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Frontal Foothills structures in central Alberta: the thin end of the intercutaneous wedge?

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ABSTRACT

The para-autochthonous section northeast of the surface expression of the triangle zone in west-central Alberta is cut in places by hinterland-vergent backthrusts and by more steeply dipping (*ca.* 40°) foreland-directed reverse faults. The geometry of the thrust faulting above the upper detachment can be mechanically explained, in principle, by a model in which the para-autochthonous section is underlain by a hinterland-verging detachment. Blind duplexes observed within the foreland basin are interpreted as part of an extremely low-angle intercutaneous wedge that extends into the foreland from the triangle zone. Given this interpretation, geometric relationships between the faulting observed above and below the upper detachment show that these structures could not have formed in sequence. Alternative models that do not involve a low-angle intercutaneous wedge require substantial layer-parallel shortening above the upper detachment and, so far, lack observational support.

RÉSUMÉ

La section para-autochtone située au nord-est de l'affleurement de la zone triangulaire des Foothills du centre-ouest de l'Alberta est interrompue par endroits par des rétrochevauchements ainsi que par des failles inverses à pendage plus accentué (*c.* 40°) orientées vers l'avant-pays. La géométrie des chevauchements situés au-dessus du décollement supérieur pourrait, en principe, s'expliquer au moyen d'un modèle selon lequel il y aurait, en-dessous de la section para-autochtone, un décollement à vergence vers l'arrière-pays. On propose que les duplex aveugles présents dans le bassin de l'avant-pays font partie d'un prisme intercutané extrêmement effilé qui s'étend dans l'avant-pays à partir de la zone triangulaire. Selon cette interprétation, les rapports géométriques existant entre les failles observées au-dessus et au-dessous du décollement supérieur montrent que ces structures n'ont pu se former en succession normale. Les modèles ne faisant pas intervenir de prisme intercutané effilé impliquent obligatoirement un raccourcissement parallèle de couche considérable au-dessus du décollement supérieur, et n'ont jusqu'à présent été corroborés par aucune observation empirique.

Traduit par Annick Geoffroy-Skuce

INTRODUCTION

The geometry of the Alberta triangle zone has been studied in great detail over the last ten years or so; its internal structure is more complex than was originally thought and significant variations have been observed along the length of the Foothills Belt (Spratt and Lawton, 1996). A common feature in all published interpretations of triangle zones is the wedge of rock that is thrust between the passively uplifted para-autochthonous section above the upper detachment and the undeformed autochthonous rocks below the lower detachment, *viz.*, the intercutaneous wedge (Fig. 1). The angle of the wedge is highly variable, both within and between individual cross-sections. So far, no fully satisfactory mechanical model has been established to account for the geometry of the wedge and it seems to be the case that the models that have been proposed have increasing difficulty in accounting for lower-angle wedges. Particularly awkward to explain are the extremely low-angle wedges, in which upper and lower detachment surfaces having opposite vergence may be parallel over tens of kilometres. In Canada, such structures have been proposed in northeast British Columbia (McMechan, 1985), the Parry Islands fold belt (Harrison and Bally, 1988) and west-central Alberta (Skuce *et al.* 1992). Elsewhere, continuous hinterland-vergent detachments more than 100 km long above highly deformed thrust wedges have been interpreted in Pakistan (Jadoon *et al.* 1994) and Papua New Guinea (Hobson, 1986). The basis for this kind of interpretation is principally geometric; the blind wedges appear to be internally shortened considerably more than the sequences above and below them. The apparent mechanical implausibility of these interpretations causes, in many other geologists, reactions ranging from scepticism to outright disbelief.

The question addressed in this article is not so much whether intercutaneous wedges exist at triangle zones but, rather, how

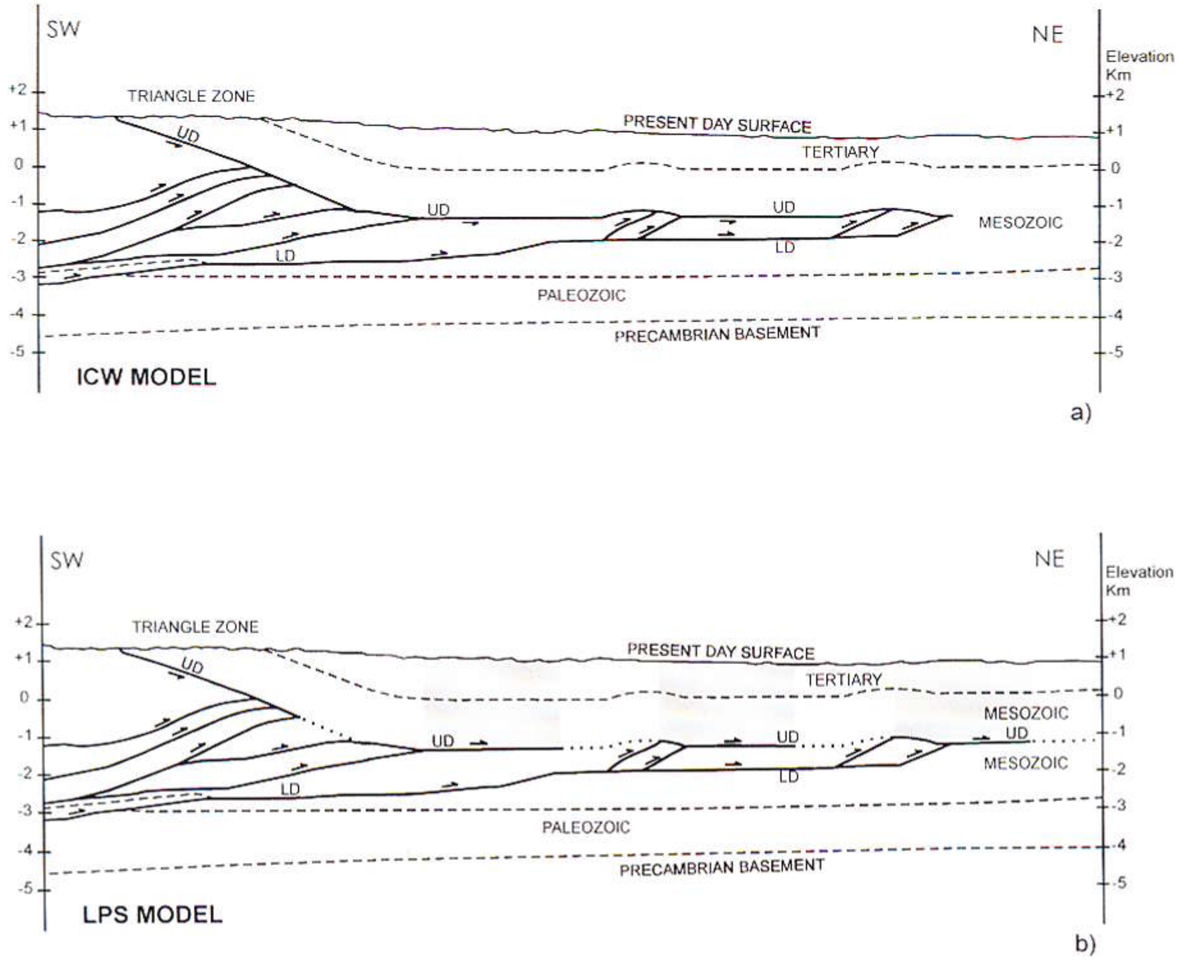

Fig. 1. Schematic cross-sections showing alternative interpretations of the relationship between the triangle zone and blind duplexes located farther into the foreland. Bold lines represent faults with the relative displacement direction shown by the arrows. **a)** The intercutaneous wedge (ICW) is the body of rock situated between the hinterland-verging upper detachment (UD) and the foreland-verging lower detachment (LD). The section above the UD has not been displaced laterally by any significant amount relative to the basement and is referred to in the text as the para-autochthonous section. **b)** Displacement above the upper detachment is taken up by distributed strain (layer-parallel shortening; LPS). This strain would be greatest above and to the northeast of the blind duplexes in areas shown by the grey tone. Note that the defining features of the models are not exclusive; the ICW model could easily incorporate some distributed strain and the LPS model features a (high-angle) intercutaneous wedge. The critical difference between the models is the direction of displacement on the upper detachment outboard of the triangle zone.

