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Influence of slump folds on tectonic folds: an example from the Lower Ordovician of the Anglo-Brabant Deformation Belt (Belgium)

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Abstract: Although it is generally accepted that buckle folds will not develop in a perfectly planar layer without the presence of some irregularity or perturbation at which the folds initiate, there are very few cases in which individual natural folds can be linked to specific irregularities. Within the Lower Ordovician Abbaye de Villers Formation, Anglo-Brabant Deformation Belt, metre-scale tectonic folds occur, of which the position and, to a certain extent, the geometry appear to be controlled by slump folds and related features. The metre-scale tectonic folds, interpreted as parasitic structures on the limb of a large-scale host fold, occur only within a stratigraphic level affected by slumping. In this level, tectonic antiforms tend to form superimposed on antiformal slump folds and on zones of abrupt, slump-related thickness increase, and tectonic synforms on synformal slump folds and on zones of abrupt thickness decrease. The rather irregular 3D geometry of sedimentary sequences suggests that many more similar cases should exist in which folds can be linked to specific irregularities. However, possibly it is also this abundance of irregularities in sedimentary sequences, in combination with fold and outcrop scale, that makes it difficult to attribute a particular fold to a particular perturbation.

Keywords: Brabant Massif, cleavage, folds, slump structures.

One of the most intriguing questions in structural geology is why particular structures form at particular localities. In the case of folding, theories and experiments have shown that buckle folds will not develop in a perfectly planar layer without the presence of some irregularity or perturbation at which the folds initiate (Cobbold 1975; Lewis & Williams 1978; Williams *et al*. 1978; Abbassi & Mancktelow 1990, 1992; Price & Cosgrove 1990; Mancktelow 1999; Zhang *et al*. 2000; Williams & Jiang 2001). In addition, perturbations may also influence the shape of the resulting buckle folds (e.g. Cobbold 1975; Williams *et al*. 1978; Abbassi & Mancktelow 1990, 1992; Mancktelow 1999). The perturbations may be an original property of the layered system, such as a local layer thickening or the presence of isolated competent bodies (e.g. channels, intrusive bodies) or may be induced by failure during initial deformation (e.g. Cobbold 1975; Abbassi & Mancktelow 1990; Price & Cosgrove 1990; Zhang *et al*. 2000; Williams & Jiang 2001). Also, local rheological variations or local reductions in cohesive strength may act as perturbations (Price & Cosgrove 1990; Williams & Jiang 2001).

The quasi-periodic form of many natural fold trains is comparable with those produced in numerical models, leading to the suggestion that irregularities in fold shape and orientation observed in natural fold trains are also determined by the location and shape of initial perturbations (Mancktelow 1999). However, judging from the literature, natural examples of perturbations demonstrated to have acted as buckle fold initiation points and to have influenced the final fold shape are very rare. For instance, in the Brabant Massif, representing the Belgian part of the Early Palaeozoic Anglo-Brabant Deformation Belt, the Asquempont synform and its slightly oblique orientation with respect to the general tectonic trend may result from the presence of a large wedge-like overturned slump sheet in the synform hinge zone (Debacker *et al*. 2001). Similarly, in the same outcrop area, the competent Fauquez volcanoclastic deposits, which show a strong eastward decrease in thickness, may have influenced the position and geometry of the large-scale Fauquez antiform, and the local, non-cylindrical parasitic folds within it (Debacker 2001). However, although both the large-scale slump sheet and the volcanoclastic deposits are likely to have acted as large-scale perturbations, it is difficult to demonstrate this supposed relationship. This is mainly due to the large scale of the tectonic folds and the poor degree of exposure.

In this study, natural examples are given of syn-cleavage buckle folds, of which the location and to a certain extent also the geometry are controlled by soft-sediment deformation features, which acted as perturbations during tectonic layer-parallel shortening. These examples are found within a specific lithostratigraphic unit, the Abbaye de Villers Formation, in the southern part of the Brabant Massif, Anglo-Brabant Deformation Belt.

Geological setting

The Brabant Massif is a poorly exposed, low-grade metamorphic, Early Palaeozoic slate belt in North and Central Belgium. It represents the southeastern part of the largely concealed Anglo-Brabant Deformation Belt, one of the deformation belts of Eastern Avalonia (Fig.1; Van Grootel *et al*. 1997; Verniers *et al*. 2002). The Brabant Massif has an anticlinal subcrop appearance, with deposits ranging in age from earliest Cambrian in the core to late Silurian along the rims, and is unconformably overlain by undeformed, diagenetic Givetian and younger sequences (Legrand 1968; De Vos *et al*. 1993; Van Grootel *et al*. 1997; Verniers *et al*. 2002). Throughout the exposed parts of the massif, there is only evidence for one single-phase progressive deformation (e.g. Sintubin 1997, 1999; Verniers *et al*. 2002; Debacker *et al*. 2004*b*, 2005,

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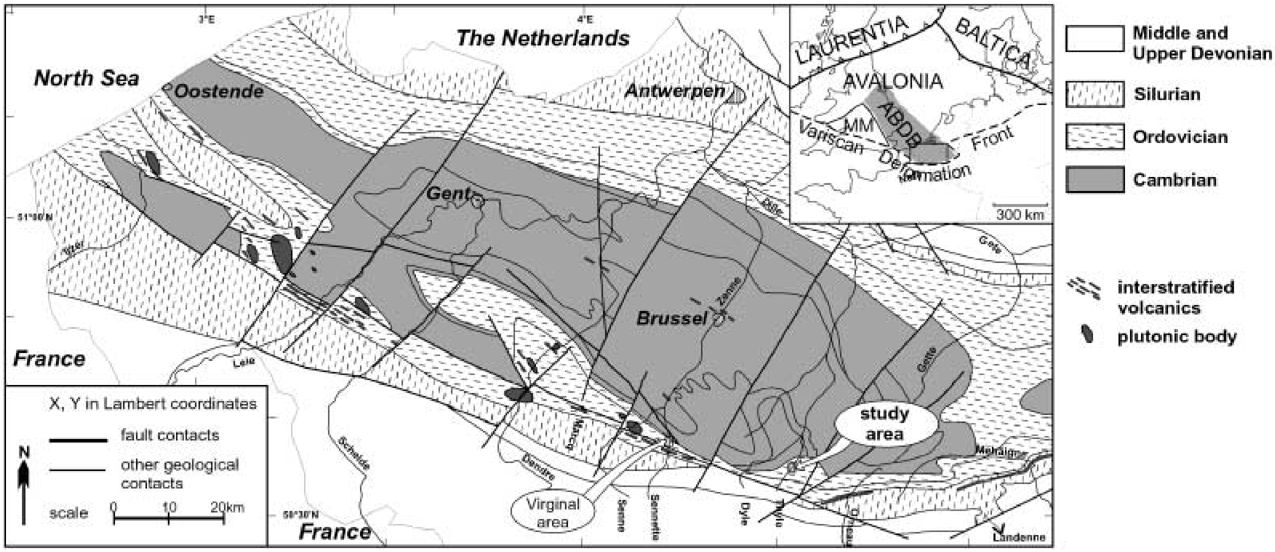

