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Folds and cleavage/fold relationships in the Brabant Massif, southeastern Anglo-Brabant Deformation Belt

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ABSTRACT: Detailed analyses of folds and cleavage/fold relationships within the Lower Palaeozoic Brabant massif have resulted in the recognition of pre-cleavage folds, three types of syn-cleavage folds, and two broad types of post-cleavage folds. These detailed analyses, and their integration with data from other sources, have led to major advances in our understanding of the Brabant Massif. The distribution of the different fold types yields significant insight into the small- and large-scale influence of pre-existing sedimentological, lithological and deformation features on the final deformation geometry. Changes in lithology and sedimentology result in different types of syn-cleavage folds, slump folds and local competent magmatic bodies control the position and to some extent also the orientation of syn-cleavage folds and on a larger scale also early basin-bounding features exert a major control on later deformation geometries. Moreover, combined with other data, the analysis of folds and cleavage/fold relationships provides valuable information on the large-scale architecture of the Brabant Massif, and on the evolutionary history of the Brabant Basin, from initial basin development to the final stages of basin inversion and gravitational collapse, and this both in time and in space. As illustrated herein, a thorough inventory of numerous detailed analyses of folds and cleavage/fold relationships is an absolute necessity for an understanding of the structural architecture, evolutionary history and tectonic significance of fold belts and slate belts.

KEYWORDS: cleavage, cleavage/fold relationship, kink band, Palaeozoic, slump fold

1. Introduction

The Brabant Massif (Fig. 1) forms the Belgian part of the Anglo-Brabant Deformation Belt, a slate belt composed of lowermost Cambrian to upper Silurian, mainly siliciclastic deposits. Although it is the best exposed area of the Anglo-Brabant Deformation Belt, even in the Brabant Massif outcrops are scarce. The outcrops are generally restricted to river valleys that cut through the overlying Cretaceous and Cenozoic deposits, and therefore outcrop areas are named after the river valleys in which they occur. The Brabant Massif formed as a result of the progressive inversion of the Brabant Basin, a basin comprising the Lower Palaeozoic deposits now found in the Condroz Inlier, in the Brabant Massif and below the Upper Palaeozoic deposits of the Namur Basin. This progressive inversion is now called the Brabantian Deformation event, and is considered to have taken place from the late Llandovery onwards until the Middle Devonian (Debacker et al., 2005a). This deformation event mainly resulted in the formation of folds and a moderately to well-developed cleavage.

Within the Brabant Massif, the presence of folds has been documented since the very beginning of Belgian geological research (e.g. Dumont, 1848; Malaise, 1873). However, these early studies mainly focused on stratigraphy. Although fold orientations were occasionally taken into account in order to determine the lateral continuation of the different stratigraphic units, three-dimensional fold geometry and cleavage/fold relationships were generally neglected.

Beyond any doubt, Fourmarier (1914, 1921) was the first geologist in Belgium who emphasized the importance of investigating cleavage and the relationship between cleavage and folds. Indeed, it was Fourmarier who introduced the concept of the relationship between folds and cleavage in Belgium, and who was the first to document the arc-shaped pattern of cleavage and folds along the southern part of the Brabant Massif. It was also Fourmarier (1921) who first documented the presence of steeply plunging folds within the Brabant Massif. Unfortunately, Fourmarier never actually focused on the cleavage/fold relationship in three dimensions, but confined himself to the axial-planar nature of cleavage as observed in fold profiles, and quite understandably was at a loss about the geological significance of the steeply plunging folds.

It comes as no surprise that the ideas of Fourmarier (1914, 1921) concerning cleavage and cleavage/fold relationships greatly influenced later geologists, both on a large, regional scale (e.g. Beugnies, 1963, 1964; Michot, 1976, 1978, 1980) and on a smaller scale (Mortelmans, 1953; Legrand, 1967). The ideas of Fourmarier (1914, 1921) inspired Mortelmans (1953) and Legrand (1967) to undertake a detailed analysis of the enigmatic convergent cleavage fans within the folds in the Silurian in the southern part of the Brabant Massif. Although their regional conclusions can be criticized (e.g. Debacker et al. 1999; Debacker, 2001), their analysis forms the basis of modern structural geology on cleavage/fold relationships in Belgium, in which emphasis is put on detailed observations of cleavage orientation, fold orientation, and cleavage/fold relationships, rather than on the large-scale, regional distribution of the different stratigraphic units. However, Mortelmans (1953) and Legrand (1967) also stuck to a rather two-dimensional analysis, focusing on cleavage

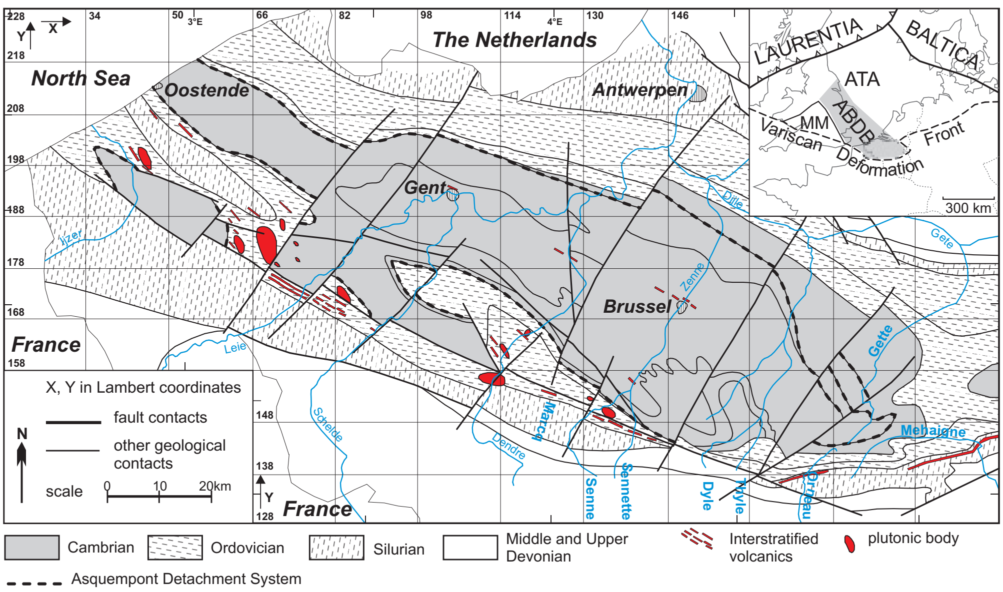

Figure 1: Geological subcrop map of the Brabant Massif, after De Vos *et al.* (1993b) and Van Grootel *et al.* (1997). Trace of the Asquempont Detachment System is taken from Debacker et al. (2004b, 2005b). The upper right inset shows the position of the Brabant Massif within Avalonia (ATA), as the southeastern part of the Anglo-Brabant Deformation Belt (ABDB), flanking the Midlands Microcraton (MM). Names of river valleys forming Lower Palaeozoic outcrop areas referred to in the text are written in bold.

fanning and an axial-planar relationship of cleavage with respect to the fold hinge line, without taking into account concepts such as cleavage refraction and the three-dimensional fold geometry.

Vandenven (1967) was probably the first to introduce a three-dimensional fold analysis in the Brabant Massif. Influenced by the work of Mortelmans (1953) and Kaisin (1933), he examined the three-dimensional orientation of small, post-cleavage kink bands in the Silurian of the Orneau valley and the Mehaigne and Burdinale valleys. He was also the first author to do a true kinematic analysis of folds (kink bands) in the Brabant Massif, and who stressed the importance of an analysis of kink bands, structures previously generally neglected in Belgian geology.

Despite the progress of Vandenven (1967) towards a detailed three-dimensional geometrical and kinematic analysis, it was not until the very last years of the 20th century that such studies were actually performed on a frequent basis in the Brabant Massif. In particular, geologists like Sintubin and Debacker and various co-workers systematically performed detailed geometrical and kinematic analyses of folds in the various Cambrian, Ordovician and Silurian outcrop areas (Sintubin, 1997a, 1999; Sintubin et al., 1998; Debacker, 1999, 2001, 2002; Debacker et al., 1997, 1999, 2001, 2002, 2003, 2004a, 2005a, 2005b, 2006, 2008, 2009; Beckers & Debacker, 2006; Debacker & Sintubin, 2008; Debacker & De Meester, 2009). In addition, these fold data and cleavage/fold relationships were successfully incorporated into the basin evolutionary history of the Brabant Massif (e.g. Sintubin, 1999; Sintubin & Everaerts, 2002; Verniers et al., 2002; Debacker et al., 2004a, 2005a).

In this paper an overview is given of the state of current knowledge on folds and cleavage/fold relationships in the Brabant Massif, mainly based on the works of Debacker, Sintubin and co-workers. On the basis of the cleavage/fold relationship three main groups can be distinguished: post-cleavage folds, syn-cleavage folds and pre-cleavage folds, some of which can further be subdivided.

2. Post-cleavage folds

2.1. Main characteristics

Post-cleavage folds are folds that fold cleavage. The axial surface of the post-cleavage folds is generally oblique to the pre-existing cleavage and, typically, the angle between cleavage and bedding remains constant across the fold. Mechanically, the instability responsible for folding may be due to strong viscosity contrasts associated with bedding, to the strong cleavage anisotropy or to a combination of both. The post-cleavage folds observed in the Brabant Massif appear to have formed primarily by buckling or by rotational shear deformation (e.g. contractional kink bands), although locally post-cleavage folds formed by bending have also been observed (e.g. extensional kink bands). Below, a distinction is made between fault-related post-cleavage folds, only observed within the damage zone of large faults, and kink bands. This distinction is somewhat arbitrary, as kink folds have also been observed locally within fault damage zones and kink bands often occur also within strongly faulted outcrop areas. Creep folds, formed by recent, gravitational shearing within the weathering top of the Lower Palaezoic basement, are not taken into account in the present work, as these merely result from a differential tilting of straight-sided, loose rock fragments under gravitational pull.

2.2. Kink bands

Here only a brief overview is given of the occurrence of the different types of kink bands in the Brabant Massif. For more detailed information on kink band terminology, kink band description and

