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Suprastructure/infrastructure transition, east-central Cariboo Mountains, British Columbia: geometry, kinematics and tectonic implications

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Abstract—In the Cariboo Mountains of east-central British Columbia, syn-metamorphic D_2 folds of the Proterozoic Kaza Group change from upright, open and symmetrical to gently inclined, tight (locally isoclinal) and SW-verging. Plunge measurements and change in metamorphic grade show that the transition occurs with paleo-depth in the orogenic belt. This 'suprastructure–infrastructure' transition is attributed to a downwardly increasing component of non-coaxial strain accumulated during northeastward underthrusting of allochthonous marginal basin sediments and autochthonous distal North American continental margin sediments beneath more proximal sediments of the continental margin.

A comparison of the structural thickness of the Kaza Group with its original thickness reveals that a greater amount of structural thickening is represented at deeper structural levels than at immediately overlying shallower levels. This geometry requires that shallow structural levels to the west be separated from deeper structural levels by thrust faults. A belt of pre-metamorphic-peak, SW-verging thrust faults in the Quesnel Highlands may be the shallow level manifestation of infrastructure thickening. Approximately 60 km of NE displacement of rocks at deep structural levels with respect to a fixed suprastructure can account for both the infrastructural thickening in the east-central Cariboo Mountains and the geometry of thrust faults in the Quesnel Highlands.

INTRODUCTION

THE CENTRAL and southern Cariboo Mountains (Figs. 1–3) are underlain by a NW-plunging structural sequence of meta-sediments of the Proterozoic Kaza Group (Campbell 1968, Campbell *et al.* 1973). Originally deposited near the base of the continental margin sequence fringing western North America (Gabrielse 1972, Wheeler *et al.* 1972, Campbell *et al.* 1973, Young *et al.* 1973), Kaza Group rocks were deformed and metamorphosed during early mid-Jurassic to Tertiary interaction of western North America with oceanic 'suspect' terranes (Davis *et al.* 1978, Coney *et al.* 1980, Monger *et al.* 1982). Penetrative deformation of these rocks occurred relatively early in this interval with strains recorded before, during, and after the mid-Jurassic peak of regional metamorphism (Nguyen *et al.* 1968, Campbell 1968, Campbell *et al.* 1973, Pigage 1977, 1978, Fletcher & Greenwood 1979, Pell & Simony 1981, Murphy & Journeay 1982, Parrish & Wheeler 1983, Archibald *et al.* 1983, Murphy & Rees 1983). Post-metamorphic uplift of the Premier Range in the south-eastern Cariboo Mountains tilted the sequence into its present NW-plunging attitude.

In mapping the Canoe River and McBride map-areas (Fig. 1), Campbell (1968) and Campbell *et al.* (1973) noted a distinct contrast in structural style between lower grade and higher grade rocks of the Kaza Group. Where the Group contains synkinematic muscovite and chlorite, it is weakly deformed, and is openly to tightly folded around steep axial surfaces. In contrast, rocks that were deformed at conditions supporting the growth of garnet and higher grade minerals are tightly to isoclinally folded around gently dipping axial surfaces. Following De Sitter & Zwart (1960), Campbell (1970, 1973) and Wheeler *et al.* (1972) referred to the change in structural style from shallow to deep levels as a 'suprastructure–infrastructure' transition and considered the difference in structural style to be a consequence of the difference in rheology between rocks deformed under different metamorphic conditions. They thought that the total amount of shortening at all structural levels was the same and limited by the 10–20% shortening apparent in the folds of the suprastructure.

As initially outlined, this change in structural style between suprastructure and infrastructure is a tantalizing feature, potentially implying much about the kinematic evolution of this part of the Canadian Cordillera but lacking the necessary detailed documentation for a full evaluation. Of critical importance are data on: (1) the geometry, kinematics, and relative timing of strains at different structural levels, (2) the nature of the coupling between structural levels and (3) the nature of the Kaza multilayer.

In consideration of these questions, three 2-month field seasons between 1981 and 1983 were spent mapping between Castle Creek and Tête Creek north of the Premier Range (Fig. 2). A series of maps and cross-sections across the structural transition was prepared at 1:20,000 scale. This scale permitted the illustration of structural geometries but precluded complete mapping of the 1400 km^2 region. However, the spectacular topographic relief (up to 2.5 km) of the Cariboo Mountains and lateral continuity of some strata within the Kaza Group made it possible to link maps and, especially, cross-sections with a reasonable degree of certainty.

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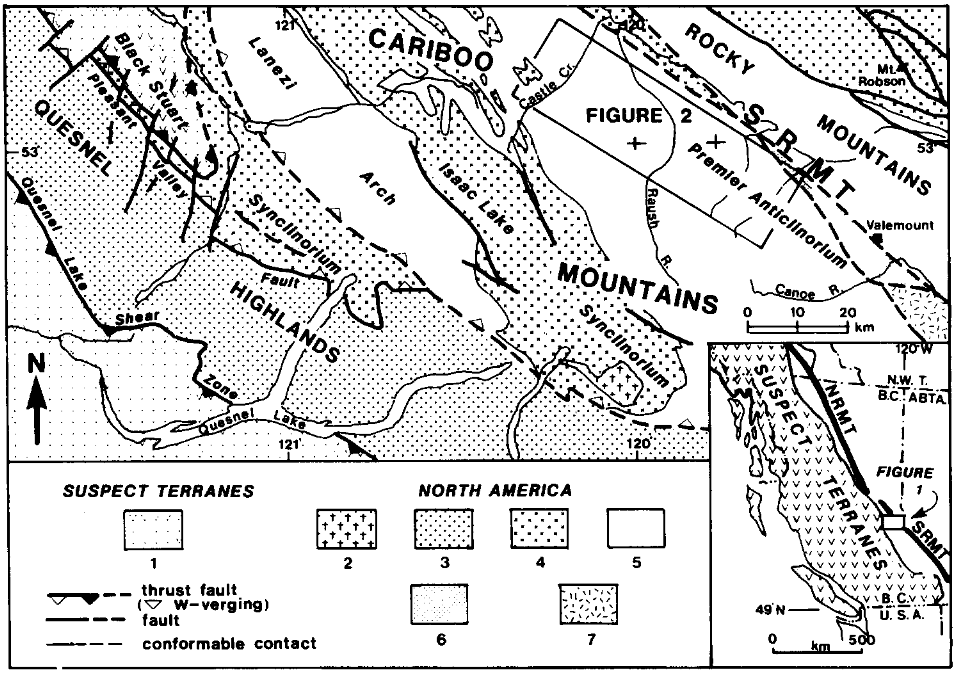

Fig. 1. Geologic map of the Cariboo Mountains and the adjoining Quesnel Highlands and Rocky Mountains. The location of the study area (Fig. 2) is shown with respect to regional and local geological features discussed in the text. 1, Terrane I, Paleozoic to Middle Jurassic volcanic and sedimentary rocks of volcanic arc and ophiolite affinities; 2, Middle Jurassic granodiorite; 3, Proterozoic to upper Paleozoic sediments and meta-sediments of the Snowshoe Group inferred to have been deposited at the distal part of the North American continental margin sequence; 4, Isaac Formation (upper Proterozoic) and younger sediments and meta-sediments of the North American continental margin sequence (Cariboo Group in the Cariboo Mountains; upper Miette Group in the Rocky Mountains); 5, upper Proterozoic Kaza Group (Cariboo Mountains), middle Miette Group (Rocky Mountains); 6, upper Proterozoic Horsethief Creek Group (Cariboo Mountains), lower Miette Group (Rocky Mountains); 7, Malton gneiss, inferred to be basement to the North American continental margin sequence; NRMT, Northern Rocky Mountain Trench; SRMT, Southern Rocky Mountain Trench.