acute can the wedge angle be? A critical element in this discussion is the nature and extent of the deformation that has occurred in the para-autochthonous section above the intercutaneous wedge (Fig. 1). This article will show some examples of structures in west-central Alberta from the foreland side of the surface expression of the triangle zone that may throw some light on this issue. First, however, current theoretical models of thrust wedges will be reviewed in order to evaluate their applicability in constraining interpretations of frontal structures.

Theoretical Models

The critical-wedge model of Davis *et al.* (1983) has provided insight into the gross macroscopic mechanics of accretionary prisms as well as foreland fold-and-thrust belts. It has been successful in quantitatively characterizing the regional wedge geometry of these belts. However, such a model does not (nor did its authors intend it to) describe the complex internal structures of thrust belts. The model is founded on the analogy between the formation mechanism of mountain ranges and the wedge of soil in front of an advancing bulldozer blade; this image is now commonplace in geological textbooks. Despite its success, the applicability of the theory to foreland belts has been challenged on geological grounds by Woodward (1987) and on theoretical grounds by Bombalakis (1994).

Jamison (1993) applied the critical-wedge model to what he termed "the backthrust wedge" (*i.e.*, the para-autochthonous rocks above the upper detachment) at triangle zones. The critical wedge in this case is modelled as a hinterland-vergent mass of rock. He determined a range of stable wedge geometries for varying boundary conditions such as basal detachment angle, coefficient of internal friction and pore-pressure ratio. The precision of the analysis is limited by uncertainties in the topographic profile at the time of deformation and by the amount of isostatic rebound that has occurred due to subsequent erosional unloading. This model, it should be noted, provides no constraints on the shape of the intercutaneous wedge, only on the dip of the upper detachment. A valuable inference that can be drawn from Jamison's analysis, however, is that overpressured rocks are probably required at the upper detachment to account for geometries in which this surface dips at angles of less than about 12° towards the foreland. Such cases are commonly observed (*e.g.*, Teal, 1983; MacKay, 1991; Lawton *et al.*, 1994). More recent work (Jamison, 1996), using finite-element modelling techniques, improves the constraints on the required boundary conditions for a range of realistic triangle zone geometries.

The upper surface of a critical thrust wedge is assumed to be a free surface where principal stresses normal to the surface and shear stresses parallel to it are zero. Any constraints on wedge geometry calculated on this basis should not be expected to apply to intercutaneous wedges where the upper surfaces are subsurface detachments with hinterland vergence. An attempt to describe mechanically the intercutaneous wedge itself has been made by Charlesworth *et al.* (1987) who proposed that the upper and lower boundaries of the wedge were conjugate shear planes and that the angle that forms between these planes should lie in the range of 30° to 40°. However, observed angles between the upper and lower detachments are commonly less than 30° and, in some cases, are much less.

A simplifying assumption made in all theoretical attempts to constrain wedge geometries is that the rocks involved are isotropic and homogeneous. In real thrust structures, the layering of rocks with very different mechanical properties (*i.e.*, competent, incompetent and zones of "superweakness"; Gretener, 1972) exerts a profound influence on the distribution and geometry of detachments and thrust ramps. Price (1986, 1994) stated that a necessary condition for tectonic wedging and delamination is a strongly layered and anisotropic body of rock. D.C. Lawton (*pers. comm.*) has pointed out that triangle zones do not seem to form in accretionary prisms, perhaps because such sequences do not possess the varied and strongly layered sedimentary units typical of foreland fold-and-thrust belts. Therefore, models that exclude the mechanical anisotropy introduced by sedimentary layering, a necessary condition for the formation of triangle zones, can hardly be expected to explain the origin or to constrain the geometry of such structures.

One possible reason that triangle zones are better developed in Alberta than in many other parts of the margins of the Rocky Mountains and other thrust belts is that the rocks comprising the Alberta Foothills were essentially undeformed prior to the compressional episode. Elsewhere, continental margins and foreland basins are commonly faulted by detached and basement-involved normal faults as well as by strike-slip faults. Had they occurred, such faults would have disrupted and offset any planes of weakness, thereby preventing the formation of the long, continuous detachments and low-angle wedge structures that are typical of the frontal portion of the fold-and-thrust belt in Alberta.

Seismic Examples

Edson Example

Skuce *et al.* (1992) described some structures in the Edson area (Figs. 2, 3), interpreting them as blind duplexes that had formed between parallel detachments which can be correlated to the detachments in the triangle zone some 30 km to the southwest. This interpretation calls for a wedge 400 m thick and 40 km long to have been thrust several hundreds of metres between essentially uncontracted units above and below it. Skuce *et al.* (1992) suggested that this very thin wedge was emplaced through lateral hydraulic fracturing by overpressured shales, akin to the intrusion of an igneous sill. Dechesne (1994) described evidence from cores in the Second White Speckled Shale (the level of the lower detachment at Edson) in the same general area that shows that these rocks were once overpressured.

Finite-element modelling shows that the coefficient of friction on the bounding detachments of a low-angle intercutaneous wedge must be very low (Jamison, 1996). This implies that pore pressures must be very high. Such pressures have not been observed in an active foreland belt but Moore *et al.* (1995) reported *in situ* measurements of pore pressures close to the lithostatic pressure in an active detachment zone in the Barbados accretionary prism. These very high pore pressures are apparently caused by lateral influx of pore fluids, causing dilation. This (nonintercutaneous) prism has a very low taper angle of about 2.5°, indicating that its basal detachment zone has a very low effective strength.

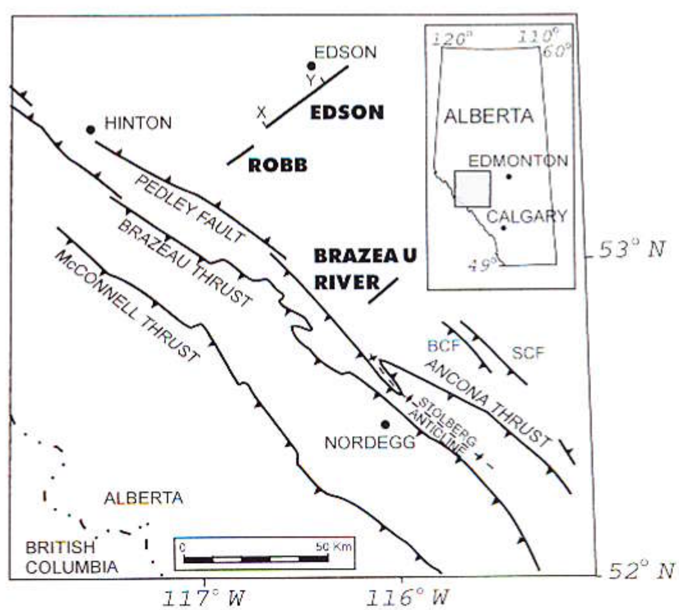

Fig. 2. Index map showing the locations of the Edson, Robb and Brazeau River examples discussed in this paper. Surface outcrops of some major thrusts (barbed lines) are taken from Price *et al.* (1977). The barbs are on the thrust hanging walls. BCF = Brewster Creek Fault; SCF = Sylvester Creek Fault (Jones, 1971). X and Y are the limits of the seismic line shown in Figure 3. The rectangular area on the inset shows the map area.