Fig. 1. Geological subcrop map of the Brabant Massif (after De Vos *et al*. 1993; Van Grootel *et al*. 1997) showing the position of the study area (see Fig. 2) and of the Virginal area (see Fig. 10). The inset shows the position of the Brabant Massif within the Anglo-Brabant Deformation Belt (ABDB) along the NE side of the Midlands Microcraton (MM) in the context of Avalonia, Baltica and Laurentia.

and references therein). This deformation phase, termed the Brabantian deformation phase, is considered to have taken place between the late Llandovery and the Emsian (Debacker *et al*. 2005) and is tentatively attributed to an anticlockwise rotation of the Midlands Microcraton (Verniers *et al*. 2002). The main features associated with this deformation are folds with a well-developed cogenetic cleavage. In the Ordovician–Silurian sequences in the southern part of the massif, subhorizontal to gently plunging folds occur, with a south-verging asymmetry, a common stepfold geometry, and sizes ranging from decimetre- to kilometre-scale. The small- and meso-scale tectonic folds generally occur within the hinge zones of the hectometre- to kilometre-scale stepfolds (e.g. Debacker *et al*. 1999, 2001, 2005; Debacker 2001).

The studied folds occur within the Lower Ordovician Abbaye de Villers Formation (uppermost middle to upper Arenig) in the Thyle valley (Fig. 1). Twenty-one outcrops were studied along a 400 m long, north–south-directed discontinuous outcrop section (Figs 2 and 3). From a structural point of view, the section is situated within the subhorizontal to gently south-dipping northern limb of a kilometre-scale antiformal stepfold (Herbosch *et al*. 2002). The Abbaye de Villers Formation consists of bioturbated, grey to dark grey, fine-grained sandstone to mudstone, with an irregular, lenticular centimetre-scale lamination (Fig. 4). Characteristically, the fine-grained sandstone laminae have rather diffuse limits (Verniers *et al*. 2001). These sediments were deposited in a shelf environment, when Avalonia was already a separate continent, drifting away from Gondwana towards Baltica (Verniers *et al*. 2002). Within the Abbaye de Villers Formation two lithological units are distinguished (Fig. 3) (Beckers 2003, 2004). The northern, stratigraphically lower unit, with a minimum thickness of 10 m (outcrops 1–12 and 14), has a well-stratified outcrop appearance, clearly reflecting the characteristic centimetre-scale lamination. The southern, stratigraphically upper unit, with a minimum thickness of 75 m (outcrops 13 and 15–20), is slightly more sandy and, although also having an irregular to lenticular centimetre-scale lamination (Fig. 4), has a more massive outcrop appearance. The transition between the two units is gradual.

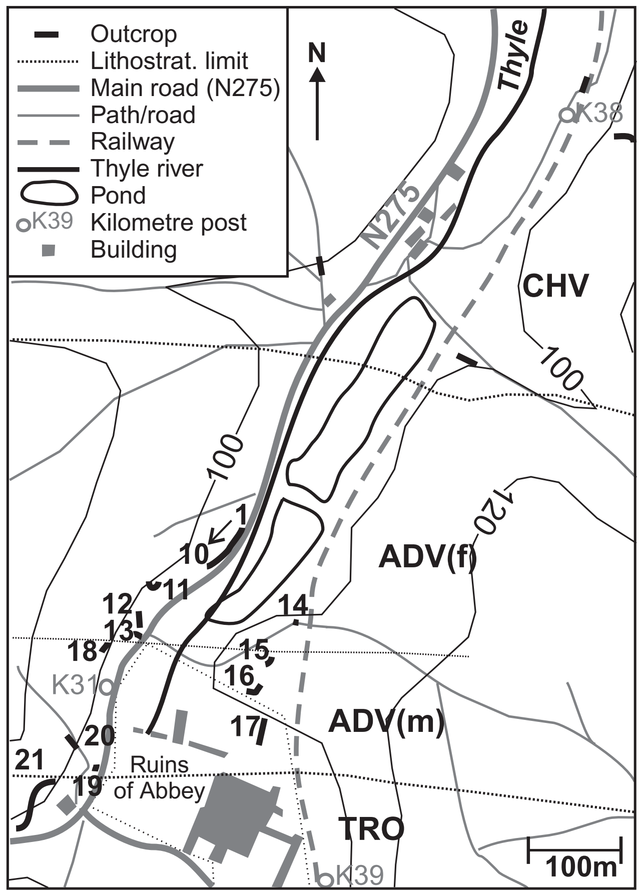

Fig. 2. Simplified topographic map of the study area around the abbey of Villers, with the position of the 21 outcrops studied. The approximate limits between the Chevlipont Formation (CHV; Tremadoc), the Abbaye de Villers Formation (ADV; Arenig; ADV(m): upper unit; ADV(f): lower unit) and the Tribotte Formation (TRO; upper Arenig) are shown, based on Herbosch & Lemonne (2000) and personal observations.

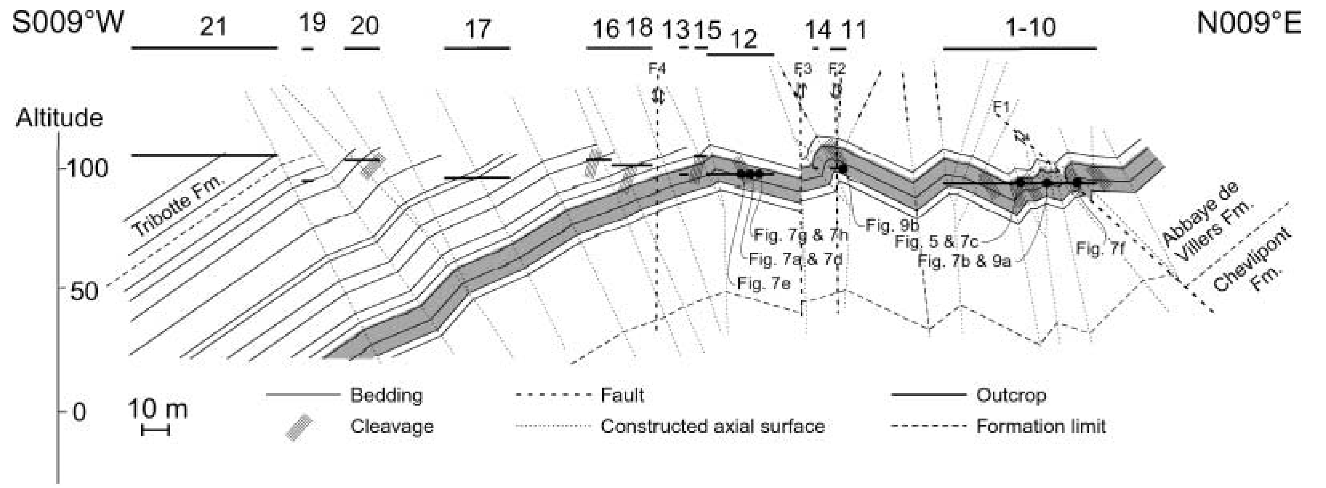

Fig. 3. Cross-section of the study area (modified after Beckers 2003, 2004), constructed by means of the kink band method, showing the position of the outcrops (1–21). The cross-section shows a hectometre-scale, gentle antiform with a hinge zone situated around the southern part of outcrop 12 and the northern part of outcrop 15. Meso-scale folds are observed only in the minimum *c.* 10 m thick lower unit of the Abbaye de Villers Formation (marked in grey). It should be noted that, because of the isolated position of the pre-cleavage folds between relatively undeformed beds, the cross-section is based entirely on the syn-cleavage folds; pre-cleavage changes in bedding orientation have not been taken into account. The poorly constrained limit between the Abbaye de Villers Formation and the underlying Chevlipont Formation (Tremadoc) is based on Herbosch & Lemonne (2000).