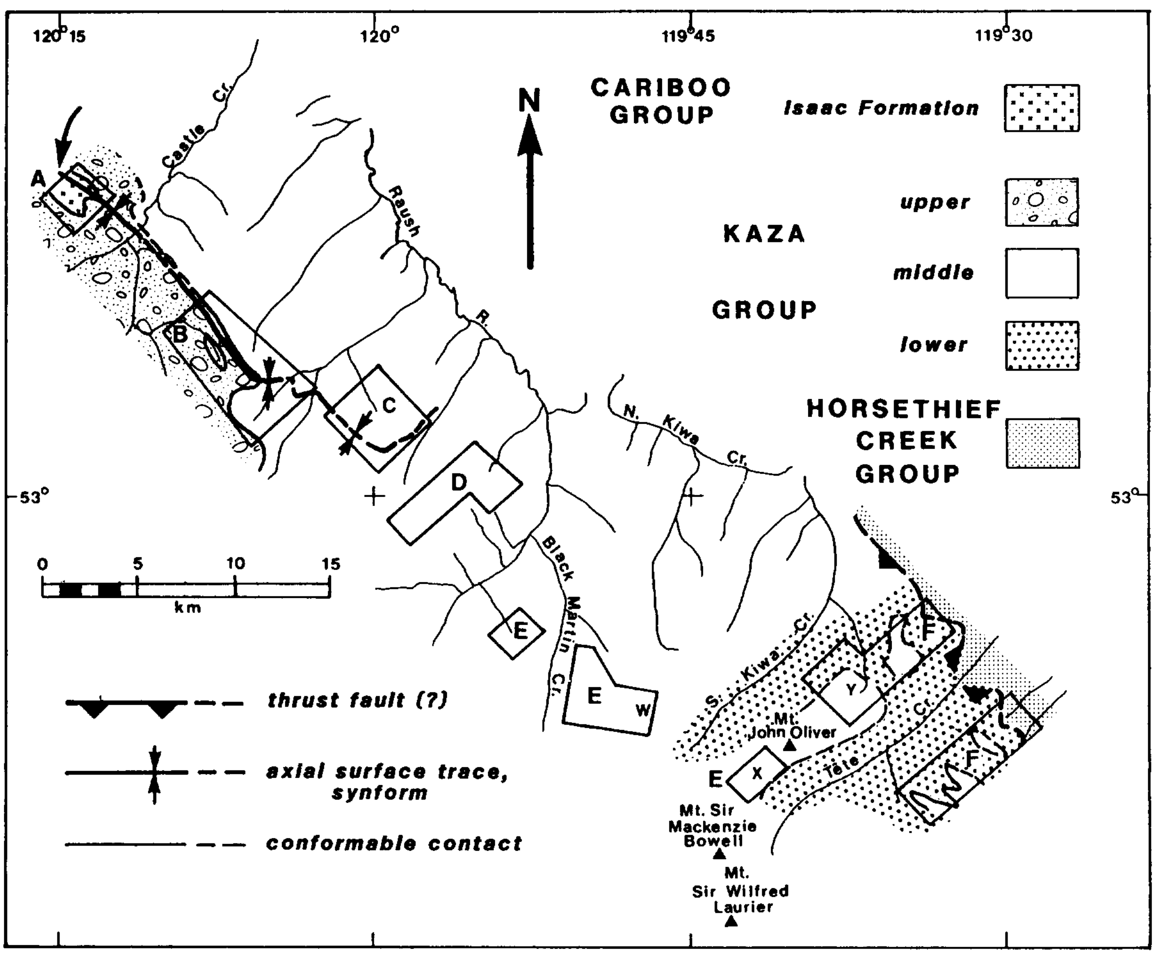

Fig. 2. Distribution of rock units in the east-central Cariboo Mountains and locations of areas (A–F) mapped in detail.

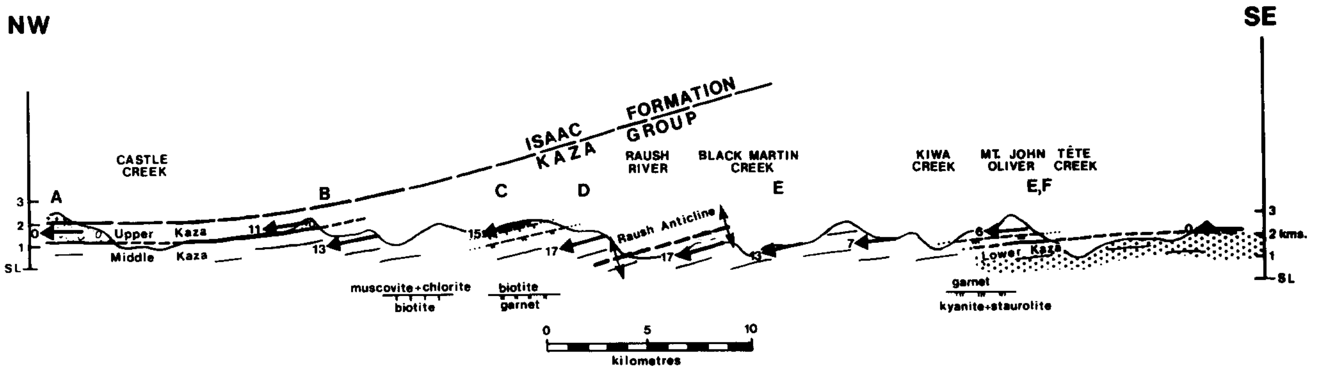

Fig. 3. Longitudinal (NW–SE) cross section, from ridge north of Castle Creek to ridge south of Tête Creek, illustrating how variation in structural plunge exposes deeper structural and stratigraphic levels in the southeastern Cariboo Mountains. A–F are areas mapped in detail shown in Fig. 2. The line of section passes through areas A, B and D; data from areas C, E and F are projected.

The purpose of this paper is to document, in detail, the nature of the structural transition from suprastructure to infrastructure and to discuss its tectonic implications. A significant finding of this study is that the downward transition from upright to gently inclined structures involves folds with SW vergence. The structural geometry at even deeper structural levels is also dominated by SW-verging regional-scale folds. Structures which verge away from the craton have long been recognized in the Canadian Cordillera and various models have been proposed to reconcile these structures with the more prominent E- to NE-verging structural features such as the Rocky Mountain fold and thrust belt. These models are discussed in light of the data presented in this study.

THE STRUCTURAL SEQUENCE OF THE KAZA GROUP

In the east-central Cariboo Mountains, the consistent NW plunge of structures allows a cross-section to deep structural levels to be constructed from the surface geology (Figs. 3–5). This cross-section illustrates the structural geometry of a 10 km thick sequence of Kaza Group rocks ranging in metamorphic grade from muscovite/chlorite zone to kyanite/staurolite zone.

The Kaza Group of the east-central Cariboo Mountains has been divided into lower, middle and upper units (Fig. 6) following the informal stratigraphy of Pell & Simony (1984). The lower Kaza Group consists of calc-schist (meta-marl or calcareous shale), orange, buff or grey marble and meta-sandstone. The middle Kaza Group consists of equal amounts of amalgamated meta-sandstone alternating with phyllite in units 5–20 m thick, and is capped by a distinctive persistent 200 m thick unit comprising green laminated meta-siltstone at the base, finely interlaminated grey and black phyllite, and black graphitic phyllite and marble at the top. The upper Kaza Group consists of metamorphosed sub-feldspathic sandstone, pebbly sandstone, minor graphitic and non-graphitic phyllite (meta-shale), and rare thin marble beds (Murphy & Rees 1983). The top of the Kaza Group is the stratigraphic contact with the overlying Isaac Formation, which is exposed on the north side of Castle Creek (Fig. 2). The base of the sequence is an inferred

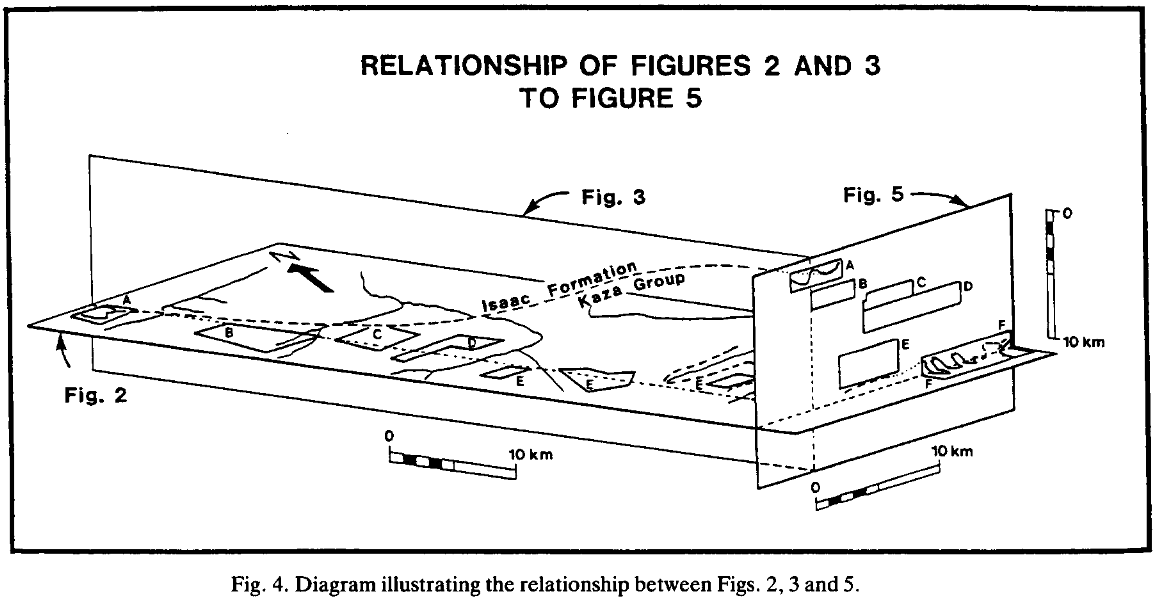

Fig. 4. Diagram illustrating the relationship between Figs. 2, 3 and 5.