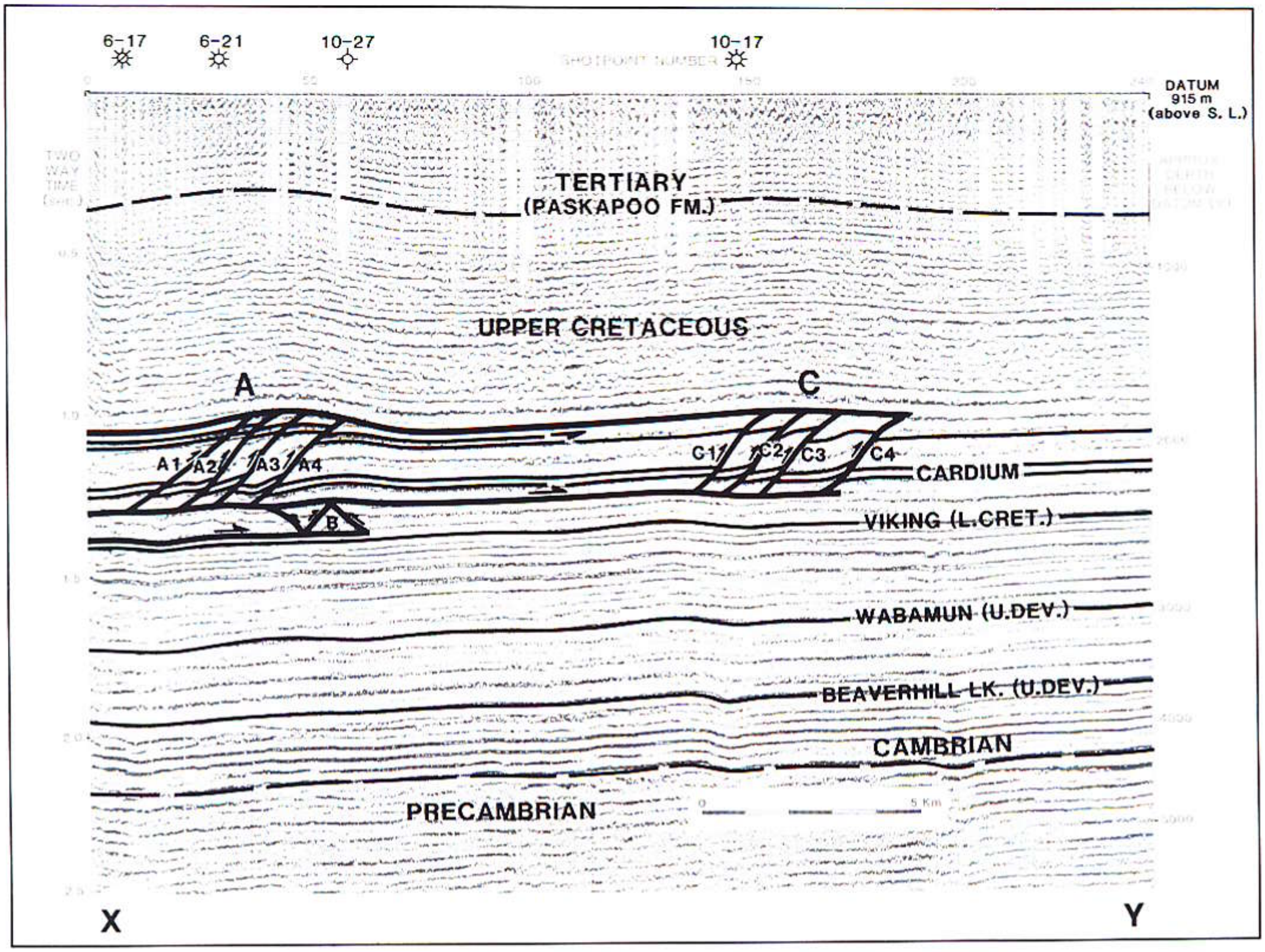

Fig. 3. Interpreted seismic section near Edson (see Fig. 2 for location). Note that the vertical exaggeration on this display averages about 3.5:1. The three wells above duplex A are located in Township 51-18W5M and all penetrate as deep as the Beaverhill Lake Formation. Well 10-17-52-17W5M penetrates duplex C and terminated more than 200 m below the top Viking marker. See Skuce *et al.* (1992) for more details.

The interpretation that the upper surface of the wedge at Edson is a hinterland-vergent detachment was based on: a) the absence of any evidence of shortening in the units above it, and b) its continuity with detachments that have well-established hinterland vergence in the triangle zone to the southwest. Alternatively, it has been proposed by Dechesne (1994) and B. Veilleux and H.A.K. Charlesworth (*pers. comm.*) that substantial layer-parallel shortening has occurred above the upper detachment outboard of the triangle zone, obviating the need to interpret a hinterland-vergent detachment there (Fig. 1b). Because such shortening could be spread over a large distance it need not manifest itself in any obvious or even observable way. Implicit in such models is that the lower detachment, and shortening above it, extend even farther into the foreland than the duplexes observed in the Edson example.

Two broad categories of layer-parallel shortening can be envisaged (Harrison and Bally, 1988). The first involves volume conservation (*e.g.*, small-scale folding, flow), the other entails volume loss (*e.g.*, horizontal compaction, stylolitization). The latter kind of layer-parallel shortening might be very difficult to establish in practice since the effects would be small and the required data (seismic velocities, well logs, cores, outcrops) are probably too imprecise or sparse to be useful. On the other hand, layer-parallel shortening with volume conservation would produce thickness changes that might be large enough to be observable by careful seismic isochron work. This work has been attempted here. Figures 4 and 5 illustrate the time structures and time thicknesses between seismic reflections on the section shown in Figure 3. Note that the seismic line extends some 15 km farther to the northeast than shown in Figure 3. The most spectacular thickness change is across the duplexes previously described. The younger intervals thicken gradually to the southwest, as predicted by models of foreland basin subsidence (Beaumont, 1981). The Wabamun (Devonian)-Cambrian interval thins to the west due to the influence of the West Alberta Ridge

(Oldale and Mundy, 1994). No anomalous thickening is observed above or to the northeast of the duplexes. Thus, the approximately 1.5 km of shortening observed in the duplexes in Figure 3 cannot be taken up in the section above the upper detachment along the length of the seismic section, as long as volume-conserving layer-parallel shortening is the mechanism, or unless this shortening is remarkably homogeneous and very widely distributed.

