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FIELD GUIDE TO THE PRE-CLEAVAGE DEFORMATION AND STRATIGRAPHY OF THE JODOIGNE AREA

Cambrian slump deformation and evidence for the Asquempont Detachment System along the N-side of the core of the Brabant Massif

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Cambrian slump deformation and evidence for the Asquempont Detachment System along the N-side of the core of the Brabant Massif

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(27 pages, 20 figures)

Cover illustration : Hinge of steeply N-plunging, E-facing fold in the Rue du Vieux Moulin at Jodoigne (stop 4, fold I10). The hammer is parallel to the fold hinge line and ripple lamination.

FIELD GUIDE TO THE PRE-CLEAVAGE DEFORMATION AND STRATIGRAPHY OF THE JODOIGNE AREA: CAMBRIAN SLUMP DEFORMATION AND EVIDENCE FOR THE ASQUEMPONT DETACHMENT SYSTEM ALONG THE N-SIDE OF THE CORE OF THE BRABANT MASSIF

Abstract. This field guide gives an insight into the Lower Palaeozoic geology of the Geete outcrop area in the surroundings of Jodoigne. Five selected outcrops or groups of outcrops are visited. These are the northern Dongelberg Quarry at Dongelberg, the Les Fosses Quarry at Opprebais (Incourt), outcrops along the Rue du Maka at Jauchelette, outcrops along the Rue du Vieux Moulin at Jodoigne and the outcrop below the town hall at Jodoigne. In each case, a detailed description is provided of the lithology, lithostratigraphy and structural architecture, followed by remarks and interpretation.

The geological observations from these field trip stops are used to illustrate lithological differences between the Blanmont Formation and the different facies of the Jodoigne Formation, to demonstrate the presence of steeply plunging and gently plunging folds, to illustrate the common occurrence of pre-cleavage folds, interpreted as slump folds, and to outline our arguments for the newly proposed stratigraphic position of the Jodoigne Formation. The cartographic proximity of the Lower Cambrian Blanmont Formation and the Middle to Upper Cambrian Jodoigne Formation is explained by means of the Asquempont Detachment System.

The observations and their implications are placed in the broader context of the Belgian Lower Palaeozoic.

Key-words: Belgium, Blanmont Formation, cleavage/fold relation, Jodoigne Formation, lithostratigraphy, Lower Palaeozoic.

1. Introduction and scope

Despite its considerable size, extending from Jodoigne in the north to Glimes in the south, the Geete outcrop area has usually been neglected by recent studies. This outcrop area, however, is the only Lower Palaeozoic outcrop area situated to the north of the central axis of the Anglo-Brabant Deformation Belt (Fig. 1), and is the only area that exposes the enigmatic, Cambrian, Jodoigne Formation, of which the stratigraphic position is highly disputed.

Two main opinions exist about the stratigraphic position of the Jodoigne Formation (Fig. 2). One group of researchers considers the Jodoigne Formation as being older than the Blanmont Formation, whereas a second group of researchers suggests a Middle to Upper Cambrian stratigraphic position. The first opinion is favoured by Dumont (1848), Malaise (1900; cf. Malaise, 1883), Kaisin (1919), de la Vallée Poussin (1931), Raynaud (1952), Mortelmans (1955, 1977), Lecompte (1957) and Verniers *et al.* (2001). The main argument for this hypothesis is the relative outcrop position within the Brabant Massif with respect to the other Cambrian formations. However, as pointed out by Michot (1980), the outcrops of the Jodoigne Formation are situated "on the northern limb of the Brabant Anticlinorium" (on the map of Legrand, 1968) and therefore should be younger than the Blanmont Formation. The second opinion, in which the Jodoigne Formation is considered as Middle to Upper Cambrian, is favoured by Malaise (1911), Fourmarier (1921), Legrand (1968), Michot (1980), Vanguestaine (1992) and De Vos *et al.* (1993). However, as correctly pointed out by Raynaud (1952), if this were the case, the magnetite-bearing Tubize Formation should be expected between the Blanmont Formation and the Jodoigne Formation in the Geete outcrop area. A magnetic field survey of Raynaud (1952) did not show magnetic anomalies between both formations, leading him to favour the first opinion. Also the most recent formal stratigraphical synthesis of the Brabant Massif (Verniers *et al.*, 2001) favours the first hypothesis, but also in this work no truly convincing arguments are put forward.

Apart from the uncertainty regarding the stratigraphic position of the Jodoigne Formation, also the structural architecture remained largely unknown until very recently. Fourmarier (1921) already pointed out the presence of steeply plunging folds within the Jodoigne Formation in the Geete outcrop area: "*Sur la rive gauche de la Geete, en face du Moulin, dans un petit chemin creux, on peut observer des plissements très aigus des couches. Les plis sont déversés vers le Sud. On voit parfaitement que tous ces petits plis ont un ennoyage vers l'Ouest, extrêmement prononcé et atteignant parfois presque la verticale*" (Fig. 3). However, although during the last decade steeply plunging, syn-cleavage folds have been documented in detail from within the Cambrian parts of both the Senne-Sennette outcrop area and the Dyle-Thyle outcrop area (Sintubin *et al.*, 1998; Debacker *et al.*, 2004a, 2005a; Piessens *et al.*, 2004), until very recently the question remained as to what extent the folds described by Fourmarier in the Geete outcrop area indeed have a syn-cleavage origin and whether the presence of these folds has regional implications. According to Fourmarier (1921) himself, for instance: "*Comme il s'agit de chiffonages très localisés, il ne faudrait pas accorder une importance trop grande à cette observation et l'ériger en règle générale pour tous le pays*".

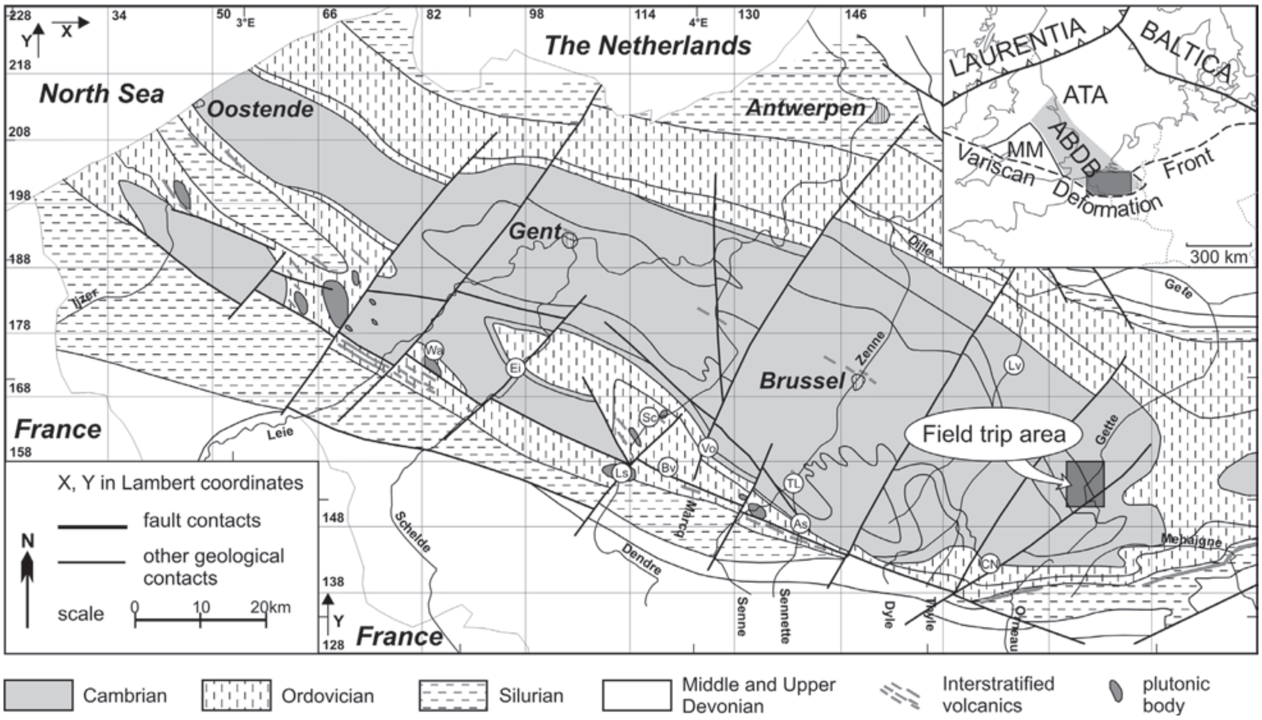

Fig. 1. Geological subcrop map of the Brabant Massif (after De Vos *et al.*, 1993 and Van Grootel *et al.*, 1997) showing the position of the visited Geete outcrop area. The upper right inset shows the position of the Brabant Massif within the Anglo-Brabant Deformation Belt (ABDB) along the NE-side of the Midlands Microcraton (MM) in the context of Avalonia (ATA), Baltica and Laurentia. Localities referred to in the text are indicated: Waregem (Wa), Eine (Ei), Lessines (Ls), Schendelbeke (Sc), Bever (Bv), Vollezele (Vo), Asquempont (As), Tubize and Lembeek (TL), Cortil-Noirmont (CN), Leuven (Lv).

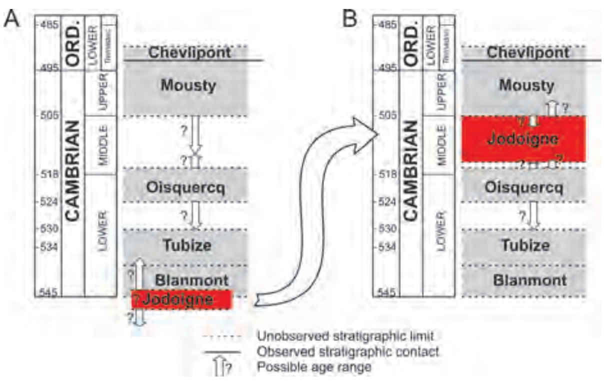

Fig. 2. Stratigraphic subdivision of the Cambrian and lowermost Ordovician of the Brabant Massif with the probable position of the Jodoigne Formation highlighted, according to (A) Verniers *et al.* (2001), and (B) according to Herbosch *et al.* (2008) as most recent work. It is the latter stratigraphic subdivision of the Cambrian that is advocated herein. Note that the possible age range of the Jodoigne Formation, as represented by the arrows, allows for an overlap with the lower part of the Mousty Formation.

During the last five years we have performed detailed fieldwork in the Geete outcrop area. This work was executed partly in the framework of the construction of the new geological map Jodoigne-Jauche at scale 1/25000 (Herbosch *et al.*, submitted), which is presented during the field trip. This detailed field work, of which the main results have been published in Debacker *et al.* (2006) and Herbosch *et al.* (2008), has resulted in a thorough revision of the Cambrian stratigraphy (see Fig. 2B), which, contrary to previous hypotheses, is backed-up by several lines of evidence, and, moreover, has resulted in an unravelling of the structural complexity of the area. These results also contributed significantly to the construction of the new, still unpublished, geological map of the Flemish part of the Brabant Massif (cf. Piessens *et al.*, 2005), presented also at the Geologica Belgica international meeting 2009 at Ghent. Furthermore, with our revised Cambrian stratigraphy, a much more straightforward link can be made between the Cambrian of the Brabant Massif and the Cambrian of the Stavelot-Venn Inlier, thus bridging the apparent gap between the Anglo-Brabant Deformation Belt and the Ardennes Inliers.

Fig. 3. Initial observations by Fourmarier (1921) of steeply plunging folds within the Jodoigne Formation at Jodoigne, at or in the vicinity of stop 4 (see text). Left: floor observation, right: section observation.

During this fieldtrip we visit outcrops of both the Blanmont Formation and the Jodoigne Formation, demonstrate the lithological differences between both formations, introduce the reader to several facies of the Jodoigne Formation, demonstrate the structural complexity of the area, show our arguments for the revised stratigraphic position of the Jodoigne Formation as proposed in Herbosch *et al.* (2008), and explain how this all fits in within the other parts of the Anglo-Brabant Deformation Belt.

2. Field trip stops

All five stops are marked on the map of Fig. 4. For all stops, detailed information is provided on lithology, lithostratigraphy, sedimentological observations and structural architecture, accompanied by personal remarks and interpretations. Much of this information has already been published (Debacker *et al.*, 2006; Herbosch *et al.*, 2008). For more detailed information and discussions, the reader is referred to Debacker *et al.* (2006), Herbosch *et al.* (2008, submitted) and references therein. Outcrop numbers are those of the archives of the Belgian Geological Survey and refer to observation points on the geological map 118W (see also Herbosch *et al.*, 2008).

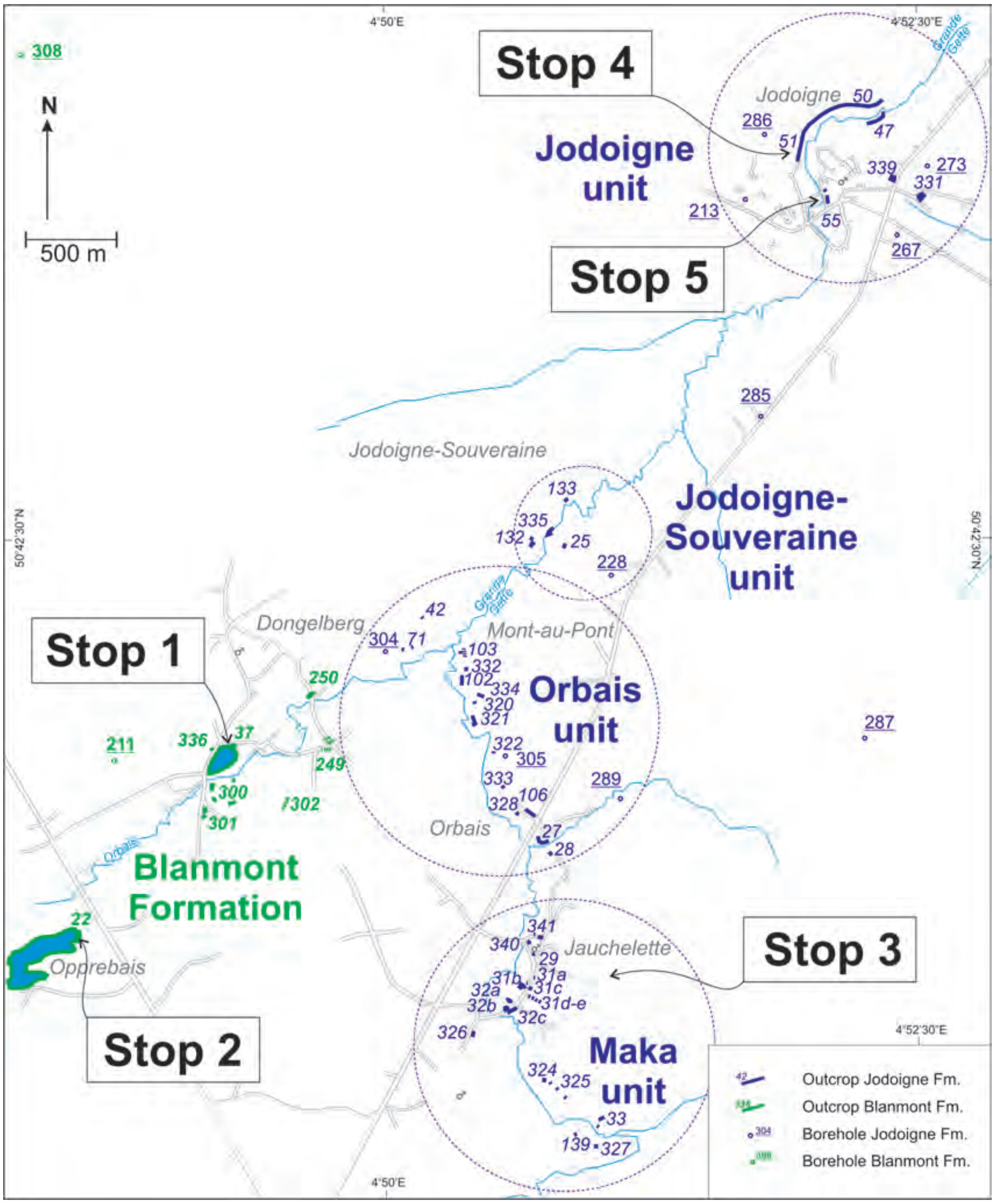

Fig. 4. Simplified topographic map of the Geete outcrop area, showing the outcrop distribution of the Blanmont Formation and of the four units of the Jodoigne Formation (after Herbosch *et al.*, 2008). Field trip stops 1 to 5 and outcrop numbers are added. The outcrop numbers are those of the archives of the Belgian Geological Survey and refer to observation points on the geological map 118W (see also Herbosch *et al.*, 2008).