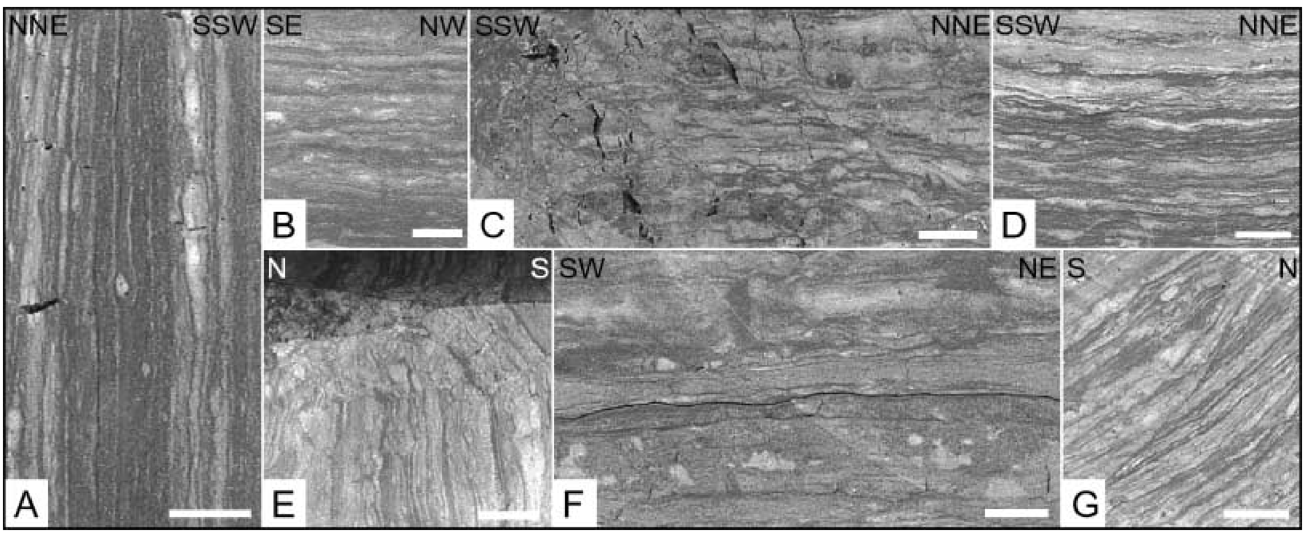

Fig. 4. The Abbaye de Villers Formation in the Thyle valley: characteristic lithologies (**b**, **d**: upper unit; **a**, **c**, **e**, **f**, **g**: lower unit) and examples of small-scale soft-sediment deformation (**c**, **e**, **g**). The scale-bar represents 1 cm. (**a**) Typical laminated, bioturbated lithology; diffuse limits and lenticular nature of sandy beds should be noted (TD1252; outcrop 6; bedding: 115/80SW). (**b**) Typical, laminated, bioturbated lithology; diffuse limits of sandy beds should be noted (TD1260; outcrop 18; bedding: 138/30SW). (**c**) Irregular, slightly deformed (bioturbated?) bedding, which becomes more fragmented towards the south, and, at 1 cm from the left border, abuts on a steep, pre-cleavage breccia, consisting of a pelitic matrix with small, isolated, sandstone and siltstone fragments (TD1255; outcrop 11; bedding: 295/22NE). (**d**) Typical, laminated, bioturbated lithology; diffuse limits and lenticular nature should be noted (TD1259; outcrop 15; bedding: 115/05SW). (**e**) Oblique view of sample, showing typical laminations that are truncated by a small NW-dipping welded fault; above the fault, bedding is tilted and brecciated (RB02-31; outcrop 12; bedding: 081/39S). (**f**) Silty to sandy, bioturbated level; some sandy levels have diffuse limits (up) whereas other have sharp limits (centre) (TD1257; outcrop 11; bedding: 317/26NE). (**g**) Bioturbated and slightly deformed laminated lithology, affected by a gently south-dipping welded detachment (upper right to lower left; RB02-29; outcrop 8; bedding: 351/10E).

Cleavage–fold relationships

General

Along the outcrop section, cleavage dip shows a large-scale change. In the northern outcrops (outcrops 1–12 and 14; Fig. 3), where the sheet dip is subhorizontal, cleavage generally dips to the north, whereas in the southern outcrops (outcrops 13 and 15–21; Fig. 3), where the sheet is gently south-dipping, cleavage dips to the south. Hence, the large-scale structure is that of a hectometre-scale gentle antiform, with a well-developed divergent cleavage fan and a hinge zone situated around the northern part of outcrop 12 and the southern part of outcrop 15 (Fig. 3). Within this gentle antiform, numerous metre-scale folds occur, as mentioned previously by several workers (e.g. Anthoine &

Anthoine 1943; Michot 1977). Significantly, these folds occur only in the northern outcrops. On the basis of the cleavage–fold relationships these can be divided into pre-cleavage folds and syn-cleavage folds.

Syn-cleavage folds

The syn-cleavage folds show a pronounced divergent cleavage fan, symmetrical about the fold hinges, with opposing senses of cleavage refraction in the two fold limbs. The folds are of decimetre- to metre-scale and usually have open interlimb angles. Most folds have a stepfold geometry with a marked south-verging asymmetry, relatively straight limbs, subangular to rounded hinges and moderately north-dipping axial surfaces (e.g. Fig. 5; see also Fig. 7c). These folds are comparable with fold types 2D and 2E of Hudleston (1973). Locally, less asymmetric folds are observed (e.g. outcrop 4), with two moderately dipping limbs and steeply north-dipping to subvertical axial surfaces. These folds, comparable with fold type 2C of Hudleston (1973), are usually better rounded than the stepfolds. In the stepfolds, cleavage usually dips to the north in both limbs, compatible with the generally north-dipping cleavage in the northern outcrops. However, because of the divergent cleavage fanning, in the moderately south-dipping limbs of the less asymmetric folds, locally south-dipping cleavage planes occur.