Horizontal compaction caused by tectonic stresses would result in increased seismic velocities. This raises the possibility that lateral changes in compaction could be observed by carefully examining interval velocities. This has not been attempted here for the following reasons. First, seismic interval velocities do not usually have sufficient resolution to detect small changes in porosity even in thick, homogeneous units. Second, any observed changes in interval velocity could not be interpreted in terms of porosity variations unless the primary lithologies of the units were tightly controlled. Third, any longer wavelength lateral porosity variations that could be isolated could not be directly attributed to horizontal compaction effects without first removing the effects of differential burial. Coincident section thickening and tectonic compaction could theoretically produce a situation in which the time interval across a thickened section remains constant due to the increased interval velocities resulting from the compaction; however, this is considered improbable.

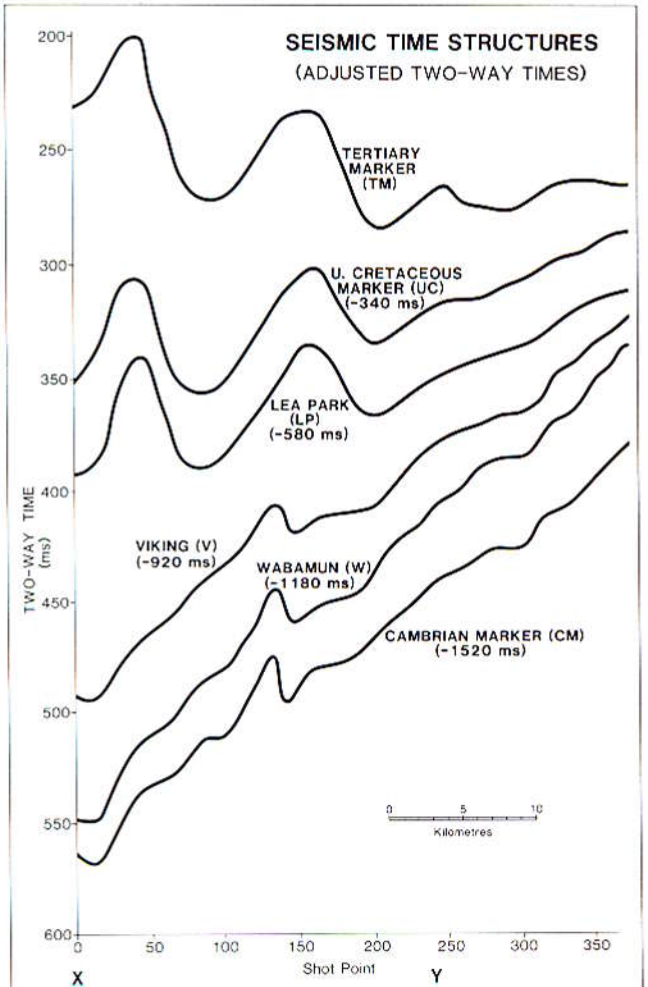

Fig. 4. Seismic time structures on selected horizons measured from the section shown in Figure 3. Points X and Y (see the base of the figure) correspond to the ends of the seismic section shown in Figure 3; the locations of these points are shown in Figure 2. Note that values on the right-hand side of the diagram are taken from a 12-km extension of the line to the northeast that is not shown in Figure 3. For illustrative purposes, the lines have been shifted together by the time offsets shown in parentheses under the horizon names. The actual two-way times on the original section can be recalculated by subtracting the time offset from the Y-axis value. Note the anticlinal structures on the three upper horizons that formed above the duplexes interpreted in Figure 3.

Robb Example

Figure 6 shows a line drawing of a seismic section that is located between the Edson example described above and the triangle zone at Robb (Fig. 2). The data quality of this line is not as good as in Figure 3. The interpretation shows a duplex in the same interval as in the Edson example. In this case, small-displacement backthrusts, dipping about 20° or less, are interpreted above the upper detachment. These features do not seem to reach the surface. To the northeast of the Pedley Fault, Lebel *et al.* (1996) have mapped an outcropping backthrust

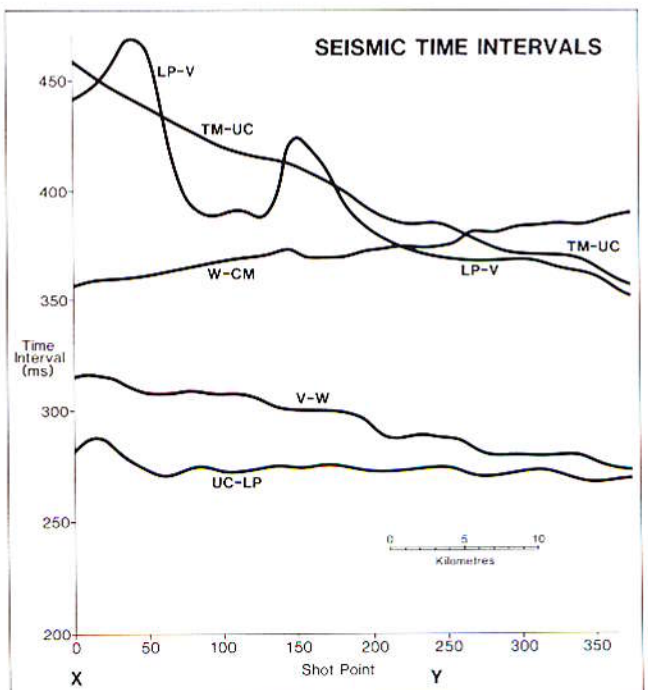

Fig. 5. Seismic time intervals between the seismic horizons in Figure 4. The intervals are labelled using abbreviations for the horizons shown in Figure 4. Note that all intervals except one thicken westward as a function of foreland basin subsidence (Beaumont, 1981). The Wabamun (Devonian)-Cambrian marker interval was deposited before the formation of the foreland basin and thins to the west. The only anomalous thickness changes occur over the interval that contains the duplexes shown in Figure 3.

(the Robb Fault) that, on seismic data, seems to be rooted in a detachment in the Brazeau Formation. It is possible that the backthrusts shown in Figure 6 flatten upwards into this detachment but the seismic data are not good enough to establish this definitively.

An interesting relationship seen in Figure 6 is the way in which the backthrusts merge with the upper detachment at the same points as the thrust ramps in the duplex. This suggests that there may be a genetic link between the faults above and below the detachment. If so, the features must have formed after the Edson duplexes (*i.e.*, out of sequence); otherwise the backthrusts would now be located about 1.5 km southwest of the underlying duplex due to this amount of contraction in the Edson structures (Fig. 10a). Alternatively, this observation could be interpreted to show that no displacement occurred on the upper detachment after the backthrusts formed (Fig. 7b). Thus, if the Edson duplexes formed after (*i.e.*, strictly in sequence with) the faults in the Robb example, the upper detachment of the Edson features could not be hinterland vergent, meaning that its displacement direction would not be consistent with the model proposed by Skuce *et al.* (1992).

Brazeau River Example

This seismic line (Fig. 8) is located about 70 km southeast of the previous two examples (Fig. 2). The triangle zone is not developed in its simplest form in this part of Alberta; the most easterly exposed major thrust fault is not a hinterland-vergent detachment but the Ancona Thrust, a foreland-directed thrust fault that outcrops about 15 km southwest of the Brazeau River example (Fig. 2). Locally (*e.g.*, at the Blackstone River; Douglas, 1958), this fault has NE-dipping Tertiary and Upper Cretaceous rocks in its footwall and can be interpreted as *post-dating* and cutting through the triangle zone monocline. A similar geometry has been interpreted by Liu *et al.* (1996) near Cache Creek, where out-of-sequence motion on the Muskeg Thrust has cut through the upper detachment and frontal monocline. Near Nordegg (Fig. 2), the Ancona Thrust is folded (Jones, 1971) and thus *predates* the blind frontal structure beneath it. The timing and nature of the frontal structures in this general area are evidently complex and are worthy of a detailed study in their own right.