2.1. Stop 1: Northern Dongelberg Quarry at Dongelberg GeoDoc n° 118W037 or outcrop 37 of Herbosch** et al. **(2008).

Lithostratigraphy and lithology

Blanmont Formation. Hard, massive, pale-coloured, greyish to greenish-grey or bluish grey quartzite. In the visited part of the outcrop (NW corner), the rock more resembles a sandstone than a quartzite. This is possibly due to weathering.

The sum of the beds present in the two Dongelberg quarries and the direct surroundings (up to ~200 m of stratigraphic thickness) has an extremely homogeneous, hard, pale-coloured quartzitic nature. Bedding is usually very difficult to observe and lithological variations are extremely vague, without obvious variations in pelitic content or very marked changes in granulometry. Neither are conglomerates or arkosic levels observed, as often mentioned in old literature (e.g. de la Vallée Poussin, 1931; Raynaud, 1952).

At the visited locality, preserved (syn-)sedimentary features include decimeter-scale elongate ball-shaped structures within the basal parts of some of the thick sandstone beds, as well as local ripple marks and current lineations. Typically, the eastern edges of the ball-shape structures is more pronounced than their western edges.

In a small abandoned quarry close to the southern Dongelberg quarry, the quartzite is characterised by decimetric beds, showing bedding-parallel and oblique stratification (private property: GeoDoc n° 118W301 or outcrop 301 of Herbosch *et al.*, 2008; see Fig. 4).

Structural architecture

Throughout the quarry, bedding has a uniform orientation, being subvertical and trending N-S to NNE-SSW (186/83W**).** Where observed, cleavage is always steeply N-dipping (260/77N). This results in a steeply plunging cleavage/bedding intersection lineation (Fig. 5). Provided folding and cleavage are cogenetic, this implies a position within the limb of a steeply plunging syn-cleavage fold (so-called type B fold of Debacker *et al.*, 2004a; see Fig. 6). Several sets of (thin) veins occur. The orientation of these veins is compatible with the inferred overall fold geometry.

Sense of younging, inferred from grading and load-casts, is towards the east.

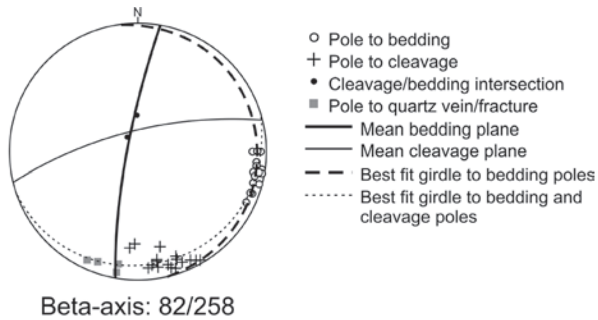

Fig. 5. Lower-hemisphere equal area projection showing bedding, cleavage and fracture data from field trip stop 1 (outcrop 37) and outcrops 300 and 301 at Dongelberg.

Remarks and interpretation

The decimeter-scale elongate ball-shaped structures within the basal parts of several of the thick sandstone beds have never been studied, and hence their origin remains uncertain. Morphologically, the structures represent load balls (pseudonodules or ball-and-pillow structures), or transition structures from load balls towards convolute bedding. As a working hypothesis, we suggest these structures to have formed by fluid pressure-assisted partial loss of sediment strength, combined with density inversion, likely as a result of overpressuring during rapid burial. In this respect, their occurrence within the basal parts of the beds, combined with their better developed western edges, are compatible with an eastward younging sense.

The beds are cross-cut by several sets of veins. The cleavage-parallel veins are considered to have formed

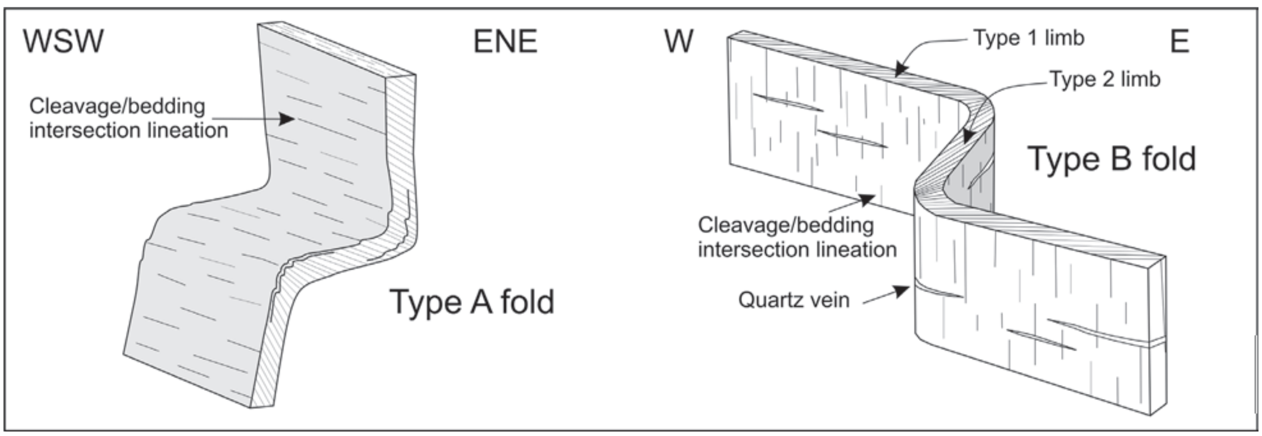

Fig. 6. The two main syn-cleavage fold types in the Brabant Massif: type A folds and type B folds (after Debacker *et al.*, 2004a).

either during or after cleavage development and were investigated in detail by Dewaele (Dewaele, 2004; Dewaele *et al.*, 2004). No alteration of the wall rock occurs around the veins. The cleavage-parallel veins consist of quartz and also contain sulphides and chlorite. Pyrite has a Co/Ni ratio > 1, suggestive of a magmatic or hydrothermal origin. Chlorite geothermometry on the Fe-rich chlorite (aphrosiderite to ripidolite), present in the vein rims, suggests temperatures in the order of 340-380°C during vein development. Two types of fluids are present within the quartz vein fluid inclusions: a syncompressional, high-temperature H_2O-CO_2-(X)-NaCl-KCl fluid, with a homogenisation temperature of 224-288°C and an apparently younger, cooler H_2O-NaCl-KCl fluid, with a homogenisation temperature of 172-189°C. Both the chlorite geothermometry and the H_2O-CO_2-(X)-NaCl-KCl fluids indicate a compressive vein formation temperature close to the maximum metamorphic temperature within the Cambrian core (~350°C), as inferred by André & Deutsch (1985). This is also compatible with the absence of an alteration halo around the veins at Dongelberg. Both types of fluids have also been encountered in compressive vein systems in younger parts of the Brabant Massif (e.g. Lower Ordovician at Marcq, Lower Ordovician at Asquempont, and in Upper Ordovician at Lessines and Bierghes). In these places, chlorite geothermometry and fluid inclusion studies suggest compressive vein development at similar temperatures as within the Cambrian core. However, in these younger units, these temperatures are often significantly higher than those to which the host rock was subjected. Dewaele (2004; cf. Dewaele *et al.*, 2004) explained this by syntectonic expulsion of hot metamorphic fluids from within the deforming Cambrian core of the massif towards the overlying younger units.

2.2. Stop 2: "Les Fosses" Quarry at Opprebais (Incourt) GeoDoc n° 118W022 or outcrop 22 of Herbosch et al. *(2008).*

Lithostratigraphy and lithology

Blanmont Formation. The outcrop part visited, situated at the eastern extremity of the quarry, consists of an alternation of decimetric to metric beds of pale-coloured quartzite and sandstone and decimetric to metric beds of laminated siltstone and darker pelites (Fig. 7).

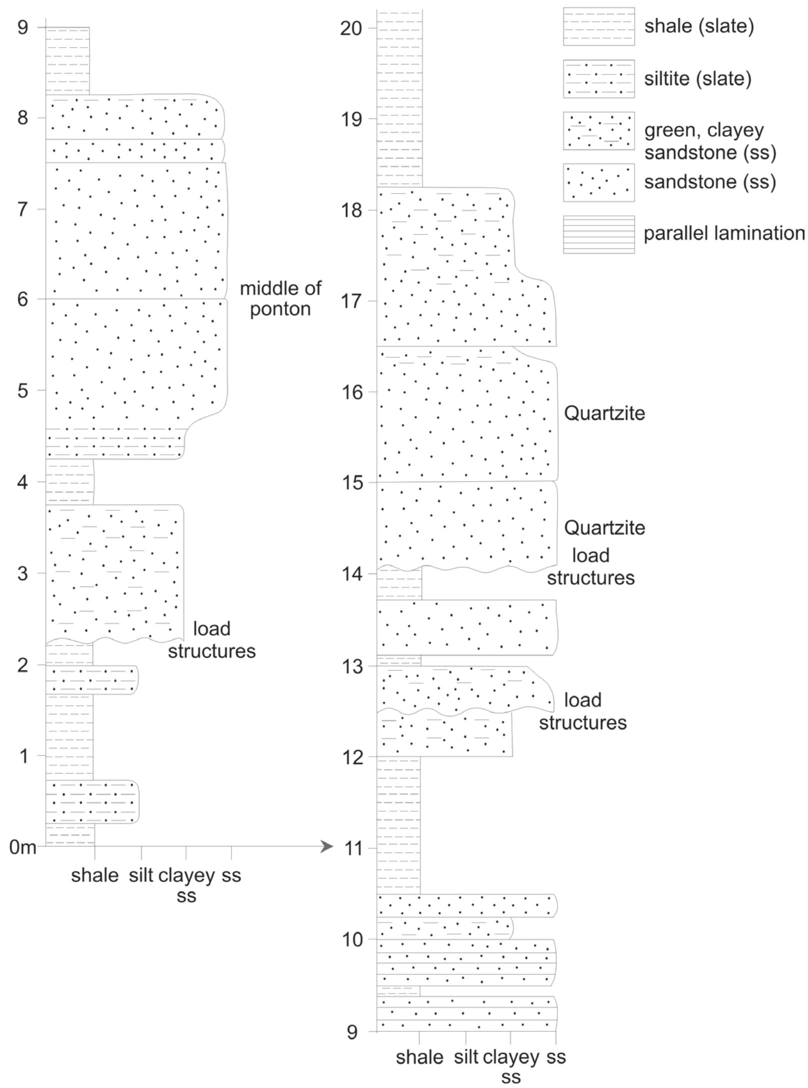

Fig. 7. Lithological log of the Blanmont formation at Opprebais, recorded at the eastern extremity of the Les Fosses quarry (stop 2; outcrop 22 of Herbosch *et al.*, 2008). Compare with Fig. 14.

The siltstone and pelites are very weathered, but when unweathered the colour is brown to grey. These fine-grained levels contain biotite, as originally described by de Magnée (1977). The sandstone, occurring in decimetric beds, ranges in colour from grey to greyish-green, depending on clay content. The greenish sandstone beds are well-stratified, whereas the thick quartzite levels usually lack any visible internal structures. Loadcasts at the base of several sandstone beds demonstrate a S-ward younging.

Throughout the quarry bedding trends E-W and dips steeply south, suggestive of an overall stratigraphic thickness of minimum 300 m. The overall lithology changes from more silty-clayey in the northern third of the quarry (the part visited) to more sandy in the southern two third parts of the quarry. This lithological change is also reflected by the top of the basement, which descends rapidly from the south (120 m) towards the north (below 100 m), thus explaining the overall asymmetry of the quarry, with steep cliffs in the southern part and sandy banks in the northern part.

Structural architecture

Bedding is steeply S-dipping (084/81S), and cleavage dips steeply to the north (259/81N). This results in a subhorizontal, E-W-trending cleavage/bedding intersection lineation (Fig. 8). Provided that cleavage development and folding occurred simultaneously, this suggests a position within the steep limb of a gently plunging syn-cleavage fold (so-called type A fold of Debacker *et al.*, 2004a; see Fig. 6). Based on the cleavage/bedding relationship, younging sense is expected to be towards the south in such a limb. This is confirmed by sedimentological observations based on loadcasts. Hence, the quarry is interpreted to be composed entirely of a steep, S-younging limb of a subhorizontal type A fold.

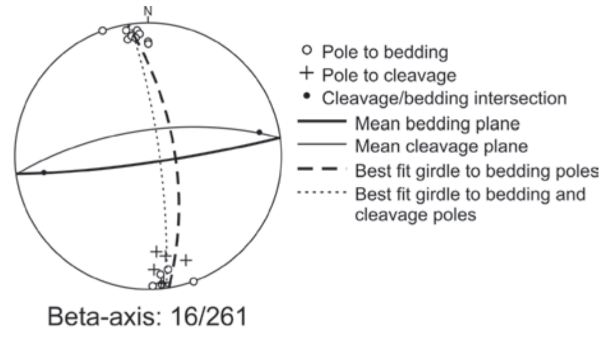

Fig. 8. Lower-hemisphere equal area projection showing bedding and cleavage data from field trip stop 2 (outcrop 22) and outcrop 250.

The finer grained parts (siltstone) show a slight crenulation of bedding, with a crenulation axis plunging gently towards the west and being situated within the cleavage plane. This is compatible with the overall cleavage/bedding intersection lineation.

Remarks and interpretation

Locally, current lineations can be observed (66/251), with an orientation subparallel to the long axis of elongated loadcasts. Hence, considering the rather straightforward structural architecture, palaeocurrent studies could be performed in this quarry. At present, no systematic analysis of palaeocurrent indicators has been performed within the Cambrian of the Anglo-Brabant Deformation Belt. Both the Dongelberg quarry and in particular the Opprebais quarry might be incorporated in such studies.