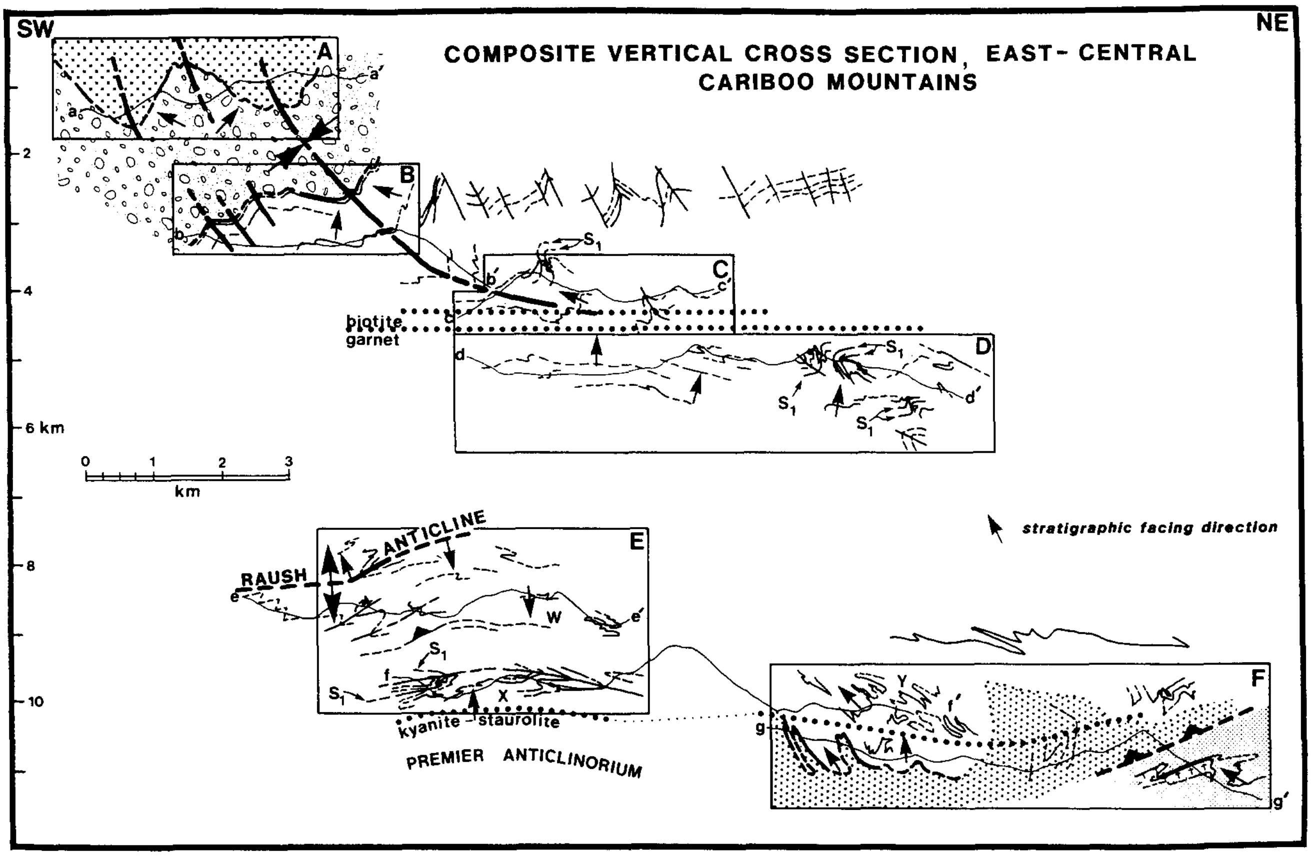

Fig. 5. Composite vertical cross-section compiled from 30 serial cross-sections (a–a′ to g–g′ are representative topographic profiles) of areas A–F. Relative positions of cross-sections are determined from lineation data (Figs. 3 and 8). Form lines are bedding traces except where indicated. Form lines outside of boxes are drawn from photomosaics of cliff faces. Ornamentation and symbols are as in Fig. 2.

fault separating right way up rocks of the lower Kaza Group from an underlying sequence of inverted rocks which are lithologically similar to the semi-pelite and amphibolite (SPA) and middle marble units of the Horsethief Creek Group (Murphy 1984). The inferred fault is located in the western slopes of the Rocky Mountain Trench west of Valemount (Fig. 2).

At all structural levels, rocks of the Kaza Group exhibit macro- and micro-structural evidence of polyphase deformation. Four sets of structures have been identified on the basis of structural superposition and relationship to the mid-Jurassic episode of metamorphic mineral growth (Fig. 7). Pre- to early-metamorphic D_1 structures comprise tight to isoclinal NW-trending folds mappable at 1:15,000 and larger scales, and an axial planar foliation (S_1) that is readily apparent at shallow structural levels but is almost completely transposed at deeper structural levels (Figs. 5 and 8). Where preserved at deeper structural levels, S_1 is defined by muscovite, chlorite, and possibly biotite, implying that D_1 was not associated with high-grade regional metamorphism. The amount of structural thickening due to D_1 deformation is not known, but the small size of the structures and lack of significant metamorphism suggest that D_1 structural thickening was minor.

Syn-metamorphic D_2 structures are the dominant structures at all levels, mappable at 1:50,000 and smaller scales. D_2 deformation is responsible for the main bulk of the structural thickness across which prograde mineral isograds are printed. Randomly oriented minerals overgrow D_2 structures, implying that metamorphism outlasted D_2 deformation.

Post-metamorphic D_3 structures are found primarily at the bottom of the sequence of Kaza Group rocks, increasing in importance near the southern Rocky Mountain Trench. Post-metamorphic D_4 structures are NE-trending, oblique to the NW trend of all earlier structures. D_4 structures appear to be responsible for the larger-scale variations in the plunge of D_2 hinge lines and lineations (Fig. 3).

For descriptive purposes, the structural sequence of the Kaza Group has been subdivided into 'Suprastructure', 'Infrastructure', and 'Transition Zone'.

Suprastructure

The upper Kaza Group and part of the middle Kaza Group are represented in the suprastructure. The metamorphic grade of the suprastructure increases down-section and to the southeast from muscovite–chlorite zone to biotite zone. The structural geometry (Figs. 5 and 8) is dominated by regional-scale, open to close, NW-plunging, SW-verging folds (Murphy & Rees 1983). At the level of the contact between the Kaza Group and the Isaac Formation, the axial surfaces of these folds dip moderately to steeply NE. Above this level, the axial surfaces are more nearly upright and folds more open (Campbell *et al.* 1973). Steep to over-