Interpretation of the seismic data in Figure 8 shows two duplexes that are confined to approximately the same stratigraphic level as the Edson and Robb examples described above. Interesting features on this line are two emergent foreland-directed reverse faults that cut the upper detachment and the entire Upper Cretaceous and Tertiary section above it. The two faults appear to extend from ramps within the westerly duplex. Note that the displacement on the reverse faults is less than that on the two corresponding faults within the duplex. Gardiner *et al.* (1990) have reported very similar structures that are important in controlling the Peco oil field in the Belly River Formation, located a few km north of the seismic line shown in Figure 8.

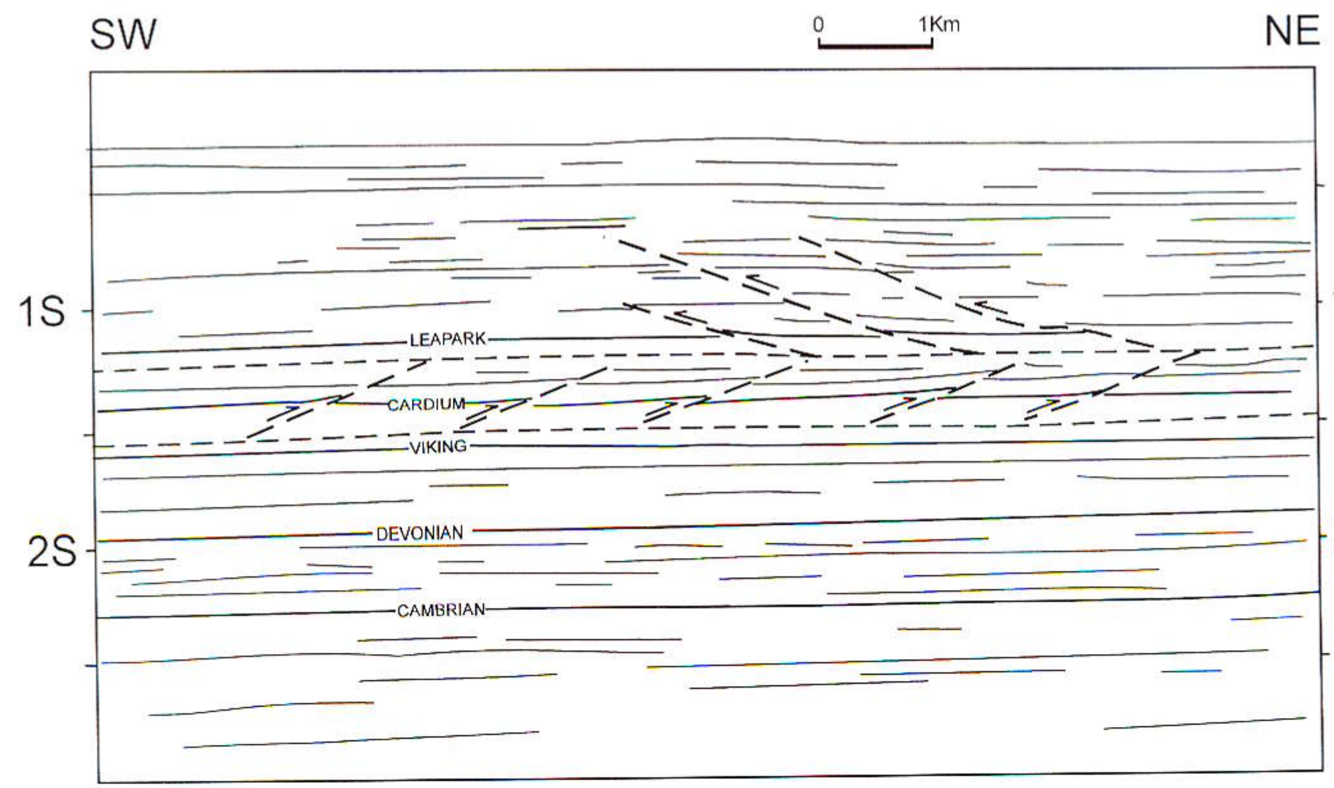

Fig. 6. Interpreted line drawing of a seismic line discussed as the "Robb example". Solid lines are coherent seismic reflections; dashed lines are interpreted thrust faults and detachment horizons. See Figure 2 for location. Little vertical exaggeration. TWT means two-way transit time.

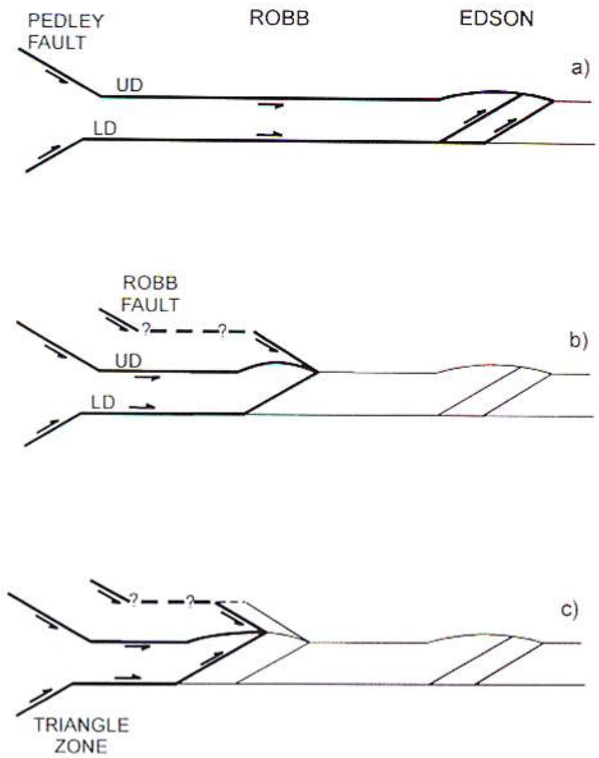

Fig. 7. Schematic cross-sections showing the sequence of deformation of the Robb and Edson examples. Bold lines show active faults, other lines represent inactive faults or detachment levels. UD – upper detachment; LD – lower detachment. The sequence **a)**, **b)** then **c)** is required to preserve the spatial relationship of the faults above and below the upper detachment at Robb. The linkage shown in **b)** and **c)** between the backthrusts above the UD and the outcropping Robb Fault (Lebel *et al.* 1996) is speculative.

As in the Robb example discussed previously, the geometric relationships in the Brazeau River example indicate an out-of-sequence order of formation to remain consistent with the intercutaneous wedge interpretation shown in Figure 8. The upper detachment is cut by the reverse faults, which means that there could have been no displacement along the upper detachment at this point after the reverse faults formed. One possible interpretation of this is that the more easterly of the two blind duplexes formed first, before the reverse faults and the western duplex (Fig. 9a-c). An alternative possibility (Fig. 9x-z) is that all the faults formed strictly in sequence. In this case, an interpretation involving no hinterland-vergent detachment is required.