This is a classical outcrop from which samples were taken for the first radiometric dating of the Brabantian orogeny. On samples from this outcrop, an age of 450 ± 21 Ma was obtained by means of Rb-Sr whole-rock dating (Michot, 1976) and an age of 401 ± 13 Ma by means of Rb/Sr on newly formed biotite (Michot *et al.*, 1973). Both ages were considered to reflect cleavage development (André *et al.*, 1991).

The 401 ± 13 Ma Rb-Sr age on syn-tectonic metamorphic biotite (Michot *et al.*, 1973) is in agreement with the recently obtained $^{40}Ar/^{39}Ar$ ages on single-grain, newly formed syntectonic muscovite/sericite from within a shear zone in the Ordovician of the Marcq area (Dewaele *et al.*, 2002; cf. Dewaele, 2004). An $^{40}Ar/^{39}Ar$ age of 426.1 ± 0.7 Ma, corresponding to the middle Wenlock according to the time-scales of Gradstein & Ogg (1996) and Gradstein *et al.* (2004), likely reflects an early stage of hot metamorphic fluid circulation (Dewaele *et al.*, 2002). The metamorphic fluids are considered to have migrated from the Cambrian core during an early stage of the Brabantian orogeny (Dewaele, 2004; Dewaele *et al.*, 2004; cf. Piessens *et al.*, 2002). Most of the $^{40}Ar/^{39}Ar$ ages in the Marcq area fall between 419 and 412 Ma (Pridoli-Lochkovian), and are interpreted as marking the most important period of alteration, deformation and cleavage development (Dewaele *et al.*, 2002). Importantly, these ages only hold true for cleavage development in (the Ordovician of) the Marcq area and should not be extrapolated to the entire Brabant Massif (Debacker *et al.*, 2005b). For instance, at 419-412 Ma only a small overburden covered the Ronquières Formation (lower Ludlow) at the southern rim of the Brabant Massif, being insufficient to cause the anchizonal metamorphism and the formation of a well-developed cleavage. In the Marcq area, and also at Kruishoutem, also younger $^{40}Ar/^{39}Ar$ ages (401-407 Ma: Emsian) have been identified (Dewaele, 2004). Also these are interpreted as reflecting syn-tectonic metamorphic fluid circulation. The Brabantian orogeny appears to have lasted for a period of at least ~30 Ma (~430–400 Ma: ~Wenlock-Emsian; Debacker *et al.*, 2005b), and this long time-span of deformation is in complete agreement with the long time-span of metamorphic fluid circulation evidenced by the ^{40}Ar/^{39}Ar dating (~25 Ma; Dewaele *et al.*, 2002; Dewaele, 2004). The 401 ± 13 Ma Rb-Sr age obtained on biotite from the Opprebais quarry (Michot *et al.*, 1973) matches both the 419-412 and the 401-407 Ma groups of ^{40}Ar/^{39}Ar ages (see Debacker *et al.*, 2005b).

The Rb-Sr whole-rock age of 450 ± 21 Ma (Michot, 1976) is more difficult to reconcile with recent radiometric ages and with the overall deformation history of the Brabant Massif (see above; cf. Debacker *et al.*, 2005b). Considering that this whole-rock age is obtained on samples from the same quarry, but strongly differs from the single grain age, it may be possible that this age represents a mixing age, influenced both by autigenic, newly-formed mica (younger) and by inherited, sedimentary mica (older).

2.3. Stop 3: Rue du Maka at Jauchelette
GeoDoc n° 118W029 – 118W032b. Series of outcrops along the Rue du Maka, between outcrops 29 and 32b of Herbosch et al. *(2008).*

Cars are parked at the church of Jauchelette (outcrop 29, see Fig. 9). As many outcrops in the area surrounding the Rue du Maka are situated in private property, and in some outcrops bedding cannot be observed, only a few of the many outcrops will be visited in detail. These are situated in the southern part of the Rue du Maka.

A detail of the area surrounding the Rue du Maka, the proposed type locality of the Maka unit, is shown in Fig. 9.

Lithostratigraphy and lithology

Jodoigne Formation, Maka unit. An alternation of massive, pale-grey to grey quartzite and pyritic black slate, the latter with intercalated pale-grey centimetric sandstone beds. The quartzitic zones, in which bedding is very difficult to observe, have a thickness of several meters to tens of metres, and are relatively well exposed, even on hill tops. In fact, these quartzites often seem to control the local topography (e.g. outcrops 31a, 31c, 31d and 31e). By contrast, the intercalated black slate and sandstone are rarely observed in outcrop. Only in outcrops 32b and 32c, the intercalated black slate can be observed, clearly distinct from the pelitic parts of the Blanmont Formation (see stop 2).

The coarse-grained beds are only slightly different from those of the Blanmont Formation. The coarse-grained beds of the Maka unit always consist of very homogeneous, massive sandstones to quartzites, that are harder than those of the Blanmont Formation observed at Dongelberg (stop 1) and Opprebais (stop 2). Within the Maka unit, the quartzites are internally extremely homogenous, usually lacking any primary sedimentary features such as stratification, cross-bedding or grading. This makes it difficult to determine younging sense and even bedding orientation. However, locally, basal bedforms, such as load casts and undulating or bulbous surfaces can be observed. These, and the only rarely observed limits between quartzite, sandstone and black slate, are the only convincing bedding indicators within the Maka unit.

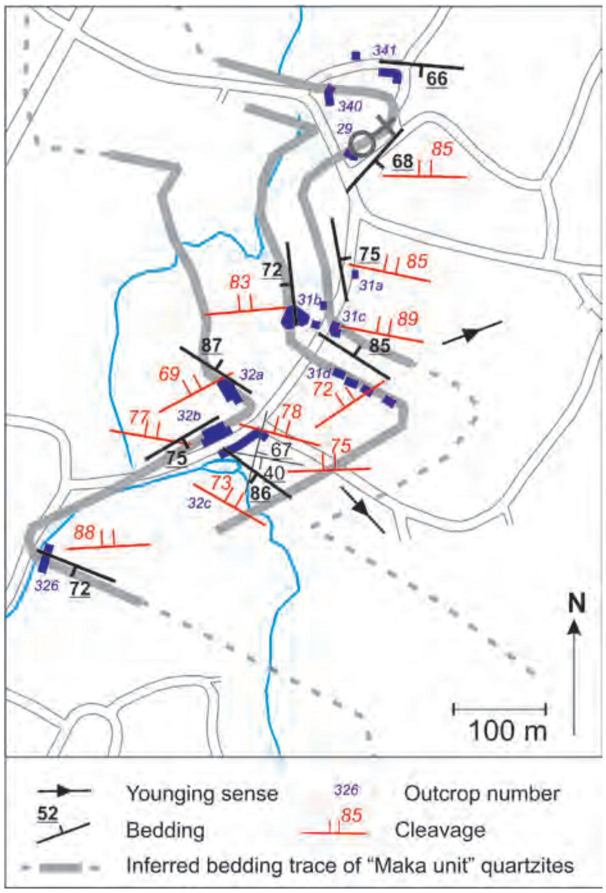

Fig. 9. Simplified topographic map of the area surrounding the Rue de Maka, type locality of the Maka unit (stop 3). Orientation of bedding and cleavage is shown, as well as local younging sense, inferred bedding traces and the inferred lateral continuation of the quartzite beds of the Maka unit.

Structural architecture

In the massive quartzites it is usually extremely difficult to find bedding. In outcrop 29, we interpreted a steeply SE-dipping foliation as bedding, and a steeply N-dipping foliation as cleavage. In outcrop 31a, inferred bedding dips steeply ENE, whereas a second foliation, interpreted as cleavage, dips steeply north. In outcrop 31c, inferred bedding is steeply NNE-dipping, whereas a second foliation, interpreted as cleavage, again dips steeply north. In all three cases the cleavage/bedding intersection lineation plunges steeply towards the SE. Provided folding and cleavage are cogenetic, this implies a position within the limb of a steeply plunging syn-cleavage fold (so-called type B fold of Debacker *et al.*, 2004a; see Fig. 6). Also the best fit girdle to the bedding poles points to a fold axis subparallel to the cleavage/bedding intersection, thus supporting the presence of type B folds (Fig. 10B). In the massive quartzites of outcrops 31d-e, situated in a meadow, no bedding was observed and inferred cleavage dips steeply towards the NW. Judging from bedding orientation in nearby outcrops 31c and 32a, the alignment of the different quartzite exposures of outcrop 31d likely corresponds to bedding, again suggestive of a steeply plunging cleavage/bedding intersection. A steeply plunging cleavage/bedding intersection lineation is also present at outcrop 31b.

Outcrops 32a to 32c are visited in detail. In the massive quartzites of outcrop 32a, bedding is steeply NNE-dipping, whereas cleavage dips steeply to the NW. This disposition is identical to that inferred for outcrop 31d-e. Within the massive quartzites of outcrop 32b, cleavage is steeply N-dipping, but bedding could not be observed. Within the W-side of the sunken road, however, a steeply SE-dipping stratigraphic contact can be observed between the massive quartzites and black shales. Again this points to a steeply plunging intersection lineation, which may reflect the presence of type B folds. Indeed, the best fit girdle to the bedding poles points to a fold axis subparallel to the cleavage/bedding intersection lineation (Fig. 10A), thus supporting the presence of type B folds. Likely, the quartzites at 32a are the same as those at 32b that are folded by a large-scale type B fold, with outcrop 32a situated in the type 1 limb and outcrop 32 b situated in the type 2 limb. Outcrop 32c is dominated by black shale, with interstratified silt- and sandstone beds. Bedding orientation, however, is highly variable, often at high angles to bedding in outcrop 32b, and locally folds can be observed. The intersection lineation within these fine-grained units varies from gently plunging to steeply plunging. Towards the east, again massive quartzites occur, but the limit with the black shales is not exposed.

In outcrop 326, which is not visited during the fieldtrip, steeply SSW-dipping bedding and steeply N-dipping cleavage result in a gently plunging intersection lineation. Provided folding and cleavage development are cogenetic, this suggests the presence of a gently plunging fold or so-called type A fold (cf. Fig. 6).

Within the quartzites of the Maka unit, at least three sets of quartz veins can be recognised. One set is perpendicular to cleavage, one set is subperpendicular to the cleavage/bedding intersection and inferred fold hinge lines, and one set is subparallel to cleavage (Fig. 10).

Remarks and interpretation

The data suggest that the Maka unit in the surroundings of the Rue du Maka is deformed by a system of steeply plunging, E-facing folds. The plunge of these type B folds is variable. In the SW (Fig. 10A), for instance, the folds plunge steeply NW, whereas in the NE they plunge steeply to the SE (Fig. 10B).

The variable bedding orientation in outcrop 32c, often oblique to that of the surrounding beds, combined with the strongly variable cleavage/bedding intersection orientation and the presence of small folds, is difficult to link to the type B folds. It may be possible that the black shales in outcrop 32c were affected by slumping and that these slumped levels were later folded by the large-scale type B folds.

The quartz veins of the Maka unit have never been studied in detail, neither in terms of geometry and kinematics, neither by means of geochemistry and fluid inclusion studies. A systematic geometrical analysis of the different vein sets would offer more insight into the complex structural architecture of the Maka unit in the area surrounding the Rue du Maka. Moreover, fluid inclusion studies, combined with a systematic geometrical analysis of the veins, may give an insight into the deformation history experienced by these quartzites.

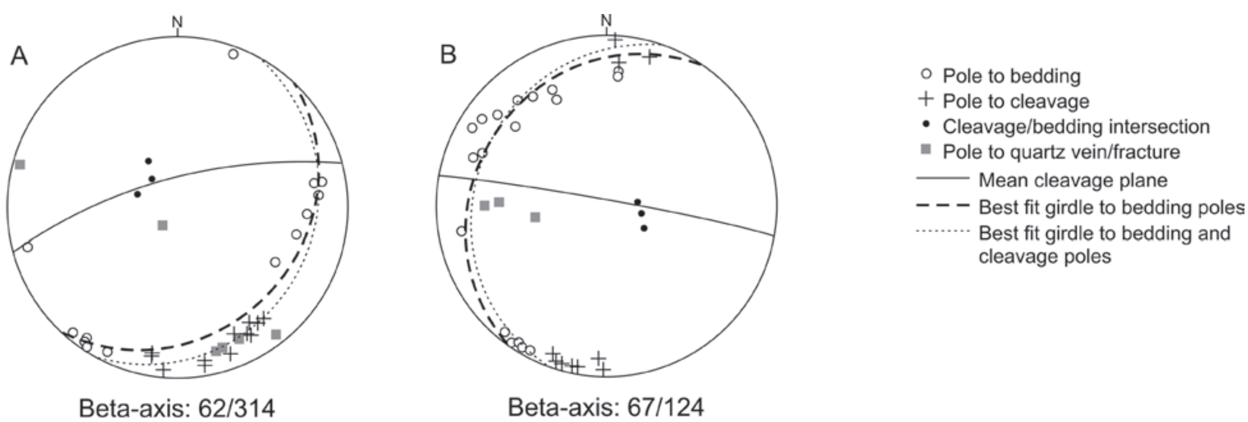

Fig. 10. Lower-hemisphere equal area projections showing bedding, cleavage and fracture data from the Maka unit outcrop area, visited during field trip stop 3. A) outcrops 33, 327, 32a-b, 31b; B) outcrops 31a, 31c, 29 and 341.

Sedimentologically, it is possible that the black shales and intercalated cm-thick sandstone beds have a turbiditic origin. If so, the adjacent massive quartzites might correspond to turbidite channels. However, this is merely a working hypothesis that should be checked by future work on better preserved outcrops or cores.

2.4. Stop 4: Rue du Vieux Moulin at Jodoigne GeoDoc n° 118W051, outcrop 51 of Herbosch et al. *(2008) and outcrop I in Debacker* et al. *(2006).*

Cars are parked at the large parking space directly west of the Town Hall of Jodoigne (outcrop 55 of Herbosch *et al.*, 2008 and stop 5 of this guide).

Lithostratigraphy and lithology

Jodoigne Formation, Jodoigne unit. Very dark grey to black mudstone sequences, with intercalated pale grey to greenish grey siltstone and sandstone beds, often rich in pyrite, with a thickness ranging from a few millimetres to three decimetres. The sandstone and siltstone beds are usually cross-bedded, with the thicker beds often containing convolutions and showing a well-developed ripple lamination. Current ripples and convolution axes are sub-parallel to the hinge lines of a system of metre-scale, steeply plunging folds. The southern part of the outcrop is almost entirely composed of greenish grey, decimetric, cross-bedded sandstones, whereas the northern part is dominated by a mudstone sequence with a few intercalated sandstone beds. Some of the latter sandstone beds are dark grey to black, a facies regarded typical for the Jodoigne Formation (cf. Verniers *et al.*, 2001).

