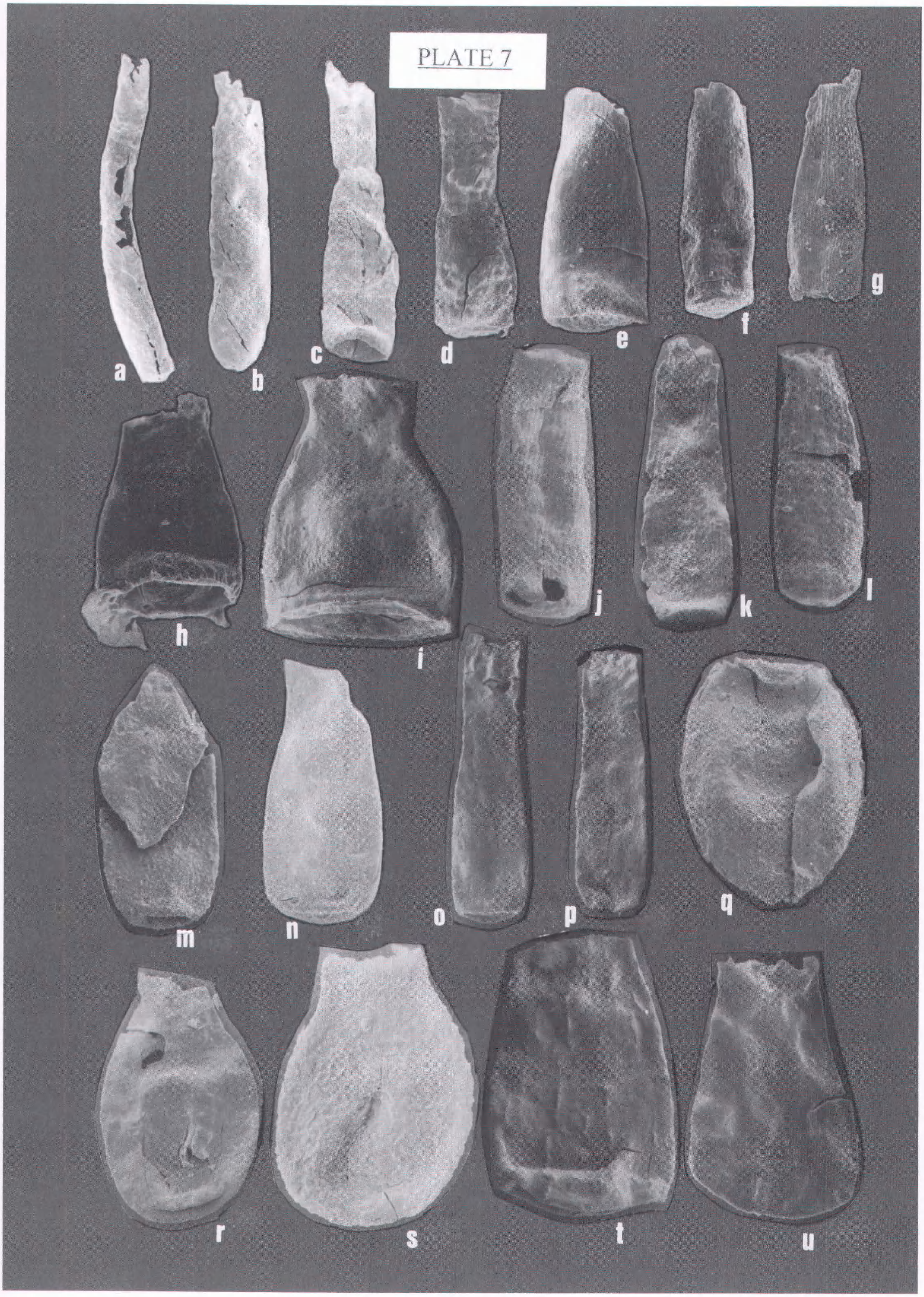
PLATE 6
a
b
c
d
e
f
g
h
i
j
k

# Plate 7

Fosses & Génicot formations

a. *Tanuchitina bergstroemi*, bb40.13, 99/0497A, 1, 0270, 140x
b. *Conochitina minnesotensis*, bb40.13, 99/0497A, 1, 0252, 175x
c. *Spinachitina taugourdeaui*?, bb33.3, 1, 0183, 210x
d. *Spinachitina bulmani*, bb33.3, 99/0448, 1, 0172, 290x
e. *Cyathochitina jenkinsi*?, bb33.5, 99/0449, 1, 0118, 290x
f. *Conochitina rugata*, bb33.5, 99/0449, 1, 0127, 235x
g. *Cyathochitina jenkinsi*?, bb33.5, 99/0449, 1, 0130, 200x
h. *Armoricochitina nigerica*, bb15.6, 99/0579B, 2, 0633, 350x
i. *Armoricochitina nigerica*, f54, bb33.12, nr. 54, 650x
j. *Conochitina rugata*, f24, bb33.6, nr. 5, 350x
k. *Conochitina rugata*, f26, bb33.6, nr. 11, 350x
l. *Conochitina rugata*, f29, bb33.6, nr. 21, 400x
m. *Belonechitina* spp., f68, bb 3/4/2000B, nr. 18, 500x
n. *Belenochitina robusta*, bb15.6, 99/0579B, 2, 0624, 290x
o. *Conochitina* spp., f47, bb33.12, nr.19, 300x
p. *Belonechitina* spp., f56, bb33.12, nr.57, 400x
q. *Desmochitina minor*, f40, bb33.12, nr.2, 850x
r. *Sphaerochitina* spp., f71, bb3/4/2000B, nr.24, 400x
s. *Sphaerochitina* spp.,f65, bb3/4/2000B, nr.15, 850x
t. *Hercochitina cf gamachiana* f52, bb33.12, nr.41, 850x
u. *Conochitina* spp., f60, bb3/4/2000B, nr.2, 850x