The reverse faults are estimated to dip at about 40°, a value close to that measured in the field (Jones, 1971) on two similar features, the Brewster Creek and Sylvester Creek faults, which are located 30 km to the southeast of the Brazeau River example (Fig. 2). Such angles are unusually steep; typical footwall cutoff angles in the Alberta Foothills are less than 30° (Spratt and Lawton, 1996). Coulomb fracture criteria (Jaeger, 1969) predict that isotropic rocks having an internal coefficient of friction of 0.85 (Byerlee, 1978) should fracture along planes that make an angle of about 25° to the direction of the maximum compressional stress. From this it can be inferred that the maximum compressive principal stress was not horizontal at the time the reverse faults formed. A mechanism for distorting the stress field in a thrust sheet was proposed by Hafner (1961), using the model of a rectangular block in lithostatic equilibrium being pushed horizontally across a frictional surface. The model predicted that forethrusts should dip at angles less than

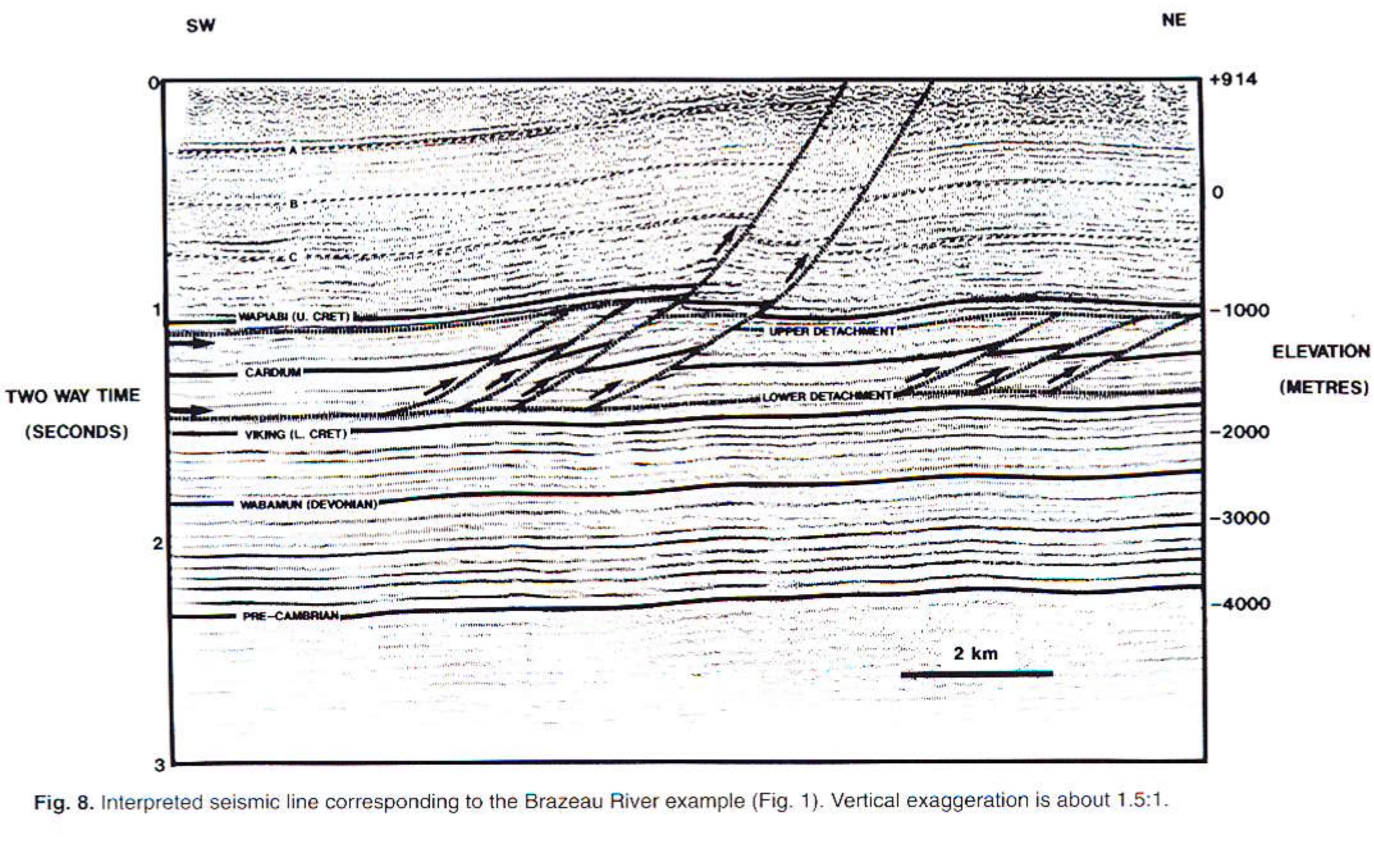

Fig. 8. Interpreted seismic line corresponding to the Brazeau River example (Fig. 1). Vertical exaggeration is about 1.5:1.

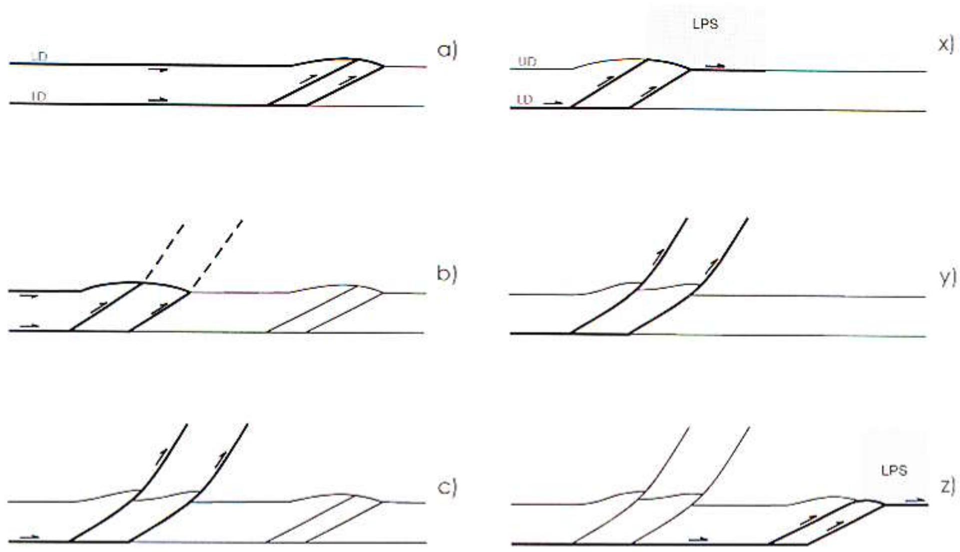

Fig. 9. Schematic cross-sections showing the sequence of deformation for the Brazeau River example. Bold lines show active faults, other lines represent inactive faults or detachment levels. UD – upper detachment; LD – lower detachment. Sections a)-c) show the sequence of deformation required for the intercutaneous wedge model. Sections x)-z) show a possible sequence of deformation involving layer-parallel shortening (LPS) above the UD.

the dips expected in the case where the maximum principal stress was simply horizontal; similarly, backthusts would have greater dips (Fig. 10a). At first sight, this seems contrary to the cases described here where the forethrusts (Brazeau River example) have a high dip and the backthrusts (Robb example) have a low dip. However, as shown by Jamison (1993), the section above the upper detachment can be modelled as thrust sheet with hinterland vergence. In this case, the nomenclature of backthusts and forethrusts can be reversed; the steep reverse fault on the Brazeau River example can be viewed as a backthrust in this context. Thus, the observed faulting in the para-autochthonous section described in this paper is compatible with Hafner's (1961) model, *providing that the upper detachment has a hinterland sense of displacement.*

However, the reality is not likely to be this simple. First, all the reservations about the simplifying assumptions in the wedge models discussed earlier will also apply here. Second, the reverse faults appear continuous across the upper detachment rather than rooted in it as the model in Figure 7a would require. It is possible that small displacements on the faults within the blind duplex resulted in bending of the para-autochthonous material above the upper detachment, locally modifying the stress field and causing the steep reverse faults to nucleate above the duplex (Fig. 10b). Subsequent movements could have resulted in the faults above and below the detachment running together and eventually cutting it. More modelling studies are required to further explore these points.