Within the mudstone-dominated northern part of the outcrop, a zone of rather large (up to 5 decimetres long and 2-3 decimetres wide), isolated, lens-shaped sandstone fragments occurs, apparently being truncated and displaced by detachments oriented sub-parallel to bedding.

Structural architecture

The outcrop is dominated by a decametre-scale, steeply N-plunging fold, with a step-fold geometry, a steeply N-dipping axial surface and a close interlimb angle (fold I10; Fig. 11, Fig. 12). In the southern fold limb, two smaller steeply N-plunging folds are present, having gentle to open interlimb angles and an S-shaped asymmetry (folds I11 & I12). The cleavage/bedding intersection lineation is parallel to the fold hinge lines. Although difficult to see in these cross-bedded sandstones, cleavage is sub-parallel to the fold axial surfaces, and shows a well-developed divergent cleavage fanning, with a symmetrical disposition about the fold hinges (Fig. 12). All three folds face towards the E (younging sense).

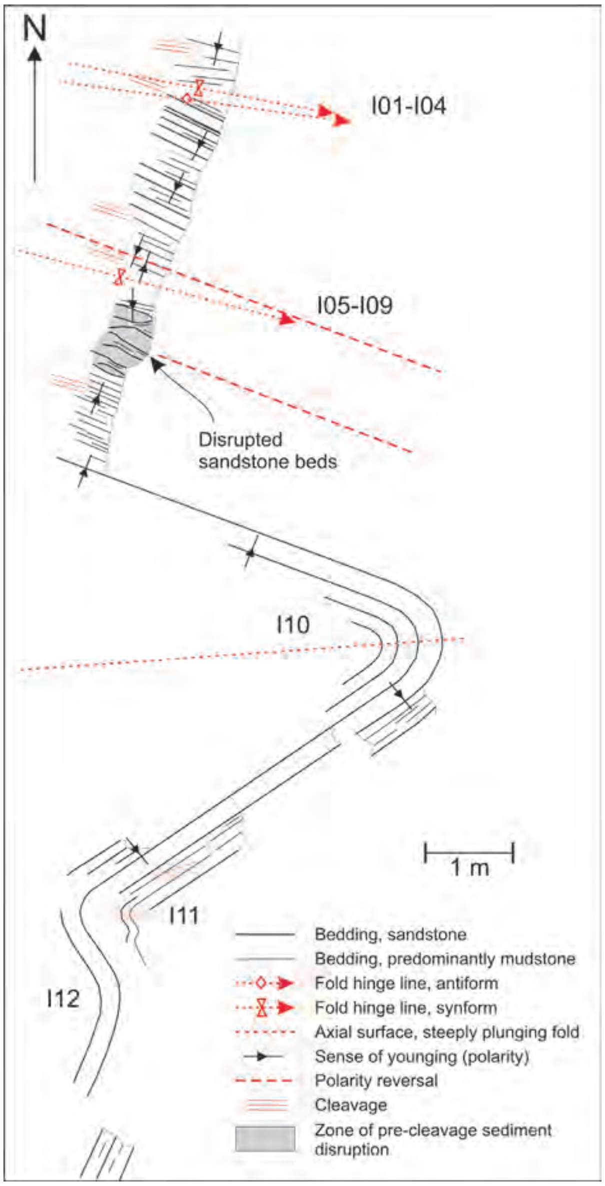

Fig. 11. Map view of outcrop 51 (field trip stop 4), showing bedding, cleavage, fold hinge lines, axial surfaces, younging sense, and zones of pre-cleavage sediment disruption (after Debacker *et al.*, 2006). Note the presence of sub-horizontal to gently plunging folds (folds I01 to I09), within the northern limb of a steeply plunging fold (fold I10).

In the more pelitic northern part of the outcrop, in the northern limb of the steeply plunging fold (I10), two zones occur containing centimetre- to metre-scale, gently to moderately plunging folds (folds I1 to I4 and folds I5 to I9; Fig. 11, Fig. 13). These folds have close to tight interlimb angles and steeply NNE-dipping, slightly curving axial surfaces. The cleavage, being at a very low angle to the mean, steeply dipping bedding, commonly shows a divergent cleavage fanning symmetrical about the fold hinges, and has an opposing sense of cleavage refraction on opposite fold limbs. In addition, cleavage only slightly transects the folds, with a transection angle that may vary from fold to fold. However, despite this seemingly almost perfect axial planar cleavage/fold relationship, the cleavage/bedding intersection lineation is strongly variable (Fig. 13). In the more or less uniform, steeply N-dipping beds in between these two fold zones, the intersection is steeply plunging, whereas in the fold zones, the intersection varies from steeply plunging to sub-horizontal. Sub-horizontal to moderately plunging intersections are only observed where a relatively large angle exists between bedding and cleavage: in the hinge zones, and in places where bedding is sub-vertical to steeply S-dipping. The sense of younging is downwards in the southern, synformal, fold zone and upwards to the south in the northern antiform-synform fold pair zone. Hence, the overall younging direction of the northern part of the northern limb opposes that of the large steeply plunging fold to the south (I10). This reversal in polarity occurs around the zone of isolated, lens-shaped sandstone fragments.

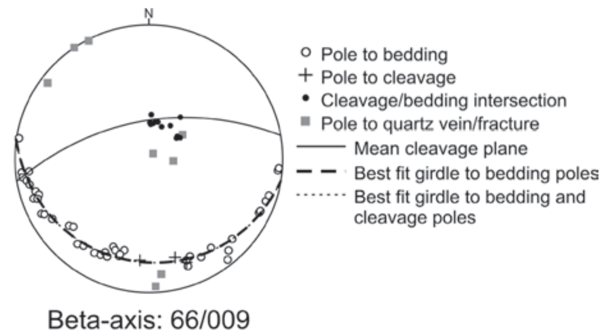

Fig. 12. Lower-hemisphere equal area projection showing bedding, cleavage and fracture data from field trip stop 4 (outcrop 51). Only bedding and cleavage data associated with the steeply plunging folds (I10-I12) are incorporated (see Fig.11 for position). For detailed projections of data related to the small-scale, gently plunging folds, the reader is referred to Debacker *et al.* (2006).

Two quartz vein sets occur. A first set consists of sub-vertical veins, of which the intersection with bedding is sub-parallel to the fold hinge line and to the cleavage/bedding intersection lineation of the steeply plunging folds. Often, these veins are lined by pyrite. Around the steeply plunging folds, vein orientation remains sub-perpendicular to bedding, suggesting a pre- or syn-folding origin. In the NW-SE-trending northern fold limb, the orientation of these veins remains the same, irrespective of the younging sense of the beds. The second set, cross-cutting the first, consists of sub-horizontal to gently SW-dipping veins, oriented sub-perpendicular to the fold hinge lines and cleavage/bedding intersection lineation of the steeply plunging folds. Vein orientation remains constant across the steeply plunging folds. These veins reflect extension along the fold axis, and probably have a late syn-folding origin. In the NW-SE-trending northern fold limb, this set is locally slightly folded, and gives rise to bone-shaped structures, thus reflecting a shortening at high angles to bedding after vein development. In the NW-SE-trending northern fold limb, vein orientation remains constant, irrespective of the younging sense of the beds. The two vein sets do not show any geometrical relationship with the gently plunging folds (folds I01-I09).

Remarks and interpretation

It is at, or in the direct vicinity of this outcrop that Fourmarier (1921) first described steeply plunging folds (Fig. 3). Unfortunately, Fourmarier (1921) never paid further attention to these folds.

The marked divergent cleavage fanning, symmetrical about the fold hinge, combined with the parallelism between the cleavage/bedding intersection lineation and the fold hinge lines and the geometrical relationship with the two vein sets suggests that the steeply plunging folds have a tectonic, syn-cleavage origin (Fig. 11). Similar steeply plunging tectonic folds have been described in the more western Cambrian outcrop areas of the Brabant Massif, where they are termed type B folds (Lembeek fold type of Sintubin, 1997, 1999 and Sintubin *et al.*, 1998; see also Debacker *et al.*, 2004a, 2005a; Piessens *et al.*, 2004). Generally, these folds have quite consistent limb orientations, one having an E-W- to NW-SE-trend, referred to as type 1 limb, and the other having a NE-SW- to N-S-trend, referred to as type 2 limb (see Fig. 6). Because of the geometrical similarities, also the steeply plunging folds observed here are considered as type B folds. The northern part of the outcrop represents a type 1 limb, and the southern part represents a type 2 limb of type B fold I10 (Fig. 11). Hence, this outcrop is one of the very few permanent outcrops of the Brabant Massif in which the hinge zones of mesoscale, steeply plunging syn-cleavage folds (so-called type B folds; Debacker *et al.*, 2004a) can be observed.

Vein set 1 formed prior to or during folding. For a pre-folding origin, vertical compaction during burial is a possible cause, whereas for a syn-folding origin, extension in the outer part (“above” the neutral surface) of a longitudinal strain fold might be invoked (e.g. Ramsay & Huber, 1987). Considering that these veins are present also in the straight limbs, the first possibility seems more likely. Vein set 2 probably has a late syn-folding origin. Considering the low dips, these veins likely result from hydraulic fracturing, when, during folding, fluid pressures exceeded lithostatic pressure. The fact that within the NW-SE-trending fold limb vein set 2 is locally folded, implies shortening after vein development, at high angles to bedding. Considering the late syn-folding origin of this vein set, this reflects a post-buckle flattening of the NW-SE-trending type 1 fold limb of at least 20%. This flattening may also partly be responsible for the extremely small angle between cleavage and bedding in the more pelitic units in the northern part of the outcrop.

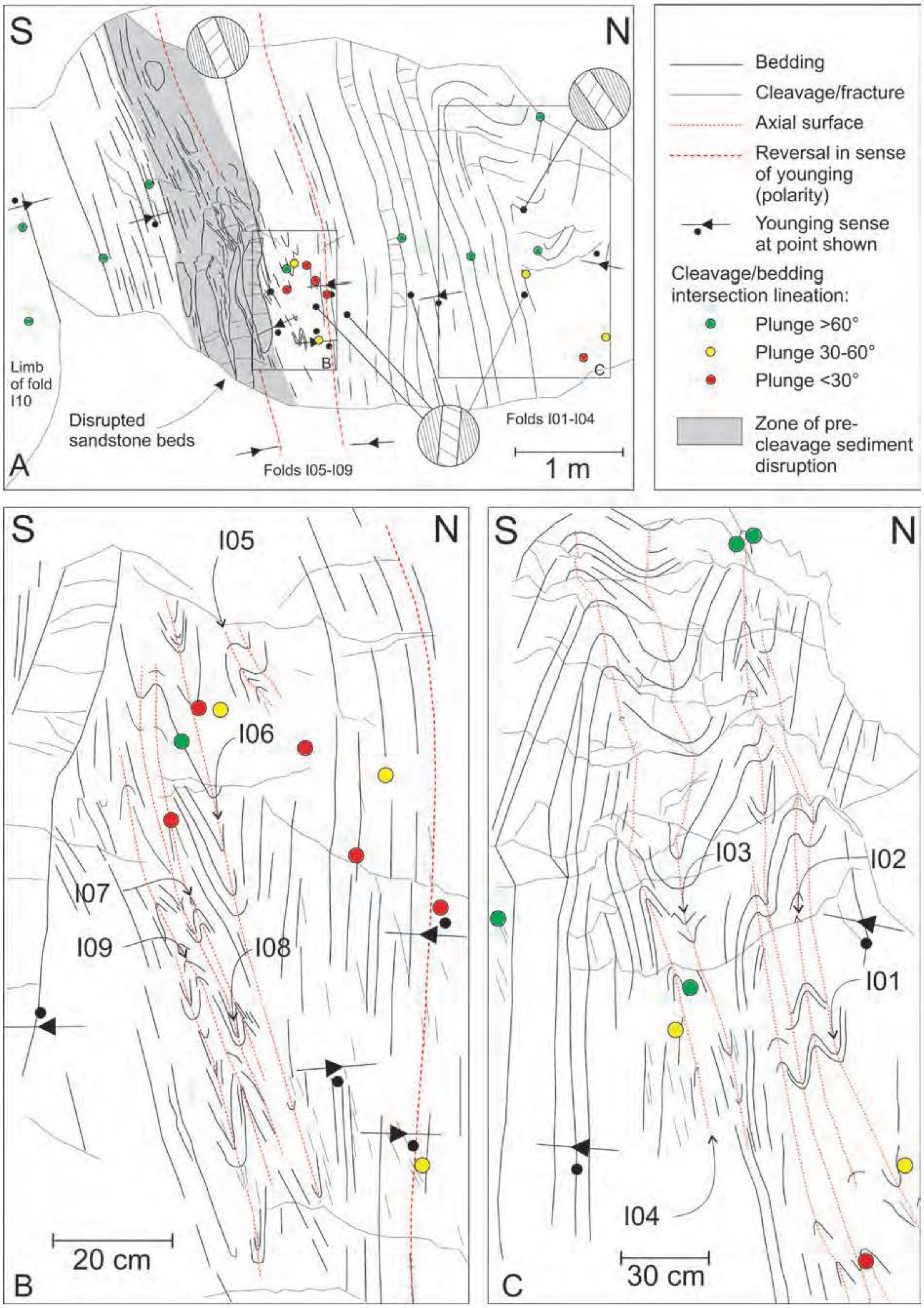

Fig. 13. Profile view of the northern part of outcrop 51 (field trip stop 4), showing the sub-horizontal to gently plunging folds (folds I01 to I09), the sense of younging, and the cleavage/bedding relationship within the northern limb of steeply plunging fold I10 (A), together with detailed profile views of fold trains I05-I09 (B) and I01-I04 (C) (adapted after Debacker *et al.*, 2006). Note the truncation of bedding by the overlying sandstone bed to the south of I09, and the bedding-parallel, probably welded nature (not recognisable in outcrop) of the inferred polarity reversal zone.

As cleavage shows a divergent cleavage fanning, symmetrical about the fold hinges of the tight to close gently to moderately plunging folds (I01-I09), and only slightly transects these folds, a syn-cleavage origin could be considered also for the gently to moderately plunging folds. However, across these folds, the cleavage/bedding intersection lineation strongly varies: within the fold hinges, the intersection lineation is sub-parallel to the fold hinge lines, but within the limbs, the intersection lineation may change over short distances from gently plunging to steeply plunging (***Fig. 13***). This change entirely depends on the orientation of bedding with respect to the N-dipping cleavage. Being situated within the type 1 limb of a large type B fold, steep intersection lineations (and fold hinge lines) would be expected. Taking into account the possibility of curvilinear hinge lines, changing N-ward from steeply plunging type B folds to gently plunging (type A) folds, the cleavage/bedding intersection lineation would be expected to reflect this change rather progressively, and not reflect the fold hinge line orientation only in the hinge of the gently plunging folds (e.g. see Holdsworth *et al.*, 2002; Debacker *et al.*, 2004a, 2005a). In addition, in the assumption of curvilinear hinge lines, a change in fold plunge well exceeding 90° would be necessary, instead of the "easier" ~40-50° (from ~60-70°N to ~50-30°E), in order to match the asymmetry of the gently plunging folds ("s-asymmetry") to that expected within the type 1 limb of the type B fold ("z-asymmetry"). Finally, within this northern outcrop part, the orientation of the two vein sets remains unmodified, and cannot be related to the gently to moderately plunging folds. All these observations imply that these gently to moderately plunging folds within the type 1 limb of the type B fold in fact pre-date cleavage, cleavage-related folding and vein development.