# PLATE 7

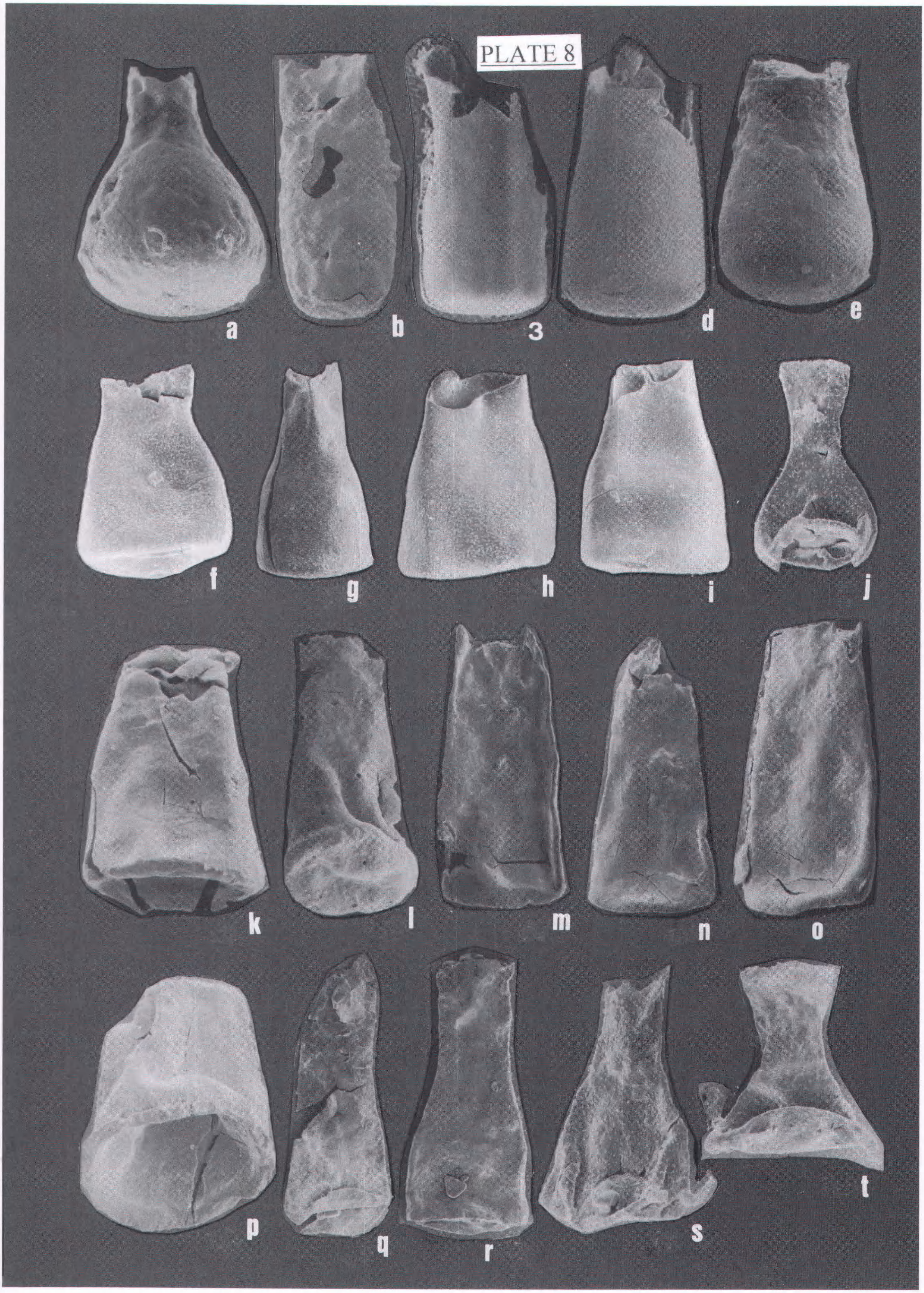


# Plate 8

Génicot Formation

a. *Ancyrochitina* spp., f32, bb33.11a, nr.4, 1200x
b. *Conochitina* spp., f36, bb33.11b, nr.7, 650x
c. *Eisenackitina* aff. *toddingensis*, f73, bb3/4/2000C, nr.1, 600x
d. *Eisenackitina* aff. *toddingensis*, f75, bb3/4/2000$D_1$, nr.3, 1000x
e. *Eisenackitina* aff. *toddingensis*, f76, bb3/4/2000$D_2$, nr.4, 1000x
f. *Eisenackitina* aff. *toddingensis*, bb33.16, 99/0467, 1, 0108, 450x
g. *Eisenackitina* aff. *toddingensis*, bb33.16, 99/0467, 1, 0099, 290x
h. *Eisenackitina* aff. *toddingensis*, bb33.16, 99/0467, 1, 0097, 450x
i. *Eisenackitina* aff. *toddingensis*, bb33.16, 99/0467, 1, 0103, 290x
j. *Angochitina* sp., bb33.16, 99/0467, 1, 0098, 450x
k. *Spinachitina* spp., f01, bb3/4/2000A, nr.6, 850x
l. *Spinachitina* spp., f00, bb3/4/2000A, nr.2, 750x
m. *Belonechitina* spp., f04, bb3/4/2000E, nr.9, 600x
n. *Spinachitina* spp., f06, bb3/4/2000E, nr.19, 500x
o. *Conochitina* spp., f09, bb3/4/2000E, nr.27, 500x
