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MEMOIRS OF THE GEOLOGICAL SURVEY OF BELGIUM
N. 49 - 2003

Palaeozoic deformation history of the Asquempont-Virginal area (Brabant Massif, Belgium)

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by

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(30 pages, 26 figures)

cover illustration: Sub-vertical fault zone separating the Oisquercq Formation from an unnamed unit and of which the crush-breccia/protocataclasite is made up mainly of the Chevlipont Formation. Central Asquempont section, canal Brussels-Charleroi between 40.114 and 40.120 km (see § 3.4.2).

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PALAEOZOIC DEFORMATION HISTORY OF THE ASQUEMPONT-VIRGINAL AREA (BRABANT MASSIF, BELGIUM): LARGE-SCALE SLUMPING, LOW-ANGLE EXTENSIONAL DETACHMENT DEVELOPMENT (THE ASQUEMPONT FAULT REDEFINED) AND NORMAL FAULTING (THE NIEUWPOORT-ASQUEMPONT FAULT ZONE)

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(26 figures)

ABSTRACT. Outcrop data are presented from the Asquempont-Virginal area, type locality of the Asquempont fault. These are used to redefine the Asquempont fault, to stress the importance of normal faulting in the Palaeozoic deformation history of the Brabant Massif and to suggest large-scale slumping during the middle Caradoc.
The Asquempont fault is redefined as a pre-cleavage and pre-folding low-angle extensional detachment that, in the study area, forms the contact between the Lower to lower Middle Cambrian Oisquercq Formation and the Ordovician and was active between the middle Caradoc and the timing of cleavage development.
The fault originally described by Legrand (1967) as the Asquempont fault, supposedly a steeply NE-dipping reverse fault with a sinistral component, is in fact a steeply dipping to sub-vertical post-cleavage normal fault that in the classical outcrop of the Asquempont fault truncates the redefined Asquempont fault, and hence only locally forms the limit between the Cambrian and the Ordovician.
The Asquempont fault *sensu* Legrand (1967) forms part of a large set of post-cleavage normal faults representing the eastern extension of the Nieuwpoort-Asquempont fault zone. Thus far, there is no evidence for strike-slip movement along this fault zone. The investigated large normal faults appear to have formed between the time of cleavage development and the Givetian.
The aeromagnetic Asquempont lineament is not related to the Asquempont fault redefined in this study, nor to the Asquempont fault *sensu* Legrand (1967).
Instead of the overturned anticline hypothesis ("*anticlinal renversé*") of Legrand (1967), the overturned beds within the Asquempont synform are attributed to large-scale slumping during the middle Caradoc.

Keywords: cleavage/fold relationship, extensional detachment, Lower Palaeozoic, normal faulting, slate belt, slump

1. INTRODUCTION

The Asquempont-Virginal area, 25 km SSW of Brussels, is situated within one of the most important outcrop areas of the Lower Palaeozoic Brabant Massif (fig. 1). This outcrop area, the Senne-Sennette area, contains most of the Lower Palaeozoic formations, ranging from the lowermost Cambrian to the upper Silurian, includes the well-known Upper Ordovician volcanic deposits of the Fauquez area, shows the angular unconformity between the deformed Lower Palaeozoic and the relatively undeformed Givetian and younger deposits (the Inclined Shiplift of Ronquières) and contains the enigmatic limit between the steep Cambrian core and the Ordovician-Silurian southern part of the massif. Legrand (1967) published a cross-section through this outcrop area in which all these features are incorporated (fig. 2). To date, this cross-section is still widely used (e.g. Sintubin, 1999; Hennebert & Eggermont, 2002). On this cross-section a number of structural complexities occur, of which two are situated within the Asquempont-Virginal area. These are the Asquempont fault and the Asquempont synform.
The Asquempont fault (Legrand, 1967; this is the Virginal fault of Mortelmans, 1955), named by Legrand (1967) after its type locality, the Asquempont area, (the Virginal area in Mortelmans, 1955), is commonly considered as one of the most important faults of the Brabant Massif. Although Dumont (1848), Malaise (1873, 1908) and Mortelmans (1955) all clearly mentioned the exis-

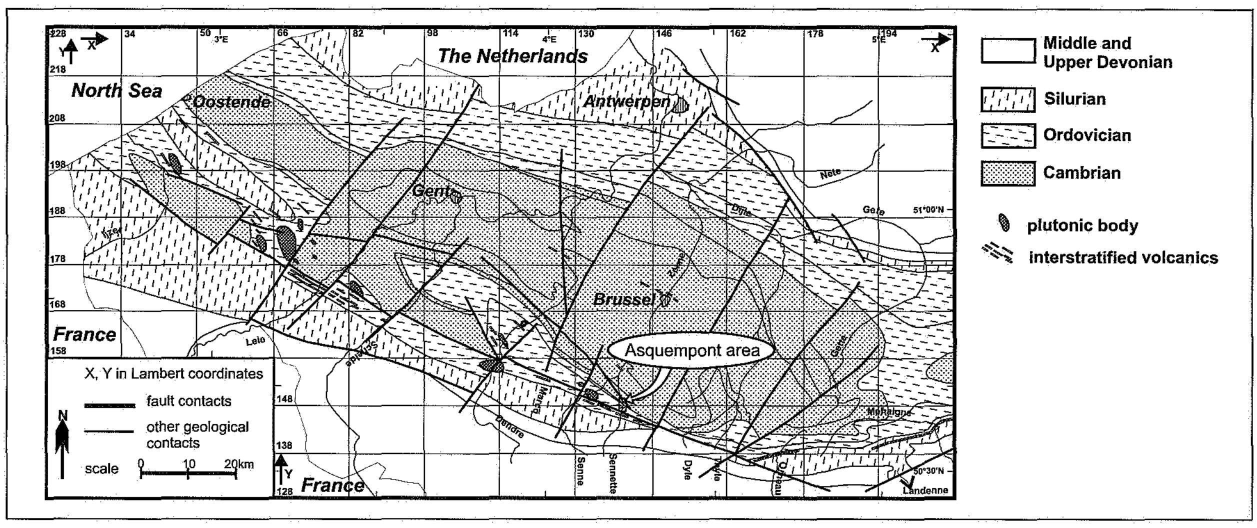

Figure 1: Geological subcrop map of the Brabant Massif (after De Vos *et al.*, 1993 and Van Grootel *et al.*, 1997), with the position of the Asquempont area.

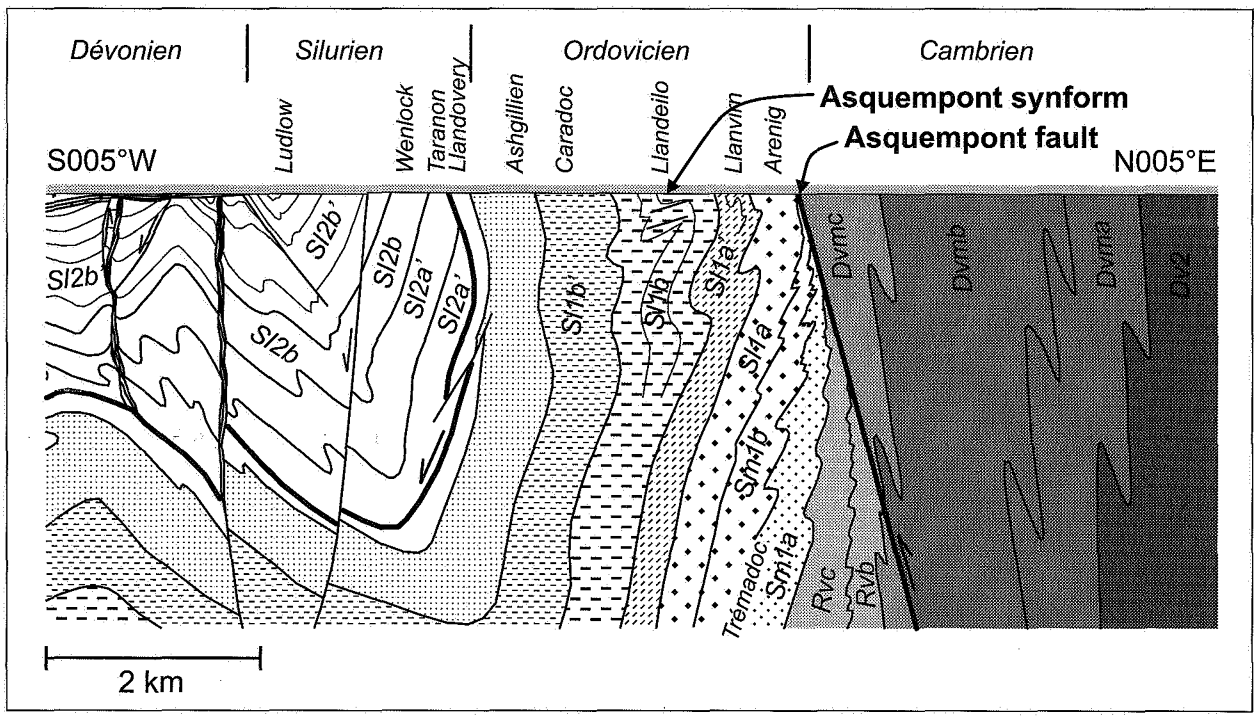

Figure 2: Part of the geological section of Legrand (1967), showing the Asquempont fault and the Asquempont synform.

tence of this fault, it was only Legrand (1967) who published a short fault description. He considered it as a steeply NE-dipping reverse fault, with a sinistral component, forming the limit between the Oisquercq Formation and the Tremadoc, hence placing the Lower to lower Middle Cambrian on top of the lowermost Ordovician (fig. 2, fig. 3). However, there is no (kinematic) evidence presented and on his geological map of the Brabant Massif, published one year later, this supposedly important fault is not depicted (Legrand, 1968).

The estimated importance of the Asquempont fault entirely depended on the stratigraphy (fig. 3). Fourmarier (1914, 1921), for instance, never saw real evidence for an important fault because a) he did not observe a fault in the Asquempont-Virginal area between the Cambrian and the Ordovician ("*on voit nettement dans la tranchée qu'il n'y a pas de faille au contact des deux formations mais qu'elles passent progressivement de l'une à l'autre*") and b) he was not convinced that the Upper Cambrian was missing in the Asquempont-Virginal area.

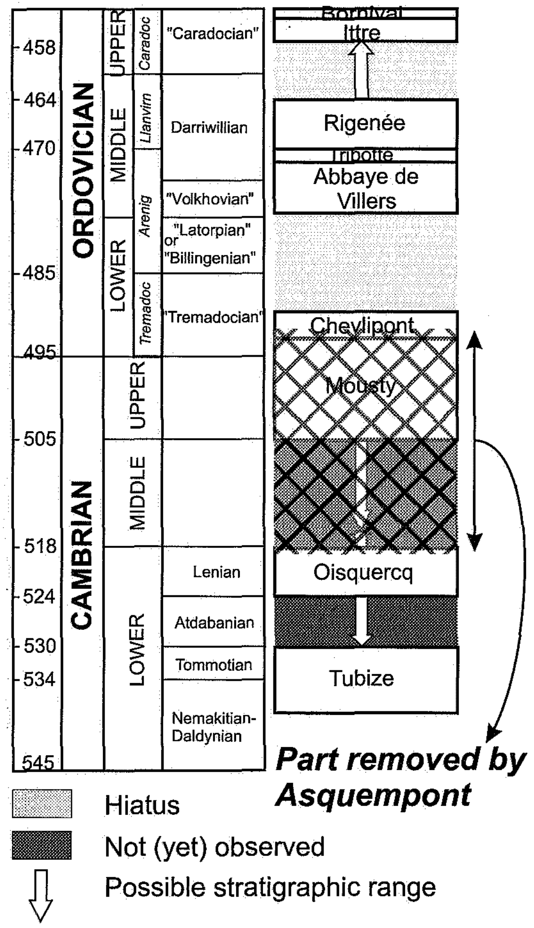

Figure 3: Stratigraphy of the Asquempont area (after Verniers *et al.*, 2001), showing the effect of the Asquempont fault (Debacker *et al.*, submitted).

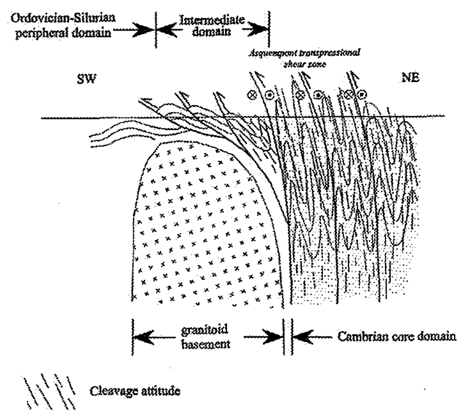

Figure 4: Conceptual cross-section across the southwestern side of the Brabant Massif (parallel to the canal Brussels-Charleroi, but 10 km more to the west), based on structural outcrop data and potential field data (after Sintubin, 1999). In this model, the aeromagnetic Asquempont lineament (the southwesternmost steep dextral transpressive fault), is equated with the Asquempont fault as described by Legrand (1967; i.e. steep reverse fault forming the limit between Cambrian and Ordovician; compare with fig. 2).

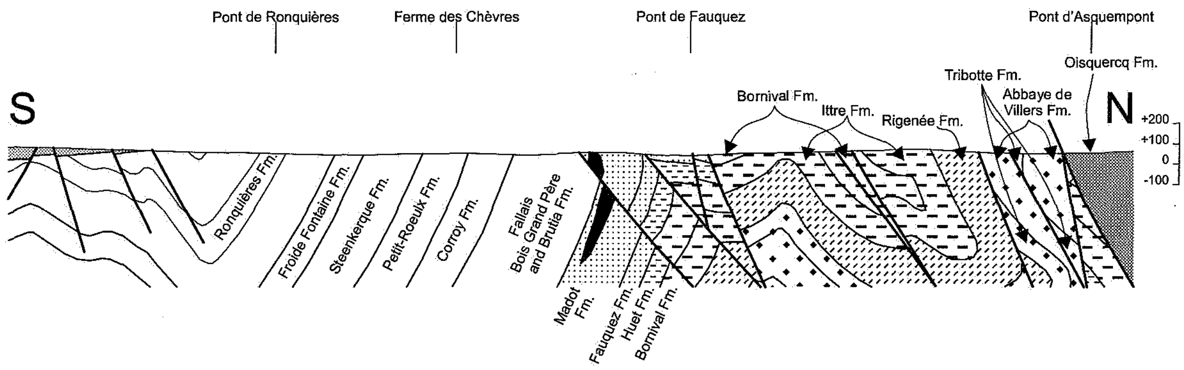

Figure 5: Part of the geological section accompanying the new geological map of the Asquempont-Virginal area (Hennebert & Eggermont, 2002). Note the similarity with the section of Legrand (1967; fig. 2) in terms of the Asquempont fault and the Asquempont synform. The Asquempont fault is still shown as a very important reverse fault that forms the limit between the Cambrian and the Ordovician (and even truncates other faults), and the Asquempont synform is still explained as an overturned anticline, although not with the same geometry as that depicted in Legrand (1967).

Indeed, several researchers (e.g. Fourmarier, 1921) argued that the Oisquercq Formation might be a lateral, westerly, equivalent of the Mousty Formation of the Dyle-Thyle area, later dated as Upper Cambrian to lowermost Tremadoc (for ages: see Lecompte, 1949; Martin, 1976; Vanguestaine *et al.*, 1989). It was only by the dating of the Oisquercq Formation in the Lessines borehole by Vanguestaine (1991; see also Herbosch *et al.*, 1991) as Lower Cambrian to lowest Middle Cambrian by means of acritarchs that there was proof of the apparent absence of the Upper Cambrian and hence the importance of this fault (fig. 3).

Later, a lot of authors used the description of the Asquempont fault by Legrand (1967) in combination with geophysical data to make large-scale interpretations of the structural architecture of the Brabant Massif (Sterpin & De Vos, 1996; Sintubin, 1997a, 1999; Sintubin *et al.*, 1998; Debacker, 1999; Mansy *et al.*, 1999) without, however, examining this fault (fig. 4). Even the most recent geological map of the Asquempont-Virginal area (Hennebert & Eggermont, 2002) is still largely influenced by the ideas of Legrand (1967) (fig. 5; cf. fig. 2). Recent structural fieldwork and mapping in the Asquempont-Virginal area, however, shows a completely different image (Debacker, 2001).