The cleavage/bedding disposition remains virtually identical across the polarity reversal zones, implying overturning prior to cleavage development. In addition, current ripples and convolution axes in the southern, N-ward younging part of the type 1 limb of the type B fold (I10) are sub-parallel to those in the northern, S-ward younging part of this limb. This indicates that the axis of overturning was either sub-parallel or sub-perpendicular to the ripple marks and convolution axes. As the type B folds are cogenetic with cleavage development, they cannot be held responsible for the change in polarity. In contrast, considering their local downward-facing nature and their pre-cleavage origin, the gently to moderately plunging folds, being oriented at high angles to the ripple marks and convolution axes, are likely candidates. Hence, the sudden changes in polarity are considered to occur around pre-cleavage detachments that are intimately related to the sub-horizontal to gently plunging pre-cleavage folds. Also the pre-cleavage disruptions and truncations of the sandstone beds are likely related to this deformation event.

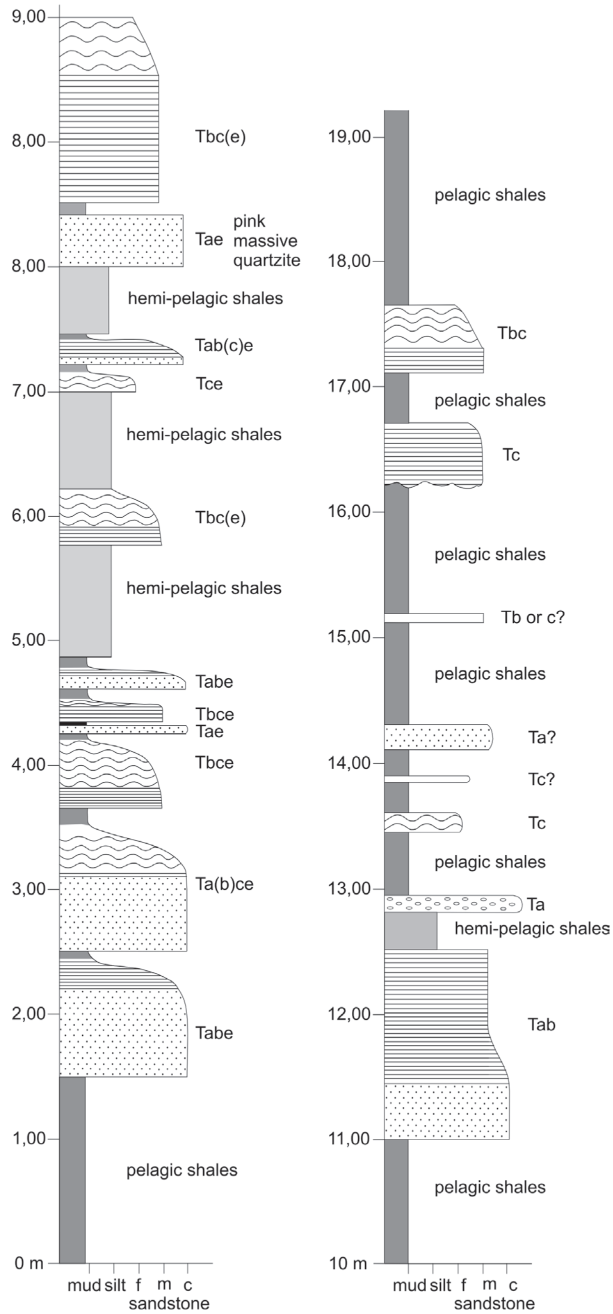

Fig. 14. Lithological log of the Jodoigne unit of the Jodoigne Formation at Jodoigne, recorded at outcrop 47 of Herbosch *et al.* (2008). Tabcde implies a turbidite sequence consisting of an a, b, c, d and e interval of Bouma (1962). F, m and c sandstone refer to fine, medium and coarse sandstone, respectively, and usually correspond respectively to c, b and a turbidite intervals of Bouma (1962). Compare with Fig. 7.

Considering the largely bedding-parallel disposition of the pre-cleavage folds, their rather isolated occurrence, being surrounded by relatively undeformed zones, and their occurrence in the direct vicinity of pre-cleavage detachments, pre-cleavage brecciation zones and zones containing disrupted and truncated sandstone beds and sandstone lenses, these folds, and the sudden changes in younging sense are attributed to slumping (cf. Debacker *et al.*, 2001 and references therein). The atypical, near-coincidence of cleavage and slump fold axial surfaces is a result of the extremely small angle between cleavage and bedding. This in turn, is due to a position within the (type 1) limb of a type B fold (Fig. 18), possibly combined with the influence of post-buckle flattening.

Sedimentologically, the convolute to obliquely stratified sandstone beds may be interpreted as being deposited by high-density turbidity currents, whereas the centimetric to millimetric siltstone beds within the black shales may be interpreted as belonging to low-density turbidites. The black pyritic shale likely corresponds to the normal, pelagic sedimentation, in between the turbidity events. The presence of pyrite and the high concentration of organic matter point to an anoxic environment. This type of environment occurs from the continental slope to the basin plain. The common presence of slumps (this stop and stop 5) is compatible with the presence of a slope and slope deposits, and hence complies with the inferred sedimentation environment.

All outcrops of the Jodoigne unit of the Joidoigne Formation show this presence of high- and low-density turbidite deposits, alternating with black pelagic to hemi-pelagic shales. Their appearance and relative abundance, however, is extremely variable. They may occur isolated within hemi-pelagic black shales (stops 4 and 5), or as thicker, volumetrically more important beds, alternating with pelagic or hemi-pelagic black shales (GeoDoc n° 118W047, outcrop 47 of Herbosch *et al.*, 2008; Herbosch, unpublished) (Fig. 14), or as ochre-coloured (i.e. oxygenated) rhythmic sequences of several tens of metres separated by relatively thin black shale horizons (GeoDoc n° 118W331; temporary outcrop 331 on Fig. 4; Herbosch, unpublished).

2.5. Stop 5: Town hall at Jodoigne
GeoDoc n° 118W055, outcrop 55 of Herbosch et al. (2008) and outcrop II in Debacker et al. (2006)

Lithostratigraphy and lithology

Jodoigne Formation, Jodoigne unit. Very dark grey to black mudstone sequences, with intercalated pale grey to greenish grey siltstone and sandstone beds, often rich in pyrite. This outcrop, however, is dominated by black mudstone, with relatively thin siltstone and sandstone beds as compared to stop 4. The sandstone and siltstone beds, usually cross-bedded, have a thickness ranging from a few millimetres to ~one decimetre and the thicker, convoluted beds seem restricted to the southernmost and northernmost parts of the outcrop. Where observed, current ripples and convolution axes are sub-parallel to the hinge lines of steeply plunging folds.

Zones occur in which the mudstone sequences contain isolated, lens-shaped fragments of disrupted siltstone and sandstone beds floating in a mudstone matrix, apparently being truncated and displaced by detachments oriented sub-parallel to bedding (Fig. 15). These lenses, which are generally smaller than in stop 4 (usually no more than 10 centimetres long and a few centimetres wide), are sometimes rotated with respect to the surrounding beds, and are associated occasionally with gently plunging, truncated folds (e.g. folds II11 and II07). At several places, these isolated lenses are observed to be cross-cut by cleavage, demonstrating their pre-cleavage origin. Also zones of pre-cleavage brecciation are recognised. The most important pre-cleavage deformation zone occurs in the central to northern part of the outcrop.

Structural architecture

The outcrop is dominated by a fold train of steeply plunging, decimetre- to metre-scale folds, with a Z-shaped fold asymmetry (Fig. 15, Fig. 16). In the northernmost part of this outcrop (folds II01 to II05) these folds have tight to close interlimb angles, steeply NE-dipping axial surfaces and steeply to moderately E-plunging fold hinge lines. The plunge of the fold hinge lines varies along strike (up to 30°), apparently reflecting curvilinear folds (see folds II04a,b,c and II05a,b,c,d). Towards the south, fold plunge decreases significantly. Despite this variation in fold plunge, cleavage is sub-parallel to only slightly oblique to the axial surfaces, and shows a divergent cleavage fanning, symmetrical about the fold hinges. The individual folds only show a small axial cleavage transection angle (< 020°), of which the sense varies between different folds. Overall, folds in this northernmost outcrop part have a small anticlockwise axial cleavage transection (013°), as well as a mean cleavage that is slightly anticlockwise (011°) with respect to the mean fold axial surface. However, the cleavage/bedding intersection lineation is sub-parallel to the fold hinge lines across the folds only in the northernmost folds (folds II01 and II02). Going towards the south, the plunge of the cleavage/bedding intersection lineation becomes highly variable (Fig. 15). Across the moderately to gently plunging folds II04 and II05, the cleavage/bedding intersection lineation shows gentle to moderate plunges in the fold hinges, but steep plunges in the steep limbs. Within this northernmost part of the outcrop, the folds face towards the E (younging sense).

Within the fine-grained zone to the south of fold II05, pre-cleavage and pre-lithification sediment deformation features occur. Within this zone, the overall NW-SE-trending bedding, essentially reflecting the southern limb of fold II05, shows a significant change in trend (zone labelled II06; Fig. 15, Fig. 17). This change in trend is cross-cut by cleavage, thus demonstrating its pre-cleavage origin. Going towards the SE, the NW-SE-trending, steeply to moderately NE-dipping beds contain several close (e.g. II07, II12) to tight (e.g. II09, II11), moderately (II08, II09a) to gently (II07, II09b,

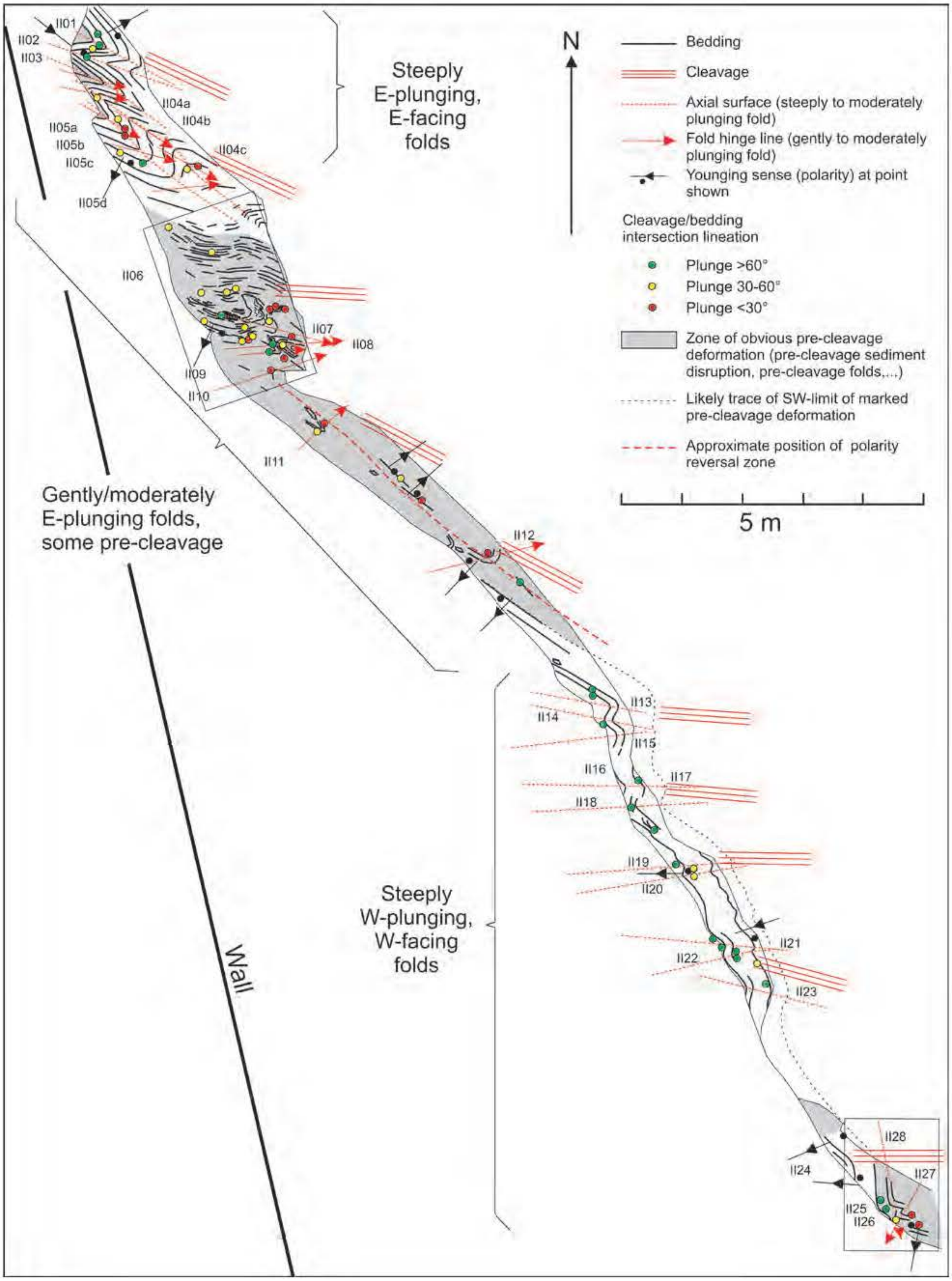

Fig. 15. Map view of outcrop 55 (field trip stop 5), showing bedding, cleavage, fold hinge lines, axial surfaces, cleavage/bedding intersection lineation plunges, younging sense, and zones of pre-cleavage sediment deformation (adapted after Debacker *et al.*, 2006). Note the occurrence of zones of sub-horizontal to moderately plunging folds (folds II03 to II12 and II27, II28), in between zones of steeply plunging folds. Details of insets are shown in Fig. 17.

II10, II11, II12) E- to NE-plunging folds, some of which appear to occur as isolated lenses in the mudstone matrix (II11). Of all these folds, only fold II12 has a large axial cleavage transection angle and a relatively large angle between its axial surface and the cleavage. However, as suggested by the variation of the cleavage/bedding intersection lineation, folds II7 and II8 also have an anomalous cleavage/fold relationship (Fig. 17). Within this zone of NW-SE-trending bedding, younging sense changes from NE-younging in the NW, to SW-younging in the SE. This polarity reversal appears to occur in the vicinity of fold II12.

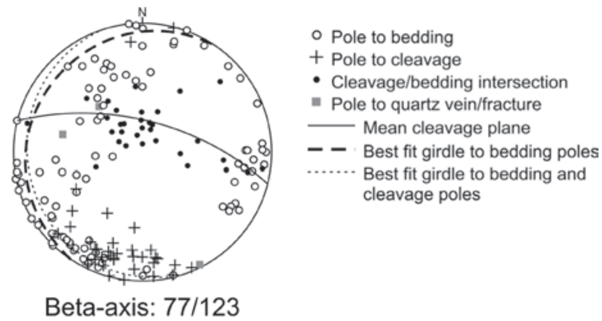

Fig. 16. Lower-hemisphere equal area projection showing bedding, cleavage and fracture data from field trip stop 5 (outcrop 55). Although data from suspected slump folds have not been incorporated, the cleavage/bedding intersection lineation still shows a lot of variation. For more detailed stereographic projection analyses of the data of this outcrop, the reader is referred to Debacker *et al.* (2006).