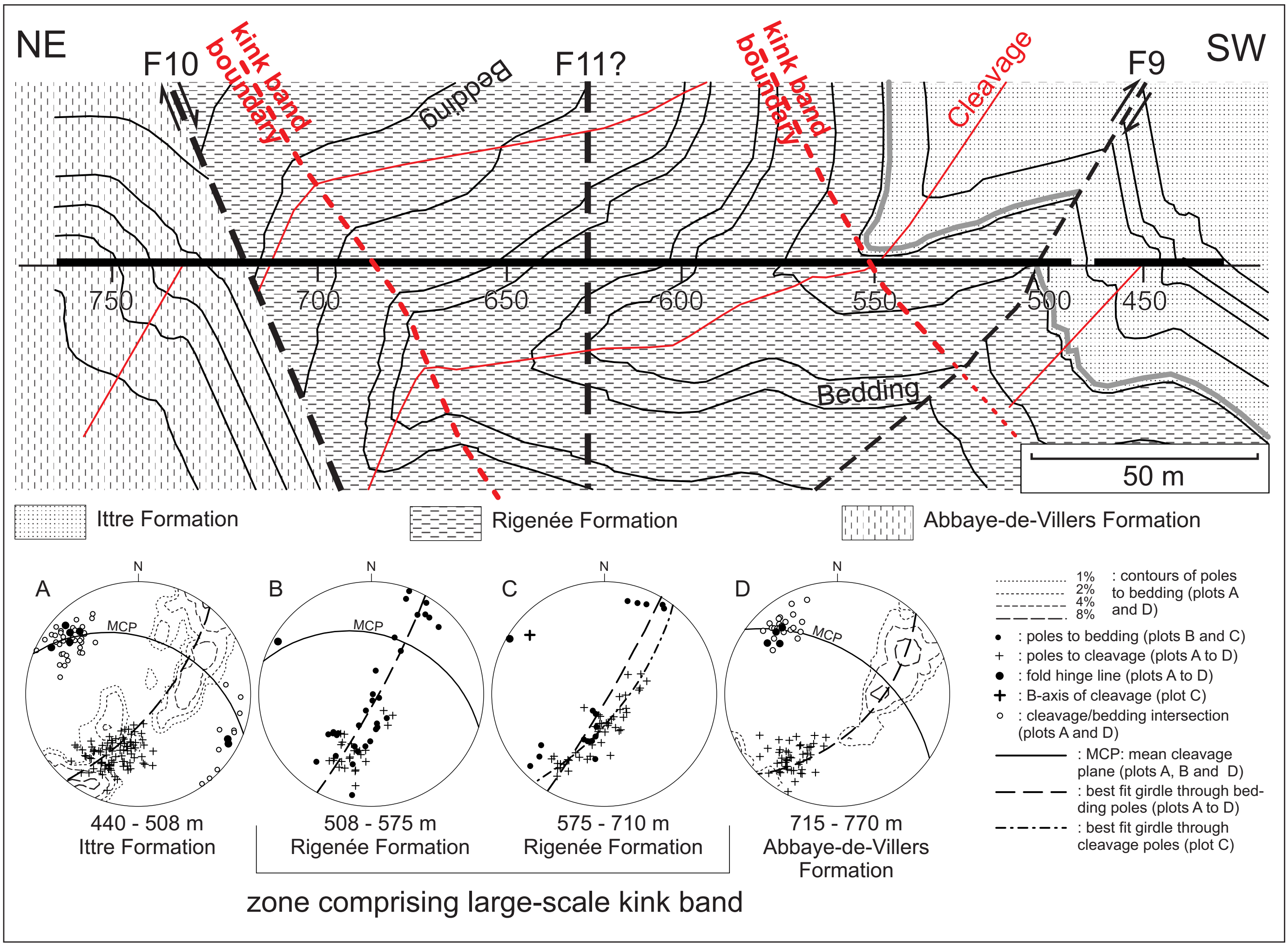

Figure 2: Large-scale kink band affecting deposits of the Rigenée Formation and pre-existing type A2 folds along the Bief 29 section, an isolated remnant of the old canal Brussels-Charleroi at Virginal (Asquempont, Ittre, Sennette valley) (after Debacker, 2001 and Debacker et al., 2003, 2004b). Projected distances are measured along the NE-side of the old canal, starting from its easternmost extremity.

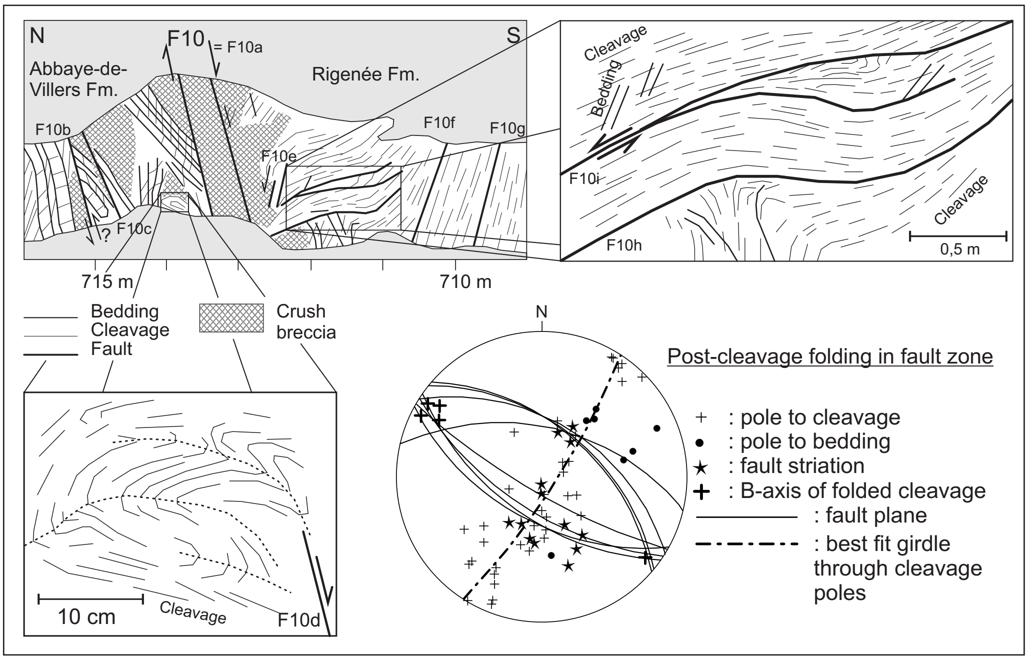

Figure 3: Fault-related post-cleavage folds and related deformation within the damage zone of normal fault F10 along the Bief 29 section at Virginal (Asquempont, Ittre) in the Sennette valley (after Debacker 2001 and Debacker et al., 2003, 2004b). See Figure 2 for position of F10.

the kinematic and dynamic interpretation of kink bands, the reader is referred to Anderson (1974) and Debacker et al. (2008) and references therein.

Small-scale kink bands in the Brabant Massif were first studied by Vandenven (1967) in the Silurian of the Orneau valley. These kink bands have a width of up to 5 cm, and rarely extend for more than one metre. These are contractional kink bands, with subhorizontal to gently dipping kink band boundaries and subhorizontal to gently plunging kink axes. Two sets occur: a top-to-the-north set and a top-to-the-south set. Locally, these form a conjugate system. Although less well developed, both sets of contractional kink bands have also been observed in the Silurian of the Sennette valley (Debacker et al., 1999, 2008). In both Silurian outcrop areas, the kink bands affect an upright to moderately dipping cleavage, and reflect development under the influence of a subvertical shortening (Debacker et al., 2008).

Also in the Silurian of the Mehaigne and Burdinale valleys small-scale contractional kink bands occur. These, however, have steeply dipping to subvertical, NNE-SSW-trending kink band boundaries and steeply plunging kink axes (Vandenven, 1967). Only dextral sets have been observed, suggestive of a WSW-ENE-directed shortening (Vandenven, 1967; Debacker, 2002). Comparable contractional kink bands, but with a sinistral asymmetry, have been observed locally in the Ordovician rocks of the Senne valley, to the NW and W of the Quenast plug (Debacker & Sintubin, 2008). These suggest a local NNW-SSE-directed shortening.

Thus far, small-scale extensional kink bands have only been observed within a high-strain zone in the Ordovician directly along the N-side of the Quenast plug (Senne valley). These kink bands are very narrow, with widths of only a few millimetres, and are very densely spaced (spacing often less than one centimetre). Kink axes are subvertical to steeply plunging and kink band boundaries are oriented subperpendicular to cleavage. These kink bands reflect a subhorizontal NE-SW-directed shortening at high angles to cleavage and to the plug – host rock interface (Debacker & Sintubin, 2008).

Also large-scale kink bands, with kink band widths of one metre or more, occur locally within the Brabant basement. The largest documented example has an estimated kink band width of ~120 m (Fig. 2). This kink band, with a subhorizontal to gently plunging, WNW-ESE-trending kink axis, occurs in the Sennette valley and affects folded rocks of both the Ordovician and the Cambrian (Debacker, 2001; Debacker et al., 2003, 2004b). The large-scale geometry suggests that this kink band is contractional, and formed under the influence of a subvertical shortening. Judging from its position amidst several normal faults of the Nieuwpoort-Asquempont fault zone, it is not unlikely that there is some genetic relationship between this large-scale kink band and fault initiation (Debacker, 2001; Debacker et al., 2003, 2004b).

2.3. Fault-related folds

Post-cleavage folds that formed due to faulting have only been observed in the Brabant Massif within the damage zone of normal faults. The folds within these damage zones are usually relatively small, with wavelengths rarely exceeding one metre.

There is a wide variation in fold geometry and orientation. The folds may be very angular, such as chevron folds and kink folds, or may be well-rounded, forming almost perfect sinusoids. Although some folds are perfectly cylindrical, many are markedly non-cylindrical (Fig. 3). Fold asymmetry may be extremely variable, ranging from symmetrical to strongly asymmetrical. Fold asymmetry does not necessarily match the fault movement sense, so in general the folds cannot be used for determining the sense of fault movement. The folds are often refolded (Fig. 3), and sometimes different fold generations may be recognized (e.g. kink bands overprinting earlier fold hinges). Characteristically, the folds are truncated by small faults or surrounded by zones of (post-cleavage) crush breccia or cataclasite, and may also fold mineralized slip planes (Fig. 3).

Fold axial surfaces may be parallel to the master fault plane, or may have a completely different orientation. Also the plunge of the fold hinge line is strongly variable. In most cases, however, fold hinge lines are situated within the master fault plane (Fig. 3).

The best examples of fault-related folds have been found in the Ordovician rocks of the Sennette and Senne valleys, where they occur within the damage zone of large-scale normal faults of the Nieuwpoort–Asquempont fault zone (Fig. 3; Debacker, 2001; Debacker et al., 2003, 2004b; Debacker & Sintubin, 2008).

3. Syn-cleavage folds

3.1. Main characteristics

Syn-cleavage folds are distinguished on the basis of a combination of four criteria. Firstly, cleavage is subparallel to the fold axial surface in the fold hinge zone, allowing for a small amount of cleavage transection (up to ~25°). Secondly, cleavage may remain parallel to the fold axial surface across the fold, or may show a convergent or divergent cleavage fanning, which ideally should be symmetrical about the fold hinge. Thirdly, the cleavage/bedding intersection lineation has a virtually constant orientation across the fold, and is always subparallel to the fold hinge line, allowing for a small amount of cleavage transection (up to ~25°). Fourthly, an opposite sense of cleavage refraction occurs on opposite fold limbs. In syn-cleavage folds, all four conditions are fulfilled. Moreover, the vast majority of the syn-cleavage folds can be regarded as buckle folds, formed by layer-parallel shortening. On some occasions, buckling is accompanied by a rather pronounced shear component (e.g. type A2 folds in the Marcq shear zone). None of the syn-cleavage folds formed by bending.

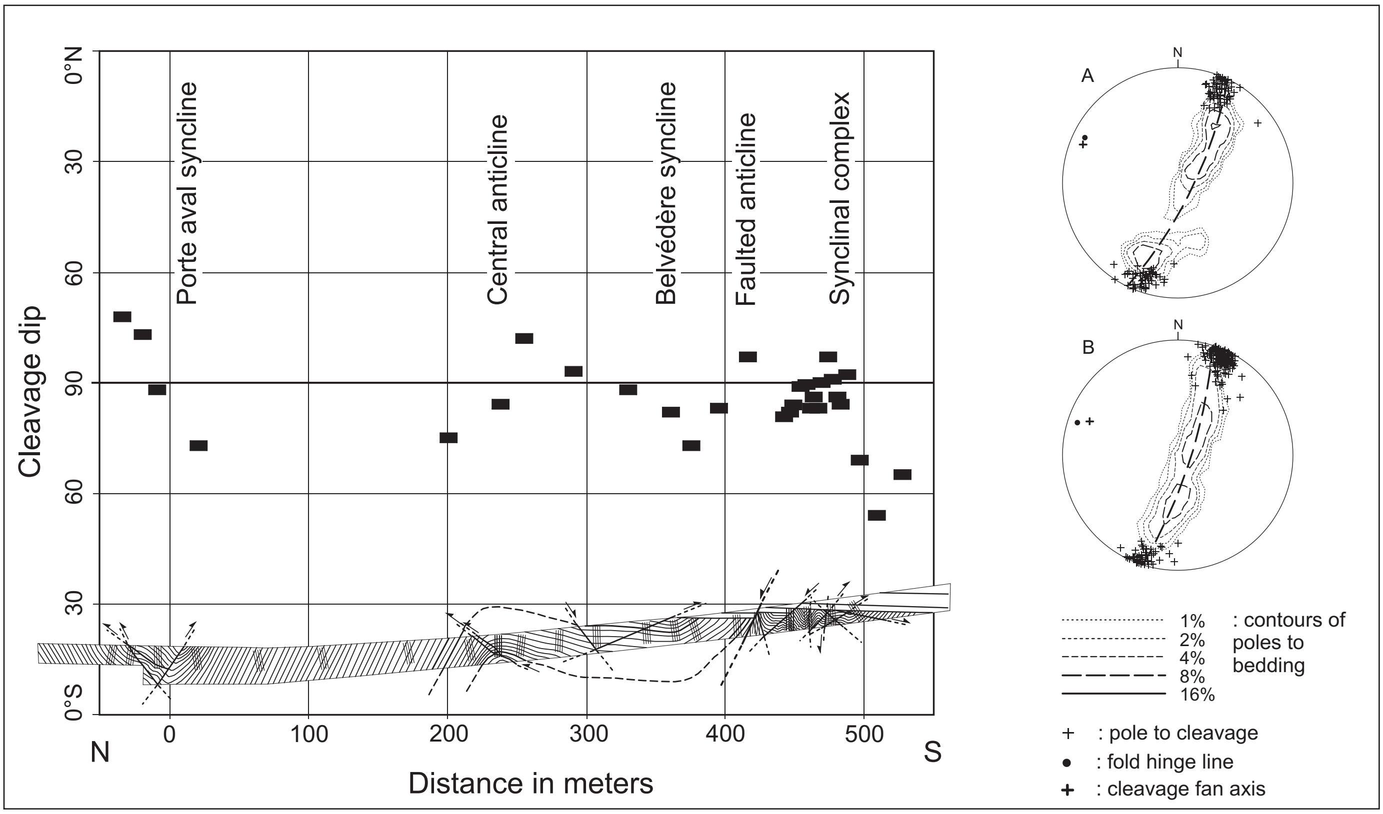

Figure 4: Type A1 folds in the inclined shiplift of Ronquières, below the Givetian angular unconformity (Sennette valley; modified after Debacker et al., 1999 and Debacker, 2001). The lower entrance to the inclined shiplift is situated at 0 m. Because of the convergent cleavage fanning, the type A1 folds can readily be recognised from changes in cleavage dip. The lower-hemisphere equal-area projections show the poles to bedding and to cleavage, the mean fold hinge line and the cleavage fan axis in the northern part of the section (A: north of 304 m) and in the southern part of the section (B: south of 304 m). Note the slightly opposite sense of cleavage transaction between the north and the south, attributed to an en-echelon periclinal fold shape (Debacker et al., 1999).