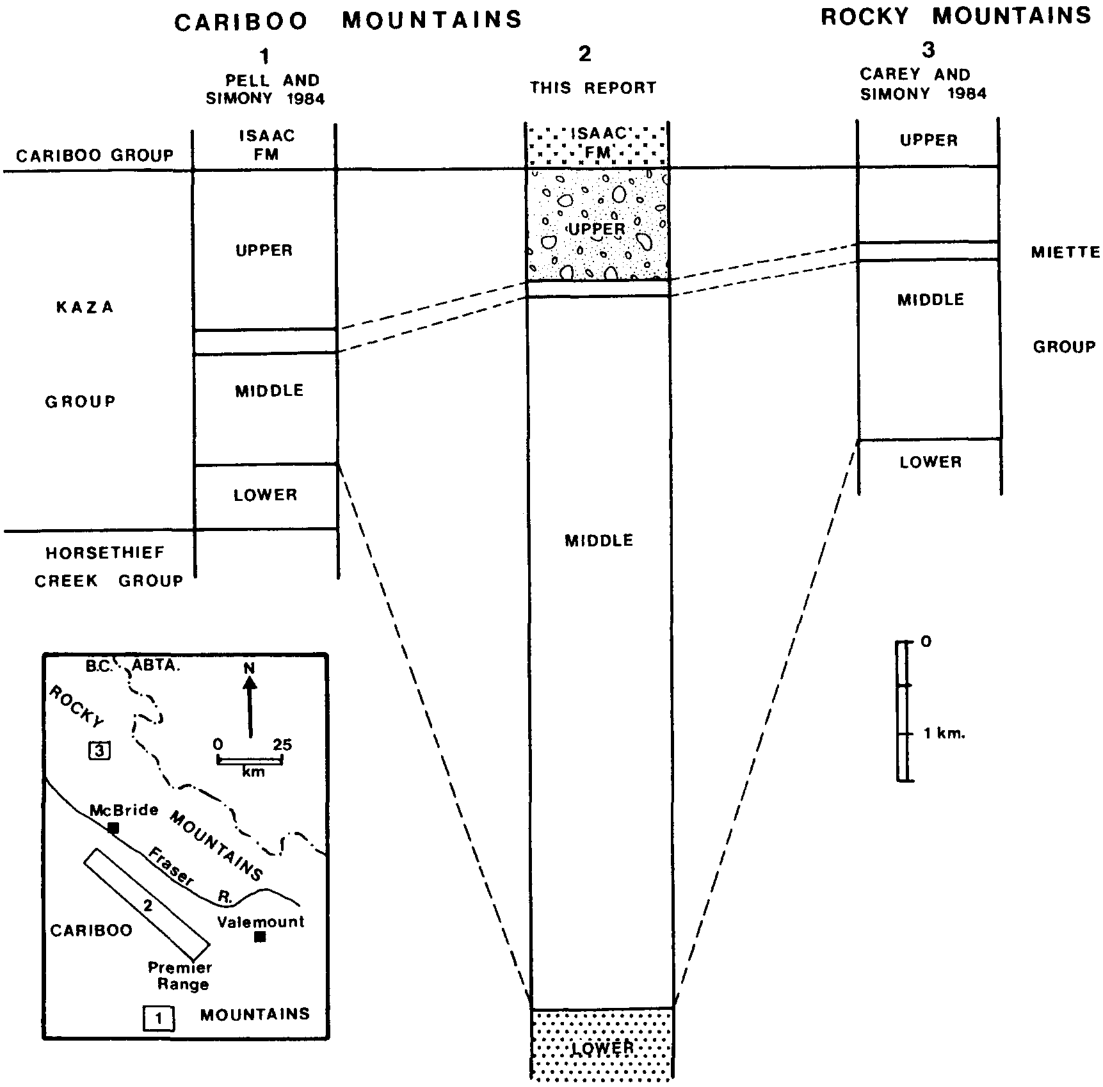

Fig. 6. Stratigraphy and regional correlation of the Kaza Group.

turned bedding is common on short limbs of second-order folds, but the suprastructure stratigraphy is dominantly right way up.

The SW-verging folds have a spaced axial-planar foliation, S_2 (Figs. 8 and 9a & b). S_2 fans in hinge zones and refracts at meta-sandstone–meta-shale contacts. Locally, where the angle of refraction is large and S_2 in meta-shale layers is subparallel to bedding, S_2 is folded by crenulations possibly related to further tightening of

SEQUENCE OF DEFORMATION AND METAMORPHISM			
youngest	D_4	post- metamorphic	NE-trending upright crenulations and regional-scale warps
↑	D_3	late- to post-metamorphic	NW-trending, NE-verging crenulations and folds; important near faulted base of structural sequence, but dying upward; includes Premier Anticlinorium
	METAMORPHIC PEAK		
TIME	D_2	syn- metamorphic	NW-trending, SW-verging folds; dominant at all structural levels; associated with significant structural thickening
oldest	D_1	pre- to early-metamorphic	Small-scale, NW-trending, NE verging folds; not associated with significant structural thickening, age uncertain but may be early Jurassic, associated with the obduction of Terrane I

Fig. 7. Sequence of deformation and metamorphism in the east-central Cariboo Mountains. 'Time' is relative to peak of regional metamorphism as recorded in individual exposures.

the fold (cf. Trayner & Cooper 1984). Graded bedding is clearly outlined by the smoothly continuous sigmoidal trace of S_2 as it passes from the coarser to the finer fraction. In phyllite, S_2 is a crenulation cleavage, produced during the folding of an earlier chlorite/muscovite foliation, S_1 (Figs. 9a & b). The planar fabric itself is defined by concentrations of insoluble residue and micas on the short limbs of asymmetric crenulations (Fig. 9a). Mica-rich and quartz-rich segregations commonly define a compositional layering at thin-section scale. In meta-sandstones and pebbly sandstones, S_2 is defined by the dimensional preferred orientation of flattened and extended quartz, and to a lesser extent, feldspar clasts. S_2 intersects bedding (S_0) to produce a lineation (L_2) which is parallel to the hinges of D_2 folds (F_2) and L_{2e}, the finite extension direction of elongate clastic grains (Fig. 9c).

S_1 is macroscopically folded by D_2 folds and appears to be related to small-scale, tight to isoclinal, NW-plunging, NE-verging folds (D_1) whose axial surfaces are clearly folded by D_2 folds (Fig. 5). On right way up D_2 limbs, where S_2 is less pervasive, S_1 is NE-verging (Fig. 9d).

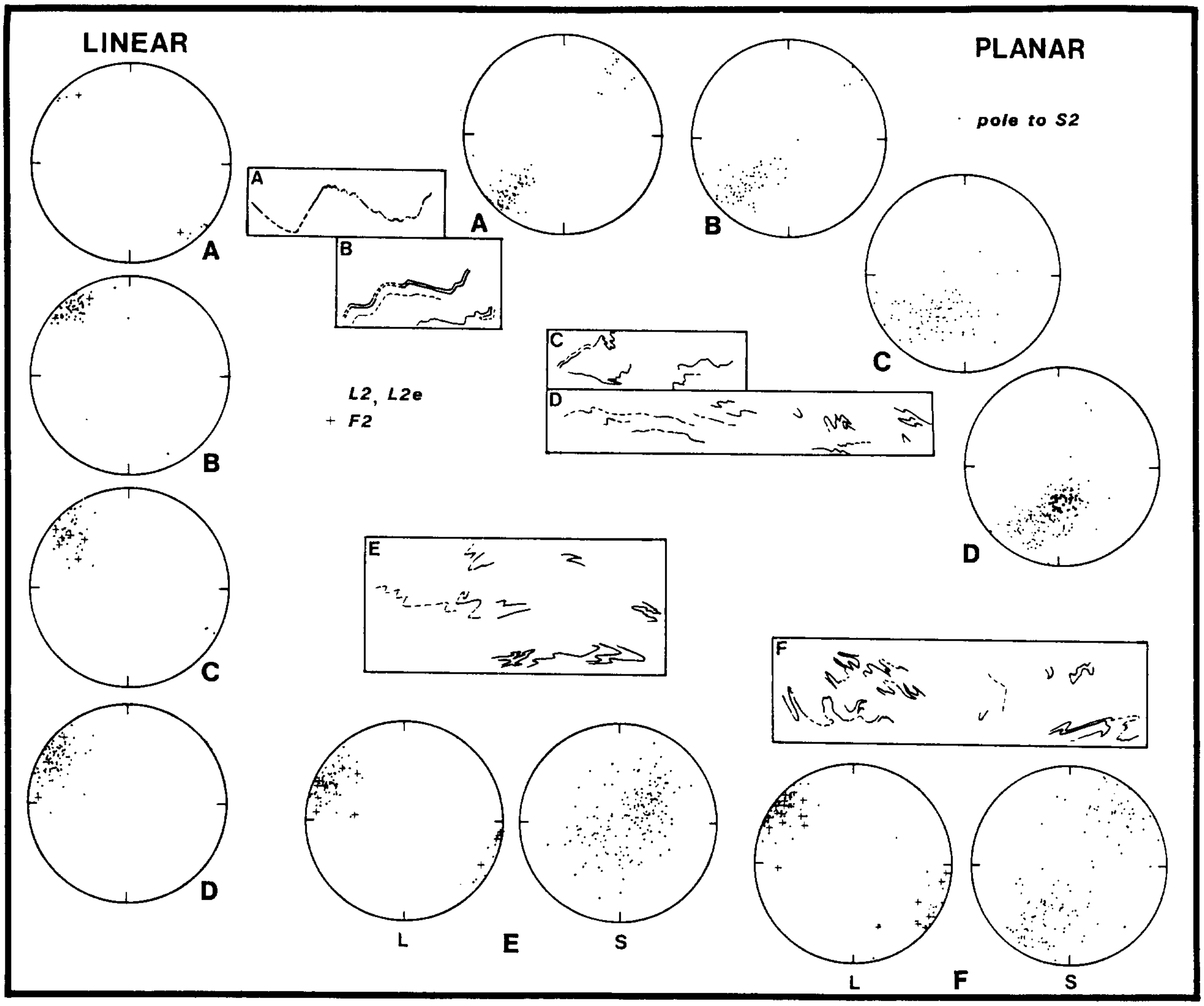

Fig. 8. Equal-area stereoplots of D_2 planar (S_2) and linear (L_2, L_{2e} and F_2) fabric elements.