In the southern outcrop part the folds (II13 to II26) have open to gentle interlimb angles, steeply N-dipping axial surfaces and steeply W-plunging fold hinge lines and younging sense is towards the W (Fig. 15). However, despite these differences, the relationship with cleavage is quite similar to that in the northernmost part of the outcrop (e.g. folds II01, II02). The cleavage/bedding intersection lineation is sub-parallel to the fold hinge lines and cleavage is sub-parallel to only slightly oblique to the fold axial surfaces, shows a well-developed divergent cleavage fanning symmetrical about the fold hinges, and shows contrasting refraction patterns on opposite fold limbs. Similarly, the individual folds show a small axial cleavage transection angle (< 020°), of which the sense varies in between the different folds, but the overall axial cleavage transection (004° anticlockwise) as well as the angular difference between cleavage and the mean fold axial surface (004° clockwise) can be neglected.

In the southernmost outcrop part, two gently S-plunging folds occur (II27 and II28), with a moderately E-dipping axial surface. These folds are cross-cut at high angles by cleavage, and hence have a pre-cleavage origin (Fig. 17).

Although, because of the predominantly finer lithology, much less well developed, the same two quartz vein sets are recognised as in stop 4. Also here, a similar relationship exists between these two vein sets and the steeply plunging folds, whereas these vein sets do not show any geometrical relationship with the gently plunging folds (folds II04-II12, II27, II28).

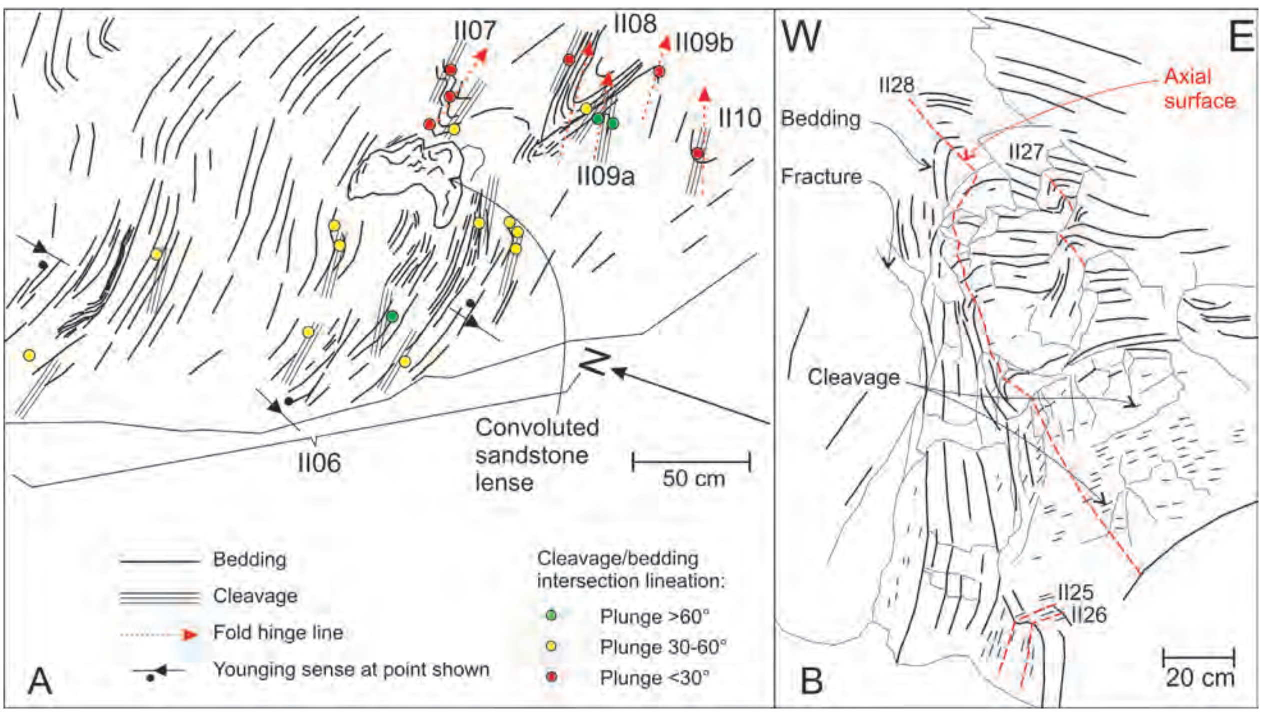

Fig. 17. Details of outcrop 55 (modified after Debacker *et al.*, 2006; see also Fig. 15). A) Plan view of the pre-cleavage deformation zone in the northern outcrop part, showing bedding, cleavage, fold hinge lines, axial surfaces, sense of younging and the plunge of the cleavage/bedding intersection lineation. From W to E, the trend of the bedding relative to that of the cleavage changes from clockwise to anticlockwise, thus demonstrating the pre-cleavage nature of the change in bedding orientation. B) Pre-cleavage folds II27 and II28, and their relationship with cleavage (southernmost outcrop part).

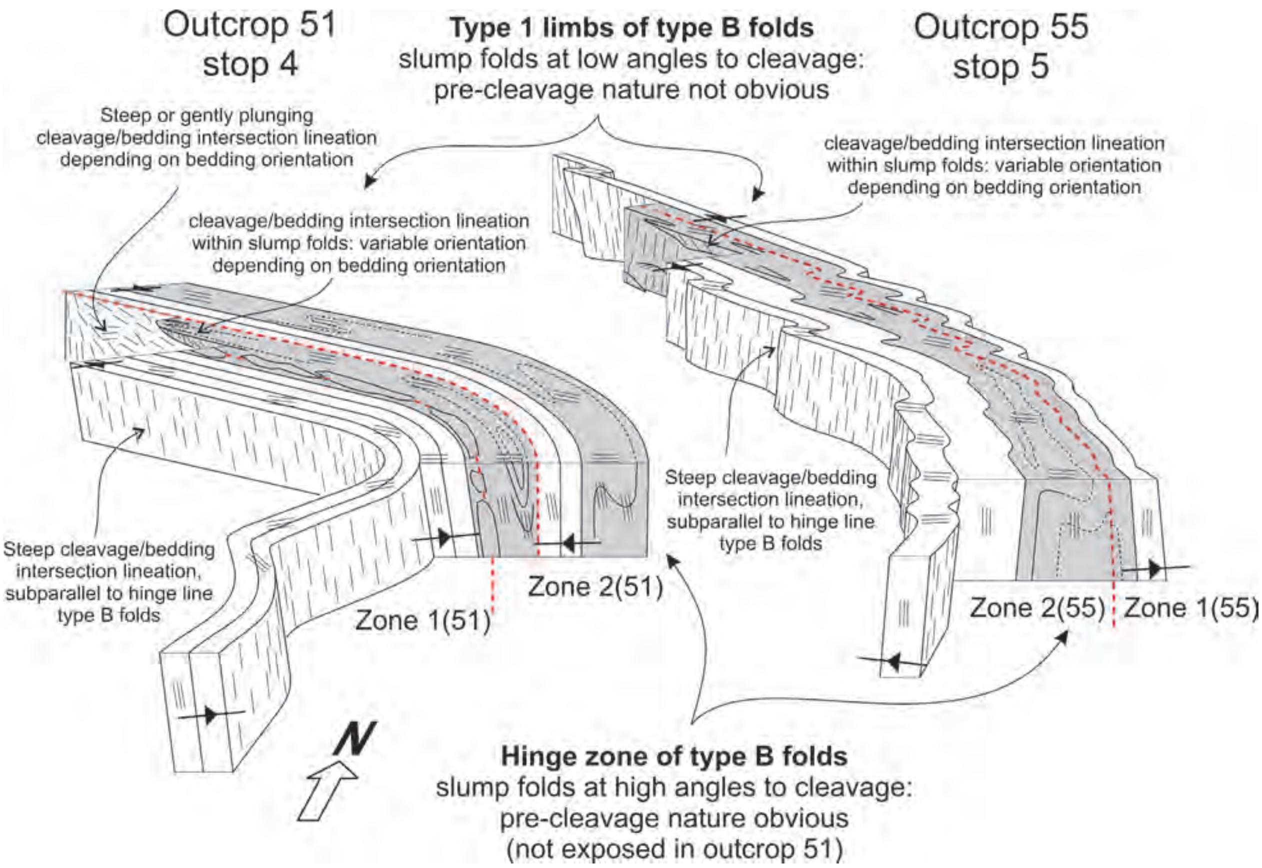

Fig. 18. Conceptual block diagrams of outcrops 51 and 55, with the pre-cleavage deformation zones marked in grey (not to scale; modified after Debacker *et al.*, 2006). In the hinge of the large type B folds the pre-cleavage nature of the sub-horizontal to gently plunging folds is obvious due to the large angle between bedding and cleavage, whereas, because of the small angle between cleavage and the slump fold axial surfaces, this is not the case within the NW-SE-trending type 1 limbs of the type B folds. Within these limbs, it is mainly (if not only) the strong variation of the cleavage/bedding intersection lineation orientation across the folds that gives away the pre-cleavage fold origin. In both outcrops, zone 1 refers to NE-ward younging beds and zone 2 to SW-ward younging beds (sense of younging shown by arrows). Note that the relative stratigraphic position of the beds of outcrop 55 with respect to those of outcrop 51 is unknown. Also note the change in overall bedding dip across the polarity reversal zones (between zones 1 and 2), which is attributed to pre-cleavage deformation (slumping) and may have a profound effect on the (local) plunge of the type B folds.

Remarks and interpretation

Also in this outcrop, the combination of the marked divergent cleavage fanning symmetrical about the fold hinge, the parallelism between the cleavage/bedding intersection lineation and the fold hinge lines and the geometrical relationship with the two vein sets, suggests that the steeply plunging folds are small-scale type B folds. The southward decrease in "z-shaped" asymmetry of these folds (compare folds II01, II02 and II03 with folds II13 to II26 on Fig. 15), together with an anticlockwise change in cleavage trend towards the south, suggests a position at the transition between a type 1 limb and a hinge zone of a large-scale type B fold.

In the southern part of the outcrop, cleavage cross-cuts the gently S-plunging folds (II27 and II28) at very high angles, indicating a pre-cleavage origin. At several places breccias occur that are cross-cut by the cleavage, as well as local changes in bedding trend that are cross-cut by cleavage (e.g. II06). Similarly, cleavage folds itself around, and cross-cuts isolated sandstone lenses and disrupted and truncated sandstone beds, without the truncation surfaces and detachments affecting cleavage (Fig. 17). All these features point to a pre-cleavage deformation event. In addition, as concerns the majority of the moderately to gently E-plunging folds (folds II04, II05, II07, II08), the same reasoning as in stop 4 can be applied. Although cleavage is almost axial planar to the folds, shows a divergent cleavage fanning symmetrical about the fold hinges and shows contrasting senses of cleavage refraction on opposite fold limbs, the strong variation in cleavage/bedding intersection lineation orientation across these folds, in combination with the occurrence within or in the direct vicinity of a pre-cleavage sediment deformation zone, suggests a pre-cleavage fold origin.

The younging sense changes from E-younging to W-younging in the pre-cleavage deformation zone in the central part of the outcrop, somewhere around fold

II12 (Fig. 15). As the structural style in the northern part of this outcrop, is, except for the sense of younging and the fold plunge, almost identical to that in the southern part, also here the change in younging sense should have a pre-cleavage origin. Hence, the change in polarity is either directly related to fold II12, having a pre-cleavage origin, or changes across an unobserved pre-cleavage detachment.

Also in this outcrop, the structural, lithological and sedimentological observations allow attributing the moderately to gently plunging folds, the sudden changes in younging sense and the pre-cleavage sediment disruption zones to slumping (cf. Debacker *et al.*, 2001). As in stop 4, the atypical, near-coincidence of cleavage and slump fold axial surfaces, observed in the central to northern part of the present outcrop, can be attributed to the extremely small angle between cleavage and overall bedding. Again, this is due to a position within the (type 1) limb of a type B fold, possibly combined with the influence of post-buckle flattening. By contrast, the large angle between cleavage and slump fold axial surfaces observed in folds II27 and II28, is due to a position in the hinge region of a large type B fold (Fig. 18).

3. Arguments for a revised stratigraphic position of the Jodoigne Formation

Below, a shortened, updated version is given of the arguments used by Herbosch *et al.* (2008) for revising the stratigraphic position of the Jodoigne Formation. For more detailed information, the reader is referred to Herbosch *et al.* (2008) and cited references.

3.1. Lithology

Although the four units of the Jodoigne Formation can be distinguished, each unit is essentially made up of an alternation of pyrite-bearing black slate and quartzite/sandstone. It is the relative amount and thickness of the black slate and quartzite/sandstone, together with more specific features such as the quartzitic nature, the presence of rhythmic, graded sequences, the presence of shale clasts and the colour of the sandstone, that allow distinguishing one unit from another.

None of the contacts between the four different units has been observed. Observational gaps of several hundred metres exist between the Maka unit and the Orbais unit, and between the latter and the Jodoigne-Souveraine unit (Fig. 4). Likely, these gaps coincide with the presence of black slate-dominated sequences. Between the outcrops of the Jodoigne-Souveraine unit and the Jodoigne unit, an apparent observational gap of about 2 km occurs. It was in this gap that several authors, who considered the Jodoigne Formation as being much younger than the Blanmont Formation (e.g. Fourmarier, 1921), placed the Tubize Formation in an attempt to explain the proximity of the Jodoigne Formation and the Blanmont Formation. However, this hypothesis was rejected by Raynaud (1952) after a magnetic survey. Also this gap likely coincides with the presence of a black-slate-dominated sequence, especially considering the particularly flat topography, the largely pelitic nature of the adjacent units, and the centimetric to decimetric, rhythmic alternation of grey sandstone, siltstone and black slate, very similar to the Jodoigne unit, observed in borehole 118W285 (Fig. 4).

Considering the lithological and sedimentological characteristics of the Jodoigne Formation (see Herbosch *et al.*, 2008) only two of the Cambrian formations known from the Brabant Massif bear some resemblance to (parts of) this formation.

The quartzites and quartzitic sandstones of the Orbais unit, and especially the quartzites of the Maka unit, may be difficult to distinguish from the quartzites of the Blanmont Formation (cf. Verniers *et al.*, 2001). This resemblance may explain why Rutot & Malaise (first geological map, 1893), Fourmarier (1921) and de la Vallée-Poussin (1931) placed the limit of the Jodoigne Formation to the east of the Maka unit. However, an important difference between the Jodoigne Formation and the Blanmont Formation can be found in the intercalated fine-grained parts. As pointed out above and observed in field trip stop 3, the fine-grained parts of the Jodoigne Formation consist of black, pyrite-bearing slate, whereas the fine-grained parts of the Blanmont Formation, observed in field trip stop 2, consist of dark brown to dark grey or green (weathered) slate (siltstone to shale) without organic matter (Verniers *et al.*, 2001; Herbosch *et al.*, submitted). Hence, provided the intercalated, fine-grained beds are observed, it is possible to distinguish the quartzites of the Jodoigne Formation (e.g. Maka unit) from those of the Blanmont Formation.