The fold hinge lines are subhorizontal to gently plunging, with a mean WNW–ESE trend (Fig. 6). However, a considerable variation in fold hinge-line orientation exists, showing a difference of 76° between the two most extreme plunge directions (076–256° and 152–332°). Outcrop observations show that this variation in fold hinge-line orientation occurs throughout the section. In some outcrops, neighbouring folds show markedly

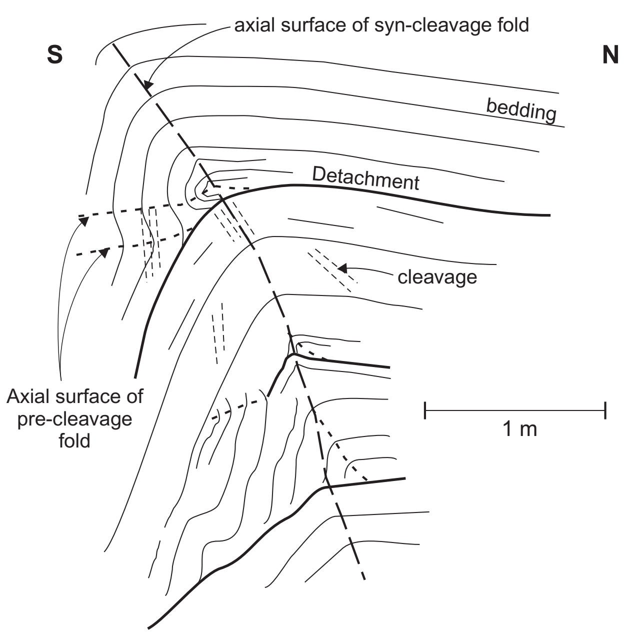

Fig. 5. Example of a syn-cleavage antiform (outcrop 6) superimposed on a pre-cleavage deformation zone. The syn-cleavage stepfold-like antiform, comparable with fold type 2D of Hudleston (1973), has a well-developed divergent cleavage fan, symmetrical about the fold hinge. In contrast, the small folds in its southern limb have an axial surface that is cut by the cleavage and hence have a pre-cleavage origin. The pre-cleavage folds are related to the pre-cleavage detachments.

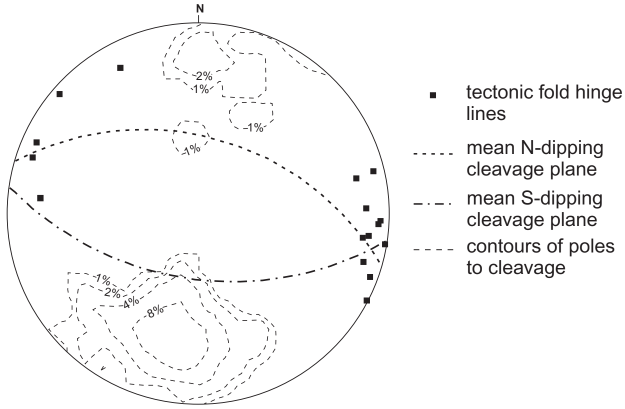

Fig. 6. Lower-hemisphere equal-area projection showing the orientation of the syn-cleavage fold hinge lines and the cleavage (mean cleavage planes and contours of poles to cleavage).

different plunge directions: a 33° difference in plunge direction occurs between the southernmost antiform of outcrop 3 and the northernmost antiform of outcrop 4, and a 25° difference in plunge direction occurs between the antiform and the synform in outcrop 11. To a large extent, this variation in plunge direction is reflected by changes in cleavage transection (Fig. 6). Although in fold profile the cleavage is axial planar (e.g. Fig. 5), an axial cleavage transection (*sensu* Johnson 1991) is common, both clockwise (up to 20°) and anticlockwise (up to 31°). Also, the plunge of the hinge lines varies slightly, not only between adjacent folds, but also within individual folds (e.g. outcrop 11). This reflects a periclinal fold shape.

Pre-cleavage folds

The pre-cleavage nature is demonstrated by the fact that cleavage is not axial planar to the folds and does not show a symmetrical fanning about the fold hinges, but crosscuts the axial surface and shows the same sense of cleavage refraction in both fold limbs (e.g. Fig. 7b, c, f-h). Obviously, this characteristic can be used only in those cases where cleavage is oblique to the axial surface of the pre-cleavage folds. However, because of the divergent cleavage fanning within the syn-cleavage folds, a pre-cleavage fold on the limb of a syn-cleavage fold, with the same axial surface orientation as the syn-cleavage fold, can still be crosscut obliquely by the cleavage, and hence can be recognized as a pre-cleavage feature (e.g. Fig. 7b).

The pre-cleavage folds have centimetre to metre sizes and interlimb angles ranging from close to gentle. 1 Both strongly asymmetric and more or less symmetric folds occur. The former all show a roughly south-verging asymmetry, ranging from SW-verging to SE-verging. Some pre-cleavage folds have fold shapes comparable with those of the syn-cleavage folds. Others, however, have different fold geometries, resembling types 1C, 1D, 2F, 3C and 3D of Hudleston (1973).

The pre-cleavage folds exhibit a significant spread in orientation (Fig. 8). The plunge ranges from subhorizontal to steeply plunging and a difference of almost 90° exists between the two most extreme plunge directions. Some axial surfaces are markedly oblique to the main cleavage trend (101–281°), whereas others are more or less parallel to the main cleavage trend.

Apart from the pre-cleavage nature, and the stronger geometric variation with respect to the syn-cleavage folds, another characteristic feature of the pre-cleavage folds is their common,