On the basis of the cleavage/fold relationship and the fold orientation, the syn-cleavage folds in the Brabant Massif can be subdivided into three fold types (Debacker, 2001).

3.2. Type A1 folds

The type A1 folds were previously called folds of the Ronquières type by Sintubin (1997a), and were renamed as type A1 folds by Debacker (2001) (Fig. 4). Type A1 folds are characterized by a gentle fold plunge (plunge < 35°) and a convergent cleavage fanning that developed within rather pelitic deposits. The type A1 folds are exceptional in the fact that the convergent cleavage fanning takes place throughout the pelitic deposits, irrespective of local changes in cleavage orientation due to cleavage refraction (Debacker, 2001, 2002). Folds have close to gentle interlimb angles and upright to steeply inclined axial surfaces. Although in the Brabant Massif the fold vergence reflected by the fold asymmetry is generally southward (e.g. Fourmarier, 1921; Sintubin, 1997a, 1999; Debacker, 2001 and references therein), the axial surfaces of the type A1 folds may be N- or S-dipping (Fig. 4). Folds are rounded to subangular. The roundness of the folds generally decreases with decreasing interlimb angle and with increasing limb dip, resulting in quite rounded gentle folds, with poorly defined hinge zones (e.g. Mehaigne valley and Burdinale valley), and subangular, close folds, with sharply defined hinge zones (e.g. Orneau valley) (Debacker, 2001, 2002).

Single-layer type A1 folds or small multi-layer type A1 folds have never been observed. Instead, type A1 folds are always quite large, with wavelengths larger than 10 metres, ranging up to several hundreds of metres, and amplitudes always larger than 2 metres. Hence, judging by the beds in which they occur, which are usually pelitic, rather incompetent, distal turbidite deposits, with a turbidite sequence thickness ranging from 1cm to 1m, the type A1 folds are typically large, multilayer buckle folds.

The type A1 folds have been defined in the Silurian of the Sennette valley in the inclined shiplift of Ronquières (lower Ludlow, upper Silurian; Debacker, 2001; cf. Legrand, 1967; Debacker et al., 1997, 1999) (Fig. 4). Type A1 fold hinges and type A1 cleavage/bedding relationships have been observed also in the Silurian of the Orneau valley (Kaisin, 1933; Mortelmans, 1953; Belmans, 2000, Debacker, 2001; Herbosch et al., 2002) and the Silurian of the Mehaigne valley and the Burdinale valley (Debacker, 2001, 2002). Type A1 folds have also been inferred on the basis of cleavage/bedding relationships (e.g. S-dipping cleavage at high angles to bedding in pelitic deposits) in the Silurian of the Landenne area (Debacker, 2001, 2002, based on data of De Winter, 1998). Type A1 folds or type A1 cleavage/bedding relationships have never been observed, nor inferred, within deposits in or older than the lower Wenlock Corroy Formation (cf. Debacker, 2001, 2002). Small-scale single-layer and multi-layer buckle folds within the Silurian deposits overlying the Corroy formation do not show a convergent cleavage fanning and belong to the type A2 folds, treated below.

3.3. Type A2 folds

The type A2 folds include the Fauquez type and the Marcq type of Sintubin (1997a). Like the type A1 folds, type A2 folds are characterized by a gentle fold plunge (plunge < 35°). In contrast to the type A1 folds, the type A2 folds never show an overall convergent cleavage fanning (Debacker, 2001). Instead, the type A2 folds are characterized by a cleavage that remains subparallel throughout the folds, subparallel to the axial surface, or shows a divergent cleavage fanning (Figs 5A & 5B). Fold axial surface orientations range from subvertical to gently N-dipping, thus reflecting an overall southward verging fold asymmetry, which locally can be very pronounced (e.g. Ordovician in the Marcq valley). Type A2 folds range from large-scale multi-layer buckle folds to single-layer buckle folds, with fold wavelengths ranging from km-scale to cm-scale (see Figs 5A, 5C, 5E & 5F). The type A2 folds are open to tight, and usually have subangular hinges in between long, relatively straight limbs (Fig. 5, see also Fig. 2).

The type A2 folds have been defined in the Ordovician of the Sennette valley at Asquempont (Figs 5A & 5B) and Fauquez (Figs. 5C & 5E) (Upper Ordovician; Debacker, 2001; Debacker et al., 2001, 2003; Verniers et al., 2005; cf. Fourmarier, 1921). Type A2 fold hinges and type A2 cleavage/bedding relationships have been observed or inferred also in the Ordovician of the Senne valley (Debacker &

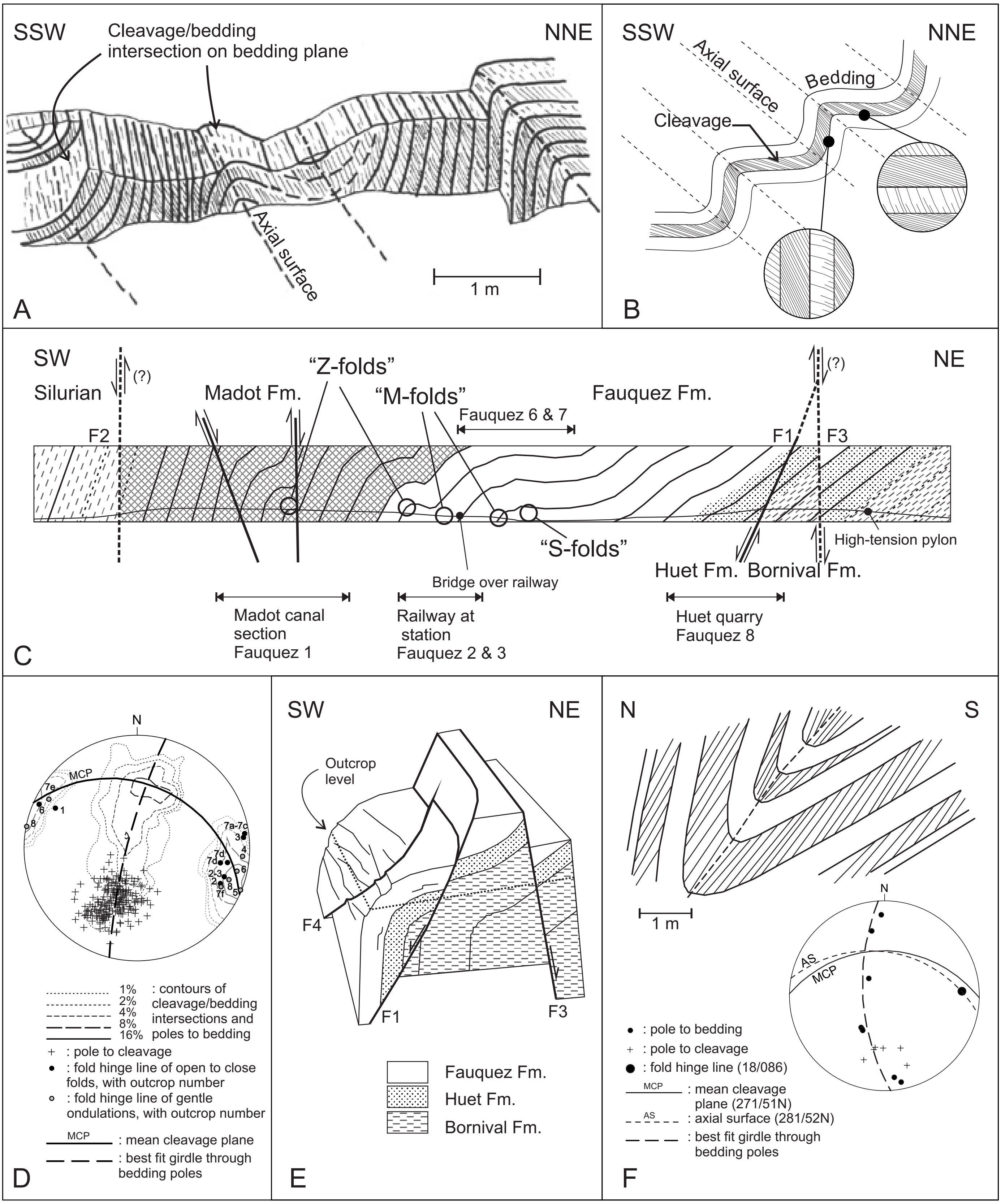

Figure 5: Type A2 folds and cleavage/fold relationships. A) Line drawing of small-scale type A2 folds in the Upper Ordovician Ittre Formation within the southern Asquempont section, along the W-side of the canal Brussels-Charleroi (Sennette valley, between 40345 and 40355 m; Debacker, 2001). B) Conceptual image of small-scale type A2 folds, based on (A) (after Debacker et al., 2001). Note the divergent cleavage fanning symmetrical about the axial surface, and the opposing sense of cleavage refraction on opposite fold limbs. Note that the folds depicted in A and B fold beds that were previously overturned due to slumping (Debacker et al., 2001). C) Simplified section across the Fauquez area (Upper Ordovician, Sennette valley), essentially consisting of a large-scale type A2 fold (after Debacker, 2001 and Verniers et al., 2005). Note the particular position of S-, M- and Z-folds. D) Lower-hemisphere equal-area projection of bedding, cleavage, cleavage/bedding intersection and fold hinge lines of the Fauquez area (after Debacker, 2001). See (C) for outcrop position. E) Schematic block diagram showing the Fauquez area as an open, asymmetric, southward verging type A2 antiform, cut by normal faults. Note the position of the smaller-scale parasitic folds close to the antiform hinge zone (after Debacker, 2001 and Verniers et al., 2005). F) Line drawing with lower-hemisphere equal area projection of type A2 fold within the Lower Cambrian Tubize Formation at Blanc Ri, Dyle valley (after Debacker, 2001 and Debacker et al., 2005b).