p. *Armoricochitina nigerica*, f02, bb3/4/2000E, nr.2, 750x
q. *Spinachitina* spp., f10, bb3/4/2000H, nr.5, 600x
r. *Cyathochitina* spp., f18, bb3/4/2000H, nr.31, 500x
s. *Ancyrochitina* spp., f13, bb3/4/2000H, nr.18, 850x
t. *Ancyrochitina* spp., f14, bb3/4/2000H, nr.20, 750x

PLATE 8



# Plate 9

Génicot Formation and Criptia Group

a. *Hercochitina* spp., f19, bb3/4/2000J, nr.1, 650x
b. *Armoricochitina nigerica*, f11, bb3/4/2000H, nr.6, 500x
c. *Belonechitina* spp., f20, bb3/4/2000J, nr.2, 650x
d. *Hercochitina* cf *gamachiana*, f23, bb3/4/2000J, nr.7, 750x
e. *Conochitina electa*, bb40.11, 99/502, 2, 0467, 400x
f. *Conochitina electa*?, bb40.11, 99/502B, 2, 0459, 500x
g. *Conochitina iklaensis*, bb40.11, 99/502B, 2, 0461, 235x
h. *Conochitina electa*?, bb40.11, 99/502A, 2, 0445, 450x
i. *Sphaerochitina* sp., bb40.10, 99/0496, 1, 0153, 400x
j. *Belenochitina postrobusta*, bb40.10, 99/0496, 1, 0160, 290x
k. *Belenochitina postrobusta*, bb40.10, 99/0496, 1, 0160, 1450x detail.

PLATE 9