In the low-grade suprastructure, S_2 is overgrown by prograde porphyroblasts of biotite, muscovite and chlorite (Fig. 9b). Thus, D_1 and D_2 folds formed prior to the peak of regional metamorphism. Post-D_2 porphyroblasts and all earlier structures are deformed by a NE-trending set of macroscopic crenulations with steep axial surfaces (D_4). These transverse structures may be responsible for the large-scale variations in the magnitude of plunge of D_2 and earlier structures (Figs. 3 and 8), although this relationship remains to be proven (Murphy & Rees 1983).

Infrastructure

Rocks of the infrastructure belong to the middle and lower Kaza Group of Pell & Simony (1984). The metamorphic grade of the infrastructure increases down-section and to the southeast from garnet zone to kyanite–staurolite zone (Campbell 1968). The structural geometry is best described with reference to the Raush Anticline, a NW-trending, gently inclined, regional-scale, SW-verging anticline (Fig. 5). The Raush Anticline and associated parasitic folds deform bedding, earlier foliation (S_1), and a set of smaller-scale NE-verging folds (D_1). They can therefore be treated, locally at least, as D_2 structures. The Raush Anticline has been identified by the inversion of stratigraphy accompanied by the reversal of vergence of second-order, though still regional-scale, folds. The long upper limb is characterized by tight, SW-verging, large-scale folds of a dominantly right way up stratigraphic sequence. The short lower limb is characterized by tight to isoclinal, NE-verging folds of a dominantly inverted sequence. The hinge zone, characterized by symmetrical folds with steep enveloping surfaces, has been mapped along a small portion of the ridge between Black Martin Creek and Raush River and has been observed, but not mapped, on the north side of the Raush River Valley (Fig. 2).

Overturned middle Kaza Group stratigraphy of the lower limb of the Raush Anticline has been traced to the ridge north of South Kiwa Creek (Fig. 2; Fig. 5, location W). On the ridge south of South Kiwa Creek between Mt. Sir John Oliver and Mt. MacKenzie Bowell (Fig. 2; Fig. 5, location X), predominantly right way up middle and lower Kaza Group crop out. The contact between the middle and lower Kaza Group can be traced east through prominent SW-verging isoclinal folds into a zone of symmetrical isoclinal folds with steep enveloping

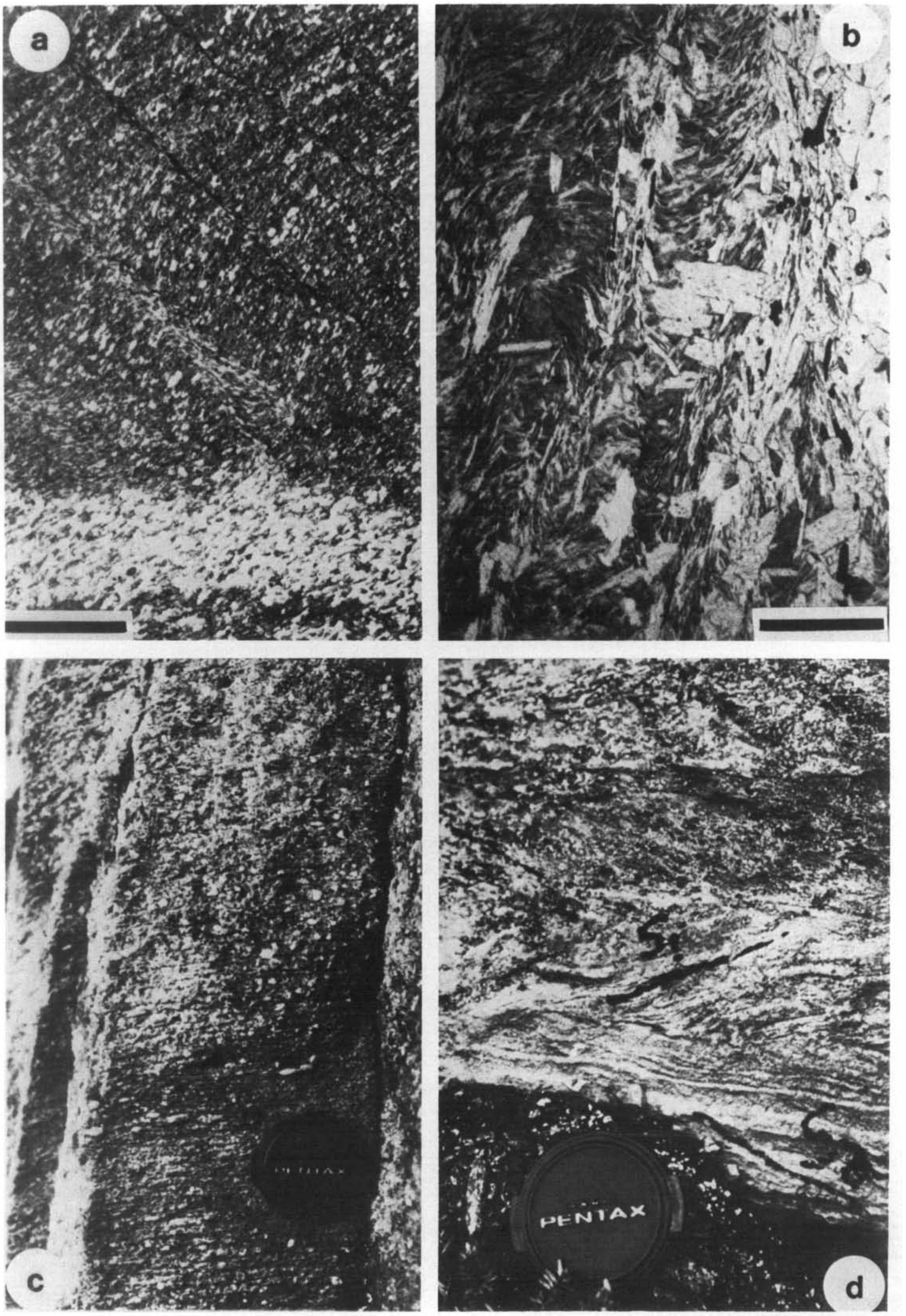

Fig. 9. (a) Spaced D_2 cleavage (S_2) produced by the crenulation of chlorite–muscovite foliation (S_2). Organic-rich seams reflect dissolution of quartz on short limbs of asymmetrical crenulations. Scale bar is 0.5 mm. (b) Randomly oriented muscovite overgrowing D_2 crenulations. Scale bar is 0.5 mm. (c) View of top of bedding surface of pebbly sandstone. Intersection of S_2 with bedding produces linear fabric L_2. The maximum elongation direction of quartz and feldspar clasts (L_{2e}) is parallel to L_2. (d) View (looking down-plunge, to the northwest) of mesoscopic planar fabrics. Bedding (right way up) is defined by darker phyllite and lighter sandstone beds. NE-verging S_1 and SW-verging S_2 are indicated.