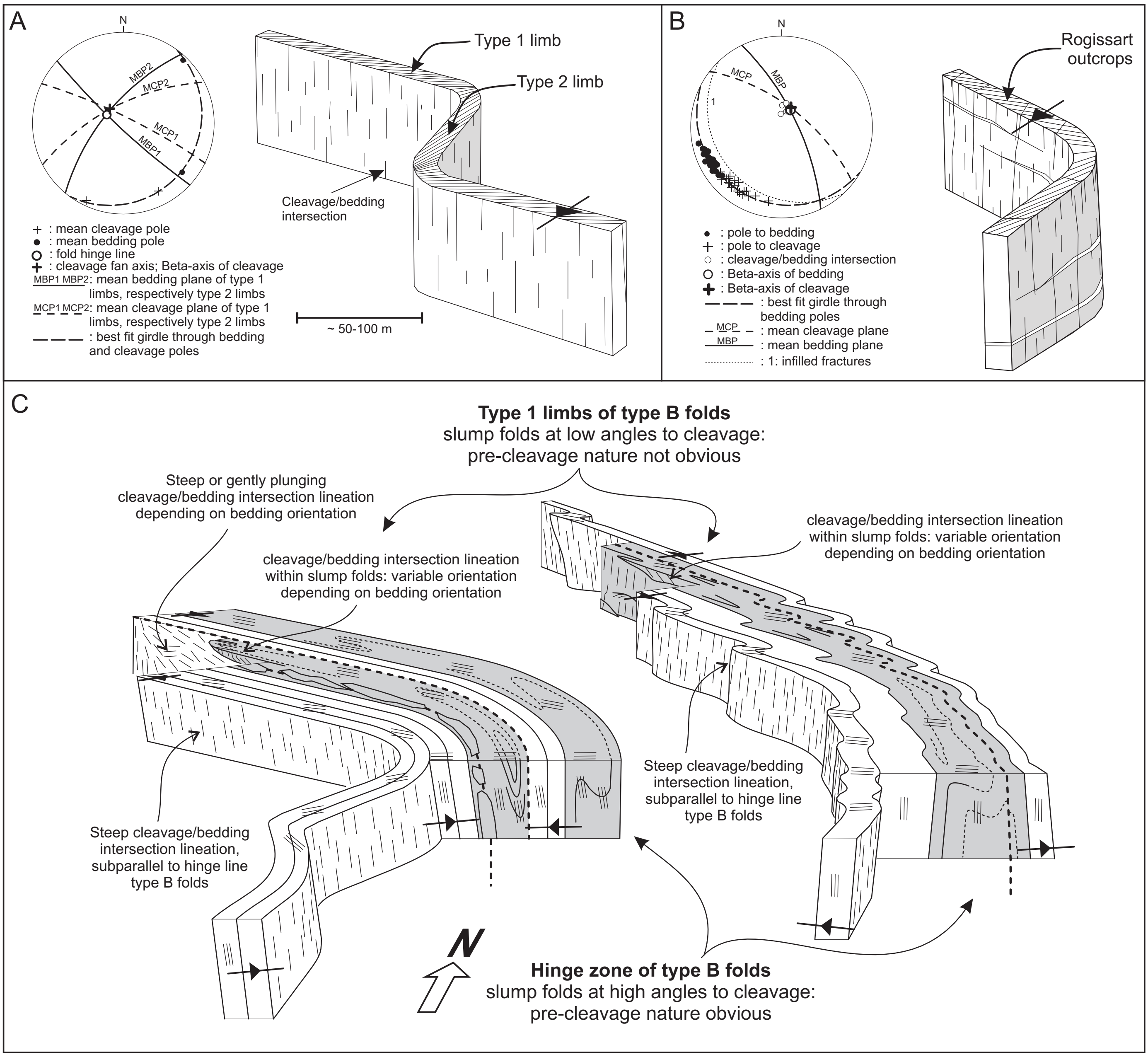

Figure 6: Lower-hemisphere equal-area projections and conceptual drawings of type B folds and cleavage/fold relationships within the Cambrian core of the Brabant Massif (after Debacker et al., 2004a, 2006). A) Type B folds in the Tubize Formation at Lembeek, Sennette valley, based on data from Sintubin et al. (1998). B) Steep, uniformly dipping beds of the Tubize Formation at Rogissart (Sennette valley), representing the type 1 limb of a type B fold. C) Conceptual block diagrams of type B folds within the Jodoigne Formation at Jodoigne (Geete valley), affecting subhorizontal pre-cleavage folds and related pre-cleavage deformation attributed to slumping (marked in grey). Younging sense is marked by arrows.

Sintubin, 2008; Debacker & De Meester, 2009), the Cambrian of the Senne and Sennette valleys (Debacker, 2001; Debacker et al., 2004a; Piessens et al., 2004), the Ordovician of the Orneau valley (Debacker, 2001; Herbosch et al., 2002), the Ordovician and Cambrian (Fig. 5F) of the Dyle and Thyle valleys (Beckers, 2003, 2004; Debacker et al., 2005b; Beckers & Debacker, 2006), the Ordovician of the Marcq valley (Debacker, 1999, 2001), the Cambrian of the Geete valley (Herbosch et al., 2008) and the lower Silurian (from the lower Wenlock Corroy Formation towards older beds) in the Mehaigne and Burdinale valleys (Debacker, 2002). Small-scale multi-layer to single-layer type A2 buckle folds have been observed also in the Silurian of the Sennette valley at Ronquières (upper Silurian) and in the Mehaigne valley in the limbs of the large-scale type A1 folds (Debacker, 2001, 2002).

3.4. Type B folds

The type B folds were previously called folds of the Lembeek type by Sintubin (1997a), and were renamed as type B folds by Debacker (2001). The type B folds are characterized by a steep fold plunge (plunge > 35°) and a steeply plunging cleavage/bedding intersection (Fig. 6). Characteristically, the type B folds often show an overall divergent cleavage fanning, even within quite competent beds (Sintubin et al., 1998). However, also type B folds with a parallel to even convergently fanning cleavage have been observed locally within competent sequences (e.g. within folded Cambrian quartzites of the Blanmont Formation and the Jodoigne Formation, Herbosch et al., 2008). Fold axial surface orientations range from subvertical to steeply N-dipping, thus reflecting an overall slightly southward verging fold asymmetry. Type B folds range from large-scale multi-layer buckle folds to single-layer buckle folds, with fold wavelengths ranging from km-scale to dm-scale. The type B folds are open to tight, and usually have subangular hinges in between relatively long, straight limbs (Fig. 6). The type B folds seem characterized by unequal limb lengths, reflecting a pronounced fold asymmetry. In the type areas in the western part of the Brabant Massif (e.g. Lembeek area, Sennette valley) the fold geometry systematically points to a dextral asymmetry, with a long, NNW-SSE-trending type 1 limb and a shorter NE-SW-trending type 2 limb (Sintubin et al., 1998; Debacker et al., 2004a) (Figs 6A & 6B). In the more eastern parts, however, a predominance of N(NE)-S(SW)-trending type 2 limbs becomes apparent, seemingly suggesting sinistral fold asymmetries

	Formation	**Marcq / Dender**	**Senne / Sennette**	**Dyle / Thyle**	**Orneau**	**Landenne**	**Mehaigne/ Burdinale**	**Geete**
Silurian	**Ronquières**		**Type A1**			**Type A1**		
	Vichenet		Type A		**Type A1**	**Type A1**	**Type A1**	
	Fumal		Type A		**Type A1**		**Type A1**	
	Vissoul		Type A		**Type A1**		Type A	
	Les Vallées		Type A		**Type A1**		Type A	
	Corroy		Type A		Type A		**Type A2**	
	Fallais		Type A		Type A		**Type A2**	
	Hosdin		Type A				**Type A2**	
	Latinne		Type A				**Type A2**	
	Bois G.-P.		Type A		Type A			
	Brutia				Type A			
Ordovician	**Madot**		**Type A2**		Type A			
	Fauquez		**Type A2**					
	Huet		**Type A2**		Type A			
	H. de Rebecq		**Type A2***		Type A			
	Bornival		**Type A2**	Type A	Type A			
	Ittre		**Type A2**	Type A	Type A			
	Rigenée		**Type A2**	**Type A2**	**Type A2**			
	Tribotte		**Type A2**	**Type A2**	Type A			
	Abb. de Villers		**Type A2**	**Type A2**	Type A			
	Chevlipont	**Type A2**	**Type A2**	**Type A2**				
Cambrian	**Mousty**			**Type A2 Type B**				
	Jodoigne							**Type A2 Type B**
	Oisquercq		**Type A2 Type B**					
	Tubize		**Type A2 Type B**	**Type A2 Type B**				
	Blanmont		**Type B**	**Type B**				**Type A2 Type B**

Table 1: Observed occurrences of type A1, type A2 and type B folds and cleavage/bedding relationships within the Lower Palaeozoic formations in the different outcrop areas of the Brabant Massif. Type A refers to cases where the distinction between type A1 and type A2 could not be made. Note that in the case of type A1 folds in the Silurian only the predominant fold type is indicated; hence, "type A1" implies that the larger folds and the majority of the cleavage/fold relationships belong to this type, irrespective of the fact that sometimes also small-scale type A2 folds have been observed (see text for explanation). Blank cells imply that the formation is not present, or that the fold type is unknown (due to absence of folds or cleavage). *: uncertain whether this formation is actually present in this outcrop area (see Debacker et al., 2011).

of regional importance (e.g. Sintubin et al., 2002; Debacker et al., 2005b; Herbosch et al., 2008).

The type B folds have been defined in a series of temporary outcrops of the Lower Cambrian Tubize Formation in the Senne-Sennette valley at Lembeek (Sintubin et al., 1998) (Fig. 6A). The only place where type B fold hinges can still be observed is in the Cambrian Jodoigne Formation in the Geete valley at Jodoigne (Debacker et al., 2006; cf. Fourmarier, 1921) (Fig. 6C). Despite the scarcity of observable type B fold hinges, type B cleavage/bedding relationships can be observed in numerous places in the Cambrian core of the Brabant Massif. The presence of type B cleavage/bedding relationships has been demonstrated in the Lower Cambrian to Middle Cambrian of the Senne-Sennette valley (Sintubin et al., 1998; Debacker et al., 2004a; Piessens et al., 2004) (Figs 6A & 6B), the Lower Cambrian to Upper Cambrian of the Dyle-Thyle valley (Sintubin et al., 2002; Debacker et al., 2005b) and the Lower Cambrian to Middle or Upper Cambrian of the Geete valley (Debacker et al., 2006; Herbosch et al., 2008) (Fig. 6C). Type B folds have never been observed in units younger than the Upper Cambrian Mousty Formation (Table 1).

3.5. Spatial occurrence of the type A1 and type A2 folds and discussion

Type A1 folds or type A1 cleavage/bedding relationships (Fig. 4) have been observed in all Silurian outcrop areas along the southern rim of the Brabant Massif, but only in deposits younger than the lower Wenlock Corroy Formation (Debacker, 2001, 2002) (Table 1). By contrast, type A2 folds or type A2 cleavage/bedding relationships have been observed in all outcrop areas. Moreover, type A2 folds or cleavage/bedding relationships have been observed in Cambrian, Ordovician and Silurian deposits (Table 1). However, in deposits younger than the Corroy Formation, only local, single-layer and small-scale multilayer type A2 folds occur (Debacker, 2001, 2002).