The Asquempont synform is a hectometre-scale synform to the south of the Asquempont fault, the core of which is composed of deposits of the Ittre Formation and the Bornival Formation (middle Caradoc; Verniers *et al.*, 1999; Samuelsson & Verniers, 2000; Van Grootel *et al.*, 1998). Several researchers noticed the overturned nature of the deposits within the core of this synform (Legrand, 1967; Legros, 1991; Servais, 1991). On his cross-section, Legrand (1967) shows this synformal structure as an overturned anticline (fig. 2; cf. fig. 5). As will be demonstrated, this explanation is difficult to reconcile with the cleavage/fold-relationship throughout this fold and with the concept of a single-phase deformation in the outcrop areas of the Brabant Massif, advocated recently in several papers (Debacker *et al.*, 1997, 1999; Debacker, 1999, 2002; Sintubin *et al.*, 1998; Sintubin, 1997b, 1999). Nevertheless, the cross-section accompanying the new geological map of the Asquempont-Virginal area (Hennebert & Eggermont, 2002), still uses the explanation of Legrand (1967).

2. AIM

2.1. FAULTS

We will demonstrate the importance of both post-cleavage and pre-cleavage normal faults in the Asquempont-Virginal area and explain our interpretation (Debacker, 2001; Debacker *et al.*, submitted) in which the Asquempont fault (i.e. the limit between the Oisquercq Formation and the Ordovician formations) is not a reverse fault but instead represents a low-angle extensional detachment, that formed prior to folding and cleavage development.

In doing so we will point to the absence of any relationship between three features that are often regarded as one fault structure. These are: the Asquempont fault according to Legrand (1967), the original anomalous Cambrian-Ordovician contact, being the Asquempont fault as redefined by us (Debacker, 2001; Debacker *et al.*, submitted), and the aeromagnetic Asquempont lineament (Sintubin & Everaerts, 2002).

The Asquempont fault *sensu* Legrand (1967). Legrand (1967) described the Asquempont fault along the new (post-1962) Brussels-Charleroi canal (stop 3.4) as : "*(la faille) présente des brèches de siltstone verdâtre du sommet de l'Assise de Oisquercq qui constitue sa lèvre Nord-Est. Elle affecte très peu le massif Cambrien mais est associée à un laminage intense de l'Ordovicien sous-jacent,...une faille quasi verticale pentée à 80° Nord-Est. La faille passe au Sud immédiat de l'écluse de l'ancien canal. Plus au Nord, on y a trouvé jadis un amas de galène, chalcopyrite et quartz. Plus au Nord encore, elle affleure à l'extrémité Nord de la boucle du canal* (stop 3. 1) *et enfin au Nord de la tranchée du chemin de fer* (stop 3. 2), *avec brèche de quartz. ...faille de charriage. Son rejet, inverse, est donc exceptionnellement important.*".

As we will show, this fault is a steep post-cleavage normal fault that, although partly forming the limit between the Cambrian Oisquercq Formation and the Ordovician along the classical outcrop at the E-side of the Brussels-Charleroi canal (stop 3.4), does not re-occur in other outcrops in the Asquempont-Virginal area. This fault is our fault F7.

The redefined Asquempont fault (Debacker, 2001; Debacker *et al.*, submitted). A pre-cleavage and pre-folding low-angle extensional detachment that, throughout the Asquempont-Virginal outcrop area, forms the limit between the Oisquercq Formation and the Ordovician. This fault can be observed along the E-side of the Brussels-Charleroi canal (stop 3.4), the Northern Virginal railway section (stop 3.2) and in the garden of one of the houses along the NE-side of the Bief 29 (Rue de l'Ancien Canal). This fault is our fault F8. On a larger scale, its trace appears to coincide with the Virginal fault of Mortelmans (1955).

The Asquempont lineament (Sintubin & Everaerts, 2002). A pronounced NW-SE-trending aeromagnetic lineament, forming the SW-limit of the aeromagnetic high of the Brabant Massif. Although, since it suggests an abrupt termination of the magnetite-bearing Tubize Formation, it is commonly interpreted as an important fault or shear zone (e.g. Chacksfield *et al.*, 1993), it bears no relationship with F7, nor with F8.

2.2. OVERTURNED BEDDING IN THE ASQUEMPONT SYNFORM

By means of a comparison of structural and sedimentological criteria, we will demonstrate that the reverse polarity within the Asquempont synform is a result of an overturning before cleavage development and suggest that this occurred by means of large-scale slumping during the middle Caradoc, at the time of deposition of the Bornival Formation (Debacker *et al.*, 2001).

3. STOPS

3.1. THE BIEF 29 SECTION

Approximately 350 m long, almost continuous series of outcrops along the NE-side of the Bief 29, an abandonned part of the old (pre-1962) canal Brussels-Charleroi, situated to the NW of the bridge of Asquempont, at Virginal (fig. 6, fig. 7). Part of the outcrop is below water level. This part of the section was studied during late summer of 1999, during extremely low water levels.

3.1.1. Lithology/sedimentology

From S to N (fig. 7): Distal turbidites of the Ittre Formation, homogeneous mudstones of the Rigenée Formation and clayey to sandy deposits of the Abbaye de Villers Formation. A more detailed description of these three formations can be found in Verniers *et al.* (2001). Millimetric to centimetric tuff layers occur in the vicinity of the limit between the Ittre and the Rigenée Formation (i.e. in silty deposits too homogeneous for the Ittre Formation, but rather coarse-grained for the Rigenée For-

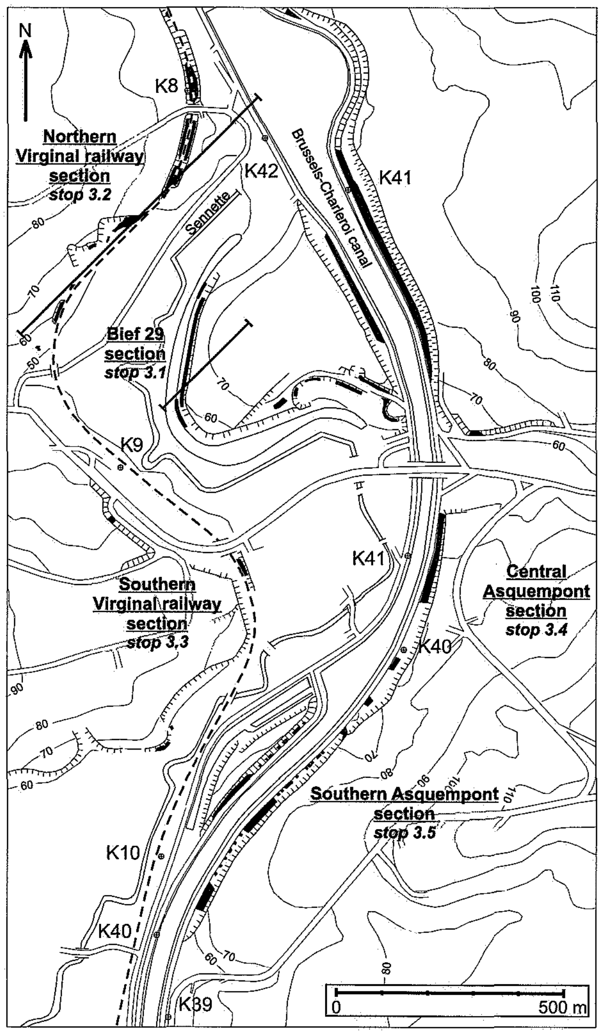

Figure 6: Schematic topographic map of the Asquempont area (after Debacker, 2001; Debacker *et al.*, submitted), showing the outcrop sections to be visited on the field trip.

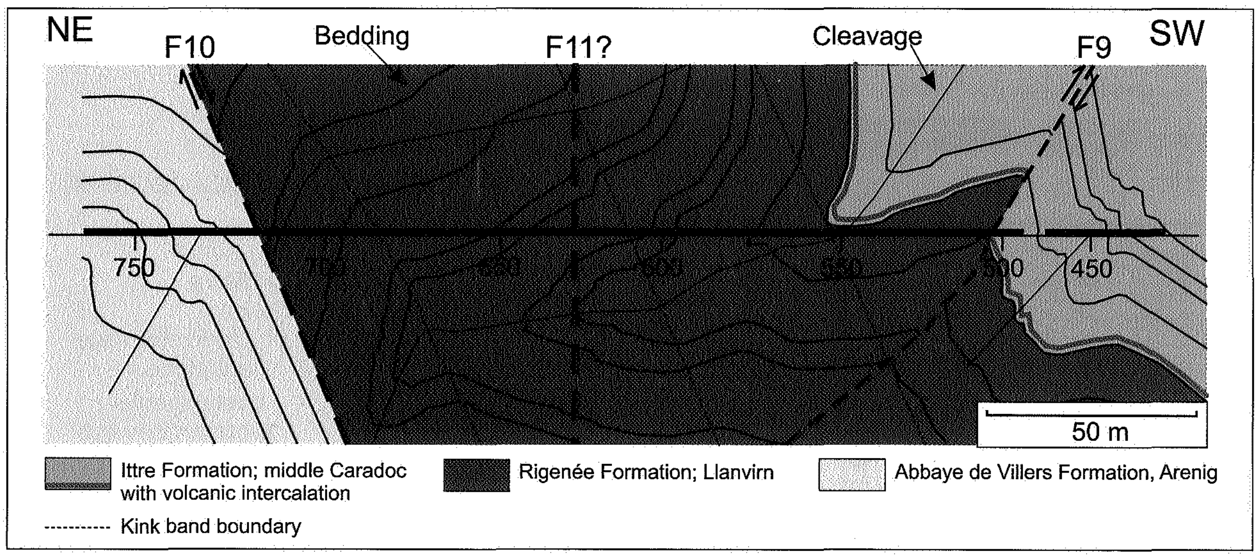

Figure 7: Cross-section along the Bief 29 section (see fig. 6 for section line), taken from Debacker (2001; Debacker *et al.*, submitted). The projected distances are measured along the NE-side of the Bief 29, starting from its easternmost extremity. Note the large-scale sigmoidal change of cleavage and bedding. Outcrops are marked in black.

mation). These tuff layers are much thinner than those described by Corin (1963) in the lowermost part of the Ittre Formation (cf. Verniers *et al.*, 2001), where they occur interstratified in fine-grained distal turbidite deposits and attain bed thicknesses of ~1 m (see 3.5; cf. 3.2), and therefore likely represent a different stratigraphic level. The three formations are separated by faults: a moderately N-dipping (stratigraphically) reverse fault between the Ittre and the Rigenée Formation (if one attributes the tuff layers and the silty deposits to the Rigenée Formation; F9) and a steeply S-dipping normal fault (zone) between the Rigenée and the Abbaye de Villers Formation (F10).

3.1.2. Structural features

Faults

The reverse (F9) and normal faults (F10, F11) have different aspects. The reverse fault (F9), post-cleavage and at a low angle to cleavage, has a subtle appearance, thin, without much wall-rock deformation (brecciation is relatively thin to absent) and is situated in the steep limb of an antiformal step fold, close to the hinge zone. In contrast, the normal faults (F10 and F11) are characterised by a several metre wide damage zone, consisting of semi-ductile (folded cleavage) to brittle (crush breccia to protocataclasite) post-cleavage wall-rock deformation and locally show evidence of fault reactivation (faulted crush breccia/protocataclasite) (fig. 8).

In places quartz veins occur. It is not clear whether or not these are related to normal faulting. Fault striations (not observed on F9) indicate a dip-slip movement.

The estimated amount of displacement along F10 is in the order of 200 metres (on basis of stratigraphy -Tribotte Formation is missing- and geometry), whereas that along F9 is unknown, but likely to be much less (geometry in combination with absence of important stratigraphic displacement).

Folds

Metre- to decametre-scale step folds occur in the three formations (those in the Ittre Formation and several of those in the Abbaye de Villers Formation occur in the part currently below water level). The folds are sub-horizontal to gently plunging and have a SW-verging asymmetry. Cleavage/fold relationships (cleavage fanning, sense of refraction, relation between cleavage and the fold axial surface...) point to a cogenetic relationship between cleavage development and folding. A small-angle cleavage transection is common, the sense of which may change from fold to fold. These changes in cleavage transection angle are related to changes in fold plunge and plunge

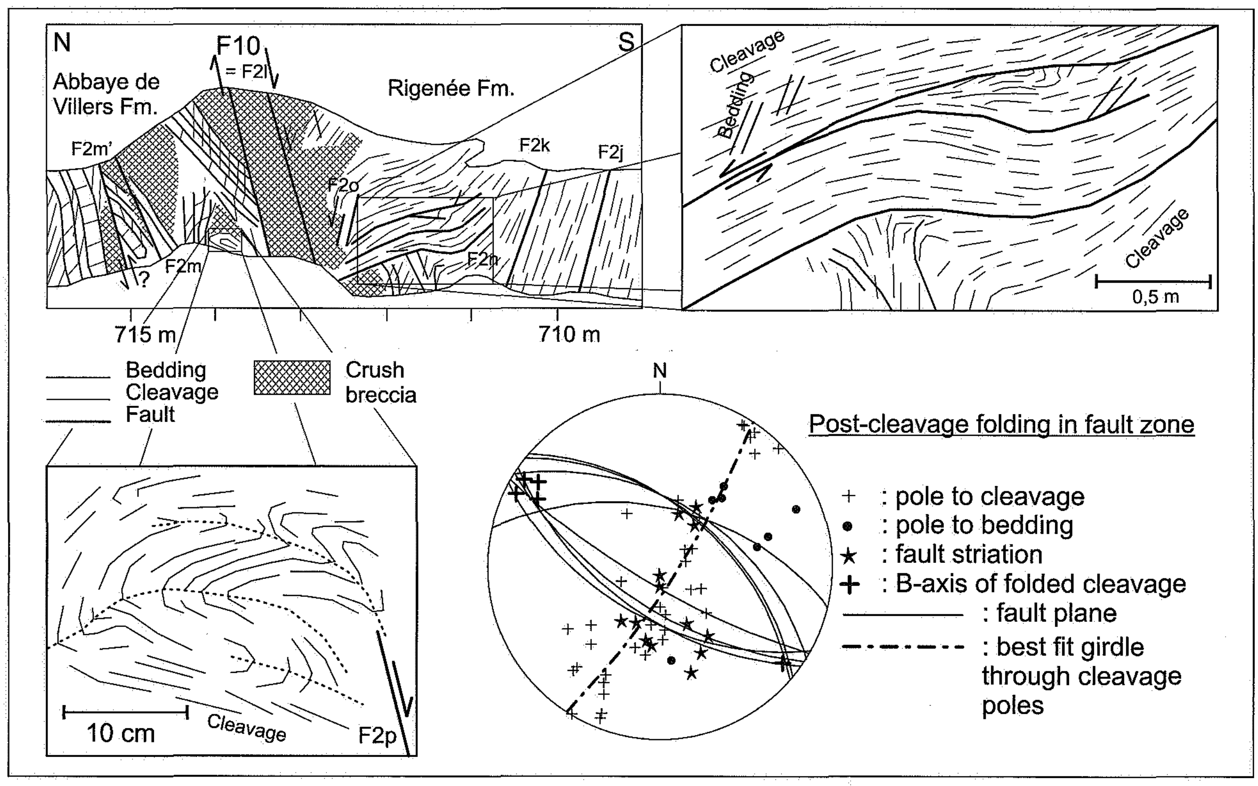

Figure 8: Structural data related to F10 (Bief 29 section), essentially forming a large post-cleavage normal fault zone, with a predominantly dip-slip displacement and a down-throw towards the south (taken from Debacker, 2001). Note the relatively wide damage zone, consisting of both brittle and semi-ductile cleavage deformation.

direction and may be attributed to a periclinal fold shape (cf. Treagus & Treagus, 1981; Debacker *et al.*, 1999). Pre-cleavage folds are observed at one location in the Rigenée Formation (between 600 and 602 m). These folds, close, with a steeply NW-dipping axial surface and wavelengths of 0.5 to 1 metre deform locally occurring coarser-grained beds (a-typical of the Rigenée Formation; cf. Verniers *et al.*, 2001). The pre-cleavage origin can be deduced from the cleavage/fold relationship: same refraction sense on both fold limbs.

Other structural features

A large-scale change in cleavage dip occurs (fig. 7): in the largest part of the exposed part of the Rigenée Formation the cleavage is sub-horizontal to gently NE-dipping, whereas both to the south and to the north, the cleavage is moderately to steeply NE-dipping. This sigmoidal change in cleavage-dip is interpreted as a large-scale post-cleavage kink band of at least 100 m wide, with steeply SW-dipping kink band boundaries.

3.1.3. Overall structure

Steeply SW-dipping beds, affected by metre- to decametre-scale tectonic step folds, locally accompanied by small reverse faults (F9), deformed by a large-scale kink band with steeply SW-dipping kink band boundaries and post-cleavage faults, predominantly steeply dipping normal faults (fig. 7).