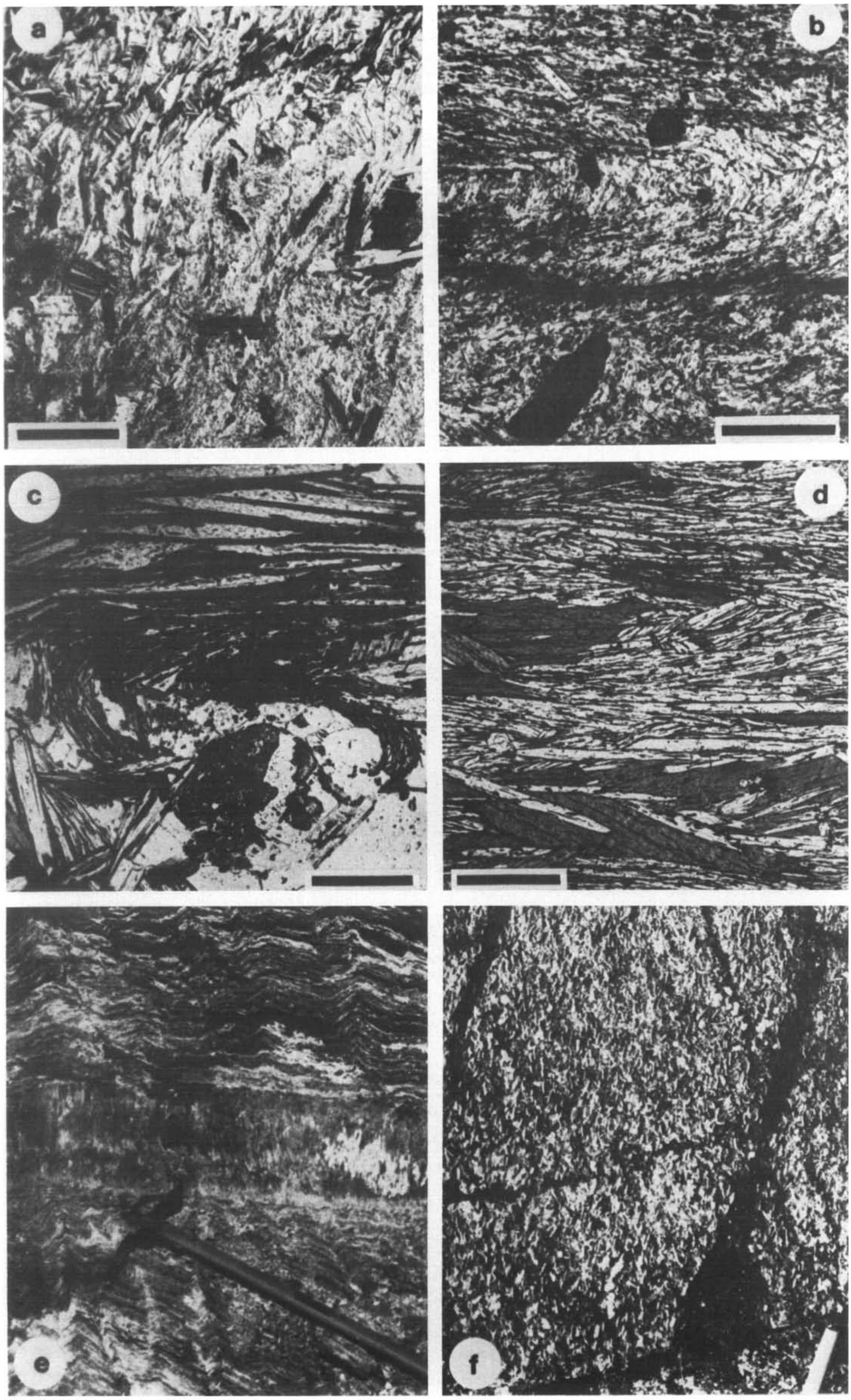

Fig. 10. (a) Syn-metamorphic (but pre-metamorphic-peak) D_2 crenulations at intermediate structural levels. Scale bar is 0.5 mm. See text for discussion. (b) Close-up of crenulations in Fig. 10(a). Note the randomly oriented muscovite laths in both hinge and limb zones of crenulations. Scale bar is 0.15 mm. See text for discussion. (c) Rare relict D_2 crenulations, from intermediate to deep levels, defined by incompletely recovered polygonal arcs of muscovite, chlorite, and graphite (?). The prominent foliation in the rock (S_2), defined by parallel laths of coarse-grained muscovite and biotite, is parallel to axial surfaces of crenulations and major SW-verging folds in the region. Scale bar is 0.2 mm. (d) At the highest metamorphic grades (deepest structural levels), D_2 crenulations are almost completely recrystallized into a foliation defined by parallel laths of muscovite and biotite. A relict hinge is possibly present at the top of the photomicrograph. Inclination of basal cleavages of micas to S_2 is attributed to nucleation of coarse micas in hinge zones of crenulations. Scale bar is 0.15 mm. (e) Upright D_4 crenulations deforming S_2. View is to the northeast. (f) Orthozoisite porphyroblasts aligned primarily parallel to L_2 and L_{2e}. A few randomly oriented (post-kinematic) porphyroblasts are also visible. Length of object in lower right-hand corner is approximately 3 cm.

surfaces (Fig. 2; Fig. 5, location Y). Although not completely mapped, right way up lower and middle Kaza Group rocks may possibly close continuously into the short inverted limb of the Raush Anticline at this locality. However, farther to the west, on the north side of South Kiwa Creek (Fig. 2; Fig. 5, location W), layer-parallel shear zones are found within the inverted limb, suggesting that the inverted limb may be detached from the underlying right way up sequence to the west and southwest.

The predominantly upright panel of middle and lower Kaza Group continues to the southern boundary of the area mapped. West of Mt. Sir John Oliver (Fig. 2), the panel and SW-verging folds affecting it arch over the Premier Anticlinorium, an open and upright D_3 fold (Fig. 5). To the east towards the southern Rocky Mountain Trench, D_3 folds are tighter and more clearly NE-verging. Just to the west of the Trench, visible from Valemount, D_2 and earlier structures steepen through the vertical in interference with D_3 folds, creating the 'fan axis' of Campbell (1968) and Wheeler *et al.* (1972).

Right way up lower Kaza Group rocks overlie, with apparent fault contact, an inverted sequence of amphibolite, quartz–plagioclase–biotite–garnet meta-sandstone, kyanite–staurolite–garnet meta-pelite, and laminated marble and calc-silicate-rich rock that has been intruded by a pre-kinematic (pre-D_1?) diorite (Fig. 5). This sequence is tentatively correlated with the semi-pelite and amphibolite (SPA) and middle marble units of the Horsethief Creek Group (Murphy 1984).

S_2, the axial surface fabric of D_2 folds, is a crenulation cleavage. The morphology of S_2 changes with rock type and metamorphic grade. Figures 10(a) & (b) are from the hinge zone of a NE-verging second-order fold on the lower limb of the Raush Anticline. In the graphitic layer (Fig. 10b), S_2 is clearly a crenulation cleavage, overgrown by coarse, randomly oriented porphyroblasts of biotite and muscovite. In the less graphitic layer, the hinges of D_2 crenulations are obscured by micas aligned in S_2, and randomly oriented post-kinematic micas. At higher metamorphic grades, the hinges of the crenulations are entirely recrystallized in the less graphitic layers and are only visible where graphitic concentrations are found (Fig. 10c). Recrystallized hinge zones are suggested where isolated coarse mica grains are elongate parallel to the foliation but with their 001 cleavages inclined parallel to the foliation at an acute angle (Fig. 10d). In coarser clastic rocks, S_2 is defined by the plane of flattening and extension of deformed quartz and feldspar clasts. S_2 intersects S_0 in an intersection lineation (L_2) that parallels the hinges of D_2 folds (Fig. 8). L_2 is also parallel to the direction of extension of quartz and feldspar clasts and the preferred dimensional orientation of prismatic and tabular porphyroblasts (L_{2e}).

At all metamorphic grades, S_2, L_2 and L_{2e} are overgrown by randomly oriented porphyroblasts of garnet, biotite, muscovite, orthozoisite (Fig. 10f), plagioclase, staurolite and kyanite. These porphyroblasts and earlier fabrics are deformed by two sets of post-metamorphic crenulations: NW-trending D_3 and NE-trending D_4 (Fig. 10e). With the exception of D_3, the relative sequence of events: D_1, D_2, metamorphic peak, D_3 and D_4, is the same as is found in the suprastructure. On this basis, SW-verging, pre-metamorphic-peak D_2 folds in the suprastructure may be tentatively correlated with D_2 folds in the infrastructure. As described in the following section, D_2 structures are physically continuous from suprastructure to infrastructure, thereby confirming this correlation.

Transition zone

The transition from upright to recumbent folding occurs totally within rocks of the middle Kaza Group, near the biotite and garnet mineral isograds (Fig. 5). To elucidate the nature of this transition, the axial surface of an individual D_2 syncline (indicated by arrow in northwest corner of Fig. 2) was mapped down-section from the suprastructure. The axial surface was traced 25 km from Castle Creek southeast to the first drainage northwest of Raush River (Fig. 2). Along its trace, the axial surface gradually decreases in dip (Fig. 5). The eastern limb of the syncline, which is the short limb of a SW-verging anticline–syncline pair, becomes shorter down-section as the axial surfaces of second-order folds become more gently dipping and more densely distributed. Thus, in profile (Fig. 5), the trace of the axial surfaces flattens downward, eventually disappearing in a zone of SW-verging folds with gently dipping axial surfaces.

The western limb of the syncline in the transition zone is only weakly folded (Fig. 5). Individual continuous meta-sandstone and phyllite–schist beds may be traced for several kilometres without passing through any major folds. Those folds that are present are relatively small-scale and consistently verge SW, apparently parasitic on the upper limb of the Raush Anticline of the infrastructure. The closure of the Raush Anticline on the north side of the Raush Valley lies approximately two kilometres structurally deeper than these relatively undeformed, continuous beds.

The transition from upright to near-recumbent folds is accompanied by a continuous change in the orientation and style of S_2. S_2 is generally parallel to D_2 axial surfaces at all structural levels; as such, S_2 undergoes the same transition from steeply to gently dipping orientations. In addition, cleavage fanning and convergence in D_2 hinge zones is less apparent at depth, cleavage refraction is less apparent at depth, and the angle between cleavage and bedding on fold limbs decreases with depth. Structure and stratigraphy may be traced continuously across the transition from suprastructure to infrastructure, uninterrupted by faults or discrete shear zones. SW-verging folds with steep axial surfaces of the suprastructure are parasitic on gently inclined deeper level structures such as the Raush Anticline. The transition occurs over a vertical profile distance of approximately 5 km and involves a structural rotation of approximately 70°.