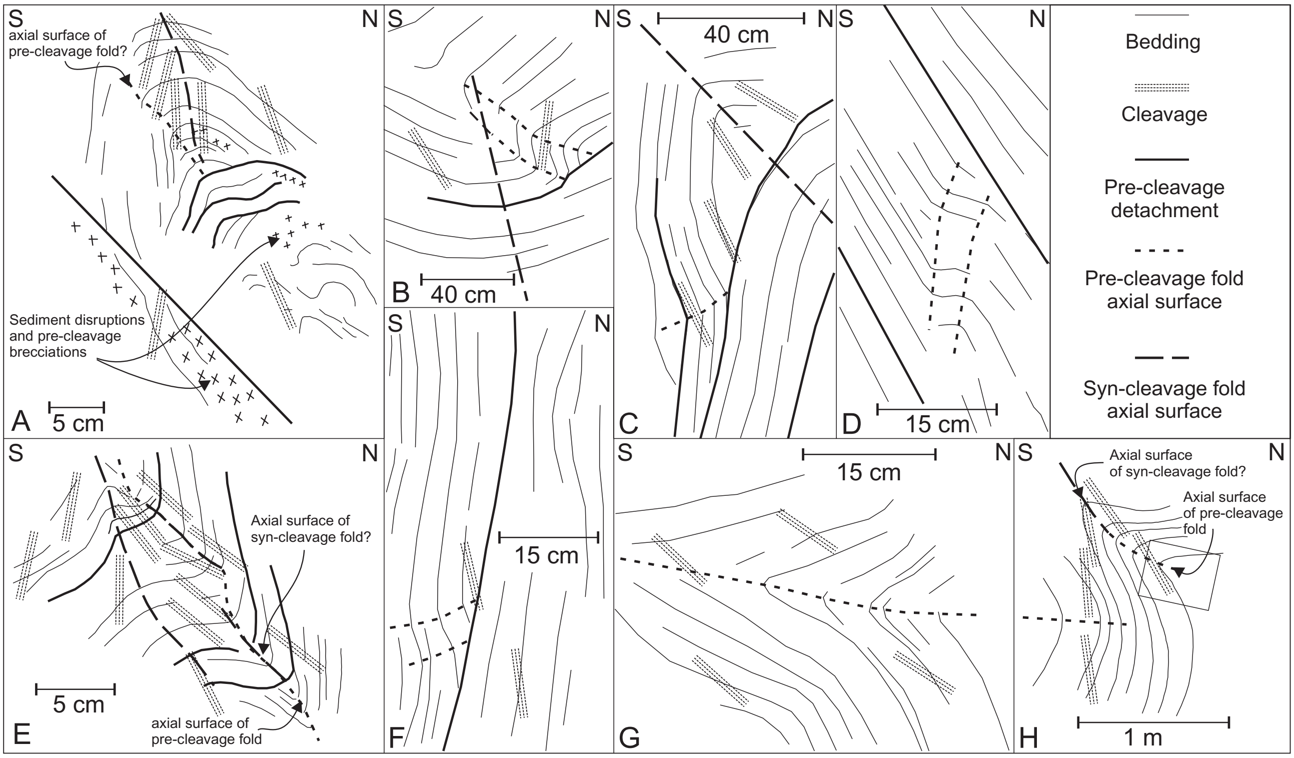

Fig. 7. Pre-cleavage deformation structures in the Abbaye de Villers Formation and their relationship with syn-cleavage deformation structures. (**a**) Pre-cleavage deformation structures (folds, detachments and related soft-sediment deformation), affected by a syn-cleavage antiform, with a well-developed divergent cleavage fan (outcrop 12, lower part). (**b**) Pre-cleavage fold pair and related pre-cleavage detachment, affected by a syn-cleavage synform (outcrop 4, lower, northern part). Cleavage shows a divergent fan, symmetrical about the syn-cleavage synform axial surface, but crosscuts the axial surface of the pre-cleavage fold pair. (**c**) Stepfold-like syn-cleavage antiform, with small pre-cleavage fold in its steep limb and related pre-cleavage detachments (outcrop 6, southern part; see Fig. 5). The cleavage shows a well-developed divergent fan in the syn-cleavage fold, but crosscuts the pre-cleavage fold. (**d**) Isolated fold pair, separated from overlying and underlying unfolded beds by pre-cleavage detachments (outcrop 12, central, lower part). (**e**) Syn-cleavage folds superimposed on pre-cleavage folds and deforming pre-cleavage detachments (outcrop 12, central, lower part). (**f**) Pre-cleavage fold pair and related detachment above unfolded beds (outcrop 3, northern, lower part). (**g**, **h**) Relationship between cleavage and pre-cleavage folds. Cleavage is only slightly oblique to the axial surface of the pre-cleavage antiform, but is almost perpendicular to the axial surface of a gentle pre-cleavage synform associated with this antiform. Towards higher levels, the cleavage–fold relationships seemingly suggest a syn-cleavage origin (outcrop 12, lower, northern part).

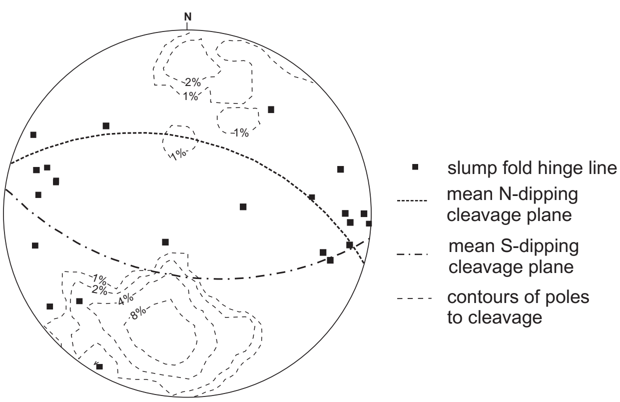

Fig. 8. Lower-hemisphere equal-area projection showing the orientation of the pre-cleavage fold hinge lines and the cleavage (mean cleavage planes and contours of poles to cleavage).

isolated, intraformational position between 'non-folded' beds (i.e. not folded by pre-cleavage folds) and their close association with pre-cleavage detachments (Fig. 7d and f). The term 'detachment' is used here for any pre-cleavage truncational surface with a low bedding cut-off angle, irrespective of whether or not there is actual evidence of slip. Hence it includes surfaces of erosional truncation. The pre-cleavage detachments are often welded, and often result in a stacking of sequences, leading to strong local thickness changes (Figs 5, 7a, c, e, and 9). In some cases, the detachments truncate the pre-cleavage folds (Figs 7e and 9a), whereas in other cases they are folded by the pre-cleavage folds (Fig. 9a). However, in all cases, the detachments are folded by the syn-cleavage folds, thus implying a pre-cleavage origin. Ideally, the pre-cleavage nature of the detachments and, where present, their associated breccias, is demonstrated by the cross-cutting relationship of the cleavage. Often, however, their pre-cleavage nature is reflected by less obvious features such as the association with zones of pre-cleavage, internal deformation of

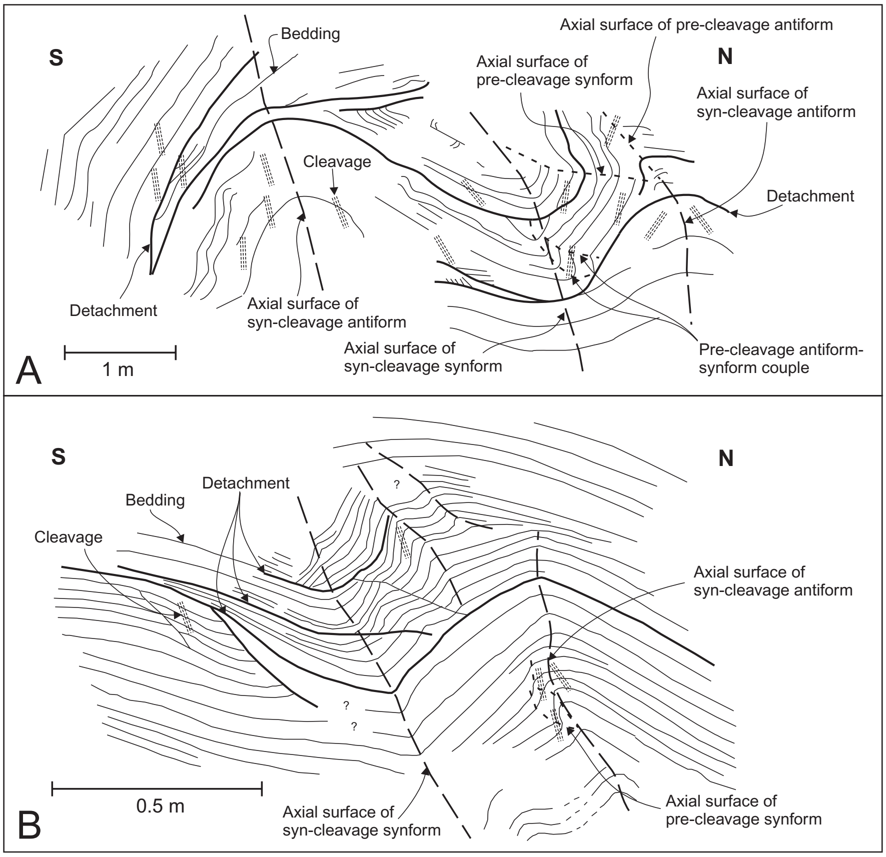

Fig. 9. Examples of the spatial relationship between syn-cleavage folds and pre-cleavage deformation structures: (**a**) outcrop 4; (**b**) outcrop 11. The syn-cleavage folds occur superimposed on pre-existing pre-cleavage folds, often amplifying these, and on zones of abrupt thickness changes related to pre-cleavage detachments.

the sediments (e.g. zones in which the sand layers are disrupted into small sand lenses within a clay matrix or zones with abrupt terminations of individual layers).