3.2. THE NORTHERN VIRGINAL RAILWAY SECTION

Approximately 900 m long discontinuous series of outcrops along the old railway at Virginal, between 7.8 and 8.7 km (expressed in railway kilometres), situated to the NW of the Bief 29 (fig. 6, fig. 9).

3.2.1. Lithology/sedimentology

From NE to SW (fig. 9): Homogeneous greenish grey mudstone of the Oisquercq Formation (Asquempont member; locally bedding is visible in outcrop), distal turbidites of the Chevlipont Formation (with typical lense-shaped silty to sandy beds), clayey to sandy deposits of the Abbaye de Villers Formation, homogeneous dark mudstones of the Rigenée Formation, and distal turbidites of the Ittre Formation, with a several-metre thick volcanic intercalation in its lowermost part, containing mudstone fragments. Note that the latter feature points to a certain viscosity of the magma, and hence a relatively close source (i.e. within the Brabant Massif). See Verniers *et al.* (2001) for a more detailed description of the formations present. Fault contacts occur between the Oisquercq Formation and the Chevlipont Formation (the Asquempont fault: F8) and probably between the Abbaye de Villers Formation and the Rigenée Formation (F10?).

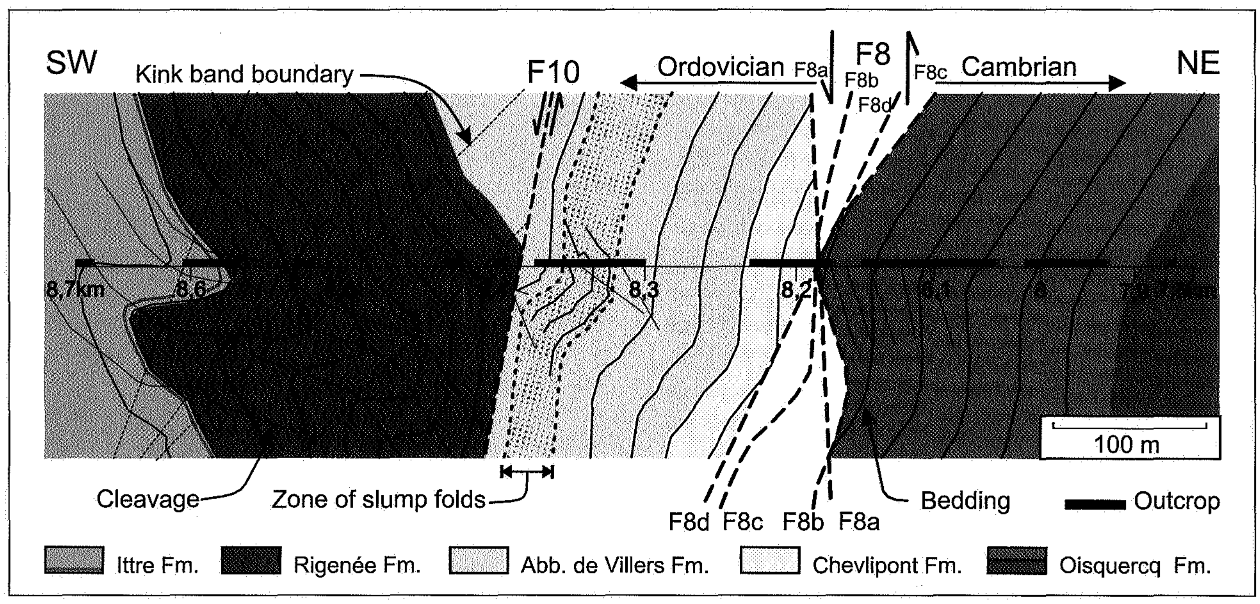

Figure 9: Cross-section along the Northern Virginal railway section (see fig. 6 for section line), taken from Debacker (2001; Debacker *et al.*, submitted), showing the Asquempont fault (F8), with four possible traces (two assuming a post-folding origin: F8a and F8d, two assuming a pre-folding origin: F8b and F8c), the likely position of F10 (cf. fig. 7), and the large-scale kinking of bedding and cleavage. Outcrops are marked in black and distances are expressed in railway kilometres.

3.2.2. Structural features

Faults

Small-displacement faults, local slip planes and brecciations occur at several places along the section. Some of these may be slump-related (pre-cleavage, without noticeable stratigraphic displacement), others reflect small post-cleavage normal slip planes, often running along the cleavage planes (also common in other outcrop areas of Brabant Massif; see Debacker, 2001, 2002, Van Grootel *et al.*, 2002). We will focus on those features that are reflected by a jump in stratigraphy.

At first sight there is no obvious fault between the Oisquercq Formation to the north and the Chevlipont Formation to the south: the nature of the outcrop remains the same, and both the joint sets and the cleavage remain largely undisturbed between the Oisquercq Formation in the north and the Chevlipont Formation in the south (cf. Fourmarier, 1914, 1921). It is only upon close examination that one notices the particular aspect of the contact between the two formations. Along the W-side of the railway, going south from 8.183 km onwards, the outcrop consists of lenses of variable sizes of the Oisquercq Formation and the Chevlipont Formation. To the north of this point only deposits of the Oisquercq Formation occur and to the south of 8.194 km (possibly a bit further north; sometimes difficult to see due to weathering and the nature of contacts) only deposits of the Chevlipont Formation are present. Within the zone between 8.183 and ~8.194 km, the contacts between the two different lithologies are usually welded, and the cleavage cuts through them, without being disturbed (except for some refraction due to competence contrasts). Only at the northern limit of this zone, some post-cleavage brittle deformation occurs. However, along the eastern side of the railway, the same pre-cleavage contact zone occurs, without containing evidence of post-cleavage deformation. Hence, the post-cleavage brittle deformation zone along the W-side of the railway can be regarded as a local phenomenon, unrelated to the limit between the Oisquercq Formation and the Chevlipont Formation. Both along the E-side and along the W-side of the railway cleavage planes are locally observed containing striations. However, such planes also occur in other places and since the cleavage cuts across the contacts between the two different lithologies, the striations cannot be related to these contacts. We interpret the pre-cleavage contact zone between the Chevlipont Formation and the Oisquercq Formation as F8, the Asquempont fault. Although in outcrop an apparent NE-dip becomes apparent, this may only be an impression resulting from the overprinting cleavage. Correlation between both sides of the railway of the northernmost occurrence of lenses of the Chevlipont Formation points to a trend of 330° (NW-SE), and an analysis of the alignment of, and contacts between, the different lenses within oriented hand specimens points to a sub-vertical to steeply SW-dipping, NW-SE to NNW-SSE-trending structure. Hence, F8 is slightly oblique to both bedding and cleavage.

As along the Bief 29 section (stop 3.1), also here the Tribotte Formation is missing. Although we cannot completely rule out a local non-deposition as a cause of this absence, a normal fault does form the contact between the Abbaye de Villers Formation and the Rigenée Formation along the Bief 29 section, thus probably removing the Tribotte Formation (fig. 7). Hence, also in the Northern Virginal railway section we propose the occurrence of this fault (F10).

Folds

Numerous folds occur along this section. However, folds that definitely have a syn-cleavage origin are rather scarce.

In the Chevlipont Formation, close to the Asquempont fault, a number of folds occur, with a normal cleavage/fold relationship (i.e. with a fold hinge line situated in the cleavage plane and contrasting senses of cleavage refraction on both fold limbs), but with strongly variable plunges, occasionally steeply plunging. Although the cleavage/fold relationship is compatible with a tectonic origin, these folds occur in zones with pre-cleavage breccias and other soft-sediment deformation features. Hence, we cannot ascertain whether these folds have a syn-cleavage or a pre-cleavage origin.

Within the Abbaye de Villers Formation numerous pre-cleavage folds occur (fig. 10). The pre-cleavage origin becomes apparent by the cleavage/fold relationships

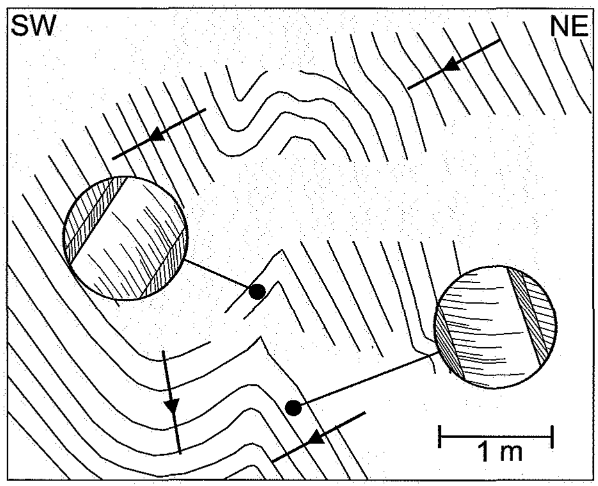

Figure 10: Pre-cleavage fold within the Abbaye de Villers Formation, between 8.339 and 8.344 km along the Northern Virginal railway section (Debacker, 2001). Note the same sense of cleavage refraction on both fold limbs, and the downward younging direction.

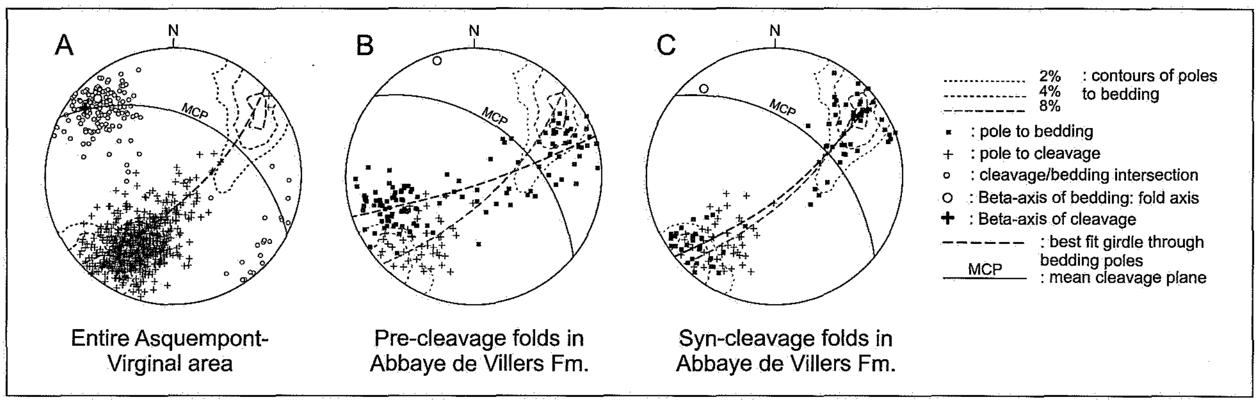

Figure 11: Lower-hemisphere equal-area stereographic projections of bedding and cleavage, demonstrating the differences in trend between pre-cleavage and syn-cleavage folds in the Abbaye de Villers Formation in the Northern Virginal railway section. A) data from the entire Asquempont-Virginal area, without the data of the overturned deposits of the Southern Asquempont section (see stop 3.5); B) data from the pre-cleavage folds in the Abbaye de Villers Formation; C) data from the syn-cleavage folds in the Abbaye de Villers Formation. To facilitate comparison, the data of plot A are shown as a grey background in plots B and C.

(same sense of refraction on both fold limbs, cleavage cross-cutting the axial surface), the difference between the tectonic polarity and the stratigraphic polarity, the pre-cleavage brecciations etc... The main pre-cleavage deformation occurs in a zone of sub-vertical to steeply SW-dipping bedding of approximately 40 metres wide (between 8.320 and 8.360 km, fig. 9). Noteworthy, bedding and fold hinge lines in the pre-cleavage deformation zones have a trend that is oriented approximately 020° clockwise of the regional trend (fig. 11). The pre-cleavage folds are not situated within the mean cleavage plane, but instead are situated within the mean bedding plane of the pre-cleavage deformation zones.
Only in the southernmost outcrops of the Abbaye de Villers Formation, between 8.360 and 8.376 km, do folds occur with a cleavage/fold relationship and an overall geometry (step folds with SW-verging asymmetry, sub-horizontal to gently plunging fold hinge lines) suggestive of a tectonic origin, cogenetic with cleavage. However, also here locally pre-cleavage deformation occurs (pre-cleavage breccia, welded truncations, zones with bedding trending clockwise with respect to the regional trend).

Other structural features

Although much poorer exposed than the Bief 29 section, also here a large-scale, sigmoidal change in cleavage dip becomes apparent within the Rigenée Formation (fig. 9). Here, this structure, which, by analogy with the Bief 29 section, we interpret as a large-scale kink band, also affects the lower parts of the Ittre Formation (with the volcanic deposits) and occupies a larger area than along the Bief 29 section. Possibly this is due to kink band bifurcation or to normal faulting (F11?).

3.2.3. Overall structure

The overall structure is very similar to that along the Bief 29 section, except that here also the Asquempont fault (F8), the Oisquercq Formation, the Chevlipont Formation and a large part of the Abbaye de Villers Formation are present, the latter with a 40 m-wide zone of pre-cleavage deformation. The southern part of the Northern Virginal railway section can easily be correlated with the Bief 29 section, both stratigraphically (same three formations) and structurally (large-scale kink band, F10, small step folds) (compare fig. 7 and fig. 9).

3.3. THE SOUTHERN VIRGINAL RAILWAY SECTION

Approximately 50 m long outcrop along the old railway line at Virginal, just south of the road Ittre-Virginal, between 9.368 and 9.320 km (expressed in railway kilometres), situated to the south of the Bief 29, just W of the paper factory (Papeterie de Virginal) (fig. 6).

3.3.1. Lithology/sedimentology

This outcrop entirely consists of steeply dipping deposits of the Abbaye de Villers Formation.

3.3.2. Structural features

Folds

Within the steeply dipping beds, several gentle, dm- to m-scale folds occur (cf. Herssens, 1957), some of which, judging from the cleavage/fold relationship, have a pre-cleavage origin (cf. Abbaye de Villers Formation in the Northern Virginal railway section).

3.3.3. Importance of this outcrop?

Because this outcrop is situated well to the south of the Northern Virginal railway section (stop 3.2) and the Bief 29 section (stop 3.1), both of them essentially consisting of steeply SW-dipping, SW-ward younging beds of the Abbaye de Villers Formation, the Rigenée Formation and the Ittre Formation, it points to a large-scale repetition of (at least) these three Ordovician formations. Towards the south of this outcrop volcanic fragments have been encountered, followed by fragments of the Ittre Formation and several small outcrops of the Ittre and the Bornival Formation, which can be correlated directly with the deposits in the large outcrops of the Southern Asquempont section along the new (i.e. post-1962) canal Brussels-Charleroi (see stop 3.5).