Previously, the type A1 folds have been interpreted as being the result of a poly-phase deformation, in which a first deformation phase formed gentle to open folds with an axial-planar cleavage, and a second deformation phase caused a further fold tightening, resulting in convergent cleavage fans (e.g. Kaisin, 1933; Mortelmans, 1953; Legrand, 1967). According to Mortelmans (1953), based on observations in the Orneau valley, the first deformation phase took place prior to the Givetian, whereas the second phase took place after the Givetian and was attributed to the Variscan Orogeny. Legrand (1967), however, demonstrated a truncation of the convergent cleavage fans in the Sennette valley by the Givetian unconformity, and concluded that both deformation phases took place prior to the Givetian, an idea followed by Michot (1978). Although not mentioned by Mortelmans (1953), a similar truncation can be observed in the Orneau valley at the locality of Les Mautiennes, (Belmans, 2000; Herbosch et al., 2002). Later, Debacker et al. (1997, 1999; see also Belmans, 2000 and Debacker, 2002) demonstrated that there is no evidence for a polyphase deformation in the Silurian of the Brabant Massif, and tentatively attributed the type A1 folds to a single, progressive deformation event, later redefined as the Brabantian deformation event (Debacker, 2001; Verniers et al., 2002; Debacker et al., 2002, 2005a). However, although such a progressive deformation, with continued fold amplification after cleavage development, can explain the formation of type A1 folds, it does not explain why

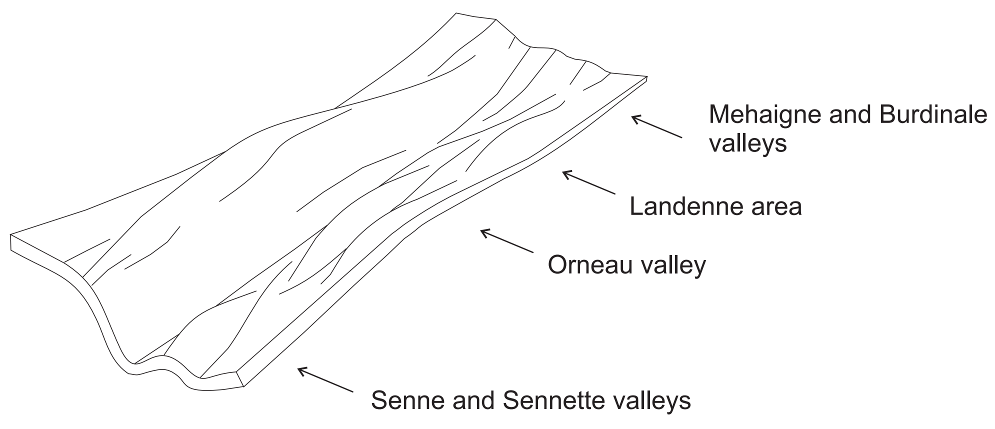

Figure 7: Conceptual drawing of the Silurian southern rim of the Brabant Massif, hosting the type A1 folds, and interpreted as a large-scale periclinal fold assemblage with a synformal shape (after Debacker, 2002). The approximate positions of the southern Sennette valley, the Orneau valley, the Landenne area and the Mehaigne and Burdinale valleys are added. Note the low amount of shortening across the Mehaigne and Burdinale valleys as compared to the southern Sennette valley and the Orneau valley. The highest amounts of shortening occur across the Orneau valley and between the Orneau valley and the southern Sennette valley.

this progressive deformation would lead to type A1 folds only in the Silurian, and not in the Ordovician, where only type A2 folds are observed. Debacker (2001, 2002) noticed that the type A1 folds always occur in a larger-scale synform that can be traced all along the southern rim of the Brabant Massif (Fig. 7), and that type A1 folds only occur above the lower Wenlock Corroy Formation. This led Debacker (2002) to suggest that the presence of type A1 folds is either due to the position within this large-scale synform, or due to the particular mechanical properties of the rather homogenous, pelitic deposits of the Silurian above the Corroy Formation (cf. Verniers et al., 2002). The former idea is difficult to evaluate. The latter idea, in which the mechanical properties of the deposits are involved, is supported by the local occurrence of small-scale type A2 within the type A1 folds, and by the observation of Belmans (2000) of gradual changes in cleavage/bedding angle within uniformly dipping type A1 fold limbs, implying changing amounts of convergent cleavage fanning with stratigraphic position in the Silurian above the Corroy Formation.

The type A2 folds and cleavage/bedding relationships (Fig. 5) are by far the most common. They have been observed in all Ordovician outcrop areas and all Silurian outcrop areas older than the Wenlock Corroy Formation. Also in many Cambrian outcrop areas, type A2 folds and cleavage/bedding relationships have been observed (Table 1). Moreover, small-scale type A2 folds have been observed also above the lower Wenlock Corroy Formation.

As outlined above, the type A2 folds include both the Fauquez type and the Marcq type of Sintubin (1997a). However, the folds at Fauquez and at Marcq are merely slightly more irregular cases of ideal type A2 folds. At Marcq, within the Marcq valley, a low-angle (~30°NE) reverse shear zone occurs (Debacker, 1999; Piessens et al., 2000, 2002). During shear zone development and propagation, intense strains developed at the frontal and lateral shear zone tips, leading to the development of folds with all the characteristics of (syn-cleavage) type A2 folds, but with a plunge direction ranging from transport-parallel to transport-perpendicular (Debacker, 1999). At present, the Marcq valley is still the only place in the Brabant Massif in which the presence of a low-angle reverse shear zone has been demonstrated and in which fold geometry can be linked directly to shear zone development and propagation. As these folds bear all characteristics of type A2 folds, there is no need for introducing an extra fold type. Hence, the use of the Marcq type of Sintubin (1997a) is discouraged.

As outlined in Verniers et al. (2005; see also Debacker, 2001), the Fauquez area (Sennette valley) is characterized by a large-scale type A2 antiform, with small asymmetrical type A2 folds on the large fold limbs close to the antiform hinge zone (so-called S or Z folds) and more symmetrical small type A2 folds within the antiform hinge zone (so-called M folds) (Figs 5C & 5E). Although this antiform is cross-cut by several large, mainly normal, post-cleavage faults, an analysis of the folds and cleavage/bedding relationship demonstrates that the change in bedding geometry across the Fauquez area is virtually entirely due to type A2 folding. This contrasts with the ideas of Legrand (1967) and Hennebert & Eggermont (2002), who show the Fauquez area as being composed entirely of fault bounded units without any folds. Some of the smaller folds in the Fauquez area are strongly asymmetrical and non-cylindrical, showing quite large cleavage transection angles and a cleavage that is markedly oblique to the bisecting plane of the fold interlimb angle. Such observations led Sintubin (1997a) to define these folds as the Fauquez type. However, folds at the other localities of the Fauquez type mentioned by Sintubin (1997a) do not show many of these characteristics. As outlined in Verniers et al. (2005; cf. Debacker, 2001), the non-cylindrical nature and strong amount of cleavage transection of some of the small folds at Fauquez may reflect a localized dextral transpression during folding, indirectly resulting from the strong eastward decrease in thickness of the competent volcaniclastic deposits of the Upper Ordovician Madot Formation. As the folds at Fauquez are merely more irregular cases of type A2 folds, the use of the name Fauquez type of Sintubin (1997a) should be avoided.

3.6. Spatial occurrence of the type A2 and type B folds and discussion

The type B folds have only been observed within the Cambrian core, from the lowermost Cambrian Blanmont Formation to the Upper Cambrian Mousty Formation (Table 1). Their presence and their occurrence restricted to the Cambrian core are puzzling. Fourmarier (1921) considered these as local features without any regional importance. By contrast, struck by the steepness of the Cambrian core (see Legrand, 1968), and in particular by the presence of a steeply plunging cleavage/bedding intersection lineation within this core, Giese et al. (1997) proposed a weak deformation event between the Cambrian and the Ordovician, consisting of a tilting or weak folding, without cleavage development (cf. Legrand, 1968). Sintubin et al. (1998) were the first to demonstrate a syn-cleavage nature of the type B folds (their Lembeek type; see Fig. 6A) and put forward two complementary models for their development. In a first model they interpret these as incongruous folds developed on the limb of a large, upright isoclinal fold structure, during a single, progressive, coaxial deformation event. Fold hinge rotation occurred because of the sub-parallelism between layering and the axial plane of the host fold structure, leading to progressive rotation of the fold hinges lines towards the regional principal extension direction. However, although borehole data indeed suggest that steeply dipping bedding dominates the largest part of the Cambrian core, isoclinal folds have never been observed (cf. Legrand, 1967, 1968). In addition, as remarked by Sintubin *et al.* (1998), significant fold hinge rotation during progressive coaxial deformation would necessitate high strains and a significant differential stretching, which is seemingly in contradiction with the open nature of the type B folds and the axial-symmetrical disposition of the cleavage fabric. In a second model, non-coaxial, dextral transpressive deformation circumstances are invoked. Either the steeply plunging folds initially developed with steeply plunging fold hinge lines, or they formed as a result of fold hinge rotation during progressive shearing of relatively competent folded layers (cf. Mawer & Williams, 1991). In both cases, shearing is considered to have occurred along dextral shear zones, thought to be reflected by the NW-SE-trending aeromagnetic gradient lineaments in the core of the Brabant Massif (cf. Sintubin, 1997b, 1999).

Debacker et al. (2004a, see also Debacker, 2001) performed a detailed analysis of the cleavage/bedding relationship in the Oisquercq Formation at Asquempont (Sennette valley) directly south of the southwesternmost NE-SW-trending aeromagnetic gradient lineament, called the Asquempont lineament (not to be confused with the Asquempont fault). This analysis revealed a gradual, but surprisingly irregular transition from type A2 folds in the south, farthest away from the Asquempont lineament, towards type B folds in the north, closest to the Asquempont lineament. Instead of a progressive N-ward increase in plunge of the fold hinge lines and cleavage/bedding intersection, plunges and plunge directions start becoming increasingly variable as the plunge becomes steeper, resulting in adjacent moderately NE- and moderately SW-plunging folds, until a true type B cleavage/bedding relationship is obtained (Fig. 8A; see also Fig. 9). This implies strongly curvilinear folds within the transition zone between the type A2 and the type B folds. Debacker et al. (2004a) attributed this transition zone between type A2 folds and type B folds and the presence of type B folds to dextral shear related to the NW-SE-trending aeromagnetic gradient lineaments, thus largely complying with the ideas of Sintubin (1997b, 1999) and Sintubin et al. (1998) (Fig. 8B). However, the absence of high strains within the transition zone, within zones of type B folds and in the vicinity of the aeromagnetic gradient lineaments were interpreted as being a result of a peripheral position with respect to the shear zones. As suggested in Debacker et al. (2004a), this absence of high strains, combined with the general absence of clear stratigraphic breaks in the wide vicinity of the NW-SE-trending aeromagnetic gradient lineaments, can be explained by considering the inferred shear zones as blind shear zones, not reaching up to the present-day erosion surface of the Brabant basement (Figs 8B & 8C). These inferred shear zones are thought to result from compression of the Cambrian core against a large, competent, low-density body at depth (roof between 2 and 5 km depth; see De Meyer, 1983, 1984; Everaerts et al., 1996; De Vos, 1997), resulting in a SE-ward escape of steepened Cambrian core material (Debacker, 2001; Debacker et al., 2004a; see also Sintubin, 1999) (Fig. 8B). This escape is considered to have happened in a discrete way at shallow depths, resulting in type B folds, without high strains or significant stratigraphic displacements (Figs 8A & 8C), and in a more pronounced, discontinuous way at larger depths, reflected by the NW-SE-trending aeromagnetic gradient lineaments (Fig. 8B).

Although this shear zone model appears to have been generally accepted, and is compatible also with more recent observations (e.g. Piessens et al., 2004; Debacker et al., 2005b), there are three kinds of observations that suggest that the situation is slightly more complicated. Firstly, type B cleavage/bedding relationships have been observed in the Cambrian core also well away from any obvious aeromagnetic gradient lineaments (e.g. Sintubin et al., 1998, 2002; Debacker et al., 2004a, 2005b; Piessens et al., 2004). This may be explained, however, by the presence of shear zones at depth that juxtapose units with a comparable magnetic susceptibility or density, or by shear zones without a significant stratigraphic displacement, thus being "invisible" on aeromagnetic and gravimetric maps. Secondly, type B cleavage/bedding relationships have been observed also in the Dyle-Thyle valley and even in the Geete valley (e.g. Fig. 6C), very far to the east of the largest low-density gravimetric anomaly body at depth, held responsible for indirectly generating the SE-ward dextral shear in the Cambrian core (Sintubin et al., 2002; Debacker et al., 2005b, 2006; Herbosch et al., 2008). The orientation of the type 1 limbs and the type 2 limbs changes in an anticlockwise fashion towards the east, from NNW-SSE trending, respectively NE-SW-trending in the west, to (N)W-(S)E- trending, respectively N(NE)-S(SW)-trending in the east (Sintubin et al., 2002; see also Debacker et al., 2005b and Herbosch et al., 2008). As pointed out by Sintubin et al. (2002; see also Sintubin, 1997b), the orientation of