Geological significance of the pre- and syn-cleavage folds

The metre-scale syn-cleavage folds, as well as the hectometre-scale gentle host antiform, are all of tectonic origin and are attributed to the Brabantian deformation phase (see Debacker 2001; Verniers *et al*. 2002; Debacker *et al*. 2005). Taking into account the asymmetry (S-shaped) of the metre-scale syn-cleavage folds and their position within the subhorizontal limb of the syn-cleavage host antiform (Fig. 3), they probably represent parasitic folds related to this antiform. In this respect, and taking into account a south-dipping cleavage in the southern antiform limb, one would expect comparable folds, with an opposing asymmetry (Z-shaped) in the southern antiform limb (southern part of Fig. 3). However, such folds have not been observed.

The pre-cleavage folds either formed during an older tectonic deformation phase or are a result of slumping (see Debacker *et al*. 2001). In the study area, as well as in the other outcrop areas of the Brabant Massif, there is no evidence for more than one tectonic deformation phase (e.g. Sintubin 1999; Debacker 2001; Verniers *et al*. 2002; Debacker *et al*. 2005, and references therein). In contrast, the pre-cleavage folds have characteristics commonly attributed to slump folds. These are the isolated intraformational position between non-folded beds (e.g. Fig. 7d), the truncation of folds by overlying, younger beds (Figs 7e and 9a), the dispersed orientation of the fold axes (Fig. 8), the often irregular fold shape (Fig. 9), the association with other soft-sediment deformation features such as welded detachments, welded faults and disrupted sediments, the absence of fold- or detachment-related veins or cleavage and the parallel or southward downcutting nature of the detachments with respect to underlying beds (see Jones 1939; Kuenen 1949; Helwig 1970; Corbett 1973; Rupke 1976; Woodcock 1976; Elliott & Williams 1988). For these reasons, the pre-cleavage folds are interpreted as slump folds. By means of the mean axis method (Jones 1939) and the separation arc method (Hansen 1965), the asymmetry of the slump folds was used to deduce the sense and direction of slumping and the probable strike of the corresponding palaeoslope (see Woodcock 1979). Both methods give similar results and suggest slumping from north (NNW) to south (SSE), and hence a probably south-dipping, east-west-trending palaeoslope (Beckers 2003, 2004).

Relative position of the pre- and syn-cleavage folds

The syn-cleavage folds and the pre-cleavage folds show a close spatial relationship. Not only do pre-cleavage folds often occur in the hinge zones of the syn-cleavage folds (e.g. Fig. 7b, c, e), but, more importantly, of the observed metre-scale syn-cleavage fold pairs (synform–antiform couple), at least one fold always coincides with smaller pre-cleavage folds and thickness changes related to pre-cleavage detachments (Figs 7a-c, e, and 9a, b). In Figure 5 (see Fig. 7c), for instance, the axial surface of the syn-cleavage antiform runs along the hinges of several pre-cleavage antiforms, probably formed by movement along, and stacked on top of one other by, pre-cleavage detachments. Similarly, in

Figure 7a, the syn-cleavage antiform occurs superimposed on a zone of strong pre-cleavage deformation. In addition, it appears that the syn-cleavage antiforms tend to coincide with pre-cleavage antiforms and zones with an abrupt thickness increase as a result of stacking along detachments, whereas syn-cleavage synforms tend to coincide with pre-cleavage synforms and zones of abrupt thickness decrease. The syn-cleavage antiforms in Figure 5 (see Fig. 7c), Figure 7a, h, and in the northern part of Figure 9a all occur superimposed on pre-cleavage antiforms or on zones characterized by a significant local thickening as a result of pre-cleavage deformation. Similarly, the syn-cleavage synforms in Figure 7e, Figure 9a (see Fig. 7b) and Figure 9b occur superimposed on a pre-cleavage synform or on zones characterized by a significant local thinning caused by pre-cleavage deformation.

Also on a large scale, an apparent spatial relationship exists between the pre-cleavage and syn-cleavage metre-scale folds. Both occur within the same part of the studied outcrop section. Because they result from slumping, the pre-cleavage folds should be restricted to particular stratigraphic levels, in this case the older unit of the Abbaye de Villers Formation (Fig. 3). However, the syn-cleavage folds, being of tectonic origin, seem also to be related to this level. Considering the common scarcity of meso-scale tectonic folds in the limbs of the large-scale host folds in the Ordovician and Silurian sequences of the Brabant Massif (Debacker *et al*. 1999, 2001, 2005; Debacker 2001), and the paucity of meso-scale folds in the Ordovician sequences of the Thyle valley, representing the subhorizontal limb of a kilometre-scale stepfold (Herbosch *et al*. 2002), it is surprising to find such a high local concentration of metre-scale tectonic folds along the northern part of the studied section. In addition, as pointed out above, although these folds probably represent parasitic folds related to the hectometre-scale host antiform, they are observed only within its subhorizontal limb (Fig. 3). In this respect, outcrop 13 deserves special attention. This is the only outcrop with pre-cleavage folds observed in the southern limb of the hectometre-scale antiform (Fig. 3). In this outcrop, decimetre- to metre-scale south-verging pre-cleavage folds occur within a 2 m wide zone between undeformed, gently south-dipping beds. No syn-cleavage folds occur within this outcrop.

Comparison with the Abbaye de Villers Formation in the Sennette valley at Virginal

In the railway section at Virginal, in the Sennette valley, *c*. 20 km to the west of the Thyle valley (Fig. 1), the lower part of the Abbaye de Villers Formation occurs in a subvertical to steeply SW-dipping, SW-younging limb of a hectometre-scale fold. Along this outcrop section, the upper parts of the Abbaye de Villers Formation and the entire Tribotte Formation are removed by faulting (Fig. 10; compare Debacker *et al*. 2004*a*, fig. 9).

Within this lower part of the Abbaye de Villers Formation, a *c*. 30 m thick zone occurs with metre-scale pre-cleavage folds and related pre-cleavage deformation structures (detachments, brecciations), which, using the same argument as above, have been attributed to slumping (Debacker 2001; Debacker *et al*. 2003). Although having an identical lithostratigraphic position to the level studied in the Thyle valley, it cannot be ascertained whether both levels have an identical age. Characteristically, bedding in this zone of pre-cleavage deformation is oriented *c*. 020° clockwise with respect to the regional trend (Fig. 10). Because of the steep bedding dip, and the oblique orientation with respect to the regional trend, the inferred slump direction varies significantly with the chosen values of the regional fold axis and mean bedding orientation. A northern, northeastern or eastern slump source is inferred (Debacker, unpubl. data), being compatible with the northern slump source inferred in the Thyle valley (see Beckers 2003, 2004).