3.4. THE CENTRAL ASQUEMPONT SECTION

Approximately 450 m long series of outcrops along the E-side of the new (post-1962) Brussels-Charleroi canal, between 40.302 (at the bridge of Asquempont) and 39.850 km (expressed in canal kilometres) (fig. 6). Only the northern part of this section, between 40.302 and 40.085 km, is continuously exposed; further south the section contains several temporary outcrops situated in a densely forested slope. The classical outcrop of the Asquempont fault, as described by Legrand (1967), is situated along this section between 40.110 and 40.130 km (fig. 12), which, at the time of the onset of this study, was heavily covered with slope scree and vegetation. Note that km-post 40 was missing and distances were measured starting from km-post 39. Hence, the distances

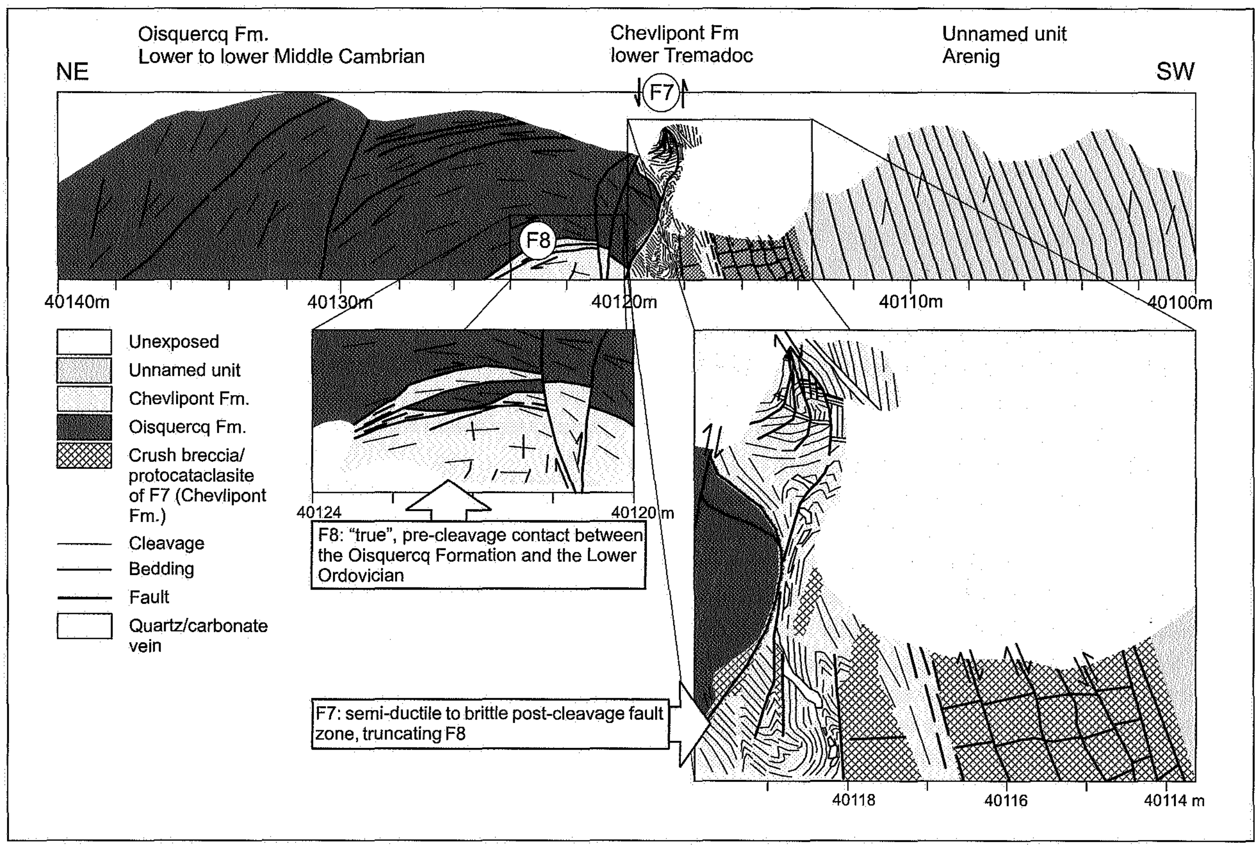

Figure 12: The classical outcrop of the Asquempont fault (Legrand, 1967), along the Central Asquempont section, E-side Brussels-Charleroi canal, after removal of debris and vegetation (taken from Debacker, 2001; Debacker *et al.*, submitted). Distances are expressed in canal kilometres. F8 pre-dates cleavage development and is affected by the post-cleavage overturning, small normal faults and F7. Locally, post-cleavage slip zones occur along parts of F8. These zones probably formed during the overturning of bedding and cleavage (flexural slip). F7 truncates F8, thus only forming an apparent limit between the Cambrian and Ordovician. F7 testifies of several episodes of movement, the early episodes accompanied by fold development, the later episodes by mainly brittle deformation. Note that, although depicted as Chevlipont Formation (dated as Tremadoc; Lenoir, 1987), the rocks within F7 do not strongly resemble the Chevlipont Formation. Only the rocks below F8, in the overturned part north of F7, have a lithology typical of the Chevlipont Formation.

used here do not match that of the newly placed km-post 40 (that in turn does not match the position on the topographical maps).

3.4.1. Lithology/sedimentology

From N to S: Homogeneous greenish grey mudstone of the Oisquercq Formation (Asquempont member; only locally bedding is visible), distal turbidites of the Chevlipont Formation (distinctive lithology is difficult to recognise because of brittle deformation; dated by means of acritarchs in fault crush-breccia/protocataclasite by Lenoir, 1987) and rather thick-bedded (~20 cm or more), predominantly silty to sandy deposits, probably of turbiditic origin, of the unnamed unit (dated with chitinizoans by Samuelsson & Verniers, 2000 and incorporated in the Abbaye de Villers Formation; see Verniers *et al.*, 2001), cut by numerous quartz veins. The small temporary outcrops in the forested talus to the south consist of the classical lithologies of the Abbaye de Villers Formation and the Tribotte Formation (cf. Verniers *et al.*, 2001). Note that this is the only place where deposits of the Tribotte Formation have been observed within the Asquempont-Virginal area (see also Hennebert & Eggermont, 2002).

3.4.2. Structural features

Faults

Between 40.114 and 40.120 km a sub-vertical fault zone occurs (F7) that separates the Oisquercq Formation (N) from the unnamed unit (S) and of which the crush-breccia/protocataclasite is made up mainly of the Chevlipont Formation (according to the dating as Lower Tremadoc by means of acritarchs by Lenoir, 1987) (fig. 12). The fault zone contains a steep fabric (microscopic observations show that this is bedding), that is locally folded (post-cleavage, as observed in thin sections) into small, irregular chevron-like folds. These folds are cut by quartz-carbonate veins, of which the quartz is often deformed and broken with the carbonate filling the gaps. The veins themselves are faulted, with a dip-slip displacement with a down-throw to the south (note: seen from above the displacement may seem dextral or oblique; this, however, is due to the fact that the veins are not vertical). In contrast, along the steep folded fabric a down-throw to the north becomes apparent from microscopic observations. In other parts this fault zone consists of a crush-breccia to cataclasite that is cut by later faults with a dip-slip displacement, usually with a down-throw to the south. These observations suggest that the fault was reactivated several times, with the later stages having a down-throw to the south. At the northern limit of F7, the crush breccia is made up of Ordovician fragments and fragments of the Oisquercq Formation (fig. 13A). According to the description of Legrand (1967), it is this post-cleavage fault zone which he considered as forming the limit between the Cambrian and the Ordovician and that he called the Asquempont fault. However, to the north of this fault zone, below an antiformal overturned part (overturning is post-cleavage) of the Oisquercq Formation, the Chevlipont Formation occurs (fig. 12). These Ordovician rocks are separated from the overlying Oisquercq Formation by a pre-cleavage contact zone of 0.5 to 1 m wide, in which, like in the Northern Virginal railway section, lenses of both the Oisquercq and the Chevlipont Formation occur (fig. 13B). This contact zone is cut by two small sub-vertical normal faults and by the post-cleavage fault zone (F7; Asquempont fault *sensu* Legrand, 1967). Parts of this contact zone contain post-cleavage crush-breccias. However, these breccias are not restricted to this pre-cleavage contact zone, but instead occur along numerous cleavage (and locally bedding) planes within the antiformal overturned part of the Oisquercq Formation and Chevlipont Formation. Hence, this post-cleavage brecciation is not considered related to the contact zone between the Oisquercq Formation and the Chevlipont Formation but instead is probably

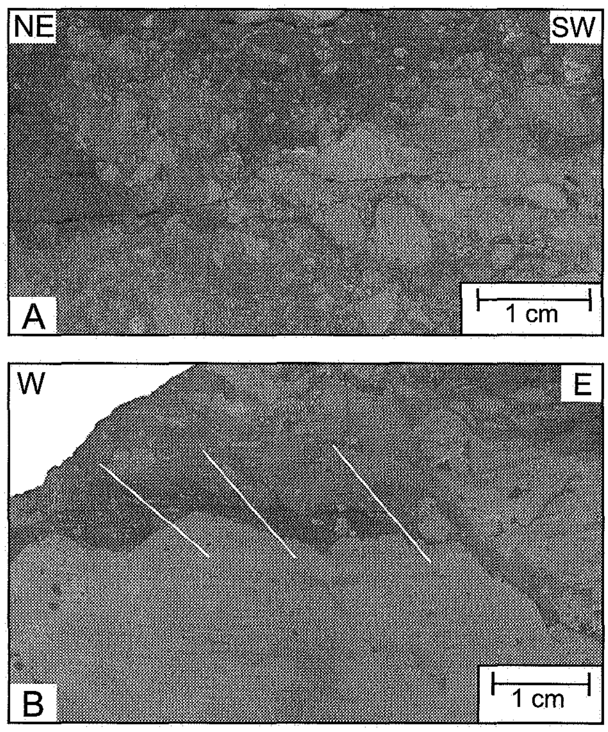

Figure 13: Difference in fault fabric between F7 and F8 (Debacker, 2001; Debacker *et al.*, submitted). A) F7: post-cleavage crush breccia/protocataclasite composed of Ordovician rocks (dark grey) and rocks of the Oisquercq Formation (light grey) (sample TD147); B) F8: pre-cleavage contact between the Oisquercq Formation (light grey) and the Ordovician (dark grey) (sample TD156). Cleavage is marked in white.

related to the post-cleavage overturning. It is the pre-cleavage contact zone that we redefine as the Asquempont fault (F8), the same structure that is observed in the Northern Virginal railway section (stop 3.2).

Around 40.085 km, a steep, semi-ductile to brittle post-cleavage deformation zone occurs, strongly resembling F7. Because of this similarity, also this structure is interpreted as a fault (F12). As argued by Legros (1991), because of the presence of the Abbaye de Villers Formation both to the north and to the south of this structure, it is likely to have had a smaller displacement than F7, and possibly represents a splay of F7.
Within the forested talus to the south of F12, the distribution of the Tribotte and the Abbaye de Villers Formation as observed in small temporary outcrops, suggests the presence of other faults (cf. Hennebert & Eggermont, 2002).

Veins
A large number of veins occur within the competent deposits of the unnamed unit, between F7 and F12. On the basis of orientation, three main vein sets can be distinguished, all of which are sub-vertical to steeply E-dipping and most of which are less than 3 cm thick. These three vein sets, two of which form conjugate sets and the third set bisecting the acute angle of the first two, can be used to deduce the stress pattern during an increment of compressive deformation: a gently NW-plunging minimum compressive stress, a steeply NE-plunging intermediate compressive stress (the intersection lineation of the three vein sets) and a gently SW-plunging maximum compressive stress, approximately coinciding with the cleavage pole. This coincidence in orientation suggests a coaxial deformation during the period between cleavage development and vein development. However, some asymmetry does come forward from the veins themselves: the dextral set occurs less frequently than the sinistral set, and sub-parallel to the sinistral set some thick veins occur (> 10 cm), filled with blocky quartz.

Folds
Along the entire section bedding is steeply dipping, on average steeply NE-dipping (slightly overturned) to sub-vertical in the part occupied by the Oisquercq Formation, and steeply SW-dipping to the S of F7. In the northernmost part of this section, the outcrop pattern suggests a metre-scale antiform-synform couple with a step fold geometry, an approximately axial planar cleavage and a gently NW-plunging hinge line. However, bedding is not evidenced in this part of the outcrop.
In the small temporary outcrops of the Abbaye de Villers Formation in the forested talus, pre-cleavage folds have been observed (Debacker, Herbosch & Verniers, unpub. data).

Other structural features
On first sight, the post-cleavage overturning adjacent to F7 (fig. 12) might be interpreted as a drag phenomenon. However, although the sense of overturning seems compatible with the down-throw to the south as suggested by the small faults deforming the quartz veins within F7, this geometry can only be formed as a drag phenomenon along F7 in the case where bedding is steeply NE-dipping and the fault is steeper NE-dipping (reverse fault), sub-vertical or SW-dipping (normal fault). This sense of movement would then also be compatible with the stratigraphic jump across F7. If, alternatively, bedding would be steeply SW-dipping, sub-vertical or steeply NE-dipping, and the fault was steeply to moderately NE-dipping, less steep than the bedding (reverse fault; situation depicted by Legrand, 1967), the sense of movement would still be compatible with the stratigraphic jump observed, but the structure could not be generated by drag. Hence, geometrically, only the first option seems possible. One may argue that it is not likely to have such a pronounced drag along one side of a fault only. However, taking into account the small angle between bedding and F7, both being steep, it takes only a change of a few degrees for the cut-off angle and its sense (e.g. 10° or -10°) and hence the deduced sense of movement and the possibility to form obvious drag features to change.
An alternative explanation for this overturning is that it represents the northern part of the large-scale kink band observed in the Northern Virginal railway section, the Bief 29 section and a group of outcrops at the N-side of the eastern part of the Bief 29. These observations indicate that this large-scale kink band may well continue SE-wards towards the Central Asquempont section where it gets truncated by F7. In that case, F7, sub-vertical to steeply NE-dipping, would have a down-throw towards the north.

3.4.3. Overall structure

The overall structure consists of a steeply SW-dipping limb, containing the redefined, pre-cleavage, Asquempont fault (F8), the Oisquercq Formation, the Chevlipont Formation, the Abbaye de Villers Formation and the Tribotte Formation, that is cut by several post-cleavage faults, of which the most important is F7 (fig. 12). It is the latter fault that was described by Legrand (1967) as the fault responsible for the contact between the Oisquercq Formation and the Ordovician (Asquempont fault *sensu* Legrand, 1967). Because it is only in this outcrop that F7 partly coincides with the limit between the Oisquercq Formation and the Ordovician, this being due to the particular position of this outcrop at the junction of F7 and F8, we redefine the Asquempont fault as F8: the original, pre-cleavage tectonic contact between the Oisquercq Formation and the Ordovician in the Asquempont-Virginal area (fig. 12).

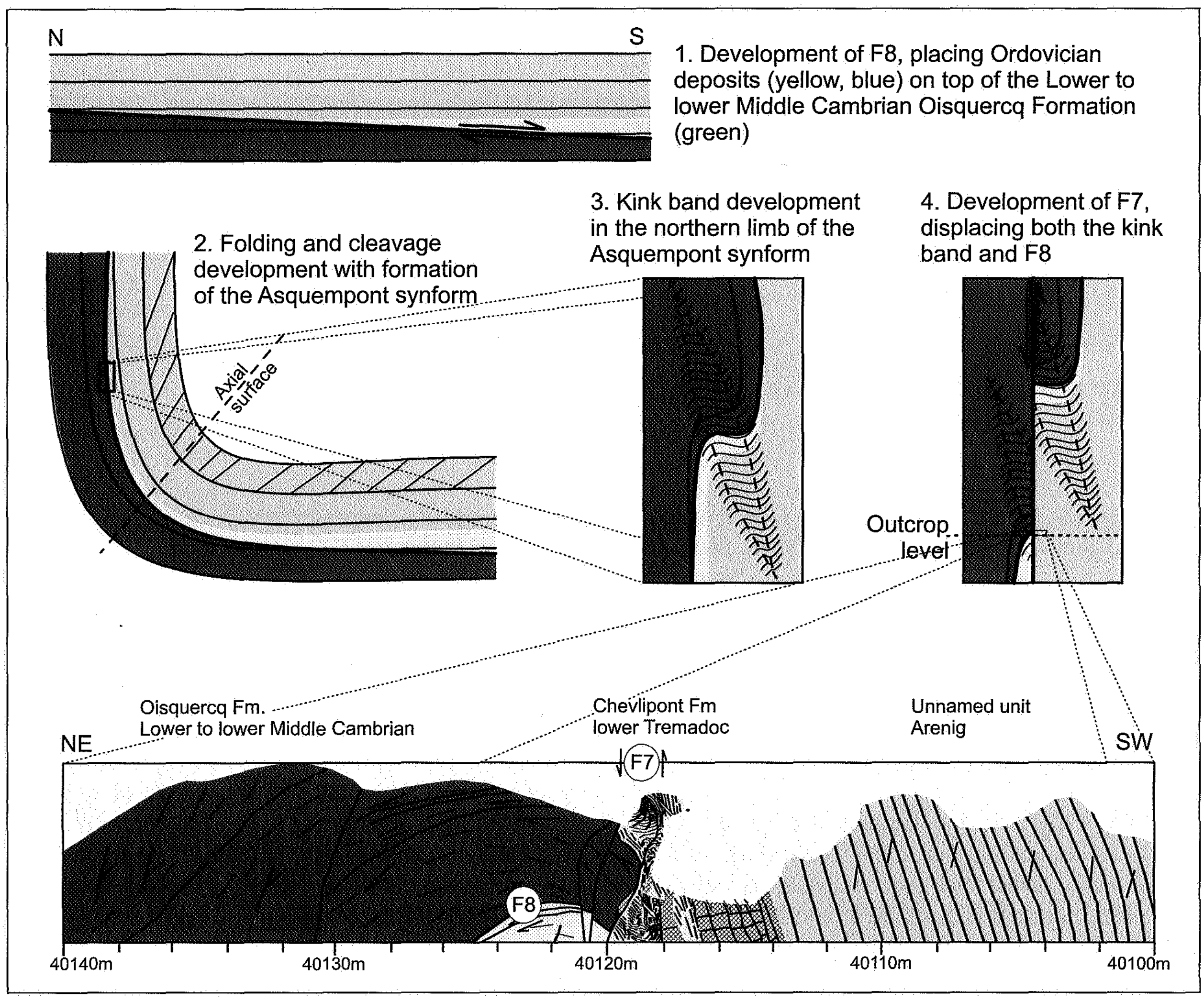

Figure 14: Schematic representation of the consecutive development of 1) the Asquempont fault as defined in this study (F8), 2) folding and cleavage (e.g. the Asquempont synform, Debacker *et al.*, 2001), 3) the large-scale kink band and 4) the steep normal fault F7, thus offering an explanation for the structural complexity of the classical outcrop of the Asquempont fault (Debacker, 2001; Debacker *et al.*, submitted).