Discussion

The interpretations of the three examples above, examined in the light of more general observations of Foothills structures, allow a number of arguments to be made in the debate between the intercutaneous wedge model (ICW model) and the layer-parallel shortening model (LPS model). The conclusiveness of many of these arguments must be tempered by the fact that the area outboard of the triangle zone has not yet been extensively studied from a structural standpoint. First, six points that generally support the ICW model:

1. *Observational absence of layer-parallel shortening.* The Edson example shows that there is no seismically observable layer-parallel shortening above the upper detachment either directly above or immediately east of the blind duplexes below the detachment.

2. *Brittle failure.* The Robb and Brazeau River examples show that the section above the upper detachment has been subject to some compressive stress and that the response to this stress is by shear failure on seismically observable but minor thrust faults. This is the kind of brittle response to compression that might be expected in a section dominated by sandstones and siltstones.

3. *Location of shallow faulting.* All the faults so far reported above the upper detachment are found to the hinterland side of deeper blind duplexes. This observation is consistent with the ICW model, whereas the LPS model predicts contraction above the upper detachment outboard of the structures below the detachment.

4. *Relative shortening and applied stress.* The magnitude of the fault-related shortening within the intercutaneous wedge in the Robb and Edson examples is about 2 km, similar to the amount of shortening observed in nearby frontal structures that involve the Paleozoic (see fig. 4a in Lebel *et al.*, 1996). In contrast, fewer faults are observed above the upper detachment and these have smaller net displacements. Measurable shortening above the upper detachment is always less than the shortening observed in both incompetent and competent units

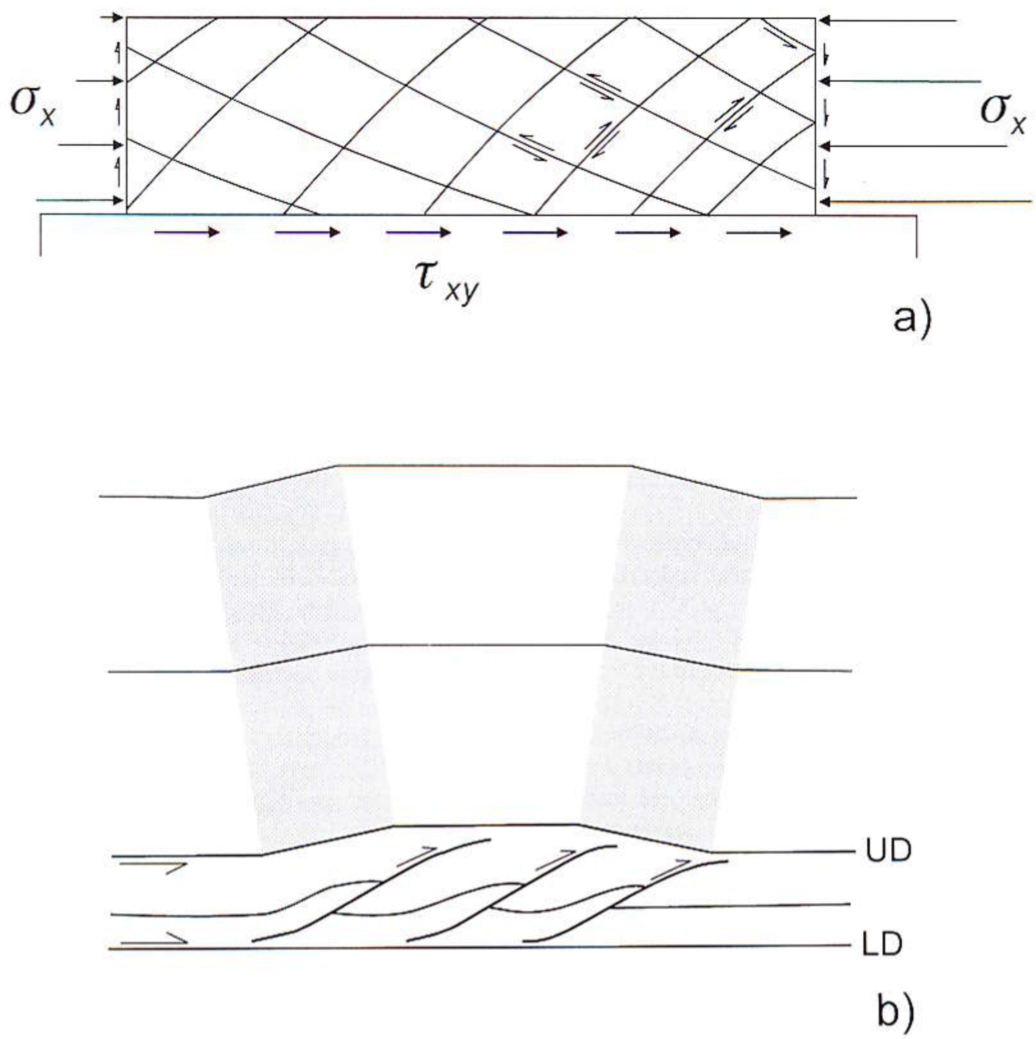

Fig. 10. a) Modified from Hafner (1961). Coulomb fracture trajectories produced in a rectangular block resting on a frictional surface and pushed from the right-hand side. The model predicts that faults that dip to the left will be steeper than the conjugate set dipping to the right. σ_x is the horizontal component of the boundary stress, the stresses on the left-hand side of the model are lithostatic, those on the right-hand side are lithostatic plus tectonic stresses; τ_{xy} represents the shear stress at the base of the model. Conventionally, this model is drawn the other way around with the tectonic push coming from the left. **b)** Sketch showing how the uplift effect of a deep duplex would cause stress modification in the shaded areas of the overlying section by bending it, perhaps providing preferential sites for subsequent faulting.

below the detachment. This is consistent with the ICW model which allows for the section above the upper detachment to be at least partly insulated from the compressive stresses associated with the mountain building.

5. *Relative competency.* The Mesozoic section below the upper detachment is dominated by marine shales, whereas the majority of the section above the detachment is composed mostly of nonmarine sandstones and siltstones. Because of this gross distribution of lithologies, the Mesozoic section below the upper detachment is generally considered to be less competent than the section above it (see, for example, fig. 3.13 in Wright *et al.*, 1994). Presumably, less competent units have a greater propensity to layer-parallel shortening, even assuming equal applied stresses above and below the upper detachment (but see point 4 above); one might therefore expect more layer-parallel shortening within the intercutaneous wedge than above it. This makes it difficult to balance the total shortening within the intercutaneous wedge (*i.e.*, faulting plus layer-parallel shortening) with layer-parallel shortening alone above the upper detachment. It would be most inconsistent to construct a regional cross-section that neglected layer-parallel shortening below the upper detachment while assuming that this mechanism is predominant above the detachment.