Significantly, within this steep limb there are no syn-cleavage folds that show a spatial relationship with individual pre-cleavage deformation structures. The only observed syn-cleavage fold pair, a metre- to decametre-scale open, rounded antiform–synform couple, occurs at the southern, upper limit of the pre-cleavage deformation zone (Fig. 10) and cannot be linked to individual pre-cleavage deformation structures. Hence, this section markedly contrasts with the section studied in the Thyle valley. First,

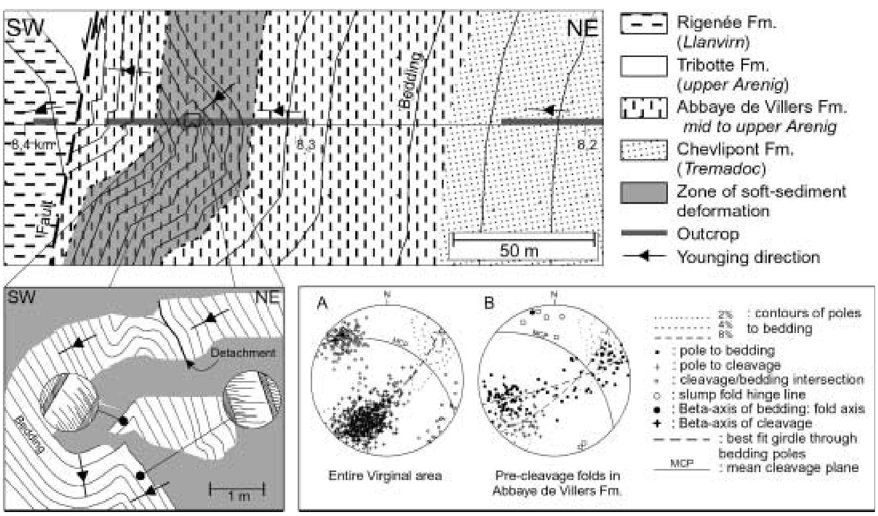

Fig. 10. The soft-sediment deformation level in the lower part of the Abbaye de Villers Formation in the railway section at Virginal, Sennette valley (see Fig. 1 for location; after Debacker *et al*. 2003, 2004*a*). The lower left inset shows the cleavage–bedding relationships within a pre-cleavage fold pair (after Debacker *et al*. 2003). The lower right frame shows lower-hemisphere equal-area projections: projection A shows bedding, cleavage and cleavage–bedding intersections associated with syn-cleavage folds from the entire Virginal area; projection B shows bedding and fold hinge lines of pre-cleavage folds as well as cleavage within the soft-sediment deformation zone in the Abbaye de Villers Formation. For comparison, bedding pole contours of projection A are added as a grey background in projection B. The different orientation of the pre-cleavage folds (projection B) with respect to the (syn-cleavage) regional trend (projection A) should be noted.

it complies with the general situation of the Ordovician and Silurian sequences of the Brabant Massif, in which the limbs of large-scale fold structures are virtually free of parasitic fold structures. Second, it does not show a clear spatial relationship between individual pre-cleavage deformation structures and syn-cleavage folds.

Discussion

One of the most cited characteristics of slump folds is the dispersed fold-axis orientation, even from within single slump sheets (Helwig 1970; Lajoie 1972; Woodcock 1976, 1979). In the soft-sediment deformation level in the Thyle valley, however, the tectonic fold orientations also exhibit a considerable spread. Taking into account observations in other Ordovician and Silurian outcrop areas of the Brabant Massif (e.g. Fig. 10), it appears that the spread in tectonic fold hinge-line orientations in this part of the Abbaye de Villers Formation is remarkably high (see Debacker *et al*. 1999, 2004*a*, *b*; Debacker 2001). It is likely that this is related to the variable orientations of the slump folds.

The spatial relationship between the two fold types and the apparent influence of the slump fold orientations on the tectonic folds suggest that it is the presence of slump folds that controls the occurrence of the metre-scale tectonic folds. Also, the apparent stratigraphically restricted occurrence of the metre-scale tectonic folds in the subhorizontal limb of the hectometre-scale gentle host antiform in the Thyle valley (Fig. 3; outcrops 1–12 and 14) may be related to the presence of the soft-sediment deformation structures. Alternatively, one might also consider lithological differences between the lower unit and the upper unit of the Abbaye de Villers Formation as an explanation for the apparent stratigraphically restricted occurrence. However, the differences between these units are not sufficient to explain the total absence of parasitic folds in the upper unit compared with the abundance of metre-scale parasitic folds in the lower unit (Fig. 4). In addition, this cannot explain the apparent absence of metre-scale tectonic folds in the lower unit in the southern, south-dipping antiform limb (outcrop 13).

As pointed out above, in outcrop 13 in the Thyle valley (Fig. 3), and in the railway section at Virginal in the Sennette valley (Fig. 10), there are no tectonic folds that can be linked to individual slump features. Possibly, this is a result of the relative asymmetry of the tectonic folds and the slump folds. As shown by experiments and numerical models, the influence of a perturbation on folding and on final fold geometry depends not only on strain, strain rate and material properties, but also on the spacing, size and asymmetry of the perturbations (Lewis & Williams 1978; Abbassi & Mancktelow 1990; Mancktelow 1999; Zhang *et al*. 2000; Williams & Jiang 2001; Jeng *et al*. 2002). Because of the northern slump source, the slump folds have a similar asymmetry to the metre-scale tectonic folds in the northern, subhorizontal limb of the hectometre-scale, tectonic antiform (Fig. 11a; compare Fig. 3). In this limb the tectonic fold axial surfaces will at least partly coincide with the axial surfaces of the pre-cleavage folds at which they originate (e.g. Figs 5 and 7c, e, h, and the northern parts of Fig. 9a and b). In the south-dipping, southern antiform limb, however, the slump fold asymmetry will oppose that of the expected metre-scale tectonic folds (outcrop 13 in Thyle valley and Virginal railway section in Sennette valley). Hence, we suggest that, within the Abbaye de Villers Formation, the parasitic tectonic folds develop on slump folds only in those cases where the slump fold asymmetry matches the asymmetry of the parasitic tectonic folds.