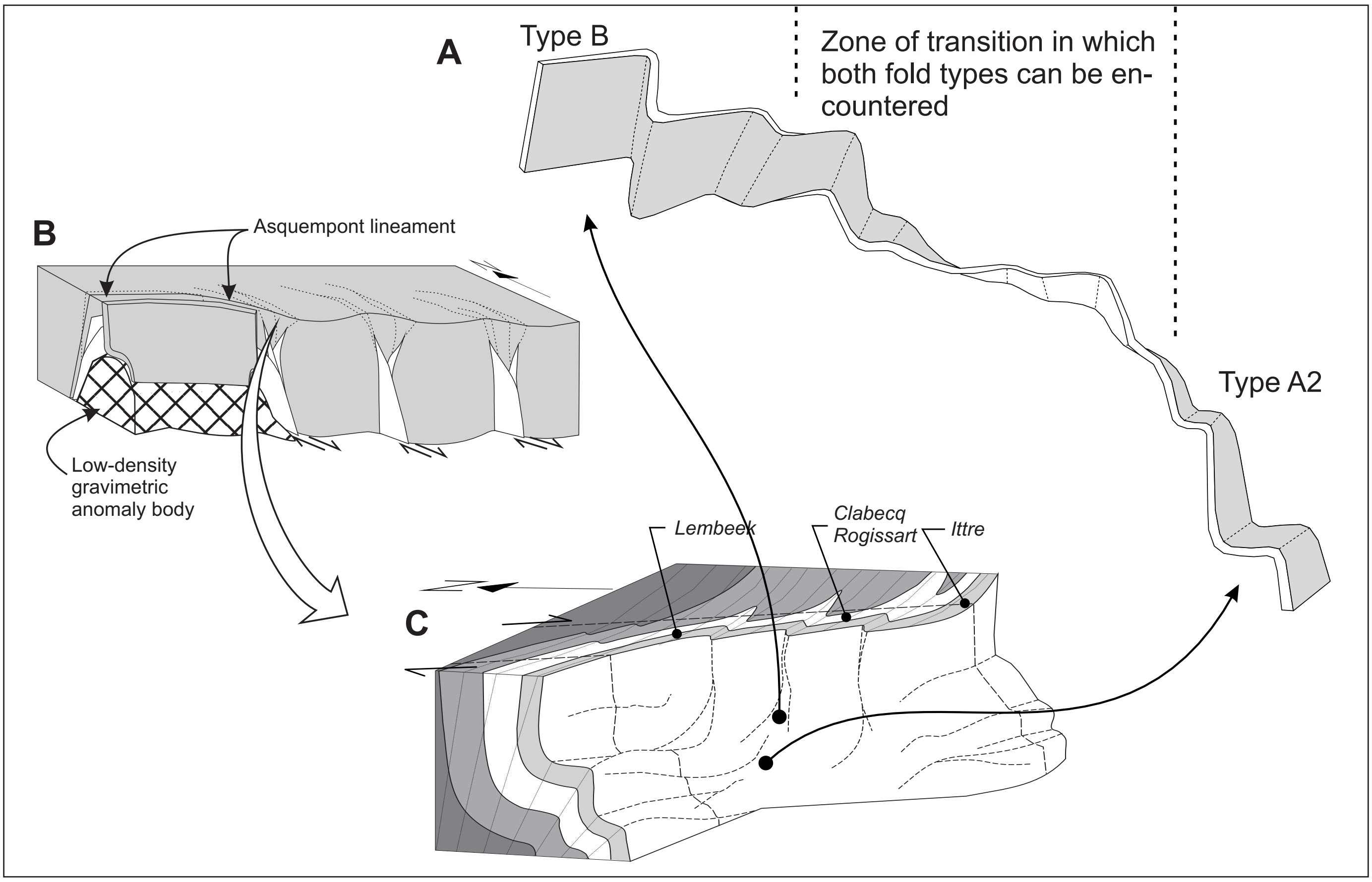

Figure 8: Transition between type B folds (north) and type A2 folds (south) in the vicinity of the aeromagnetic Asquempont lineament in the Lower Cambrian of the Sennette valley (after Debacker, 2001 and Debacker et al., 2004a). A) Schematic representation of bedding geometry within a transition zone between type A2 folds to the south and type B folds to the north, based on observations across a ~500 m long outcrop of the Oisquercq Formation at Asquempont along the E-side of the canal Brussels-Charleroi (outcrop N.A.S. 2 in Debacker et al., 2004a). B) Conceptual image of the entire southwestern margin of the Lower Cambrian core of the Brabant Massif, viewed from the SW. The core was compressed against the low-density gravimetric anomaly body, thus experiencing a steepening and giving rise to a steep belt (Sintubin, 1999). Parts of the core were thrust over the roof of this low-density body (Debacker, 1999) and a large part of the steep core experienced a dextral transpressive movement along the NE side of this body, with the development of dextral shear zones, visible as aeromagnetic lineaments (see Sintubin et al. 1998; Sintubin 1999; Verniers et al. 2002). At depth, the Asquempont lineament partly coincides with the NE margin of the low-density body. At more shallow levels the dextral movement was accommodated in a more gradual fashion, by means of Z-shaped type B folds (predominance of type 1 limbs), showing a gradual transition towards the type A folds. C) Conceptual image of the fold transition zone within the Ittre–Asquempont–Lembeek area (Sennette valley), with the approximate positions of the outcrop areas and of the observed fold train depicted in (A). It should be noted that, for convenience, all outcrop areas are placed along the Asquempont lineament, whereas in reality Lembeek appears to be situated along another lineament.

aeromagnetic gradient lineaments changes in a similar fashion (see aeromagnetic maps in De Vos et al., 1993a and Chacksfield et al., 1993), and hence also in the eastern parts the type B folds may be related to the aeromagnetic gradient lineaments. More importantly, however, the predominance of type 1 limbs in the western parts (Senne-Sennette valley; there NNW-SSE-trending) apparently gives way to a predominance of type 2 limbs towards the east (Dyle-Thyle valley and Geete valley; there N(NE)-S(SW)-trending) (see Sintubin et al., 2002; Debacker et al., 2005b; Herbosch et al., 2008). Whereas a predominance of type 1 limbs is compatible with a regional, NW-SE-directed dextral shear, the apparent predominance of type 2 limbs in the eastern parts of the Brabant Massif suggests a N-S-directed sinistral shear of regional importance (e.g. Geete valley). Thirdly, the Geete valley is situated to the north of the central axis of the Brabant Massif (e.g. Michot, 1980), and the overall younging sense of the deposits is towards the northeast, instead of towards the southwest (Herbosch et al., 2008). This implies that, apart from their change in orientation, the type 1 limbs and the type 2 limbs of the type B folds have completely opposite younging senses as compared to the type 1 limbs and type 2 limbs in the Senne-Sennette and Dyle-Thyle valleys. Although not contradicting the shear model proposed for the type B folds (e.g. Fig. 8), the second and third kind of observations hitherto remain unexplained, mainly due to the lack of exposure.

3.7. Similarities and differences in orientation between type A1, type A2 and type B folds

Fleuty (1964) devised a graph for describing fold orientation, based on the dip of the fold axial surface and the plunge of the fold hinge line. Such a graph, with all available fold data from the Brabant Massif, is shown in Fig. 9. Only folds with both known axial surface orientation and fold hinge line are shown. Hence, the documented fold transition zone between type A2 and type B folds, which is based on cleavage/bedding relationships rather than on fold observations (Fig. 8A; see Debacker et al., 2004a), does not become obvious. On this graph, the three different fold types occupy different areas.

Although a large overlap is observed between the type A1 and the type A2 folds, the latter generally have lower axial surface dips. This, in particular holds true for the type A2 folds in the Ordovician. Along the southern part of the Brabant Massif, the Ordovician often shows a gently to steeply N-dipping cleavage, axial planar to the type A2 folds, whereas the type A1 folds (Silurian above the Corroy Formation) are characterized by upright to steeply dipping axial surfaces. As the type A2 folds always have N-dipping axial surfaces, whereas type A1 folds sometimes also have S-dipping axial surfaces, a modification of the Fleuty diagram allowing for a distinction between N- and S-dipping axial surfaces would result in less overlap between the type A1 and the type A2 folds. The most gently N-dipping axial surfaces are found in the Ordovician of the Marcq area (Marcq valley), the Quenast area (Senne valley) and the Asquempont-Fauquez area (Sennette valley).

Although less pronounced, another difference between both fold types is the plunge of the fold hinge line. Type A2 folds have a larger variation in plunge than the type A1 folds. This is either due to the more irregular bedding geometries within many of the Ordovician units (because of shallow, sandy deposits or of frequent occurrence of pre-cleavage deformation e.g. Beckers & Debacker, 2006; Debacker et al., 2009) or due to the vicinity of the type B folds in the Cambrian (e.g. Debacker et al., 2004a).

The type B folds occupy a separate area on the Fleuty diagram (Fig. 9). All type B folds are characterized by steeply dipping axial surfaces. Many of these folds are reclined, implying that the plunge direction of the fold hinge line is at low angles to the dip direction of the fold axial surface.

Going from the southern, Silurian rim of the Brabant Massif towards the Cambrian core, axial surface dip decreases whereas fold hinge line plunge remains fairly similar when passing from type A1 folds to type A2 folds. The transition from type A2 folds towards type B folds involves a northward (re-)steepening of the fold axial surfaces. During this steepening, fold hinge line plunge becomes more variable, and even extremely variable (e.g. Fig. 8A), until true type B folds are obtained. This shift, from type A1 folds in the south, towards type B folds in the north, is shown schematically on Fig. 9 by the thick grey arrow. Note that the shift took place the other way round in time, with first the development of type B folds (and type A2 folds) in the core, giving rise later to type A1 folds in the Silurian rim (see below).

4. Pre-cleavage folds

4.1. Recognizing pre-cleavage folds

The pre-cleavage folds that are easiest to recognize are those cross-cut with a high three dimensional obliquity by cleavage. A three-dimensional obliquity implies that cleavage is both oblique to the axial surface in fold profile and oblique to the fold hinge line. In such cases, cleavage cross-cuts the fold axial surface and does not show a fanning symmetrical about the fold hinge (e.g. Figs 10A, 10C & 10H). Moreover, the cleavage/bedding intersection will be markedly oblique to the fold hinge line, and will change in orientation across the fold. In addition, cleavage refraction sense will not be opposite on opposite limbs, and instead of refraction sense changing at the fold hinge, this will happen somewhere within a fold limb, away from the fold hinge. As long as bedding and cleavage are well-developed, such pre-cleavage folds cross-cut obliquely by cleavage are fairly easy to distinguish from syn-cleavage folds.

However, in many cases cleavage is only slightly oblique to the fold axial surface. When folds are tight, even a small obliquity between cleavage and the fold axial surface may still be recognized in fold profile, but with increasing fold interlimb angle this becomes increasingly difficult. Moreover, it may also happen that cleavage appears axial planar in fold profile, but is in fact oblique to the fold hinge line. In such cases, cleavage may even show a fanning (divergent in incompetent beds, convergent in competent beds) that appears symmetrical about the fold hinge in the fold profile (e.g. Fig.

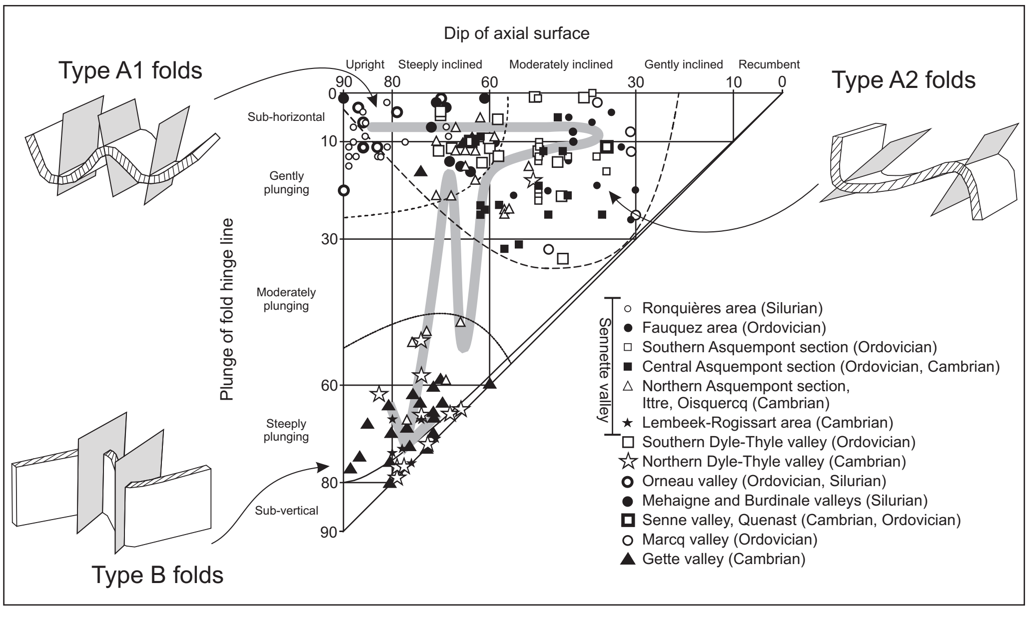

Figure 9: Fleuty diagram (Fleuty, 1964) with data of the type A1, type A2 and type B folds within the main outcrop areas of the Brabant Massif. Note that only data are shown from folds on which we have information about the orientation of the fold hinge line and of the axial surface. The grey arrow schematically marks the change in fold orientation going from the Silurian southern rim of the massif towards the steep Cambrian core.