KINEMATIC ANALYSIS

The structural geometry of the east-central Cariboo Mountains results from the superposition of four phases of deformation. However, the transition from upright to near-recumbent folding and the structural thickness of the Kaza Group are primarily the results of D_2 deformation. In this section, the structural geometry of D_2 deformation is approached from two perspectives: (1) the regional strain involved in the transition from upright to near-recumbent folding and (2) the amount and distribution of structural thickening at different structural levels.

Transition from upright to near-recumbent folding

Upright, relatively symmetrical folds and steeply dipping foliations at shallow structural levels represent approximately 10–20% horizontal shortening (Campbell 1973). The gently inclined folds and foliations found at deeper structural levels cannot be formed purely by coaxial horizontal shortening; consequently, the finite strain at deep levels includes a significant component of non-coaxial shear strain. The continuous overturning of folds and foliations with depth reflects the increase in importance of the shear component with depth. Indeed, this gradient in non-coaxial shear strain is similar to that expected at the margins of shear zones of any scale (Ramsay & Graham 1970, Sanderson 1982).

The variation in the orientation of foliations in a thrust sheet or nappe undergoing both boundary-parallel shortening and downwardly increasing non-coaxial shear strain has been modelled by Sanderson (1982). Upright and symmetrical folds are predicted for shallow structural levels where boundary-parallel shortening strain is more important than the non-coaxial component. With depth and increasing non-coaxial strain, tighter, more asymmetrical, and progressively more overturned folds and foliations result. At very high non-coaxial strains, rotation of structures into the extensional field of the incremental strain ellipsoid causes extension of folded units and reversal of fold asymmetry (Ramsay *et al.* 1983). Using the relationship presented in Sanderson (1982) between θ' (the angle between the shear direction and the trace of the XY kinematic plane in the plane containing the shear direction and the normal to the shear plane) and R, the strain ratio, for different shear strains (ν) and stretches parallel to the direction of transport (α), the magnitude of the non-coaxial component at different structural levels has been estimated (Fig. 11). This exercise assumes a shear plane that originally dipped gently to the east relative to bedding. θ' is measured from the horizontal because of the near-horizontal orientation of axial surfaces and foliations of the tightest folds in areas where post-D_2 deformation is minimal. The sense of rotation of D_2 axial surfaces and foliations suggests that the displacement or shear direction is approximately SW, parallel to the line of section of Fig. 5.

The variation in the magnitude of the component of non-coaxial shear strain has been determined for 10% ($\alpha = 0.9$) and 20% ($\alpha = 0.8$) bulk horizontal shortening. For 10% shortening, shear strain increases downward from 0.14 to 3.85; for 20% shortening, shear strain increases downward from 0.23 to 4.0. The total NE displacement of the infrastructure with respect to a fixed suprastructure ranges from 6.2 km, for 10% bulk shortening, to 7.1 km for 20% bulk shortening.

Amount and distribution of structural thickening of the Kaza Group

In the infrastructure portion of the structural sequence of the Kaza Group, stratigraphy is duplicated by isoclinal folding around gently dipping axial surfaces. This geometry implies a significant amount of structural thickening that is not present in the suprastructure. The small size and low metamorphic grade of pre- and post-D_2 structures suggest that most of this structural thickening was produced by D_2 deformation. If one assumes that most of the structural thickening is due to D_2 deformation, the amount and distribution of structural thickening can be estimated by comparing the present structural thickness of the Kaza Group with its undeformed stratigraphic thickness.

Although the original thickness of the Kaza Group at this locality is not precisely known, it can be estimated from the measured thicknesses of a less highly folded, continuously right way up section of the Kaza Group located south of the Premier Range (Fig. 2) (Pell & Simony 1984, Pell 1984) and a stratigraphic section of the probably equivalent middle Miette Group of the Rocky Mountains (Carey & Simony 1984, Carey 1984). These two sections (Fig. 6) are quite similar and it is reasonable to assume that the stratigraphic thickness of the Kaza Group in the east-central Cariboo Mountains, which lies between the other two sections, is within the range of thickness provided by the other two sections. This assumption may not be strictly correct and original thickness variations within the Kaza Group may have been present prior to deformation, although they are not likely to be very large given the proximity of the measured sections to one another.

As is shown in Fig. 6, the structural sequence of the Kaza Group of the east-central Cariboo Mountains is more than double the thickness of either of the other two sections. It is also apparent that structural thickening of the Kaza Group is unevenly distributed throughout the sequence. The upper Kaza Group is similar in thickness in all three sections. In contrast, the middle Kaza Group, where deformed by the Raush Anticline and related structures, is 5.5 times thicker than the middle Kaza Group south of the Premier Range and 3.7 times thicker than the equivalent part of the middle Miette Group.

In order to produce the observed distribution of structural thickening, middle Kaza Group rocks must have undergone a greater amount of regional shortening than immediately overlying upper Kaza Group rocks. 'Shortening' in this case is accommodated by overriding and accreting middle Kaza Group rocks which originally lay

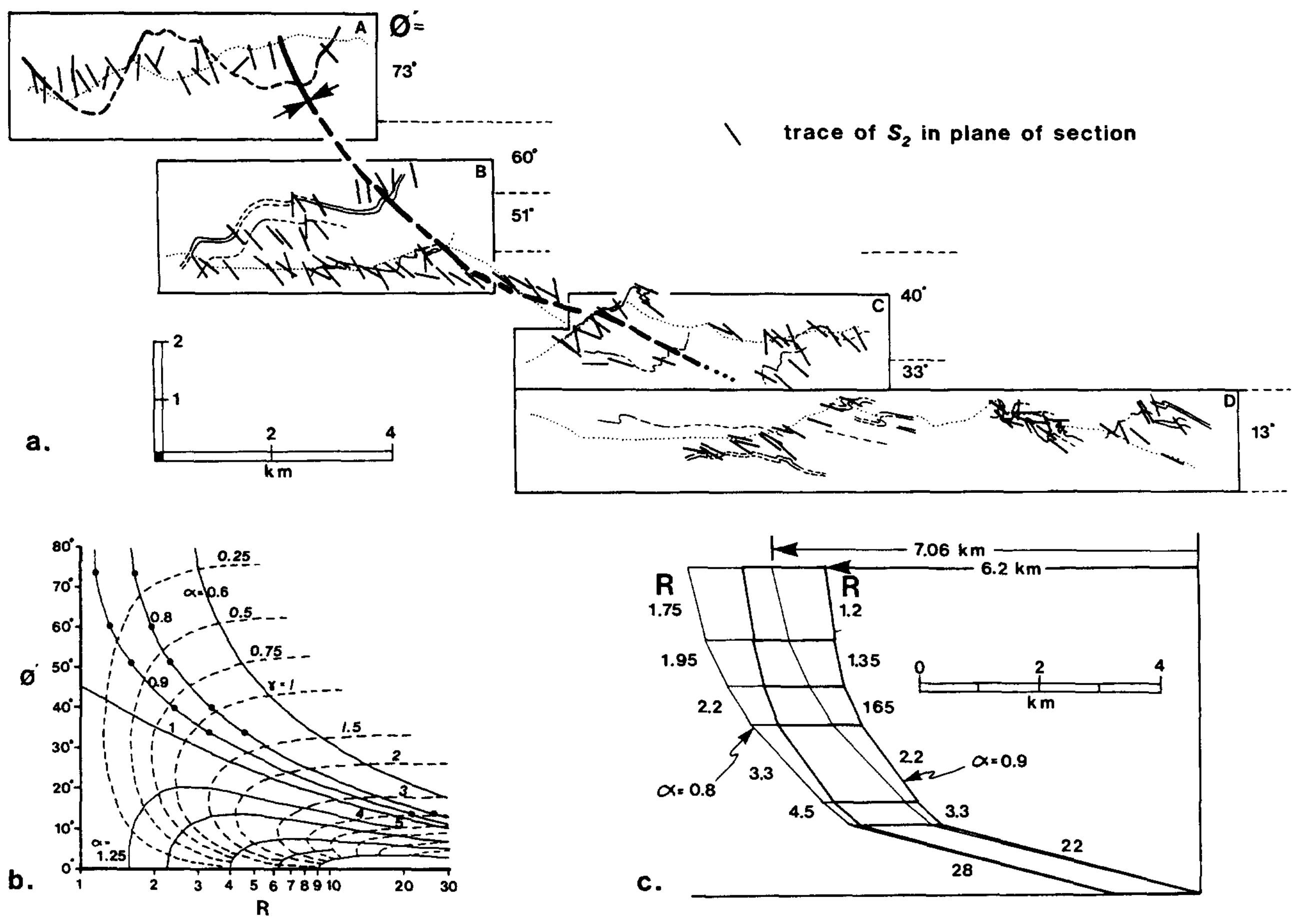

Fig. 11. Kinematic analysis of transition from upright folds and steeply dipping foliations to near-recumbent folds and gently dipping foliations. (a) Change in orientation of S_2 with depth. Angle at right is the angle between the average foliation and a horizontal plane of shear (θ'). (b) Graph showing the relationship between θ', the angle between a foliation and its associated plane of shear, and R, the strain ratio for different values of the components of ν (component of simple shear strain), and α, the longitudinal elongation or shortening ($\alpha = 1$ is simple shear strain). Large dots on $\alpha = 0.8$ and 0.9 curves are the values of θ' shown in Fig. 11(a). Curves from Sanderson (1982). (c) Values of R and horizontal displacement associated with the values of ν determined in Fig. 11(b). The heights of individual sheared boxes include vertical extension associated with α; as such, markers would have different original heights.