6. *Foothills kinematics.* Lebel *et al.* (1996) have made a detailed study of the surface and subsurface geology of the thrust belt immediately southwest of the Robb and Edson examples. Their model of the structural development of this part of the Foothills involves very acute intercutaneous wedges with relict hinterland-vergent upper detachments located in the upper part of the Wapiabi Formation. This kinematic interpretation therefore requires a structural geometry that is very similar to the ICW model throughout the development of the thrust belt.

The following three points reveal some weaknesses in the ICW model.

7. *Polarity of duplex ramps.* Given that the maximum compressive stress would have to be nearly parallel to the bedding to cause failure with opposite vergence on the lower and upper detachments east of the surface expression of the triangle zone, one might expect at least some of the blind duplexes to be dominated by ramps which are foreland dipping and hinterland vergent, whereas none of the observed examples have this geometry (Dechesne, 1994).

8. *Mechanics.* The ICW model requires a new dynamical model that has not yet been, and may never be, formulated. In contrast, the gross features of the LPS model are easier to reconcile with established models of thrust belt formation.

9. *Sequence of deformation.* Geometric factors require that the structures in the ICW model could not have formed in the normally assumed hinterland-to-foreland sequence. This is a complication that requires further independent supporting evidence before it could be fully accepted. The LPS model is consistent with the conventional sequence of deformation.

As discussed above, the simplifying assumptions made in currently accepted mechanical models of thrust belt formation severely limit their usefulness in constraining structural interpretations within triangle zones. Thus, although points 7 and 8 are serious outstanding problems, I do not believe they are fatal to the ICW model. I think it preferable to admit to unresolved problems in the area of structural dynamics than to make an *ad hoc* hypothesis of layer-parallel shortening above the upper detachment just to preserve a theoretical model that is, in any case, demonstrably oversimplified and is unable to account for even well-accepted triangle zone geometries.

The question of the sequence of deformation within the intercutaneous wedge mentioned in point 9 above may have some additional consequences for interpretations of low-taper wedge geometries. Harrison and Bally (1988) questioned the validity of their own low-taper triangle zone interpretation of the Parry Islands fold belt because of the coincidence of structures above their upper detachment with structural culminations in the intercutaneous wedge below. If their features developed in sequence then the structures above the detachment, on all but the most frontal structure, should not coincide exactly with the structures below but should have been displaced as the intercutaneous wedge continued to deform and advance. However, if their structures formed synchronously or in reverse sequence, this objection to their model disappears.

Jamison (1993) observed that McMechan's (1985) low-taper triangle zone model in northeast British Columbia involves an active undulating upper detachment, effectively requiring, in places, negative intercutaneous wedge angles that would be particularly difficult to justify mechanically. However, if the structures within the intercutaneous wedge in McMechan's model developed out of sequence then these mechanical objections to her interpretation are considerably eased. Similarly, the interpretation of Skuce *et al.* (1992) is likely to be mechanically more acceptable if the structures did not form from west to east. There is considerable field evidence for widespread out-of-sequence and synchronous thrusting in Foothills belts (Boyer, 1992); analogue modelling experiments (Dixon and Liu, 1992) also demonstrate this phenomenon.

The reverse faults with the relatively steep dips, described previously in the Brazeau River example, may have analogues within the inner Foothills Belt. An example is the steeply dipping thrust fault that cuts the west flank of the Williams Creek Syncline, shown on cross-section G-H on the Calgary 1:250 000 scale geological map sheet (Ollerenshaw, 1978). The footwall cutoff geometry of this fault is well imaged on proprietary seismic data and is similar to that observed in the Brazeau River example. Dechesne and Muraro (1996) interpreted the Williams Creek Syncline as forming part of a relict triangle zone. As discussed previously, steeply dipping foreland-directed reverse faults can be explained, in principle, by viewing them as backthrusts within the section above a hinterland-vergent detachment. Therefore, accepting this model, foreland-vergent reverse faults with unusually high footwall cutoff angles might be diagnostic of relict triangle zones within the Foothills.

Consequences of the Debate

Understanding the nature of frontal structures has economic significance since these structures are critical in the formation of oil fields in the Belly River Formation (Gardiner *et al.*, 1990) and the Second White Speckled Shale Formation (Podruski *et al.*, 1988). Recently, commercial gas fields have been discovered in thrusted structures in the Cardium Formation within the intercutaneous wedge in the area immediately southwest of the Robb and Brazeau River examples (Fig. 2).

Although this paper contains arguments for local synchronous and out-of-sequence movements on faults, the general regional development of the thrust belt is still assumed to be from hinterland to foreland. Consequently, all parts of the thrust belt were, at one stage, frontal. Studying the present leading edge of the fold belt, where structures are relatively simple, should therefore help in interpreting the geometry, kinematics and mechanics of the more complicated structures in internal parts of the thrust belt.

An important general issue arising from this debate is the manner in which layer-parallel shortening is handled in the construction of balanced cross-sections in the Foothills. Layer-parallel shortening is difficult to quantify; usually some kind of expedient assumption is required. Most recent balanced cross-sections through the Foothills assume conservation of line length and area; in other words, any kind of layer-parallel shortening is neglected. If future research on frontal structures favours the LPS model, then hard-to-observe layer-parallel shortening would have been shown to be capable of absorbing displacements of the order of kilometres. This shortening would have occurred in a relatively competent part of the section that is apparently otherwise little contracted. Clearly, neglecting layer-parallel shortening in the more intensely deformed, less competent sections within the main Foothills Belt would then no longer be a viable option.

Conclusions

Mechanical models of intercutaneous wedges in frontal Foothills structures do not, so far, adequately consider the strongly layered mechanical stratigraphy that is an essential ingredient in the initiation and development of triangle zones in particular and blind duplexes in general. Therefore, no *a priori* constraints on acceptable intercutaneous wedge geometries should currently be assumed. The upper and lower detachment surfaces are probably both associated with overpressured shales. The para-autochthonous section above the upper detachment is observed to have been deformed by relatively low-angle (*ca.* 20°) backthrusts and steep (*ca.* 40°) foreland-directed reverse faults. These fault geometries are consistent with a model that requires the upper detachment to be hinterland vergent. Hard evidence for layer-parallel shortening above the upper detachment is currently lacking. Geometric relations suggest that some features formed out of sequence. This sequence of deformation also has an advantage from a mechanical standpoint since negative wedge angles, which would be particularly difficult to justify, are no longer required.

Based on current evidence, I maintain that the upper detachment of the blind duplexes described in this article has a backthrust sense of vergence. The question posed in the title of this paper can thus be answered: yes, frontal structures in west-central Alberta can be viewed as representing the thin end of the intercutaneous wedge. Nevertheless, two caveats are required. First, the lack of evidence of significant layer-parallel shortening above the upper detachment does not constitute proof that such effects have not occurred; distributed strain is expected to be difficult to observe. Second, the current inadequacy of mechanical constraints on acceptable intercutaneous wedge geometries does not imply that "anything goes"; there must be limits and it is acknowledged that the model preferred here, while geometrically plausible, remains something of a mechanical puzzle.

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