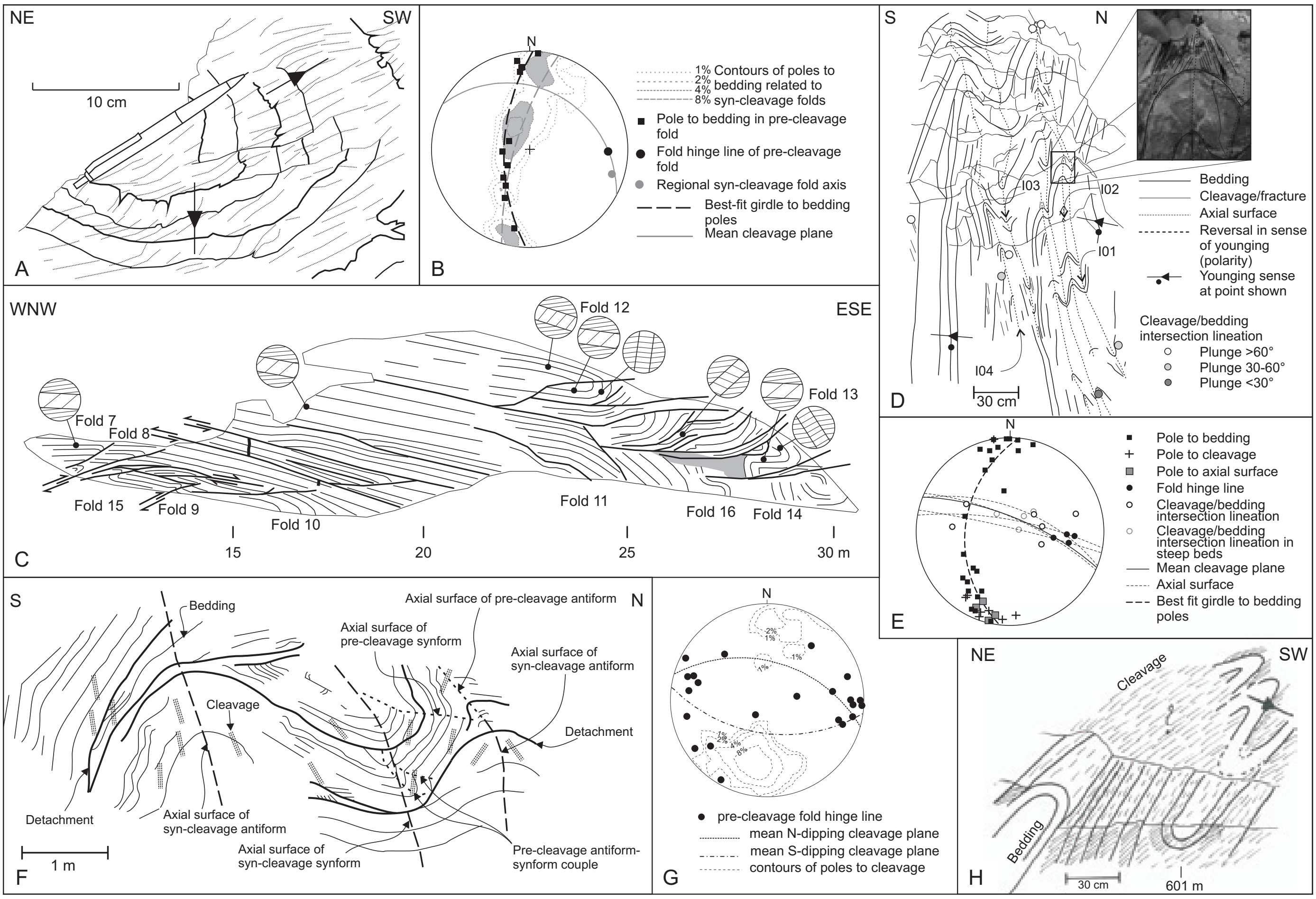

Figure 10: Pre-cleavage folds within the Brabant Massif. A-B) Pre-cleavage fold within the Ordovician Ittre Formation to the south of Asquempont (Sennette valley), cross-cut at high angles to its axial surface by cleavage (A; arrows mark younging sense) and lower-hemisphere equal-area projection of pre-cleavage fold data compared with data from regional syn-cleavage folds (B) (after Debacker et al., 2001, 2003). C) Pre-cleavage folds and related detachments within the Ordovician Abbaye-de-Villers Formation, at Quenast (Senne valley). Circular insets show local cleavage/bedding relationships (after Debacker & De Meester, 2006). D-E) Small-scale, pre-cleavage folds within the Cambrian Jodoigne Formation at Jodoigne (Geete valley), cross-cut at very low angles by cleavage (D; arrows mark younging sense) and lower-hemisphere equal-area projection of pre-cleavage fold data (E) (after Debacker et al., 2006). Note that although cleavage appears axial planar in fold profile, the cleavage/bedding intersection lineation is often not parallel to the fold hinge lines, thus giving away the pre-cleavage fold origin. F) Pre-cleavage folds and related detachments folded by syn-cleavage type A2 folds (two antiforms and one synform) within the Ordovician Abbaye-de-Villers Formation in the direct vicinity of the abbey of Villers (Thyle valley; after Beckers & Debacker, 2006). G) Lower-hemisphere equal-area projection of pre-cleavage fold hinge lines and poles to cleavage from within the Ordovician Abbaye-de-Villers Formation in the direct vicinity of the abbey of Villers (Thyle valley; after Beckers & Debacker, 2006). Note the strong variation in pre-cleavage fold hinge line orientation. H) Pre-cleavage folds within the Ordovician Rigenée Formation along the Bief 29 section at Virginal (Asquempont, Ittre, Sennette valley), cross-cut at high angles by cleavage (see Fig. 2 for position; after Debacker, 2001; arrows mark younging sense).

10D; Debacker et al., 2006). If so, the only way of recognizing the true pre-cleavage nature is by means of a detailed analysis of the orientation of the cleavage/bedding lineation around the fold (Figs 10D & 10E). Still, if in these cases cleavage is only slightly oblique to the fold hinge, it may become extremely difficult to distinguish these from axially transected syn-cleavage folds (*sensu* Johnson, 1991). Indeed, syn-cleavage folds have been reported with an axial cleavage transection of up to ~25°, and this also in the Brabant Massif (e.g. Debacker, 2001). In such complex cases, recognizing pre-cleavage folds also relies on other criteria, such as cross-cutting relationships with phenomena related to syn-cleavage or post-cleavage folds or the association with specific features that are not associated with the syn-cleavage and post-cleavage folds.

4.2. Main characteristics

The pre-cleavage folds in the Brabant Massif typically occur in bedding-parallel zones, as is well illustrated in the Ordovician of the Thyle valley (Beckers & Debacker, 2006). Typically, these folds are bounded above and below by "non-folded" beds (i.e. folded on a larger scale by syn-cleavage folds, but not by pre-cleavage folds). This boundary with the "non-folded" beds is rather sharp, and often coincides with a bedding-parallel detachment surface (e.g. Figs 10C & 10F). Where observed, also these bedding-parallel detachments have a pre-cleavage origin. Moreover, unlike most of the post-cleavage faults which clearly stand out in outcrop, these detachments often have a welded nature (e.g. Debacker et al., 2009).

Individual pre-cleavage folds generally have small wavelengths of usually less than 3 metres, and small amplitudes. Fold shape and orientation are, however, extremely variable. Some pre-cleavage folds are strongly asymmetrical (e.g. Fig. 10H), whereas others are virtually symmetrical (e.g. Fig. 10F). Some pre-cleavage folds are upright with respect to the surrounding "non-folded" beds (e.g. Fig. 10F), whereas others are recumbent (e.g. Figs 6C, 10C, 10D & 10H). A large variety in fold interlimb angles exists, even for adjacent folds, ranging from tight to gentle (e.g. Fig. 10C). Some pre-cleavage folds are perfectly cylindrical, whereas others are markedly non-cylindrical. In some outcrops, the different pre-cleavage folds have comparable fold hinge line orientations (e.g. Beckers & Debacker, 2006; Debacker & De Meester, 2009), whereas in other outcrops fold hinge line orientation is strongly variable (e.g. Debacker et al., 2009) (Fig. 10G). Still, irrespective of the variation in fold hinge line orientation and the strong variation in cylindricity, fold hinge lines are generally situated within the bedding plane of the overlying or underlying "non-folded" beds (Figs 10B, 10E & 10G).

Unlike in the case of the syn-cleavage folds, fold vergence of the asymmetrical pre-cleavage folds in the Brabant Massif is variable. However, when compensating for the occasionally strong variations in fold hinge line orientations (e.g. Debacker et al., 2009), it appears that the overall fold vergence is quite constant within specific stratigraphic

	Formation	Occurrence	Vergence	Source
Silurian	Ronquières	/	/	
	Vichenet	/	/	
	Fumal	/	/	
	Vissoul	/	/	
	Les vallées	/	/	
	Corroy	/	/	
	Fallais	/	/	
	Latinne	/	/	
	Hosdin	/	/	
	Bois Grand-Père	/	/	
	Brutia	/	/	
Ordovician	Madot	/	/	
	Fauquez	Very local	To N ?	Lammertyn (2005)
	Huet	/	/	
	H. de Rebecq	/	/	
	Bornival	Local (to frequent?)	?	Debacker (2001), Debacker et al. (2001, 2003)
	Ittre	Frequent	? and to N*	Debacker (2001), Debacker et al. (2001, 2003), Herbosch et al. (1991); * N-directed folds in Ombret Fm, Condroz Inlier (Valcke, 2001).
	Rigenée	Very local	? and to N*	Debacker (2001), Debacker et al. (2003); * N-directed folds in Chevreuils Fm, Condroz Inlier (Debacker & Vanmeirhaeghe, unpub. data; cf. Vanmeirhaeghe, 2006)
	Tribotte Abbaye de Villers	(very) Frequent	To S	Debacker (2001), Debacker et al. (2003), Beckers (2003, 2004), Beckers & Debacker (2006), Debacker & De Meester (2009); cf. Michot (1977) and Legrand (1968)
	Chevlipont	Local (to frequent?)	To N	Beckers (2003, 2004), Herbosch et al. (1991), Debacker (2001), Debacker et al. (2003)
Cambrian	Mousty	/	/	
	Jodoigne	Frequent	To N(?)	Debacker et al. (2006) and unpub data
	Oisquercq	Local (?)	?	Debacker (2001), Legrand (1968)
	Tubize	Local (to frequent?)	?	Debacker (2001), Legrand (1968), Mortelmans (1955), Giese et al. (1997)
	Blanmont	/	/	

Table 2: Observed occurrences of pre-cleavage folds within the different formations of the Brabant Massif, with indication of mean fold asymmetry. Note that for some Ordovician stratigraphic units, also data of the Condroz Inlier are incorporated, as these deposits formed part of the Brabant Basin during the Ordovician and the Silurian.

intervals. Within the Lower to Middle Ordovician Abbaye-de-Villers Formation, for instance, fold vergence is roughly towards the south (Beckers & Debacker, 2006; Debacker & De Meester, 2009; Debacker et al., 2009), whereas within the Lower Ordovician Chevlipont Formation fold vergence is towards the north (Beckers, 2003, 2004; Debacker, 2001; Debacker et al., 2003) and also in the Upper Ordovician Ittre Formation pre-cleavage fold vergence appears to be northward (Valcke, 2001; cf. Debacker, 2001; Debacker et al., 2001, 2003) (Table 2). In the case of a northward pre-cleavage fold vergence, the overprinting N-dipping cleavage usually results in a clear cross-cutting relationship, clearly demonstrating the pre-cleavage fold nature (Fig. 10A) (see Debacker et al., 2001). For the southward verging pre-cleavage folds, however, the cross-cutting relationship of the N-dipping cleavage is much more subtle (Fig. 10F) (e.g. Beckers & Debacker, 2006). In addition, for the majority of the pre-cleavage folds in the Brabant Massif there is no large angle between cleavage trend and the trend of the fold axial surface (e.g. Fig. 10E), making it sometimes impossible to find anomalous cleavage/fold relationships, in particular regarding the southward verging pre-cleavage folds.