Whereas the coarse-grained parts of the Jodoigne Formation resemble the Blanmont Formation, the fine-grained parts resemble the Mousty Formation. The latter formation is described in Verniers *et al.* (2001) as: "*Shale or slate, sometimes mudstone, of grey-blue to grey-black colour, graphitic and pyritic. Massive bedded or finely laminated…; stratification can also be marked by light or greenish coloured, more silty beds or laminae, or by banded, layer-parallel colour variations. Sometimes grey more or less clayey siltstone with pyrite occurs, and occasional centimetric to decimetric fining upward sandstone or siltstone bands, interpreted as distal turbidites. The middle part of the formation is clearly more silty with grey-black pyritic shale gradually passing downwards into a grey pyritic siltstone and sometimes a sandstone*". Such a description also applies to the fine-grained parts of the different units of the Jodoigne Formation, and in particular to the Jodoigne unit. In addition, also the sedimentological interpretation of at least some parts of the Mousty Formation (distal turbidites) matches that proposed for the most fine-grained parts of the Jodoigne Formation. Recent observations at Court-St-Etienne within the Dyle-Thyle outcrop area have demonstrated the presence of numerous decimetre-scale, intensely veined sandstone to quartzite beds within the supposedly homogeneously fine-grained Mousty Formation (Vandorpe, 2007). Starting from the deposits of the Mousty Formation, one can imagine that, if sediment supply were to increase, and the deposition area would shift towards a more proximal position, the resulting Mousty Formation would become very difficult to distinguish from the deposits of the Jodoigne Formation.

Hence, if one were to place the Jodoigne Formation somewhere between the other Cambrian formations, purely on the basis of lithology and sedimentology, a logical choice would be to place it directly below, or at the same level of, the Mousty Formation.

3.2. Map, structural observations and younging sense

In most outcrops across this area bedding is steeply dipping (> 60°). However, the strike of the predominantly steep bedding is highly variable, seemingly suggestive of steeply plunging folds (Fig. 19). In contrast to the strongly variable bedding trend, cleavage is virtually always steeply N-dipping and has a mean E-W-trend. In those outcrops where the steep bedding is at a high angle to cleavage, the resulting cleavage/bedding intersection is steeply plunging, again suggestive of steeply plunging folds. In addition, in some outcrops, due to the low angle between bedding and cleavage trend, and/or due to the low bedding dip, the cleavage/bedding intersection is gently plunging, suggestive of gently plunging folds.

On the basis of bedding orientation, cleavage/bedding-relation, orientation of the cleavage/bedding intersection, the relationship between bedding, cleavage/bedding intersection and quartz-filled fractures and comparison with folds from other outcrop areas (e.g. Debacker *et al.*, 2004a and references therein), it becomes clear that in the Geete outcrop area the steeply plunging cleavage/bedding intersection reflects type B folds, whereas the gently plunging cleavage/bedding intersection usually corresponds to type A folds (Fig. 6, Fig. 19). Note that the use of “type A folds” and “type B folds” implies a syn-cleavage origin, as originally defined in Debacker *et al.* (2004a). Exceptions occur in the Jodoigne unit at Jodoigne, where locally a gently plunging cleavage/bedding intersection occurs. As demonstrated by Debacker *et al.* (2006) and shown during this field trip, in at least two outcrops at Jodoigne (outcrops 51 and 55), the gently plunging intersection is due to the local presence of slump folds. Hence, although gently plunging, these folds and their associated cleavage/bedding intersection cannot be regarded as type A folds or type A cleavage/bedding relationships.

Type B cleavage/bedding relationships dominate the area. Especially within the Maka unit, in the southern part of Jauchelette, the presence of decametre- to hectometre-scale type B folds becomes apparent. Type A cleavage/bedding relationships are much less common. Clear examples of type A cleavage/bedding relationships are observed in the Opprebais quarry (stop 2, outcrop 22) and in outcrops 250, 27 and 328. A type A fold hinge can be observed in a private property in the vicinity of the Bordia castle, at the northern limit of Jodoigne (outcrop 47).

Where possible, the younging sense of the beds was determined on the basis of sedimentological criteria (stratigraphic polarity: grading, oblique lamination, loadcasts,…) and this was compared with the younging sense suggested by the cleavage/bedding relationship (structural polarity). In four outcrops of the Jodoigne unit at Jodoigne, local mismatches are observed between the structural and stratigraphic polarity. As pointed out by Debacker *et al.* (2006) and shown during this field trip, in two of these outcrops (outcrops 51 and 55, stops 4 and 5), this can be attributed to the local presence of overturned bedding due to slumping. Also in the other two outcrops (outcrops 331 and 339), local mismatches might be due to the presence of overturning by slumping (Debacker, Similox-Tohon, van Noorden, Kenis & Sintubin, unpub. data). In and between the other outcrops, however, a consistent image becomes apparent. Across the area, the overall younging sense is towards the eastnortheast, and this younging sense is reflected by observations from the Maka unit, the Orbais unit and the Jodoigne unit of the Jodoigne Formation and from the Blanmont Formation. This implies that the relative age within the Jodoigne Formation increases from the Jodoigne unit to the Maka unit and also suggests that the Jodoigne Formation is younger than the Blanmont Formation (Fig. 19).

3.3. Temperature-dependent variation of magnetic susceptibility

The first author analysed small temperature-dependent changes of magnetic susceptibility at the “room temperature interval (i.e. between 0 and 40°C)” for a large number of lithostratigraphic units of the Brabant Massif. This was done at the K.U.Leuven, using a KLY3S Kappabridge (AGICO; Jelinek & Pokorny, 1997). For each sample the percentage of change in magnetic susceptibility was (re-)calculated for a temperature change of 20°C within the ~0-30°C temperature-interval. For more information on this method, the reader is referred to Herbosch *et al.* (2008) and Debacker *et al.* (2009, 2010).

Several samples of both the Jodoigne and the Mousty Formation show a very large temperature-dependent susceptibility change (9 to 17%), being much more pronounced than that observed within any of the other investigated lithostratigraphic units, and which cannot be attributed to a paramagnetic behaviour. Although the cause of this remains unknown (possibly diamagnetic behaviour?), it does suggest a comparable magnetic (*s.l.*) mineralogy for both formations, something which is confirmed by thermal demagnetisation analyses (see Herbosch *et al.,* 2008 and Debacker *et al.*, 2009; 2010).

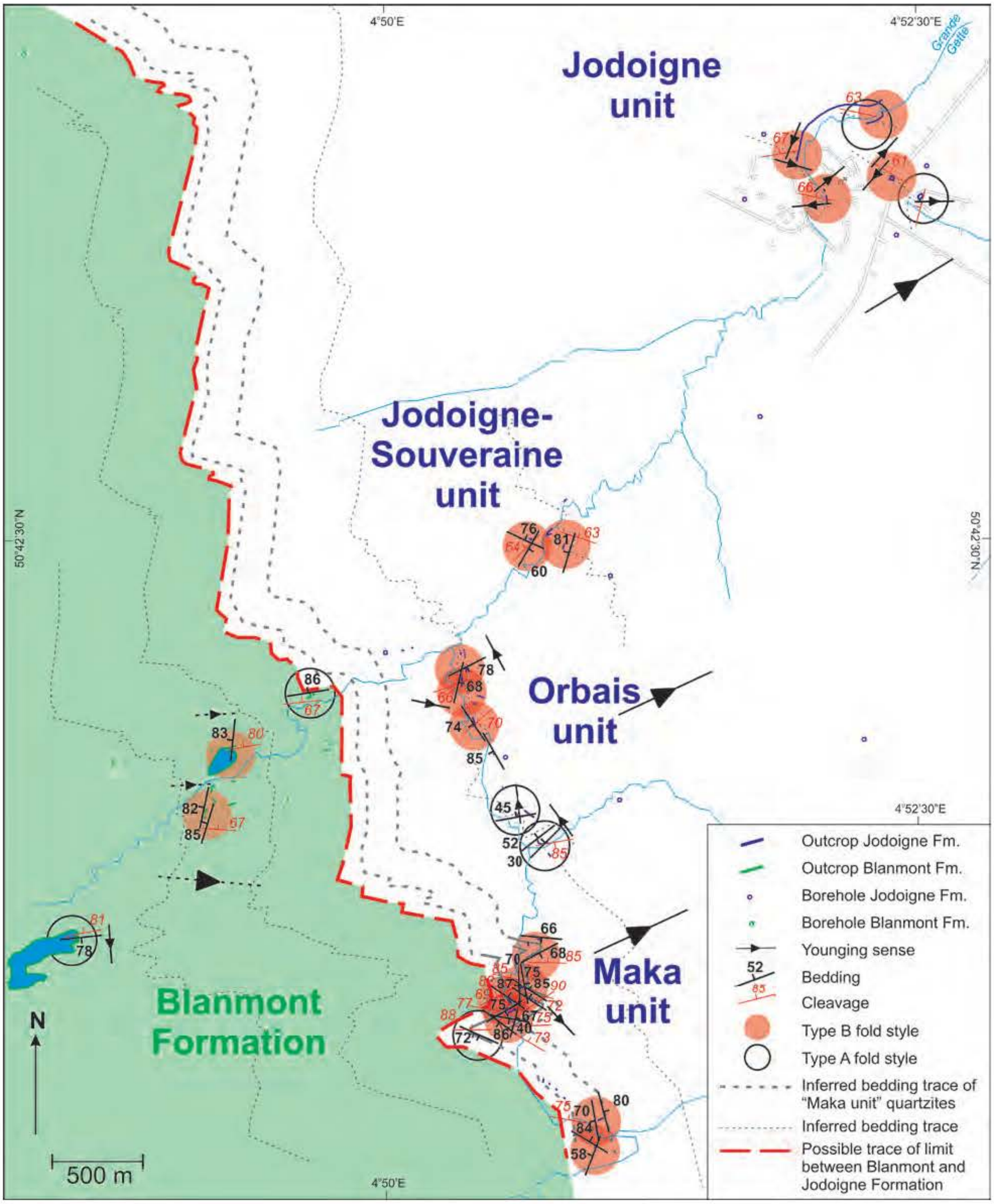

Fig. 19. Simplified topographic map of the Geete outcrop area, showing the outcrop distribution of the Blanmont Formation and the Jodoigne Formation, together with the mean bedding and cleavage orientation per outcrop and the stratigraphic younging sense (where observed; small arrows represent local observations, large arrows correspond to the overall younging sense). Also added are fold type (type A or type B), as deduced from the cleavage/bedding relationship, as well as the probable bedding traces, as inferred from structural outcrop data (modified after Herbosch *et al.*, 2008).

3.4. Biostratigraphic age

The Jodoigne Formation never yielded any biostratigraphic information (Verniers *et al.*, 2001). Also several dating attempts by means of acritarchs on the newly described units proved unsuccessful (Vanguestaine, pers. comm.).

The upper parts of the Mousty Formation were dated in outcrop as belonging to the base of the Tremadocian (Lecompte, 1948, 1949; Martin, 1968; Vanguestaine *in* André *et al.*, 1991). The older parts of the Mousty Formation were dated in boreholes by means of acritarchs. The biostratigraphic ages of the Mousty Formation at Eine (84E1372) and Vollezele (100E010) correspond to the lower and middle parts of the Upper Cambrian (Vanguestaine, 1992 and pers. comm.), whereas the age obtained at Cortil-Noirmont (130W539) corresponds to the Upper Cambrian (Vanguestaine *in* Delcambre & Pingot, 2002 and pers. comm.). Hence, judging from these borehole and outcrop data, the Mousty Formation extends from the base of the Upper Cambrian to the base of the Tremadocian. However, at Leuven (89E01), to the north of the central axis of the Brabant Massif, a completely different biostratigraphic age was obtained. The rocks in this borehole were described as Silurian by Rutot & Van den Broeck (1890; "*schistes noirâtres passant au grès*" and later as "*Revinien supérieur*" (Mousty Formation) by Legrand (1968; "*phyllades satinés, noir dense, un peu pyriteux. A la base quelques lits de quartzophyllades*"). By means of acritarchs, these deposits were dated as belonging to the lower part of the Middle Cambrian (Vanguestaine, 1973, 1992), which is significantly older than the age obtained from the other boreholes containing the Mousty Formation (see above). However, boreholes in the surroundings of Leuven (Leuven, 89E01; Heverlee 89E363) consist of a facies described as "*quartzophyllades*", with "*passées gréseuses*" (89E01). In addition, an examination by Herbosch (unpub. data) of a core of the Heverlee borehole (89E363) revealed the presence of centimetre-scale turbidite sequences, a facies which is rather different from the classical facies of the Mousty Formation, but quite similar to that of the Jodoigne unit of the Jodoigne Formation. Considering the older biostratigraphic age and the lithological and sedimentological resemblance with the Jodoigne Formation, it is likely that the aforementioned boreholes in the Leuven area contain the Jodoigne Formation instead of the Mousty Formation. Also cartographically this seems possible, as Leuven is situated only ~20km from Jodoigne, in a NW-SE-direction subparallel to the bedding trend.

3.5. Suggested stratigraphic position

A combination of different arguments suggests that the Jodoigne Formation is not older than the Blanmont Formation, but is instead much younger, and probably has a Middle to Upper Cambrian age, close to or slightly below that of the Mousty Formation (Fig. 2B). These arguments are: a) the lithological similarities between the Mousty Formation and the fine-grained parts of the Jodoigne Formation; b) the identical temperature-dependent variation of magnetic susceptibility of the Mousty Formation and the Jodoigne Formation, being exceptional with respect to that of other formations of the Brabant Massif; c) the younging sense within both the Blanmont Formation and the Jodoigne Formation in the study area, suggesting a younging from the former (WSW) towards the latter (ENE); d) its position in the northern part of the Brabant Massif, to the NE of the central axis occupied by the Blanmont Formation (Michot, 1980); e) the apparent stratigraphic hiatus, yet unexplained, between the basal Middle Cambrian (Oisquercq Formation) and the Upper Cambrian (Mousty Formation; see Verniers *et al.*, 2001); and f) the Middle Cambrian biostratigraphic age of deposits previously tentatively attributed to the Mousty Formation (base of Upper Cambrian to base of Tremadocian), but containing a facies quite distinct from that of the classical Mousty Formation and strongly resembling that of the Jodoigne unit of the Jodoigne Formation within the Geete outcrop area. Although separately each of these arguments may be criticised, together they plead a case for a Middle (to Upper?) Cambrian stratigraphic position of the Jodoigne Formation, situated below or (partly) overlapping with that of the Mousty Formation (Herbosch *et al.*, 2008).