Comparison with boreholes and outcrops situated to the NW of the Asquempont-Virginal area suggests that this fault is a pre-cleavage and pre-folding low-angle extensional detachment (fig. 14; Debacker, 2001; Debacker *et al.*, submitted; see 4.4).

The sense of movement along F7 is difficult to determine, due to the fact that both the fault and the bedding are steeply dipping. In the case that the post-cleavage overturning indeed represents the northern part of the large-scale kink band, then F7 is a steeply NE-dipping to sub-vertical normal fault with a down-throw towards the north (fig. 14). Alternatively, if this post-cleavage overturning is a drag phenomenon, then F7 should be considered as a sub-vertical to steeply SW-dipping normal fault, with a down-throw towards the south. Both options are possible, judging from the observations in this outcrop. It is only the large-scale repetition of the Ittre, Rigenée, Tribotte and Abbaye de Villers Formations, on average steeply SW-dipping, that shows that it is probably the first possibility (see 4.3). On the basis of this large-scale repetition, displacement along F7 (down-throw to the N) can be estimated between 700 and 1000 m (Debacker, 2001; Debacker *et al.*, submitted).

3.5. THE SOUTHERN ASQUEMPONT SECTION

Approximately 520 m long series of outcrops along the E-side of the new (post-1962) Brussels-Charleroi canal, between 39.767 and 39.246 km (expressed in canal kilometres; distances measured from km-post 39 onwards along E-side of canal), consisting of 3 main outcrops (A: 39.767-39.750 km; B: 39.550 -39.550 km; C: 39.317-39.246 km), and 18 small, temporary outcrops, newly created in the densely forested talus between the three main outcrops (fig. 6, fig. 15). Facing outcrop B, an-

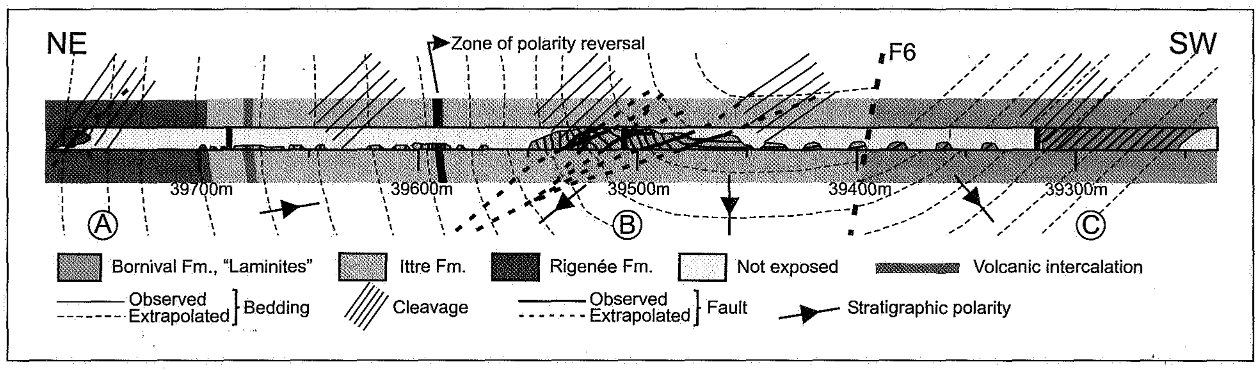

Figure 15: Section along the Southern Asquempont section, showing a large south-verging synform with overturned bedding (Debacker *et al.* 2001).

other large outcrop (B') is present along the W-side of the canal (between 40.280 and 40.410 km; distances measured from km-post 40 onwards along W-side of canal). To the north of the latter outcrop, 11 small new, temporary outcrops were created in the forested talus to allow correlation with the deposits along the E-side of the canal (between outcrops A and B) (fig. 6).

3.5.1. Lithology/sedimentology

From N to S (fig. 15): Homogeneous dark mudstones of the Rigenée Formation (outcrop A; weathered; bedding only observed on cut hand specimens), distal turbidites of the Ittre Formation (outcrops B and B'; studied by Legros, 1991 and Servais, 1991 in outcrop B), with a several-metre thick fine-grained volcanic intercalation in its lowermost part (cf. Corin, 1963), and deposits of the Bornival Formation, consisting of centimetric siltstone-mudstone alternations, probably of distal turbiditic origin (outcrop C). A more detailed description of these three formations can be found in Verniers *et al.* (2001).

The fine-grained volcanic level in the lowermost part of the Ittre Formation, described by Corin (1963) as tuff, consists of several interstratified layers, having a thickness of ~1 metre in the lower parts and becoming thinner upwards as fine-grained distal turbidites become dominant. These interstratified deposits were also found along the W-side of the canal, and occur at the same stratigraphic level as those in the southern part of the Northern Virginal railway section (stop 3.2). The marked difference in texture (more coarse-grained, with incorporated mudstone fragments in the Northern Virginal railway section; fine-grained, tuff-like here) and the aspect of the deposits in the Northern Virginal railway section, points to a rather viscous magma, and hence a rather proximal source, suggesting active volcanism within (this part of) the Brabant Massif already during the middle Caradoc (cf. Ashgill ignimbrites in Fauquez area, André, 1991; Van Grootel *et al.*, 1998).

An exceptional feature of the deposits of the Bornival formation in outcrop C is the pronounced small-scale soft-sediment deformation (fig. 16). Some of these can be shown to be of biogenic origin (bioturbations), others are of non-biogenic origin (intraformational, recumbent folds; disrupted silty beds with vague, irregular limits, anomalous grain size distributions within the silty beds: e.g. vague patches of coarser-grained material; loadcasts –sometimes on the youngest sides of the coarser beds!-, flame structures, ball-and-pillow structures). It should be noted that in many cases the biogenic and non-biogenic soft-sediment deformation features are difficult to distinguish (e.g. a cross-section will generally not allow distinguishing a burrow from a ball-and-pillow structure). These non-biogenic soft-sediment deformation features do not occur in the younger members of the Bornival Formation present in the Fauquez area to the south (Verniers *et al.*, 2001).

3.5.2. Stratigraphic polarity

The stratigraphic polarity within the turbidites of the Ittre Formation (fig. 16) can be determined by means of truncations of fine-grained turbidite intervals by younger coarser-grained intervals, graded bedding, cross-bedding, turbidite sequences (usually Tc(d)e; cf. Servais, 1991), loadcasts (sometimes risky), etc... (Debacker *et al.*, 2001). As previously recognised by Legrand (1967), Legros (1991) and Servais (1991), in the large-scale synform in outcrops B and B' these turbidites are overturned: younging towards below in the synform hinge (fig. 17) and towards the north in the northern limb. However, on a larger scale the latter limb, of which also the Central Asquempont section forms part, youngs towards the south (from N to S: Oisquercq Formation, Chevlipont Formation, Abbaye de Villers Formation, Tribotte Formation, Rigenée Formation, Ittre Formation; cf. Martin & Rickards, 1979), indicating that somewhere within the Ittre Formation in the forested talus the stratigraphic polarity has to

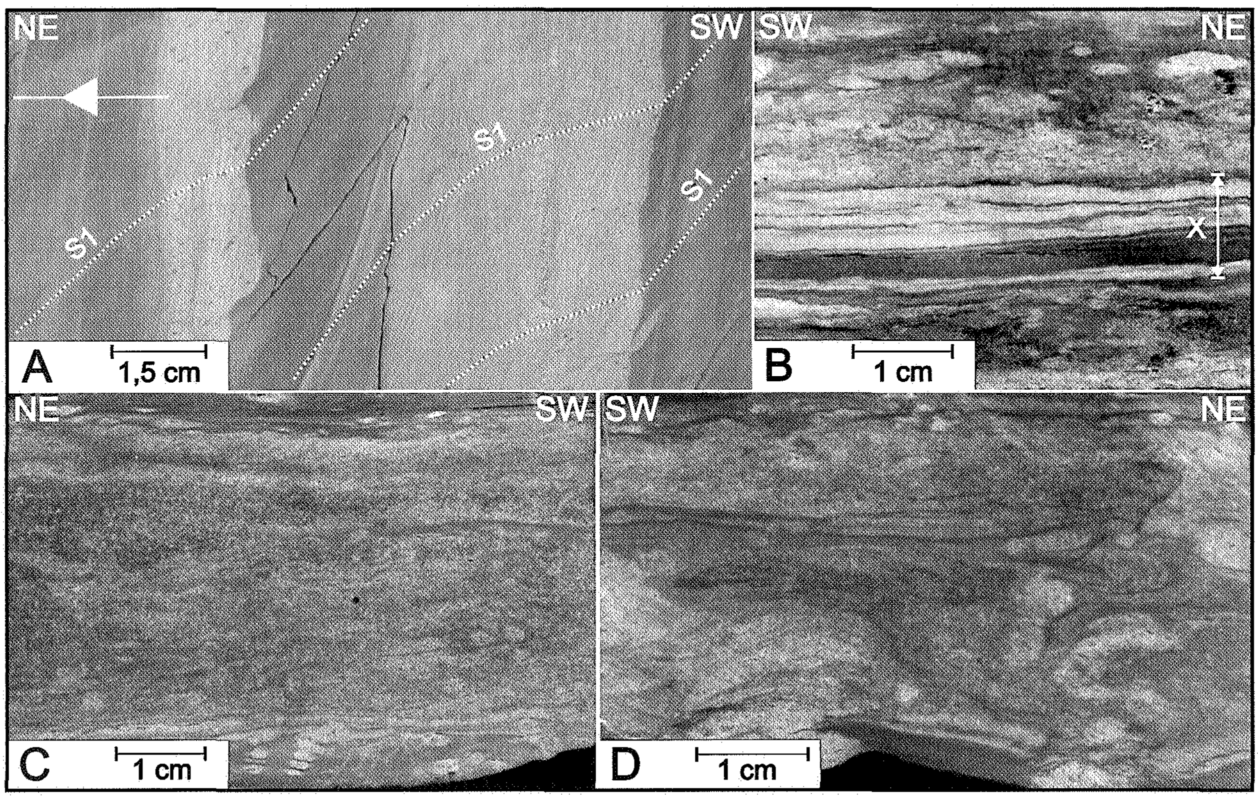

Figure 16: Small-scale sedimentological observations in the overturned beds of outcrops B and C (Debacker *et al.*, 2001). A) Relatively undisturbed turbidites of the Ittre Formation in outcrop B, younging towards the NE (shown by arrow) as indicated by graded bedding, small load casts, sharp contacts and turbidite sequences (Tcde). Also the cleavage refraction pattern (S1) can be used to determine the younging direction (39549.5 m). Photos B to D show soft-sediment deformation of biogenic (B, C) and non-biogenic (D) origin in outcrop C. B) Relatively undisturbed beds (X) with a reverse stratigraphic polarity (younging towards below) between partly to completely disrupted beds (39247 m). C) Silt to fine sandstone showing internal disruption of layers into apparently lenticular bodies (2D!), best visible along the edges of the bed (39286 m). D) Silty bed with internal remobilisation of fine silt to fine sand forming irregular bodies of coarser grains (pale-coloured) between zones of finer material (dark coloured); note the small recumbent folds (39257m).

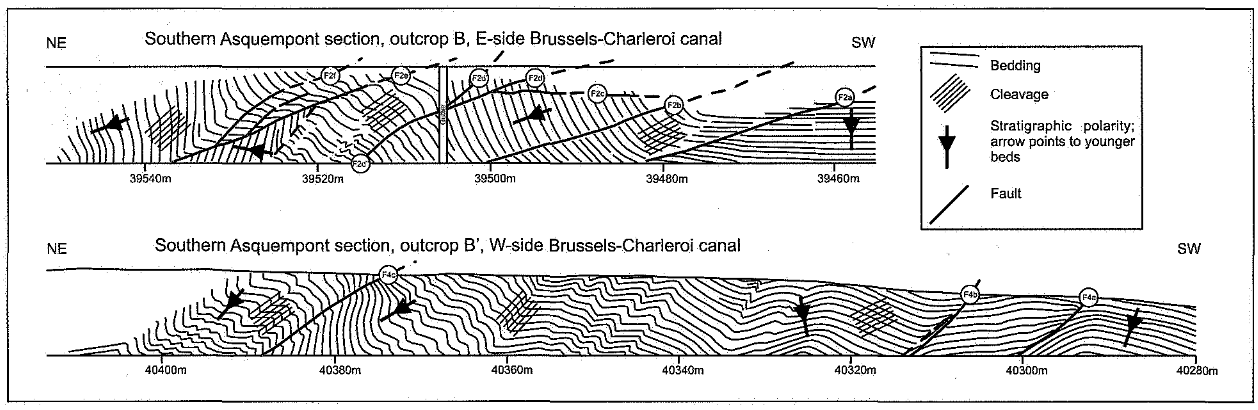

Figure 17: Outcrops B and B' along the Southern Asquempont section (Debacker, 2001), both containing the hinge zone of the Asquempont synform. For convenience, both outcrops are represented as if one was looking to the south-east; outcrop B' is thus mirrored with respect to the actual outcrop situation. The outcrops entirely consist of overturned turbidites of the Ittre Formation (cf. Servais, 1991; arrows point to younger beds). In both outcrops, the more energetic (Tcde) turbidites, with c-interval thicknesses often exceeding 5 cm, are situated in the northern parts, whereas the less energetic, more homogeneous and more fine-grained turbidites are found in the southern parts.

change (from S-ward younging in the north, towards N-ward younging in the south). This polarity reversal (found by means of digging out small outcrops in the forested talus) occurs between 39.589 km and 39.593 km, within the turbidites of the Ittre Formation in the uniformly, steeply dipping limb of the synform, just above (to the south of) some thick, sandy turbidites, that appear to be deposited under highly energetic conditions and are a-typical of the Ittre Formation (cf. Verniers *et al.*, 2001).
Also the deposits of the Bornival Formation in outcrop C, situated in the southern limb of the synform, are overturned (Debacker *et al.*, 2001; Hennebert & Eggermont, 2002), although there, because of the finer grain size, the finer laminations and the soft-sediment deformation, polarity is very difficult to determine in outcrop and should preferably be done on oriented cut hand specimens.

3.5.3. Structural features

Faults

In several localities along the Southern Asquempont section bedding- and cleavage-parallel slip zones are encountered, with a direction of slip sub-perpendicular to the trend of the fold hinge lines. In general, the sense of slip could not be determined.
The synform hinge zone (outcrops B and B') is cut by a set of gently to moderately NE-dipping faults, post-dating cleavage development (fig. 15, fig. 17). The faults generally have a random cohesive fabric, in some cases a crush breccia or protocataclasite, in other cases a cemented breccia. Striations indicate a displacement vector sub-perpendicular to the trend of the fold hinge lines. Because of the lack of marker horizons and the low angle to cleavage, the sense of movement could generally not be determined. Locally, small, moderately to steeply NE-dipping reverse faults are encountered in outcrop B, and small-scale observations in the matrix of some of the faults suggest a reverse movement, whereas small-scale observations in adjacent faults suggest a normal movement. Because there is no marked jump in stratigraphy across these faults, they are considered to have relatively small displacements. The same probably accounts for a moderately NE-dipping post-cleavage fault observed in outcrop A.
A zone of pronounced post-cleavage deformation occurs around 39400 m and apparently coincides with the limit between the Ittre Formation and the Bornival Formation. Probably this deformation and the juxtaposition of both lithological units is fault-related (F6).

Cleavage/fold relationship

The Southern Asquempont section essentially consists of a gently E-plunging, moderately N-dipping, open to close hectometre-scale synform, with a step fold geometry and a S-verging asymmetry (fig. 15). The hinge of this synform is situated within outcrops B and B' (fig. 17). Outcrop C, consisting of uniformly, moderately N-dipping deposits of the Bornival Formation, represents the southern synform limb, whereas the northernmost part of outcrop B, outcrop A and the forested talus between both outcrops represents the northern synform limb. The hinge of this synform contains numerous open to close, sub-horizontal to gently plunging metre-scale folds with a step fold geometry and a S-verging asymmetry (best developed in outcrop B') (fig. 17, fig. 18).
A well-developed cleavage occurs throughout the section. Apart from a small angle axial cleavage transection, the cleavage is approximately axial planar to the large-scale synform, showing a small-angle divergent cleavage fanning. This cleavage is also axial planar to the small folds in the synform hinge zone, showing a divergent fanning in the fine-grained beds, convergent fanning in the coarser-grained beds, with a symmetrical disposition with respect to the axial surfaces, and different senses of cleavage refraction on both fold limbs (fig. 18). Hence, the cleavage/fold relationships point to a cogenetic relationship between cleavage development and folding. Note that, although both the small folds in the synform hinge zone and the large-scale synform appear to be tectonic folds, a remarkably small angle occurs between cleavage and bedding in the southern synform limb (outcrop C; cleavage only a few degrees steeper than bedding) (fig. 19). The explanation for this lies in the geometry of the overturned sedimentary pile (see 3.5.5.).