Fortunately, there are other characteristics of the pre-cleavage folds that clearly allow them to be distinguished from syn-cleavage and post-cleavage folds. These are the common association with pre-cleavage detachments and pre-cleavage breccias and zones of disrupted sediments, suggestive of sediment failure prior to lithification (e.g. Figs 6C, 10C & 10F). In numerous cases we also found isolated fold hinges within zones of disrupted sediments (e.g. Debacker et al., 2006), as well as zones characterized by obvious pre-lithification deformation in which certain beds are overturned with respect to the overlying and underlying beds, separated from the latter by pre-cleavage detachments (e.g. Fig. 6C). Also these, more or less bedding-parallel, zones of overturned beds are attributed to recumbent pre-cleavage folds, of which the hinges are not visible or have been truncated by pre-cleavage bedding-parallel detachments (e.g. Debacker et al., 2001, 2006) (e.g. Fig. 6C).

4.3. Occurrence and geological significance of the pre-cleavage folds

The pre-cleavage folds and associated detachments and bedding-parallel zones of overturned bedding either result from an older, pre-cleavage deformation phase or they are the result of slumping. As explained at length by Debacker (2001), Debacker et al. (2001, 2006, 2009), Beckers & Debacker (2006) and Debacker & De Meester (2009), and contrary to the views of Legrand (1968) and Giese et al. (1997), there are no arguments for an older, pre-cleavage and post-lithification deformation phase in the Brabant Massif. Instead, all observations, and in particular the association with zones of disrupted sediments and other obvious pre-lithification deformation features (e.g. liquefaction features, ball-and-pillow structures), pre-cleavage breccias, welded detachments and the intraformational nature of the pre-cleavage folds, suggest that the pre-cleavage folds are a result of slumping and should be considered as slump folds (see Debacker, 2001 and Debacker et al., 2009 and references therein).

The occurrence of slump folds within the Brabant Massif has already been suggested by several previous workers. Legrand (1968), for instance, stresses the abundance of slump folds in the lowermost Ordovician (cf. Michot, 1977), and also mentions numerous slump folds in boreholes in the Lower Cambrian (cf. Mortelmans, 1955; Giese et al., 1997). Also in the Upper Ordovician of the Lessines borehole, numerous small folds were attributed to slumping (Herbosch et al., 1991). However, arguments for a slump origin were generally not provided, and none of the authors actually demonstrated a pre-cleavage origin of these folds. This is unfortunate, as slump folds should not be distinguished from post-lithification folds merely on the basis of fold morphology (e.g. Woodcock, 1976; Maltman, 1994).

Table 2 lists the occurrence of documented pre-cleavage folds in the Brabant Massif, all of which are interpreted as slump folds.

5. What can we learn from such a detailed fold analysis in the Brabant Massif?

The detailed fold analyses have given us much insight into the basin evolutionary history of the Brabant Massif, from development of the Brabant Basin to final inversion under influence of the Brabantian Deformation event (e.g. Verniers et al., 2002; Debacker et al., 2005a and references therein). Moreover, these detailed analyses have also provided valuable information on fold development.

Stratigraphic intervals with a high incidence of slump folds and related soft-sediment deformation record periods of basin instability. Worth mentioning here are the Jodoigne Formation (Debacker et al., 2006; Herbosch et al., 2008; see Figs 10D & 10E), the Abbaye de Villers Formation (Beckers & Debacker, 2006; Debacker & De Meester, 2009; Debacker et al., 2009; see Figs 10C, 10F & 10G) and the Ittre Formation to lower part of Bornival Formation (Debacker et al., 2001; see Figs 10A & 10B), all of which have a high concentration of slumping-related deformation features (Table 2). The cause of slumping within the Jodoigne Formation still remains unknown (Debacker et al., 2006), whereas slumping in the Ittre Formation to lower part of Bornival Formation has been attributed to fault activity (Debacker et al., 2001). One of the possible candidates for causing this slumping is the Asquempont Detachment System, which was active between the Caradoc and the timing of cleavage development (Debacker et al., 2004b, 2005b; Herbosch et al., 2008). Slumping in the Abbaye-de-Villers Formation is either due to fault activity or to a high meteorite influx (Debacker et al., 2009; cf. Parnell, 2009).

The slump fold geometry and slump fold orientation can be used for determining slump sense and hence paleoslope orientation, and can as such provide valuable information on the basin geometry, in particular when combined with paleocurrent data (e.g. Woodcock, 1979). As can be seen in Table 2, during most stratigraphic intervals slumping took place from south to north, suggesting a N-dipping palaeoslope of regional extent. This is fully compatible with the depositional turbidite palaeocurrent during the Silurian (Verniers & Van Grootel, 1991), and also matches the Middle Ordovician to Silurian facies distribution within the Brabant Basin, with the southern deposits (Condroz Inlier) generally thinner and more shallowly deposited than their northern equivalents (Brabant Massif) (e.g. Vanmeirhaeghe, 2006). During deposition of the Abbaye-de-Villers formation, however, southward slumping took place, suggestive of a S-dipping palaeoslope of at least 30 km long (Debacker et al., 2009 and references therein).

The occurrence of the syn-cleavage folds appears to be controlled by pre-existing features. Hence, the geometry and position of the syn-cleavage folds provides information on initial basin architecture and, moreover, also provides insight into the progressive deformation history of the Brabant Massif.

Small-scale type A2 folds within the Abbaye-de-Villers Formation in the Thyle valley are considered to have formed on pre-existing slump folds (Beckers & Debacker, 2006; see Fig. 10F). Within the Sennette valley, the position and the shape of the Fauquez antiform (type A2 fold) are believed to be controlled by the thick, irregularly shaped volcaniclastic deposits of the Madot Formation (Beckers & Debacker, 2006; cf. Verniers et al., 2005). Similarly, within the same valley, the position of the Asquempont synform (type A2 fold) is likely related to the presence of a ~200m thick wedge-shaped overturned mass within the synform hinge (Beckers & Debacker, 2006; cf. Debacker et al., 2001). The overall steepness of the Cambrian core and the steeply plunging type B folds being restricted to this steep core, are considered to be related to the overall basin architecture of a deep Cambrian basin surrounded by rigid basement blocks (Sintubin, 1999; Sintubin & Everaerts, 2002; Verniers et al., 2002; Debacker et al., 2005a).

When combined with geophysical data and age constraints, the occurrence of the different types of syn-cleavage folds allows a visualization of the progressive deformation of the Brabant Massif during the Brabantian Deformation event. As suggested in Debacker et al. (2005a), by the end of the Llandovery, the Cambrian core of the massif was already under compression whereas the Silurian deposits along the rims were still being deposited. During this compression, the Cambrian basin was shortened, and steepened, resulting in the build up of a compressional wedge (Sintubin & Everaerts, 2002). In places where shortening was oblique to the edges of the rigid basement blocks, this resulted in the formation of dextral shear zones within the Cambrian core (Figs 11 & 8). As this deformation spread outwards, type A2 folds formed in the Ordovician deposits above the rigid basement blocks (Fig. 11). The weight of the growing compressional wedge started flexing down the lithosphere, thus giving rise to Silurian foreland basin development on the edges of the Brabant Massif. During the middle Silurian to early Devonian, this progressive deformation gradually spread towards the Silurian rims of the massif, where it resulted in the development of type A1 folds in the pelitic deposits above the Corroy Formation and type A2 folds at lower stratigraphic levels (Fig. 11).

The steeply plunging kink bands reflect continued shortening after cleavage development (Debacker, 2002; Debacker & Sintubin, 2008). Both the subhorizontal kink bands and the normal faults, with their associated post-cleavage folds, result from a roughly subvertical shortening after cleavage development (Debacker et al., 2008). As the development of kink bands requires high confining pressures (Anderson, 1974), the kink bands likely formed at deeper levels than the normal faults, or earlier in the unroofing process. The analysis of the subhorizontal kink bands within the Silurian rim indicates a strong influence of pre-existing anisotropies (here bedding and cleavage) on kink band geometry and orientation (Debacker et al., 2008). In this respect, the common occurrence of two sets of subhorizontal kink bands within the Silurian rim of the Brabant Massif may be related to the presence of the type A1 folds, as only within these folds critical angles between cleavage and bedding, necessary for the development of these kink bands, occur throughout the folds.

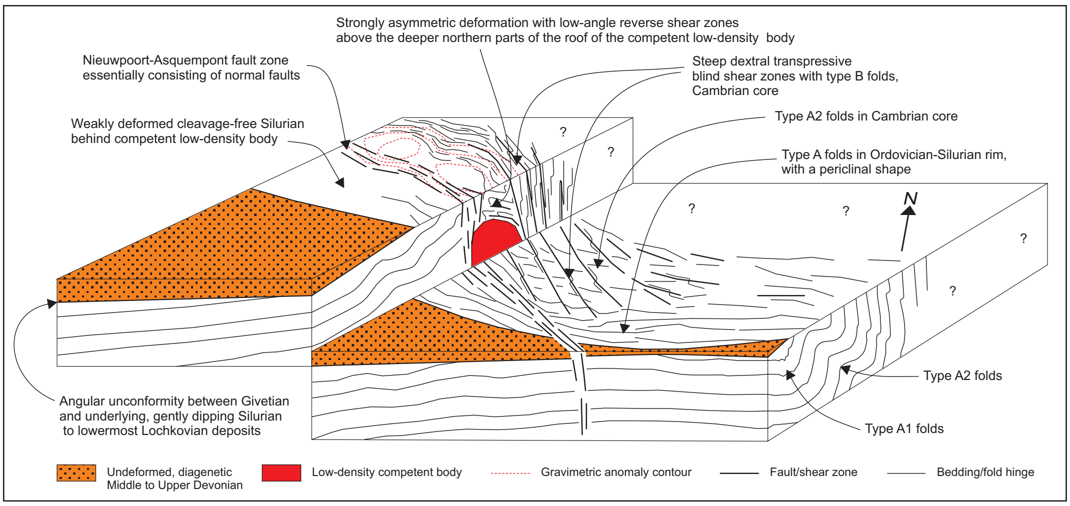

Figure 11: Conceptual block diagram of the deformation geometry and kinematics in the southern part of the Brabant Massif, highlighting the influence of the competent low-density gravimetric anomaly body and the occurrence of the type A1, type A2 and type B folds. The western part of the block diagram is shown uplifted in order to demonstrate the deformation style in the direct surroundings of the low-density gravimetric anomaly body (after Debacker, 2001 and Debacker et al., 2005a).

6. Conclusions

Despite the early pioneering works of Fourmarier (1914, 1921) and Vandenven (1967), detailed, modern analyses of folds and cleavage/fold relationships within the Brabant Massif have been executed on a systematic basis only since the last years of the 20th century. These analyses have resulted in the recognition of several types of folds of regional importance. These folds consist of pre-cleavage folds, all interpreted as slump folds, three types of syn-cleavage folds (type A1, type A2 and type B folds), formed during the progressive Brabantian Deformation event and two broad types of post-cleavage folds, being kink bands and (normal) fault-related folds.

The cartographic and stratigraphic distribution of the different types of folds has yielded significant insight into the influence of pre-existing sedimentological, lithological and deformation features on the final deformation geometry, and this on scales ranging from microscopic to basin-wide. Combined with data from other sources, these studies provided valuable information on the large-scale architecture of the Brabant Massif, and on the evolutionary history of the Brabant Basin, from initial basin development to the final stages of basin inversion and gravitational collapse.

The huge advances in our understanding of the Brabant Massif, resulting from these detailed fold and cleavage/fold relationship analyses, and from their integration with pre-existing data from other sources, serve as an illustration of the importance of detailed cleavage/fold relationship studies in fold belts and slate belts. The structural architecture, evolutionary history and tectonic significance of fold belts and slate belts cannot be understood without a thorough inventory of many detailed analyses of folds and cleavage/fold relationships.

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