4. How to explain the proximity of the Jodoigne Formation to the Blanmont Formation in the Geete outcrop area?

The close proximity in the eastern part of the Brabant Massif between the Jodoigne Formation and the Blanmont Formation can be explained by means of the Asquempont Detachment System (Fig. 20). The Asquempont Detachment System, named after the Asquempont fault at Asquempont, is a low-angle extensional detachment system that formed prior to folding and cleavage development (Debacker *et al.*, 2004b, 2005a; cf. Debacker, 2001).

In the Senne-Sennette outcrop area, where it was defined, this detachment system places Lower Ordovician strata on top of the Oisquercq Formation (Debacker, 2001; Debacker *et al.*, 2004b; cf. Legrand, 1967; Mortelmans, 1955). In the unexposed parts of the Brabant Massif to the WNW of the Senne-Sennette outcrop area, this detachment system places the Lower to Upper Ordovician on top of the Oisquercq Formation, as observed in several boreholes (from E to W: Bever 114W73, 114W93; Lessines 113E1015; Schendelbeke 100W181; Eine 84E1372; Waregem 84W1385; at Eine: contact between Mousty Formation and Oisquercq Formation) (Piessens *et al.*, 2005; cf. Debacker *et al.*, 2004b). To the east, in the Dyle-Thyle outcrop area, the presence of the Asquempont Detachment System forms the most plausible explanation for the enigmatic contact between the Mousty Formation (detachment hanging wall) on the one hand and the Tubize and Blanmont formations (detachment footwall) on the other hand (Debacker *et al.*, 2005a).

On the basis of the stratigraphic position of the hanging wall and footwall rocks in boreholes in the unexposed western part of the Brabant Massif, Piessens *et al.* (2005) deduced an initial (i.e. pre-folding) gentle NNE-dip for the Asquempont Detachment System. The apparent eastward stratigraphic ageing of both the hanging wall and footwall rocks from the western unexposed areas to the Dyle-Thyle outcrop area along the S-side of the Brabant Massif is fully compatible with the results of Piessens *et al.* (2005; Fig. 20). Extrapolating this trend towards the Geete outcrop area, a hanging wall composed of the Mousty Formation or older, and a footwall composed of the Blanmont Formation or older would be expected. As indicated by our observations in the Geete outcrop area, the presence of eastward younging deposits of the Jodoigne Formation, seemingly directly overlying eastward younging deposits of the lowermost Cambrian Blanmont Formation, necessitates the presence of a discontinuity (hiatus, fault or unconformity) between both formations, responsible for removing the Tubize Formation and the Oisquercq Formation. This discontinuity is the Asquempont Detachment System, with the Jodoigne Formation in the hanging wall and the Blanmont Formation in the footwall. Hence, the Asquempont Detachment System, the NNE-dip of this detachment system and the newly suggested stratigraphic position of the Jodoigne Formation (Fig. 2B) are fully compatible (Fig. 20).

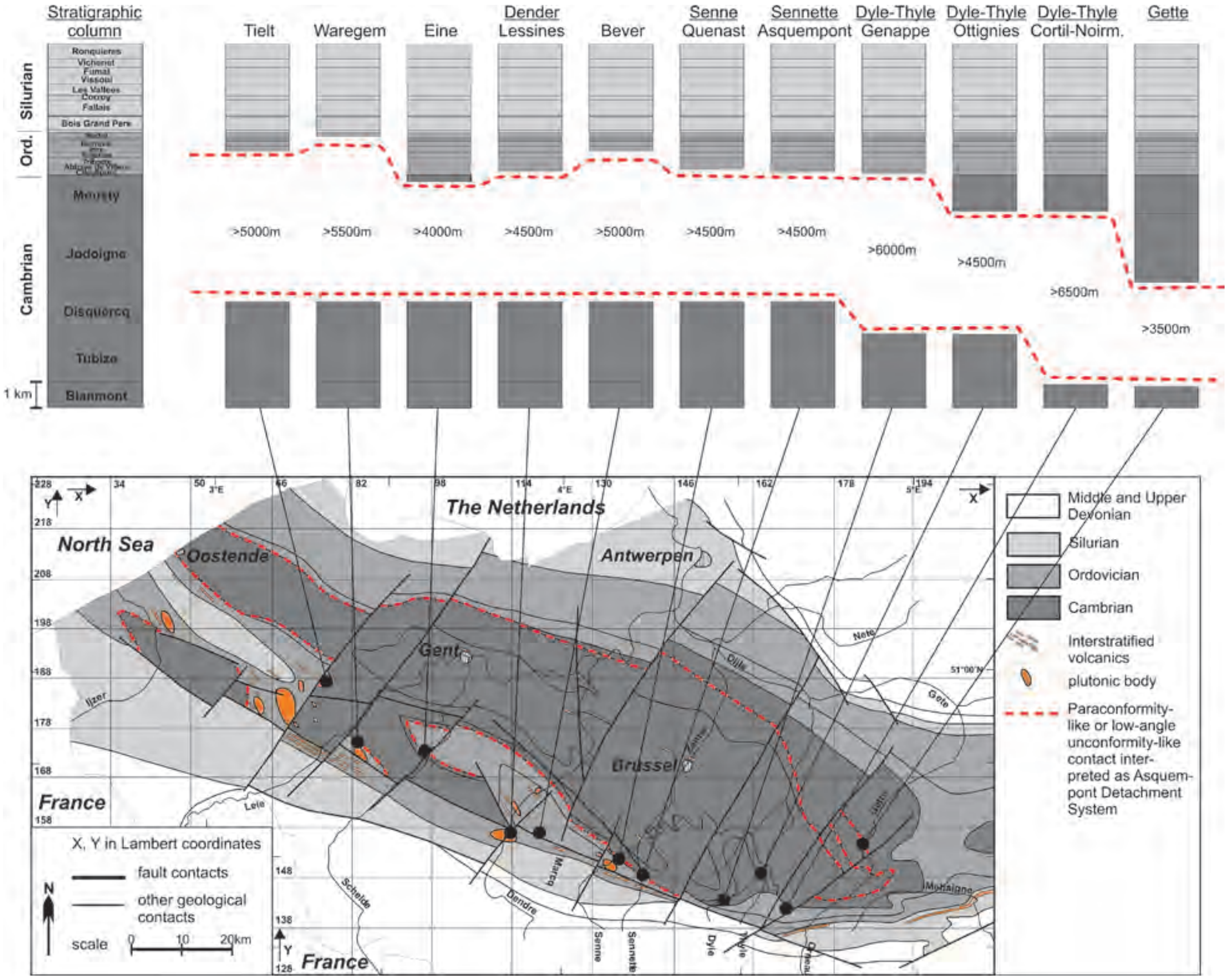

Fig. 20. Geological subcrop map of the Brabant Massif (after De Vos *et al.*, 1993 and Van Grootel *et al.*, 1997), showing the effect of the Asquempont Detachment System on the stratigraphy. For comparative purposes, for each of these points the same composite stratigraphic column is used as shown in the upper left corner. On these stratigraphic columns, the part considered to be removed by the Asquempont Detachment System is shown in white, together with a minimum thickness estimate. The thickness of the individual formations is based on Verniers *et al.* (2001), Piessens *et al.* (2005) and Herbosch *et al.* (in press) and for the Jodoigne Formation on Herbosch *et al.* (2008, submitted). Tielt, Waregem, Eine, Lessines (Dender valley) and Bever are borehole observations, whereas the columns for the Senne, the Sennette, the Dyle-Thyle and the Geete are based on outcrop observations. Data are taken from Debacker (2001), Debacker *et al.* (2004a, 2004b, 2005a), Piessens *et al.* (2005) and Herbosch *et al.* (2008). See text for explanation.

In addition, besides seemingly confirming the NNE-dip of the Asquempont Detachment System, the new geological interpretation of the Geete outcrop area forms the first strong indication for the presence of the Asquempont Detachment System along the N-side of the Brabant Massif, an idea previously already suggested on cartographic grounds by Debacker *et al.* (2004b, 2005a) and Piessens *et al.* (2005).

For further discussion on this matter the reader is referred to Herbosch *et al.* (2008).

5. Discussion

5.1. How to explain the quartzitic nature of parts of the Jodoigne Formation?

The intensely veined, quartzitic nature of parts of the Jodoigne Formation, in particular the Maka unit, is quite exceptional within the Brabant Massif, and is only paralleled by parts of the Blanmont Formation. Although we found no records of this, it is quite possible that this strongly quartzitic nature was considered as an extra argument by early workers for placing the Jodoigne Formation at the bottom of the Cambrian stratigraphy of the Brabant Massif (e.g. Dumont, 1848; Malaise, 1900; Kaisin, 1919, de la Vallée Poussin, 1931).

The local occurrence of these intensely veined quartzites can be explained by regarding (parts of) the Jodoigne Formation as a high fluid pressure cell, quite comparable to the mullion-bearing Lower Devonian Ardenne-Eifel Slate Belt (cf. Kenis *et al.*, 2005). According to the revised Cambrian stratigraphy of Herbosch *et al.* (2008), followed and presented also in this work, the quartizic parts of the Jodoigne Formation are overlain by pelitic sequences of the upper parts of the Jodoigne Formation and the Mousty Formation, and underlain by homogeneous, pelitic sequences of the Oisquercq Formation. This implies that the quartizic parts of the Jodoigne Formation are completely surrounded by low-permeability pelites. During burial and subsequent deformation, the heated fluids within these coarse-grained parts of the Jodoigne Formation were unable to escape because of the presence of low-permeability pelitic sequences above and below, thus resulting in a fluid pressure build up. Likely, these trapped, hot fluids transformed the original sandstone into quartzite, whereas the increased fluid pressure resulted in the formation of veins.

Possibly, this may also have implications for the relative timing of the Asquempont Detachment System as compared to veining and sandstone-to-quartzite transformation within the Maka unit of the Jodoigne Formation. The hypothesis outlined above seems more compatible with a sandstone-to-quartzite transformation prior to formation of the Asquempont Detachment System, as only then thick pelitic sequences were present both above (fine-grained parts of Jodoigne Formation) and below (Oisquercq Formation) the coarse-grained parts of the Jodoigne Formation (e.g. Maka unit). If the Asquempont Detachment System would have formed earlier, pressure build-ups of hot fluids are expected to have been less significant, as then in the Geete area a direct (faulted) contact existed with the underlying, much more permeable, coarse-grained Blanmont Formation. This hypothesis definitely deserves thorough investigation by future studies.

5.2. Restricted occurrence of the Jodoigne Formation within the Brabant Massif?

As pointed out by Piessens *et al.* (2005) in their inventory of all existing data for the construction of a new geological subcrop map of the Brabant Massif, facies resembling the Jodoigne Formation have only been encountered in the eastern parts of the massif. Indeed, facies characteristic of the Jodoigne Formation have only been encountered in the Geete outcrop area and in the subsurface around Leuven. Further towards the west, the Middle to Lower Cambrian is only represented (in boreholes) by a facies resembling that of the Mousty Formation (Piessens *et al.*, 2005). Only occasionally these rocks have been dated biostratigraphically.

The apparent restriction of the Jodoigne Formation towards the easternmost parts of the Brabant Massif is either due to the presence of the Asquempont Detachment System, responsible of removing this formation (Fig. 20), or due to a westward fining, causing the deposits of the Jodoigne Formation to resemble those of the Mousty Formation. The latter option is one of the reasons why we considered it possible that parts of the Jodoigne Formation and the Mousty Formation overlap in time, as to some extent they may form lateral equivalents (Herbosch *et al.*, 2008). At present, both options or a combination of both remain possible.

5.3. Similarities with the Middle and Upper Cambrian of the Ardennes Inliers?

Despite the fact that the Middle Cambrian within the Brabant Massif was unknown at that time, Verniers *et al.* (2002) concluded a similar Cambrian basin evolution for the Brabant Massif and the Ardennes Inliers, especially for the Middle and Late Cambrian. If so, and considering the newly proposed stratigraphic position favoured herein (see Herbosch *et al.*, 2008), the Jodoigne Formation should bear some resemblance to the Middle to Upper Cambrian deposits of the Ardennes Inliers.

As pointed out by de la Vallée-Poussin (1931), the lithology of the Jodoigne Formation strongly resembles that of the Revin Group of the Ardennes Inliers. This group, consisting of several formations, is mainly composed of alternations of black slate, mudstone, grey sandstone and grey to occasionally black quartzite of variable thickness (e.g. Verniers *et al.*, 2001). This general description is very similar to that of the Jodoigne Formation. According to the new stratigraphic position, the Jodoigne Formation should be time-equivalent to the Rocher de l'Uf Formation (Rv1) and/or the La Roche à 7 heures Formation (Rv2) in the Rocroi Inlier. In addition, considering the possible time-equivalence with parts of the Mousty Formation, the Jodoigne Formation possibly overlaps also with the Anchamps Formation (Rv3) and maybe even also with part of the Petite-Commune Formation (Rv4-Rv5) (e.g. Vanguestaine, 1992; Verniers *et al.*, 2001). Similarly, for the Stavelot Inlier, the Jodoigne Formation is expected to be time-equivalent to the Wanne Formation (Rv1-Rv2) and, because of the possible time-equivalence with parts of the Mousty Formation, possibly overlaps with the La Venne Formation (Rv3-Rv4) (e.g. Vanguestaine, 1992; Verniers *et al.*, 2001). Several of these formations have facies resembling the different facies of the Jodoigne Formation. Both the La Venne Formation and the La Roche à 7 heures Formation, for instance, contain massive grey quartzite, resembling the Maka unit of the Jodoigne Formation, whereas black sandstone and quartzite like in the Jodoigne-Souveraine unit of the Jodoigne Formation are present also in the Anchamps Formation, the Petite-Commune Formation and the Wanne Formation (e.g. Verniers *et al.*, 2001). Hence, although a detailed correlation between the different units of the Jodoigne Formation and the different lithostratigraphic units of the Revin Group in the Ardennes Inliers is not possible yet, the lithological similarities between the Jodoigne Formation and the lower and middle parts of the Revin Group are compatible with the proposed stratigraphic position. These similarities also seem to support the similar Middle to Late Cambrian basin evolution for the Ardennes Inliers and the Brabant Massif suggested by Verniers *et al.* (2002).

Judging from the literature, the estimated stratigraphic thickness of the Middle and Upper Cambrian is about three times less in the Ardennes Inliers (~1500 m) than in the Brabant Massif (~4500 m). In combination with results of palaeocurrent analysis and good stratigraphic constraints, such a consistent difference in thickness could provide valuable information for the basin evolution during the Cambrian. Unfortunately, because of the locally very intense deformation, stratigraphic thickness in the Ardennes Inliers is much less well constrained than in the Brabant Massif, implying that the Cambrian in the Ardennes Inliers may be thicker than estimated. In addition, to the author's knowledge, despite some local attempts (e.g. unpublished works of Boone, 2008 and Braspenning, 2009 in the La Venne Formation at the N-side of the Stavelot Inlier in the vicinity of Eupen), no large-scale palaeocurrent analyses have been performed in the Cambrian of the Ardennes Inliers and the Brabant Massif.

In summary, the possible link in terms of sedimentology and stratigraphy between the Brabant Massif and the Ardennes Inliers definitely deserves future detailed studies in which stratigraphy, sedimentology and structural geology are fully integrated.

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