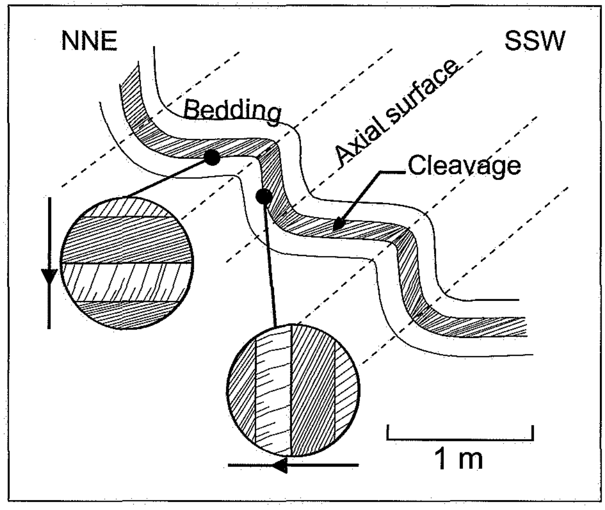

Figure 18: Cleavage/fold relationships in small asymmetric step folds in the fold hinge of the Asquempont synform (Debacker *et al.*, 2001). The fold sequence shown is taken from outcrop B', between 40340 and 40350 m (cf. fig. 17). The arrows point towards the younger beds.

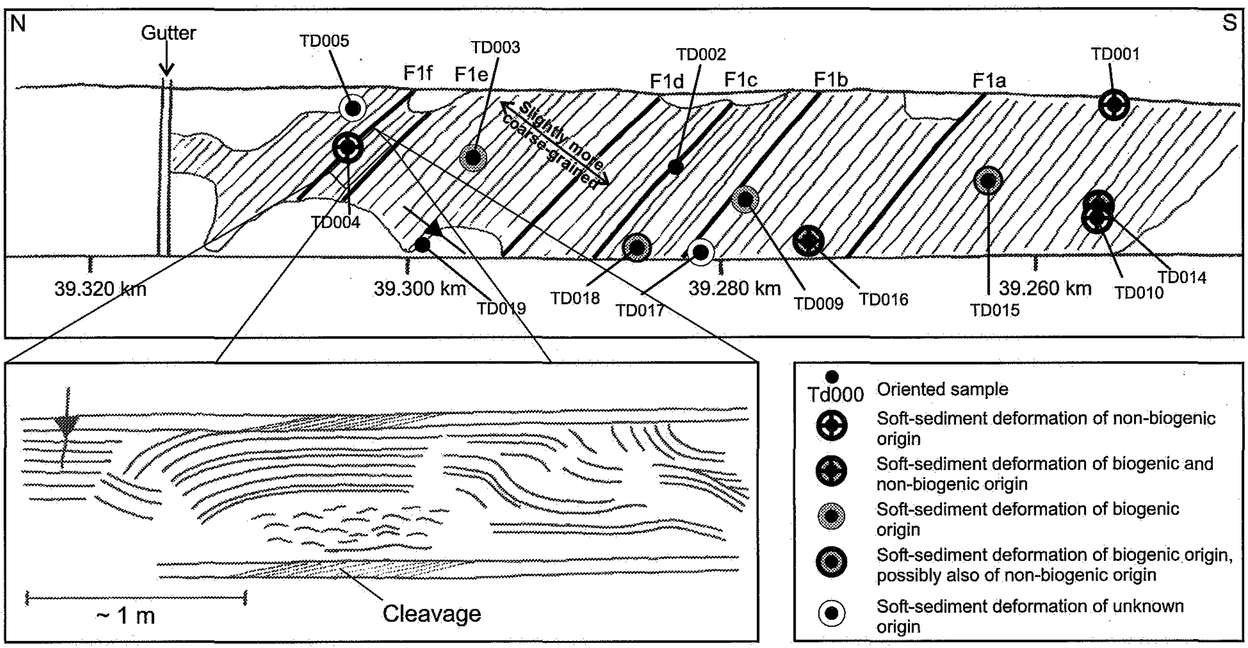

Figure 19: Outcrop C along the Southern Asquempont section (after Debacker, 2001). The deposits, probably of distal turbiditic origin and attributed to the lower member of the Bornival Formation (Verniers *et al.*, 2001), contain numerous levels of small-scale soft-sediment deformation, both of biogenic and non-biogenic origin (see also fig. 16). Also meso-scale slump folds occur (see inset). The beds young towards below (see arrows), and cleavage is at a very low angle to bedding, only dipping a few degrees steeper.

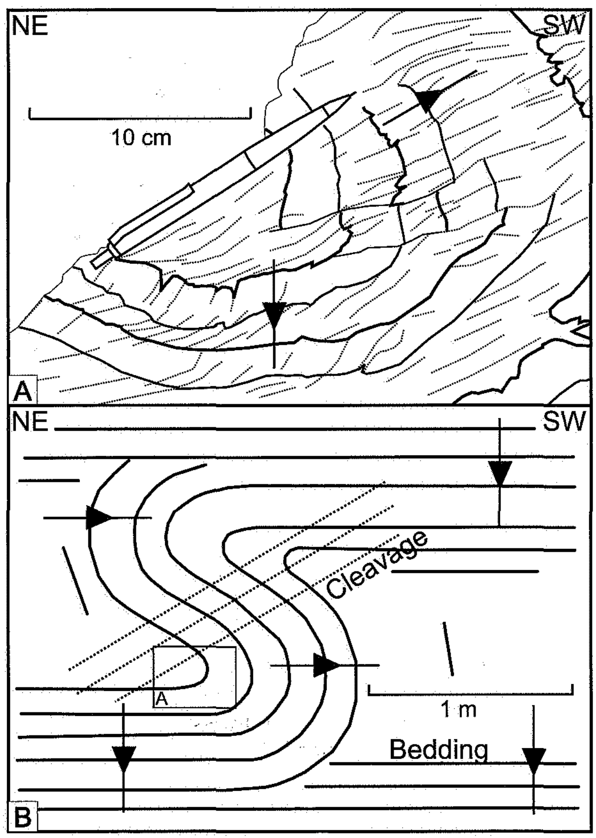

Figure 20: Pre-cleavage deformation interpreted as slump folds (Debacker *et al.*, 2001). A) Small asymmetric fold structure in the turbidites of the Ittre Formation, cut by a tectonic NE-dipping cleavage (39422 m). B) Schematic representation of slump folds between 39420 and 39440 m: an antiform-synform couple, cut by a NE-dipping cleavage, sandwiched between relatively undisturbed turbidites with a reverse stratigraphic polarity. The approximate position of the synform depicted in A is shown in B. The arrows point towards the younger beds.

Apart from these tectonic folds, also pre-cleavage folds are observed, both in the Ittre and the Bornival Formation (fig. 19, fig. 20). The best examples occur in the Ittre Formation (between 39420 and 39440 m) (fig. 20): in a zone of sub-horizontal bedding, approximately 2 metres thick, an intraformational synform-antiform couple occurs with a gently to moderately S-dipping axial surface, being cross-cut by the gently to moderately N-dipping cleavage.

Changes in trend

Within outcrops B, B' and C, the fold hinge lines and cleavage/bedding intersection plunge gently E-ward, with a plunge direction between 100 and 110° (280-290°). However, towards the north from 39.600 km onwards (forested talus), the cleavage/bedding intersection plunges gently to the NW, with a plunge direction between 300 and 320° (120-140°) (fig. 21), which is more compatible with the regional trend, as observed in the Central Asquempont section, the Bief 29 section and the Northern Virginal railway section (compare with fig. 11A). An evaluation of the bedding trend going from S to N along the northern limb of the large-scale synform shows the same image: a clockwise change in trend around 39.600 km. In contrast, neither the cleavage nor the transverse fractures reflect this change in trend (fig. 21).

3.5.4. Structural polarity and the polarity reversal zone

Because of the cogenetic relationship between the large-scale synform and the small-scale folds in its hinge on the one hand and the cleavage on the other hand, the cleavage/bedding relationship can be used to determine the structural polarity, being normal in the limbs where cleavage dips steeper than bedding, and reverse where bedding dips steeper than bedding. In all the tectonic folds and throughout the large-scale synform (outcrops B, B' and C) the structural polarity opposes the stratigraphic polarity. This indicates that overturning occurred before folding and cleavage development, thus questioning the overturned anticline hypothesis of Legrand (1967; see also Hennebert & Eggermont, 2002). This also becomes apparent from a study of the polarity

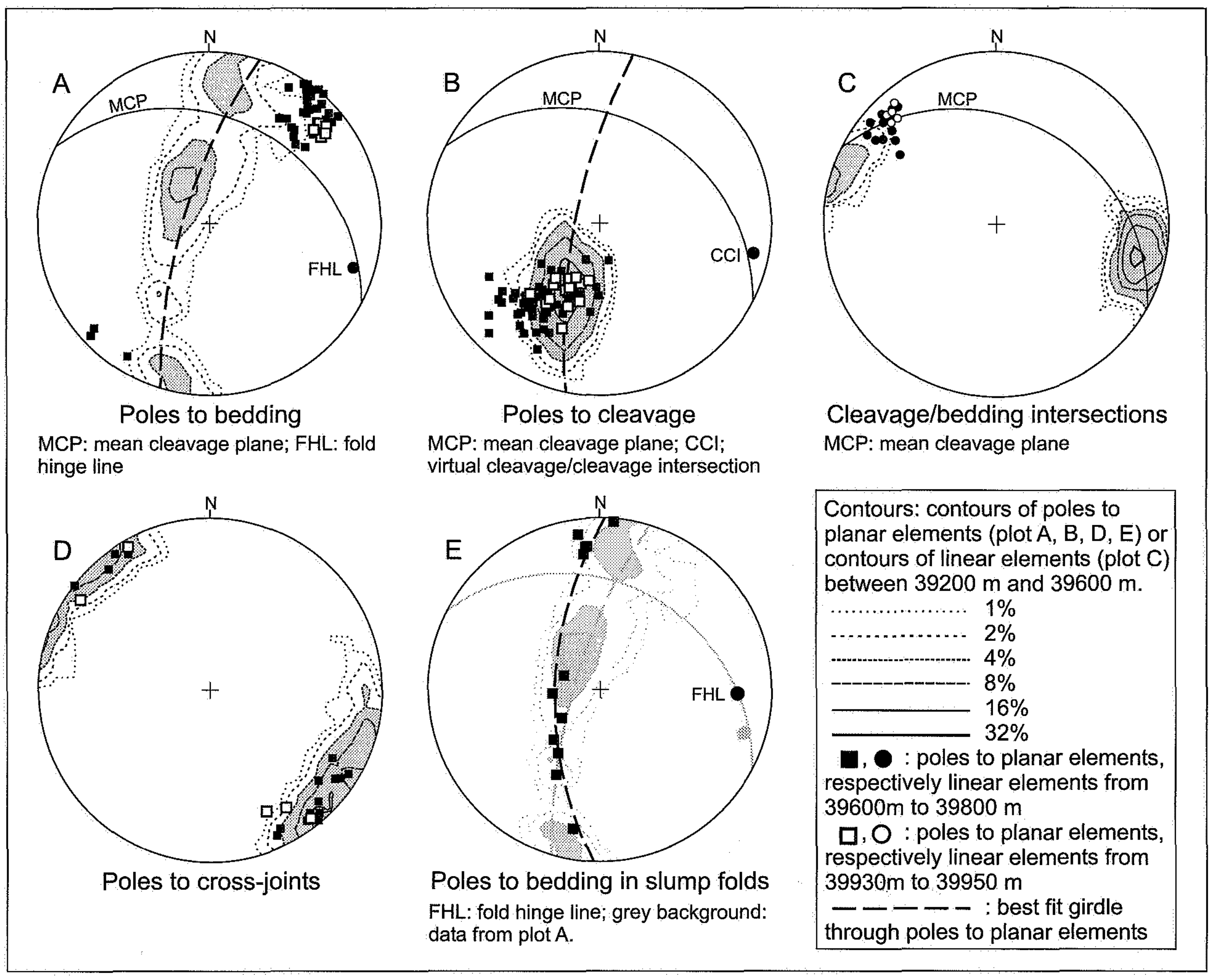

Figure 21: Lower-hemisphere equal-area stereographic projections of poles to bedding (A), poles to cleavage (B), cleavage/bedding intersections (C), poles to transverse fractures (D) and poles to bedding in slump folds (E), taken from Debacker *et al.* (2001). In projections A to D the data points from 39200 m to 39600 m, contoured, are overlain by data points from between 39600 to 39800 m (black circles and squares) and data further north (39930 to 39950 m white circles and squares), in order to check if a change in orientation occurs along the section. A change in orientation between contoured data and overlain data points is apparent from the poles to bedding (A) and the cleavage/bedding intersections (C).

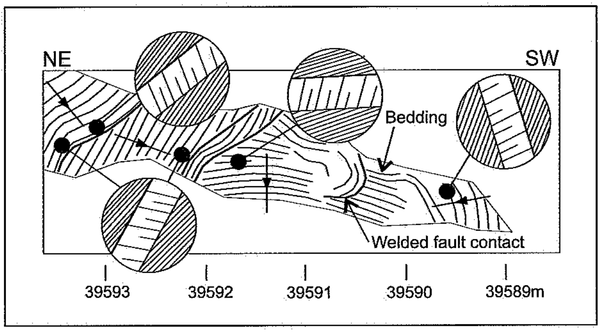

Figure 22: Polarity reversal zone between turbidites younging towards the NE (SW-part) and turbidites younging towards the SW (NE-part) in the steep northern limb of the Asquempont synform (see fig. 15 for location) (Debacker *et al.*, 2001). The cleavage/bedding relationships (circular insets) show that the contact and the overturning predate cleavage development. Truncations and fault contacts are welded. The arrows point towards the younger beds.

reversal zone (fig. 22). Both at the northern side and the southern side of this zone bedding is steeply dipping, with a moderately NE-dipping cleavage, indicative of a structurally reverse limb, and thus expected to be younging towards the south. However, only those beds at the northern side of this zone young towards the south. To the south of this, bedding is strongly disturbed, affected by welded faults and small folds. Both the welded faults and the folds are cross-cut by the cleavage that remains NE-dipping, confirming the pre-cleavage nature of the overturning. By means of the position of the polarity reversal zone, the stratigraphy (Verniers *et al.*, 2001) and the fold geometry, the thickness of the overturned deposits is estimated at minimum 200 metres (Debacker *et al.*, 2001).

3.5.5. Overall structure and interpretation

The Asquempont synform is a hectometre-scale, moderately inclined, sub-horizontal to gently plunging, open to close tectonic synform with a step fold geometry and a S-verging asymmetry, containing overturned deposits of the Bornival Formation in the southern limb, overturned deposits of the Ittre Formation in the hinge and the northern limb, and right-way-up deposits of the Ittre, the Rigenée and older formations in the steep northern limb (fig. 23).
As indicated by the cleavage/fold relationship and the stratigraphic polarity, the deposits of the Ittre Formation to the south of the polarity reversal zone and the Bornival Formation were already overturned prior to folding and cleavage development. Either this occurred during an older, thus far not recognised tectonic phase, or it occurred by large-scale slumping. In the first case, this overturning should have occurred by means of large-scale recumbent folding, possibly accompanied by thrusting. However, during such a deformation one would expect the development of a tectonic fabric, which would afterwards get crenulated by the cleavage fabric related to the observed folds. Such an older fabric is not observed. In addition, if the deposits of the Ittre and Bornival Formation got deformed by a thus far unknown deformation phase, then also older deposits should be affected by it. Also for this there is no evidence. In contrast, there are much more elements supporting the second possibility. 1) Although from the Rigenée Formation to the Ashgill calm conditions appear to have prevailed (except for the Huet Formation, upper Caradoc, possibly a tempestite deposit; Verniers *et al.*, 2001), there is the sudden appearance of turbidites of the Ittre Formation and the lower member of the Bornival Formation. 2) Small-scale slump folds and intraformational breccias are common within these deposits, not only in the section studied but also in other parts of the Brabant Basin (Lessines borehole, Herbosch *et al.*, 1991; northern Condroz Inlier, Valcke, 2001). 3) The youngest sediments of the overturned sedimentary pile (Bornival formation, outcrop C) contain numerous levels of small-scale, non-biogenic soft-sediment deformation, which appear to be caused by recurring triggers (fig. 19, fig. 16) and 4) also the coarse-grained, sandy turbidites at the base of the overturned sedimentary pile, a-typical of the Ittre Formation (cf. Servais, 1991; Verniers *et al.*, 2001) indicate a sudden change in basinal conditions. Hence, we propose large-scale slumping during the middle Caradoc as the cause of the overturned sediments in the Asquempont synform (Debacker *et al.*, 2001).
The difference in trend of the bedding and the cleavage/bedding intersection within the slump sheet, relative to those below the slump sheet, the general stratigraphy (cf. Verniers *et al.*, 2001) and the obliquity between the overturned bedding and the polarity reversal zone, point to a wedge-shaped geometry of the slump sheet (fig. 23). This may explain the low angle between cleavage and bedding in outcrop C (cf. fig. 19).
As a cause of slumping, middle Caradoc seismic activity is put forward. Possibly this seismic activity was caused by movement along the nearby Asquempont fault (F8) (Debacker, 2001; Debacker *et al.*, 2001).