Whether or not a specific slump fold or slump-related irregularity will give rise to a tectonic fold depends also on the distance between the adjacent slump folds, relative to the dominant wavelength of the tectonic folds (Abbassi & Mancktelow 1990; Mancktelow 1999; Williams & Jiang 2001). In Figure 11b, if the half-wavelengths of the tectonic folds were similar to the spacing between the synform and antiform of each slump fold pair, a tectonic fold would be expected in the four positions marked (two synforms and two antiforms). However, often only one fold of each slump fold pair develops into a tectonic fold. In Figure 5 and the northern part of Figure 9a and b, a tectonic antiform forms along the antiform of the slump fold pair, whereas the adjacent synformal slump fold is not used. Similarly, the tectonic synform in Figure 9a develops predominantly along the synform of the slump fold pair, seemingly without any influence of the adjacent antiformal slump fold. In addition, once a tectonic antiform (synform) initiates on a pre-existing slump fold or related structure, which forms a suitable perturbation, an adjacent tectonic synform (antiform) will also develop, the position of which may be influenced more by the dominant fold wavelength and by the development of the adjacent antiform (synform) than by the presence of a perturbation. Possibly, this is what happens in Figure 9b. Unlike the adjacent tectonic synform, which has a very distinct hinge zone centred on a significant pre-cleavage deformation zone, the tectonic antiform apparently

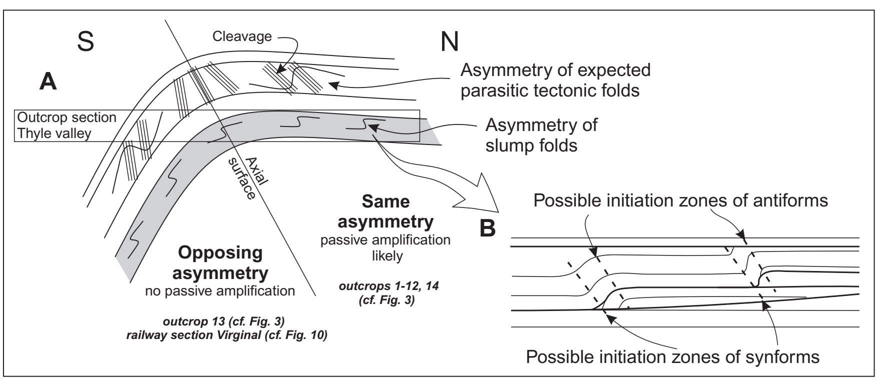

Fig. 11. Conceptual representation of (**a**) the asymmetry of the slump folds and the expected tectonic folds throughout a hectometre-scale south-verging antiform in the Abbaye de Villers Formation (see Figs 3 and 10) and (**b**) the slump-related deformation geometry in the subhorizontal, hectometre-scale fold limb in the Thyle valley, with indication of the possible initiation zones for tectonic folds. In the southern antiform limb, passive amplification of slump folds during tectonic shortening is unlikely, because of the high angle between the slump fold axial surfaces, on the one hand, and the cleavage and the expected tectonic fold axial surfaces, on the other hand. (See text for discussion.)

developed on a pre-cleavage fold of very limited extent and, although the tectonic antiform shape seems to continue towards higher levels, its axial surface is irregular and very difficult to trace. Possibly, this antiform formed primarily as a result of the adjacent synform. The above observations indicate that the dominant wavelength of the tectonic folds is larger than that of many of the slump folds, and suggest that, besides perturbation asymmetry, size and spacing, the material properties also play a significant role.

Finally, the question can be raised of whether the tectonic folds entirely result from active folding after initiation on slump-related perturbations, or whether a significant amount of passive amplification of pre-existing slump folds was involved. On the one hand, during layer-parallel tectonic shortening and cleavage development, a further, passive tightening and amplification can be expected for slump folds of which the axial surface is subparallel to the cleavage (e.g. synform in Fig. 7e). Considering the similar asymmetry of the slump folds and the tectonic folds in the subhorizontal hectometre-scale antiform limb in the Thyle valley, and the observation that tectonic antiforms (synforms) tend to develop on antiformal (synformal) slump folds, passive amplification is likely to have occurred. On the other hand, however, the marked divergent cleavage fanning implies active folding. Also, the size difference between the tectonic folds and many of the slump folds suggests that active folding took place. Hence, it is likely that both phenomena occurred, their relative importance varying from fold to fold. Interestingly, although, regardless of their asymmetry, the slump folds in the southern, south-dipping antiform limb do represent an irregularity of comparable size to the other slump folds and hence are potential perturbations during tectonic shortening, no tectonic folds are observed that can be linked to individual soft-sediment deformation structures. Because of their asymmetry, opposing that of the expected tectonic folds, passive amplification of these slump folds cannot occur. Possibly, the absence of tectonic folds at this locality is a direct result of the absence of passive amplification. Conversely, it may be possible that within the Abbaye de Villers Formation, because of the strain, strain rate and material properties, the process of tectonic fold development on pre-existing slump folds initiates by means of passive amplification, and only later changes into active folding.

Conclusions

In the southern Brabant Massif, both pre-cleavage and metre-scale syn-cleavage folds occur within a particular stratigraphic level of the Lower Ordovician Abbaye de Villers Formation. The pre-cleavage folds and related structures are attributed to slumping from a northern source, whereas the syn-cleavage folds formed during the Brabantian deformation phase (late Llandovery–Emsian).

Both the position and, to a certain extent, the geometry of the metre-scale tectonic folds appear to be controlled by the slump folds. Hence, the slump folds are considered as perturbations at which the tectonic folds initiated. However, not all soft-sediment deformation structures give rise to tectonic folds. In those cases where the slump folds have an asymmetry opposing that of the expected tectonic folds, there are no tectonic folds that can be linked to individual soft-sediment deformation structures. In addition, even though having a similar asymmetry to the expected tectonic folds, not every fold of an antiform–synform slump fold pair gives rise to a tectonic fold and apparently the tectonic fold forms on the more 'suitable' soft-sediment deformation structure. Some observations also suggest that, once a tectonic fold forms on a suitable slump structure, an adjacent fold develops, the position of which may or may not be controlled by pre-existing slump structures. Probably, the latter depends on whether or not the pre-existing perturbation is more important than the material properties controlling the dominant wavelength of the tectonic folds. The fact that the dominant wavelength of the tectonic folds is larger than that of the slump folds suggests that, although initially the perturbations (slump features) appear to control the initiation of the tectonic folds, during tectonic fold amplification the material properties become more important. This is in agreement with the results of experiments and numerical models, which indicate that the influence of a perturbation on folding and on final fold geometry depends not only on strain, strain rate and material properties, but also on the perturbation spacing, size and asymmetry (Lewis & Williams 1978; Abbassi & Mancktelow 1990; Mancktelow 1999; Zhang *et al*. 2000; Williams & Jiang 2001; Jeng *et al*. 2002).

Although in the specific case of the relationship between perturbation and buckle folding, experiments and numerical models may adequately describe and predict the development of buckle folds at particular localities, there are, judging from the geological literature, very few documented natural examples of this. Taking into account the common occurrence of slump folds and related features, the rather irregular 3D nature of sedimentary sequences and the abundance of lenticular sedimentary (e.g. channels) or volcanic bodies (e.g. rhyolitic lava flows), many more cases are expected in which individual natural buckle folds may be linked to specific perturbations. Probably, this discrepancy is a result of the complexity of geological materials, as compared with the materials used in numerical and experimental studies. This complexity of natural layer systems, in which a large number of possible perturbations coexist, makes it difficult to link a specific natural buckle fold to a specific irregularity. In addition, because of the degree of exposure and relative scale of the folds, the natural perturbation usually remains unknown.

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1. Should 'close' be changed to 'steep'? (Or 'closed'?)
2. Debacker et al., submitted - please update if possible.
3. Fleury - where in text should this ref. be cited