4. MAIN CONCLUSIONS AND IMPLICATIONS

4.1. THE TWO MAIN FAULTS IN THE ASQUEMPONT-VIRGINAL AREA: F8 AND F7

The Asquempont fault (F8 of Debacker, 2001) is redefined as a pre-cleavage and pre-folding, low-angle extensional detachment, forming the limit between the

Oisquercq Formation and the Ordovician in the Senne-Sennette outcrop area. It is exposed in the northern part of the Northern Virginal railway section, in one outcrop in the "Rue de l'Ancien Canal" along the NE-side of the Bief 29 (private property), and in the Central Asquempont section. The fault originally described by Legrand (1967) in the Central Asquempont section as the Asquempont fault is a post-cleavage normal fault (F7 of Debacker, 2001). It is this fault that we hold responsible for the repetition of the Ordovician formations to the west of the Central Asquempont section. On the basis of this repetition, an estimated displacement along F7 (down-throw to the N) of 700 to 1000 m is proposed (Debacker, 2001; Debacker *et al.*, submitted). As observed in outcrop, F7 (post-cleavage) truncates F8 (pre-cleavage). Hence, the field evidence contradicts the geological map of Hennebert & Eggermont (2002) on which, possibly influenced by Legrand (1967), F8 truncates F7.

4.2. THE ASQUEMPONT SYNFORM

As indicated by the cleavage/fold relationships, the beds in the Asquempont synform were overturned before folding and cleavage development. Hence, the Asquempont synform should not be considered as an overturned anticline (cf. Legrand, 1967; Hennebert & Eggermont, 2002). We attribute this pre-folding and pre-cleavage overturning to a large-scale slump during the middle Caradoc, probably caused by seismic activity (Debacker *et al.*, 2001). Possibly this seismic activity was generated by movement along the Asquempont fault (F8).

4.3. GEOLOGICAL MAP

Fig. 24 shows a geological map of the Asquempont-Virginal area that is more compatible with the structural observations than the new official geological map of the area (Hennebert & Eggermont, 2002). The map is based on the new structural observations and mapping results presented above and on the most recent litho- and biostratigraphic data (e.g. Verniers *et al.*, 2001). In the unexposed central parts of the map, the traces of the limits between the different formations are constrained locally by observations in former outcrops that have disappeared as a result of the construction of the new canal (i.e. pre-1962: Dumont, 1848; Malaise, 1873, 1908; Fourmarier, 1914, 1921, Herssens, 1957). Faults are kept to a strict minimum, in order to keep the map as simple as possible.
Considering the E-ward plunge of the Asquempont synform, one would expect, to the NW of the southern Asquempont section (outcrop 3.5), the synform closure with overturned deposits of the Bornival Formation (cf. fig. 23). However, because of several factors, we feel that there are not enough constraints to represent this on the map. These factors are: the difficulty of determining the exact depth to the polarity reversal zone within the synform hinge (fig. 23), the fact that although the overturned deposits plunge to the east (E-ward plunging folds and cleavage/bedding intersection), on a larger scale the synform probably has a W-ward plunge (consistently W-ward plunging cleavage/bedding intersection in the northern limb of the Asquempont synform to the north

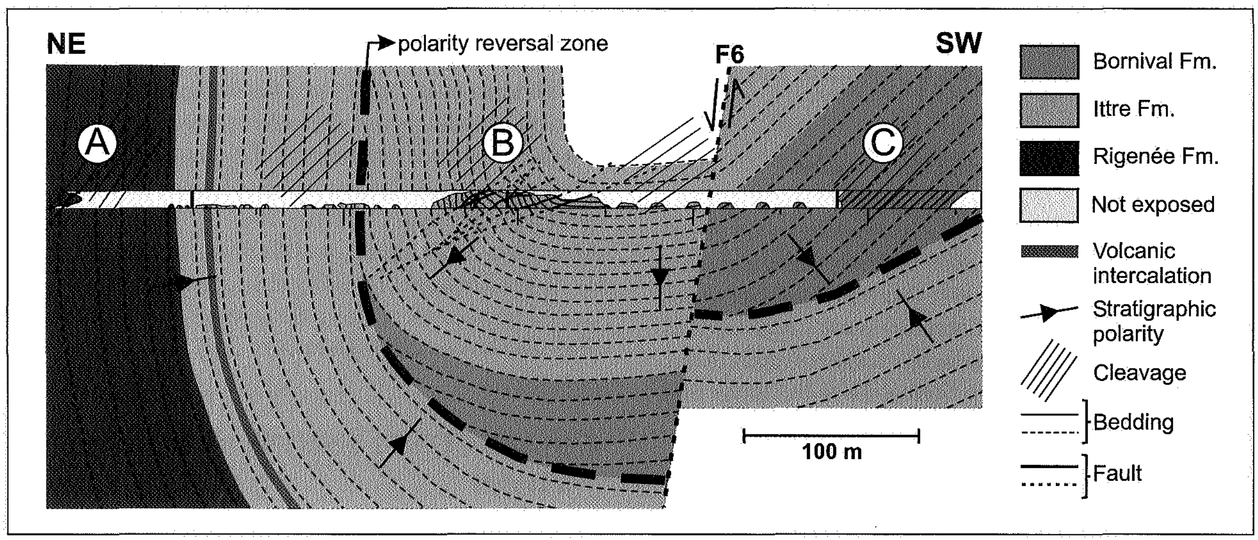

Figure 23: Inferred geometry of the overturned, slumped sequence within the Asquempont synform (Debacker *et al.*, 2001). A combination of the stratigraphic, structural and sedimentological data implies a wedge-shaped geometry. Note that the exact displacement along the fault (F6) is unknown.

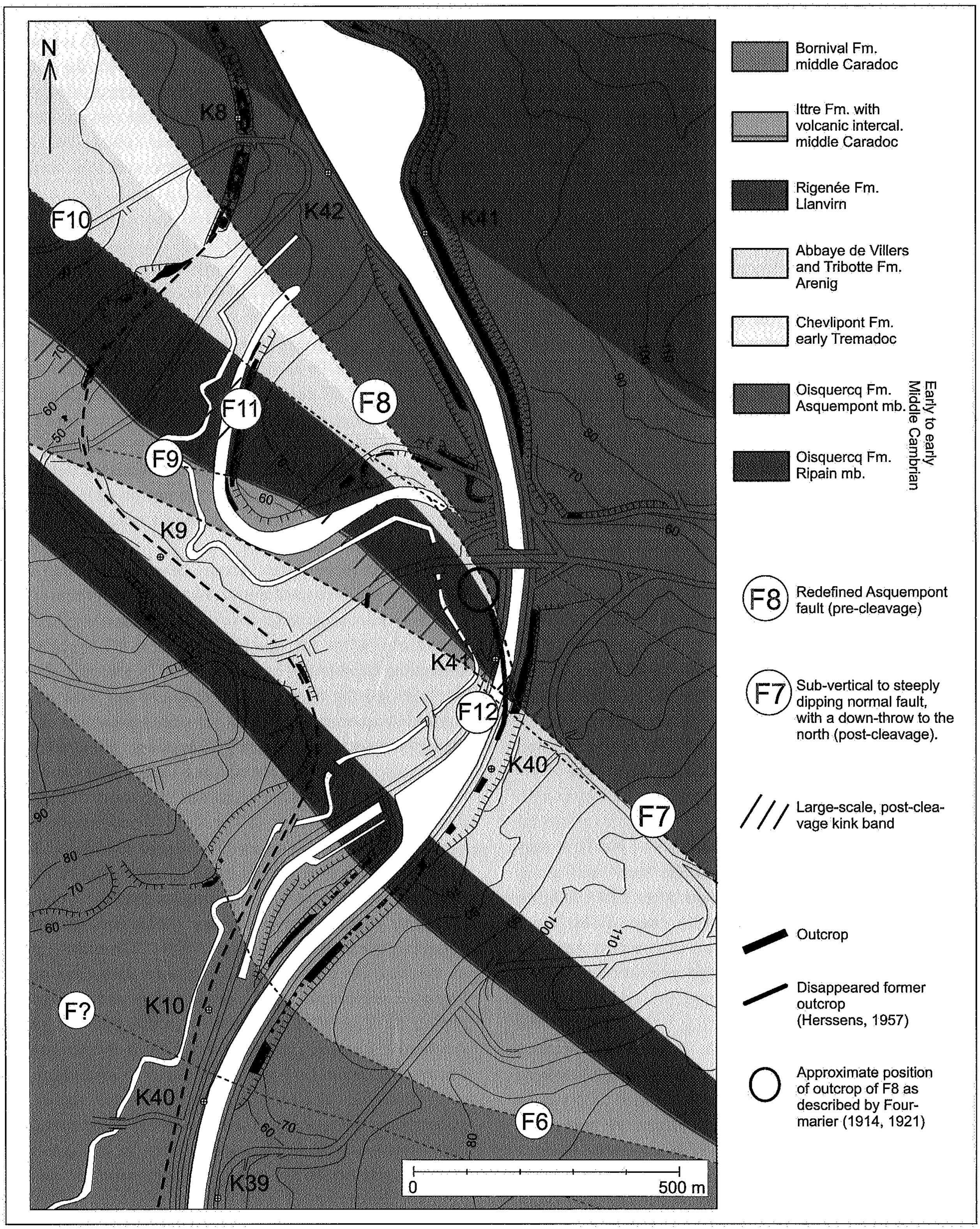

Figure 24: Geological map of the Asquempont area (Debacker, 2001; Debacker *et al.*, submitted). In order to further constrain the likely traces of F7 and F8, also observations were used from former outcrops, now disappeared (Herssens, 1957; Fourmarier, 1914, 1921). The repetition of the Ordovician beds is attributed to F7, whereas the Cambrian-Ordovician contact is formed by F8. Note the exceptional position of the classical outcrop of the Asquempont fault, situated at the junction of three large, apparently unrelated structural features: F8, F7 and the large-scale kink band. Together with F6, F10, F11 and F12, F7 forms part of a large set of normal faults in the Asquempont-Virginal-Fauquez area (Fauquez: ~ 2 km to the south of Asquempont; cf. Van Grootel *et al.*, 2002).

of the polarity reversal zone; see Debacker *et al.*, 2001 and Debacker, 2001), the wedge-shaped nature of the overturned sedimentary pile, the likely along-strike variation in plunge (irregularities due to slumping, possibly in combination with periclinal fold nature), and the absence of data from the westernmost part of the map. Similarly, in the western part of the map, to the south of F6, we would expect the re-occurrence of the Ittre Formation below the polarity reversal zone, but again the absence of data from the western part of the map, in combination with changes in plunge and differences in plunge below and above the polarity reversal zone do not allow to represent this. Also to the south of the Southern Asquempont section (outcrop 3.5), the Ittre Formation should reach the surface, judging from fig. 23. However, the Ittre Formation is not encountered in outcrops to the south (Fauquez area; see Debacker, 2001; Van Grootel *et al.*, 1998, 2002). Instead, the few outcrops further south all show deposits of the Bornival Formation, younging upwards. Loose, deformed rock fragments of the Bornival Formation in the canal slope just south of the Southern Asquempont section, at 39.220 km, in combination with a constant local water flow are compatible with the presence of a fault. This, hypothetical, fault (F? on fig. 24) should have a down-throw to the south.

The Asquempont fault (F8) should be affected by the post-cleavage normal fault F10. The trace of the latter is extrapolated towards the east from the eastern part of the Bief 29 onwards. Since both faults are steep, and F10 has a dip-slip displacement, the map trace of F8 should not be displaced much. The trace of F8 just south of F10 can be further constrained by observations of Fourmarier (1914, 1921) and Herssens (1957) in former, disappeared outcrops. Close to the "new" lock along the old canal (a lock which was built after 1907, judging from its absence on the topographic map of 1907, and largely destroyed by construction of the new canal in 1962), just to the south of the present bridge of Asquempont (fig. 24), Fourmarier (1914, 1921) observed the contact between the Oisquercq Formation and the Chevlipont Formation or Abbaye de Villers Formation, a contact with a similar appearance as that in the Northern Virginal railway section (Fourmarier, 1914: "*Près de la nouvelle écluse d'Hasquempont, un peu au sud de la route de Lillois, on peut observer le contact des deux mêmes séries, mais les couches sont ici un peu plus disloquées*"; Fourmarier, 1921: "*Nous retrouvons ces mêmes phyllades verts compacts au bord du canal, à la nouvelle écluse d'Hasquempont. On y voit donc le contact du phyllade vert et du quartzophyllade de Villers, et il nous a paru que, comme le long du chemin de fer, le quartzophyllade est intimement uni à la roche sous-jacente. Nous n'avons pas vu les filons de quartz renseignés par Dumont, ni les filons couchés d'eurite schistoïde blanche qu'il signale en cet endroit dans les couches qu'il nomme coblenciennes*". Herssens (1957) is one of the only researchers presenting a map with the actual observation points. His map shows an outcrop along the E-side of the old canal, just south of the outcrop at the "new" lock described by Fourmarier (1914, 1921) in which only Ordovician rocks are observed, thus further constraining the trace of F8.

As can be observed in the Central Asquempont section (outcrop 3.4), F7 cross-cuts F8. Taking into account the dip-slip down-throw to the north along F7 in combination with the steep ENE-dip of F8 at the N-side of F7, F8 should re-appear south of F7 somewhere to the east of the Central Asquempont section (sinistral apparent displacement of F8 along F7). The position of the trace of F8 south of F7 depends on the amount and direction of displacement along F7 and on the orientation of F8 and F7. Considering the uncertainties regarding the amount of displacement along F7 (estimated between 700 and 1000 m) and the exact orientation of F7 and F8, this position is very difficult to determine, and therefore not depicted on the map.

An apparent discrepancy occurs between the map and the observations in the Central Asquempont section (outcrop 3.4): although the Chevlipont Formation is observed in outcrop below F8, and the breccia in F7 is dated as Tremadoc (acritarchs by Lenoir, 1987), the map shows the Ittre and the Rigenée Formation to the west of the Central Asquempont section (E-ward extension of the deposits observed in the Bief 29 section). However, observations of previous authors are compatible with our map. Along the old canal, in an outcrop south of the one described by Fourmarier (1914, 1921; disappeared at the time of his study), Dumont (1848) and Malaise (1873, 1908) describe a deformed zone of dark grey to black rocks (i.e. Ordovician), containing magmatic levels, adjacent to greenish or pale grey rocks (i.e. probably Oisquerq Formation) (Dumont, 1848: "*On trouve, près de l'écluse de Voiricher, du phyllade irrégulier noir-bleuatre, subluisant, qui s'appuie sur un petit massif en presqu'ile de phyllade gedinien compacte, d'un gris pale, semblable à celui qu'on observe entre Beaurieux et Court-st-Etienne. La limite entre les deux phyllades est assez tranchée, et l'on voit dans le dernier une veine quartzeuse de quelques centimetres, brusquement interrompue par le phyllade coblenzien, ce qui annonce quelque glissement ou quelque disparition de roches entre les deux systemes. Le phyllade coblenzien renferme, près de sa limite, deux filons couchés d'eurite schistoide blanche de plusieurs décimetres d'épaisseur*"; Malaise, 1873 : "*On observe à Asquempont des phyllades quartzifères et aimantifères, verdatres. Il nous a également paru qu'il y avait contact anormal entre ces deux systèmes de roches et qu'elles étaient limitées par une faille. Nous avons rencontré, à leur contact, des frag-*

ments d'une roche prophyrique, qui, très probablement, n'est pas étrangère à ce dérangement"; Malaise, 1908 : "*A Hasquimpont (Ittre), nous arrivons à des roches très tourmentés à stratification confuse: schistes quartzeux de couleur noiratre et verdatre. Dumont les avait considerée comme étant de meme age que celle de Tubize et formant donc une voute. Certaines contournements simulent des voutes, mais on ne trouve pas la meme formation au nord et au sud. Il doit y avoir de petites failles indiquées par des filonnets de quartz et d'eurite*").
To our knowledge, the only dm-thick magmatic beds in this area are situated in the lower part of the Caradoc Ittre Formation (see Debacker *et al.*, 2001; Verniers *et al.*, 2001; cf. Corin, 1963). This is compatible with Malaise (1908), who, on the basis of *Primitia simplex*, considered the black, Ordovician rocks as "Llandeilo", now upper Llanvirn-lower Caradoc. Hence, these observations clearly support a local, fault-related proximity of the Ittre and/or Rigenée Formation to the Oisquercq Formation. Unfortunately, although compatible with our map, these observations cannot be used on their own because of the absence of a good outcrop location. Herssens (1957), however, shows the exact position of a long outcrop along the E-side of the old canal, in which he describes, from N to S, rocks that likely correspond to the Rigenée Formation, separated by a fault zone from rocks that resemble the Abbaye de Villers Formation ("*A la nouvelle écluse de Hasquempont on voit au Nord des phyllades bleu-noir très compacts; plus vers le Sud, ils passent progressivement au gris-bleu pendant une centaine de metres; vers la fin de ces schistes gris-bleu, on voit un petit banc de brèche tectonique. Ceci fait supposer un nouveau dérangement dont le rejet ici ne semble pas être très grand, puisqu'il met en présence 2 roches identiques. Plus loin il y a un petit hyatus, au Sud duquel on voit des quartzophyllades noirs très phylladeux. On détecte dans ces quartzophyllades de nombreux plis. On voit également un nouveau banc de brèche qui semble assez peu incliné vers le Nord. Au Sud, et en-dessous de cette brèche on voit des quartzophyllades inclinés au Nord et des phyllades noirs.*"). In addition, he points out another former outcrop, situated approximately 250 m to the west, in which he describes rocks that correspond to the Ittre Formation ("*L'affleurement 57 nous montre des phyllades noirs à bancs de quartzites.*"). Hence, his observations seem to confirm the presence of the Ittre and Rigenée Formation directly to the west of the Central Asquempont section, support the large-scale repetition of the Abbaye de Villers, Tribotte, Rigenée and Ittre Formation and can be used to constrain the trace of F7. The apparent discrepancy (Chevlipont Formation in outcrop versus Ittre and Rigenée Formation on map) may be explained by arguing that along the Central Asquempont section the outcropping four square metres of the Chevlipont Formation below F8, do in fact represent a lens of Tremadoc rocks trapped in F8. Considering the large width of F8 in the Northern Virginal railway section, the presence of such a lens is possible.
In addition, in contrast to what is depicted in figs. 12 and 14, the dating of the fault rock of F7 as Tremadoc (Lenoir, 1987) does not imply that F7 only contains Tremadoc rocks. Instead, it is more likely that the fault rock of F7 consists of both hanging wall and footwall rocks (i.e. Oisquercq Formation and Lower and Middle Ordovician rocks), but that in the samples investigated by Lenoir (1987) only Tremadocian acritarchs were recognised.

4.4. LARGE-SCALE IMPLICATIONS OF THE OBSERVATIONS IN THE ASQUEMPONT-VIRGINAL AREA AFTER CORRELATION WITH DATA FROM OTHER OUTCROPS, BOREHOLES AND/OR FROM OTHER SOURCES (E.G. GEOPHYSICAL DATA)

The Asquempont fault (F8) also crops out at Quenast and a similar pre-cleavage contact has been observed in boreholes at Lessines (113E1015: Chevlipont Formation on top of Oisquercq Formation; cf. Herbosch *et al.*, 1991) and Bever (114W93: Caradoc on top of Oisquercq Formation; cf. Debacker, 1998, Verniers, unpub. data). Also the occurrence of Upper Ordovician deposits directly on top of the Oisquercq Formation in the boreholes at Schendelbeke (100W181), Tielt (68E169) and Waregem (84W1382) implies an anomalous contact which may be caused by a pre-cleavage low-angle extensional detachment (De Vos, pers. comm. 1999-2003). As can be seen on fig. 25 the resulting trace, approximately coinciding with the trace of the Virginal fault of Mortelmans (1955), does not coincide with the trace of the aeromagnetic Asquempont lineament (cf. Sintubin & Everaerts, 2002). In fact, because of the pre-folding nature, the low cut-off angle and the very low difference in magnetic susceptibility (De Vos *et al.*, 1992) between the hanging wall and the footwall rocks, the Asquempont fault is not likely to show up as a marked aeromagnetic lineament.
The youngest deposits affected by the Asquempont fault belong to the Ittre Formation (e.g. borehole at Bever; dated by means of Chitinozoa; Verniers, unpub. data). Hence, its activity can be constrained between the middle Caradoc and the timing of cleavage development. Taking into account this time-gap and its position, it is possible that movement along this low-angle extensional detachment generated the seismic activity which caused the large-scale slumping during the middle Caradoc.
On the geological subcrop map of De Vos *et al.* (1993), the redefined Asquempont fault (F8) corresponds to a contact resembling a very low-angle unconformity or

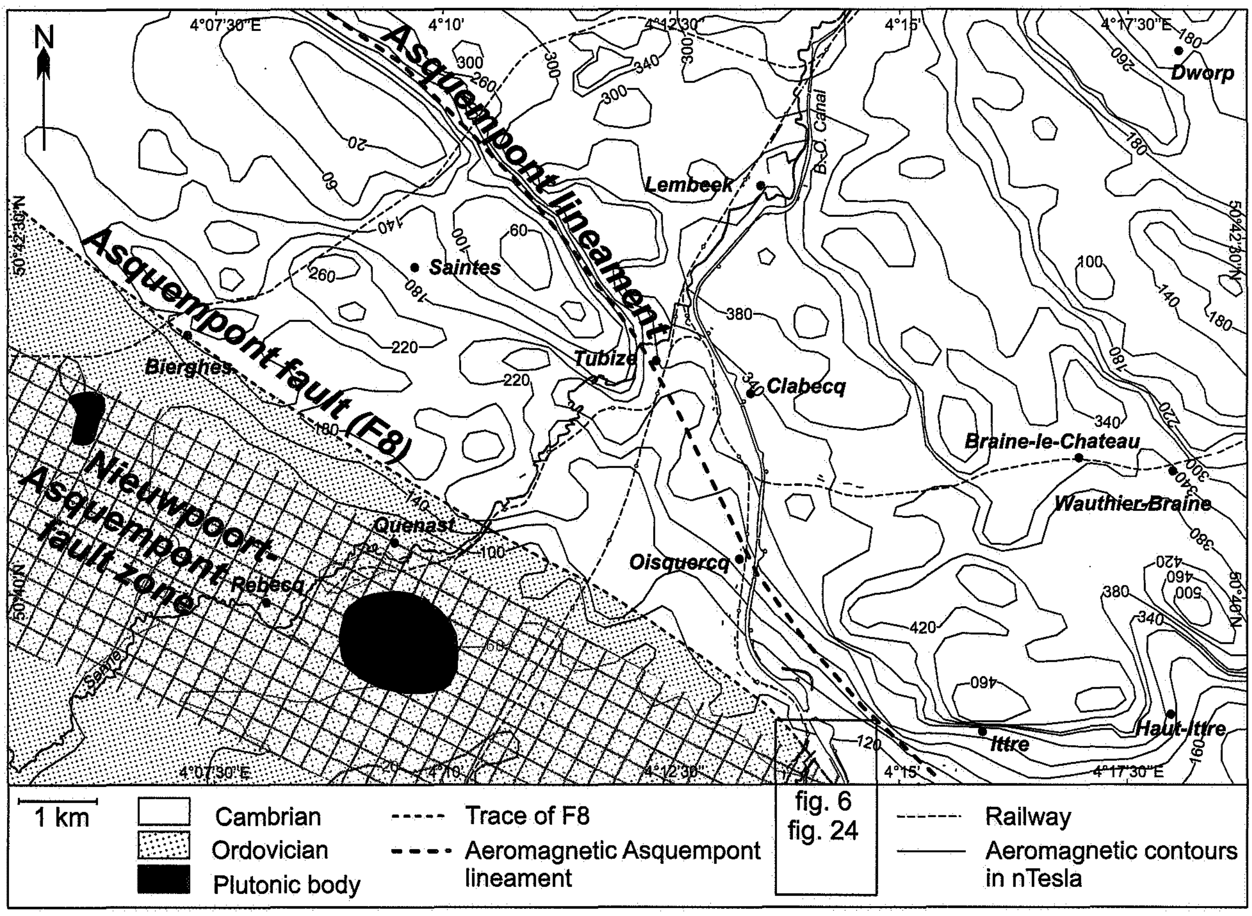

Figure 25: Simplified topographic map of the Rebecq-Ittre area, showing the aeromagnetic contours (taken from B.G.S., 1994), the trace of the aeromagnetic Asquempont lineament, the probable trace of the Asquempont fault (F8) between the Asquempont area to the SE (cf. fig. 24) and the Quenast area to the NW, and the several kilometres wide zone of normal faults between Bierghes and Asquempont, apparently forming the southeastern extension of the Oudenaarde-Bierghes fault zone, which was incorporated in the Nieuwpoort-Asquempont fault zone by De Vos *et al.* (1993; cf. Sterpin & De Vos, 1996). The trace of the Asquempont fault to the NW of Quenast is based on the distribution of Cambrian and Ordovician deposits depicted on the geological map of De Vos *et al.* (1993). Note the discrepancy in orientation and position between the Asquempont fault and the Asquempont lineament. After Debacker (2001; Debacker *et al.*, submitted).

para-conformity between the Oisquercq Formation and the Ordovician, that can be traced along almost the entire southern part of the Brabant Massif (Debacker, 2001; Debacker *et al.*, 2002, submitted). Towards the east of the Sennette valley, the map of De Vos *et al.* (1993) shows a similar contact between the Lower Cambrian Tubize Formation and the Upper Cambrian Mousty Formation, thus marking the absence of the Oisquercq Formation. In the Dyle-Thyle valley, the absence of the Oisquercq Formation has classically been attributed to low-angle thrusting (Anthoine & Anthoine, 1943; Delcambre & Pingot, 2002). However, a recent evaluation of existing geological and geophysical data by the present authors (unpub. data) shows that there is no evidence for such a large thrust. Possibly, this enigmatic limit between the Mousty and the Tubize Formation in the Dyle-Thyle valley is an Asquempont fault-like structure. A contact with a similar map appearance as the Asquempont fault also occurs along the northern part of the Brabant Massif, forming the limit between the Oisquercq Formation and the Ordovician in the west and between the Tubize Formation and the Mousty Formation in the east (cf. De Vos *et al.*, 1993). Hence, also there an Asquempont fault-like low-angle extensional detachment may be postulated.

Sintubin *et al.* (1998; cf. Sintubin, 1997a, 1999) suggested a spatial and genetic relationship between 1) the Asquempont lineament and other NW-SE-trending aeromagnetic lineaments in the core of the massif, which they interpreted as important dextral transpressive shear zones, 2) the limit between the Cambrian and the Ordovician (hereby erroneously equating the Asquempont lineament with the Asquempont fault as described by Legrand, 1967), and 3) the occurrence of steeply plunging folds within the Lower Cambrian deposits. New outcrop observations (Debacker, 2001; Debacker *et al.*, in prep.; see also Verniers *et al.*, 2002) show that, although seemingly reflecting an important tectonic break, at the present-day surface the Asquempont lineament corresponds to a gradual transition zone between steeply plunging folds, considered typical of the central steep belt ("Cambrian core domain" on fig. 4), and the gently plunging folds of the peripheral flat belt ("Ordovician-Silurian peripheral domain" on fig. 4), without there being any evidence for a marked tectono-stratigraphic break.

The post-cleavage normal faults in the Asquempont-Virginal area (F7, F10, F11, F12, F6) are situated within the eastern extension of the Nieuwpoort-Asquempont fault zone (De Vos *et al.*, 1993; cf. Oudenaarde-Bierghes fault zone of Legrand, 1968; André & Deutsch, 1985) (fig. 25). The results of the Asquempont-Virginal area, in combination with new observations at Bierghes and Quenast (Debacker, 2001; Debacker *et al.*, submitted), question the strike-slip movement along this fault zone advocated by Legrand (1968), André & Deutsch (1985), De Vos *et al.* (1993), Everaerts *et al.* (1996) and Sterpin & De Vos (1996). Instead, we consider this fault zone as essentially consisting of N- and S-dipping normal faults, deforming the Lower Palaeozoic basement in the southwestern part of the Brabant massif into a horst-and-graben geometry.

The nature of the large post-cleavage normal faults (quartz veins, semi-ductile cleavage deformation) suggests development at a certain depth. Hence, it does not seem likely that these faults originated after the Givetian unconformity. This pre-unconformity initiation is compatible with the Lochkovian to Namurian age of normal faulting in the eastern part of the Brabant Massif (Poty, 1991), can be reconciled with the Rb-Sr isotope resetting between the middle Eifelian and the late Frasnian (time-scale of Gradstein & Ogg, 1996), considered to mark movements along parts of the Oudenaarde-Bierghes fault zone (André & Deutsch, 1985), and also matches the data from Ronquières, four kilometres to the south of the Asquempont-Virginal area, where normal faults formed prior to the Givetian conglomerate deposition (Debacker *et al.*, 1999) as well as during and after deposition of the Givetian (Legrand, 1967; Debacker *et al.*, 1999). As evidenced by outcrop observations many of these normal faults were later reactivated (e.g. F10, F7; fig. 8, fig. 12). According to hydrochemical studies (Sterpin & De Vos, 1996) and recent earthquakes (Camelbeeck, 1993, 1997) parts of the Nieuwpoort-Asquempont fault zone appear to be still active (cf. Everaerts *et al.*, 1996; De Vos, 1997).

4.5. THE REDEFINED ASQUEMPONT FAULT DURING THE PROGRESSIVE UNROOFING OF THE BRABANT MASSIF

Currently, shortening across the Brabant Massif is considered to have started relatively early, possibly as early as Caradoc (Verniers *et al.*, 2002). According to Van Grootel *et al.* (1997) foreland basins were developing from the late Llandovery onwards. If truly representing foreland basins, then these basins imply the presence of a tectonic load within or in the direct vicinity of the Brabant Massif, and hence ongoing deformation, at least from the late Llandovery onwards. On the basis of stratigraphic, sedimentological, structural, geophysical and metamorphic data, a compressional wedge model for the Brabant Massif has been proposed (fig. 26), in which continued shortening caused the progressive inversion of an infilled Cambrian extensional basin, situated between cratonic basement blocks, thus giving rise to a compressional wedge (Verniers *et al.*, 2002; Sintubin & Everaerts, 2002; Debacker, 2001; Debacker *et al.*, 2002; Sintubin *et al.*, 2002). This progressive building-up of a compressional wedge resulted in a tectonic loading, in turn giving rise to Silurian foreland basins from the late Llandovery onwards. As deformation progressed outwards from the Cambrian core, finally also the Silurian foreland basins became deformed (see Verniers *et al.*, 2002; Sintubin & Everaerts, 2002; Sintubin *et al.*, 2002; Debacker, 2001; Debacker *et al.*, 2002). The Asquempont fault and possible other low-angle extensional detachments situated between the Lower Cambrian and the Ordovician can easily be fitted into this model, as shown in fig. 26 (Debacker, 2001; Debacker *et al.*, 2002).

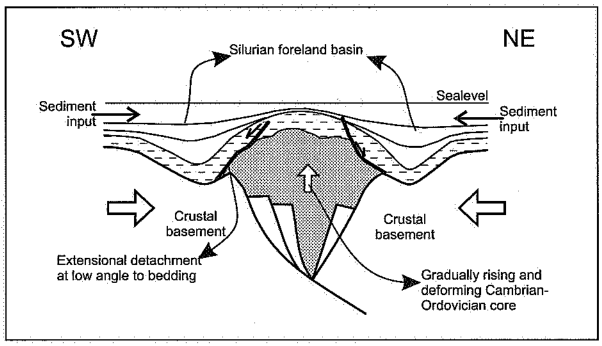

Figure 26: Model for the progressive deformation of the Brabant Massif (Debacker *et al.*, 2002). The core of the massif is being compressed between approaching crustal basement blocks, giving rise to a compressional wedge (cf. Sintubin & Everaerts, 2002). As the wedge builds up, its weight starts flexing down the lithosphere, resulting in the development of foreland basins along both sides of the core from the end of the Llandovery onwards. The steepening of the core may cause low-angle extensional detachments to form, such as the Asquempont fault.

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