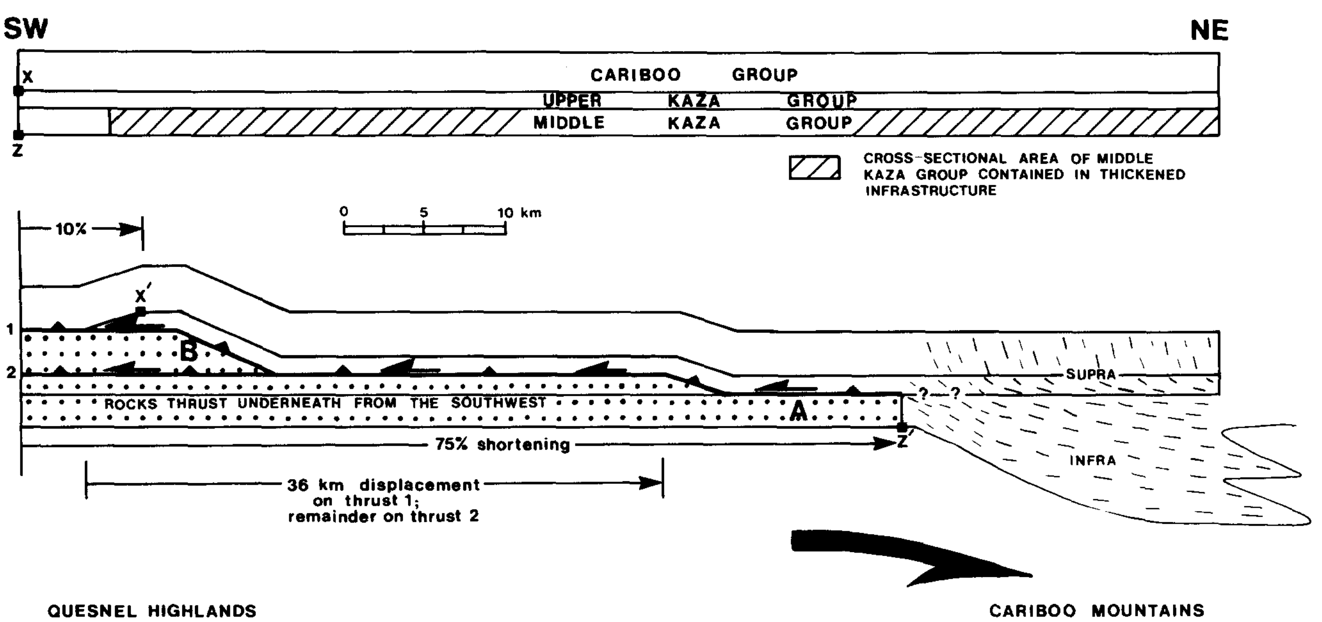

Fig. 12. Diagrammatic model illustrating the relationship between infrastructural ductile thickening and thrust faulting in the Quesnel Highlands. Section balanced and to scale.

to the southwest of the upper Kaza Group rocks of the suprastructure. The difference in the amount of regional shortening and thickening between rocks of the infrastructure and immediately overlying rocks of the suprastructure requires that they be detached from each other, to the southwest, in the direction of relative hanging wall transport (Fig. 12). Kaza Group rocks to the southwest are predicted to be in reverse fault contact with either stratigraphically younger rocks as faults ramp into high structural levels (location B, Fig. 12) or with stratigraphically older rocks on the basal flat (location A, Fig. 12).

The former relationship has been observed near Quesnel Lake (Fig. 1) where Kaza Group rocks are in the hanging wall of the structurally highest (yet identified) thrust of a SW-verging, pre-metamorphic-peak fold and thrust belt (Struik 1981, 1982a,b, 1983a,b). The nature of the strain in these rocks is not completely understood and the expression 'fold and thrust belt' may be an oversimplification, but pre-metamorphic-peak, SW-verging, low-angle reverse faults have been documented. It is proposed that the structural thickening and shear strain evident in the east-central Cariboo Mountains links laterally and up-section into this belt of low-angle faults.

Magnitude of crustal displacements

The change in orientation of structures between suprastructure and infrastructure represents approximately 7 km of NE displacement of the infrastructure with respect to a fixed suprastructure (Fig. 11). The difference in the amount of 'shortening' between suprastructure and infrastructure requires additional NE displacement. To determine the magnitude of this additional displacement, the cross-sectional area of the thickened middle Kaza Group in Fig. 5 has been restored to an idealized pre-D_2 configuration (Fig. 12). This scaled restoration establishes that approximately 40 km of NE displacement of the infrastructure can account for the structural thickness of the middle Kaza Group. Thus, the structural geometry shown in Fig. 5 represents approximately 50 km of NE underthrusting of middle Kaza Group and structurally deeper rocks with respect to a fixed suprastructure of upper Kaza Group and younger rocks.

In the Quesnel Highlands (Fig. 1), the geometry of thrust slices in the hanging wall of the structurally lowest Pleasant Valley Fault records approximately 70% shortening (Struik 1980). A comparison of the deformed and restored sections above the Pleasant Valley Fault (Struik 1981) suggests that 25–30 km of NE displacement of the footwall is associated with the shortening. The displacement on the Pleasant Valley Fault is not known due to the lack of a well-defined footwall cut-off. However, the structural topography of regional-scale culminations and depressions such as the Lanezi Arch, the Isaac Lake Synclinorium, and the Black Stuart Synclinorium (Fig. 1) may be due to changes in the trajectory of the Pleasant Valley Fault at depth. On preliminary analysis, a reasonable location of a footwall cut-off would be beneath the Lanezi Arch, suggesting a displacement on the Pleasant Valley Fault of approximately 35 km. Thus a total of 60–65 km of NE underthrusting is associated with the fold and thrust belt in the Quesnel Highlands.

The way in which relatively discrete displacements in the Quesnel Highlands transform into structural thickening in the infrastructure farther to the east is not known. However, both the fold and thrust belt and the zone of infrastructural thickening lie at the interface between a relatively uninvolved hanging wall (suprastructure) and a footwall that has been underthrust to the NE by 50–65 km. The hanging wall has been thickened by successive accretion of material stripped off the footwall and underplated by large-scale infrastructure folds and shallow-level thrust faults (Fig. 12). The fold and thrust belt and the zone of infrastructure folding may be viewed as a broad crustal shear zone accommodating shear strains associated with northeastward underthrusting.

TECTONIC IMPLICATIONS

Mid-Jurassic, syn-metamorphic W- to SW-verging structures have been recognized in many localities in the southeastern Canadian Cordillera (Fig. 13). In most of these localities, these structures fold an earlier E- to NE-verging set of structures (D_1?), including the Quesnel Lake Shear Zone (Brown *et al.* 1986), which marks the obduction boundary between Terrane I, the easternmost of the Cordilleran 'suspect' terranes (Monger *et al.* 1982), and North America. This boundary was active after the Sinemurian (late Early Jurassic), but before metamorphism and W- to SW-verging deformation. Additionally, in many of these localities, W- to SW-verging structures are deformed by late- to post-metamorphic E- to NE-verging structures which are spatially associated with early thrust faults which feed into the Rocky Mountain foreland.

In light of the importance of E- to NE-verging structures in the kinematic evolution of the southeastern Canadian Cordillera (up to 450 km of total displacement on faults in the Rocky Mountain fold and thrust belt, on the Monashee décollement, and on the Quesnel Lake shear zone; Brown *et al.* 1986), regionally distributed W- to SW-verging structures are somewhat enigmatic, falling into the categories of backfolds and backthrusts. Two tectonic models (Figs. 14 and 15) have been proposed to account for backfolding and backthrusting in the Cordillera.

Model 1 (Fig. 14)

The early Jurassic obduction of Terrane I was followed by the insertion of the basement of Terrane I along the basement/cover contact beneath the North American continental margin prism, delaminating the cover from its basement and inducing W- to SW-verging deformation and metamorphism in the overlying sediments (Price 1983, Archibald *et al.* 1983, Price *et al.* 1985). Continued E to NE displacement of the indenter on its

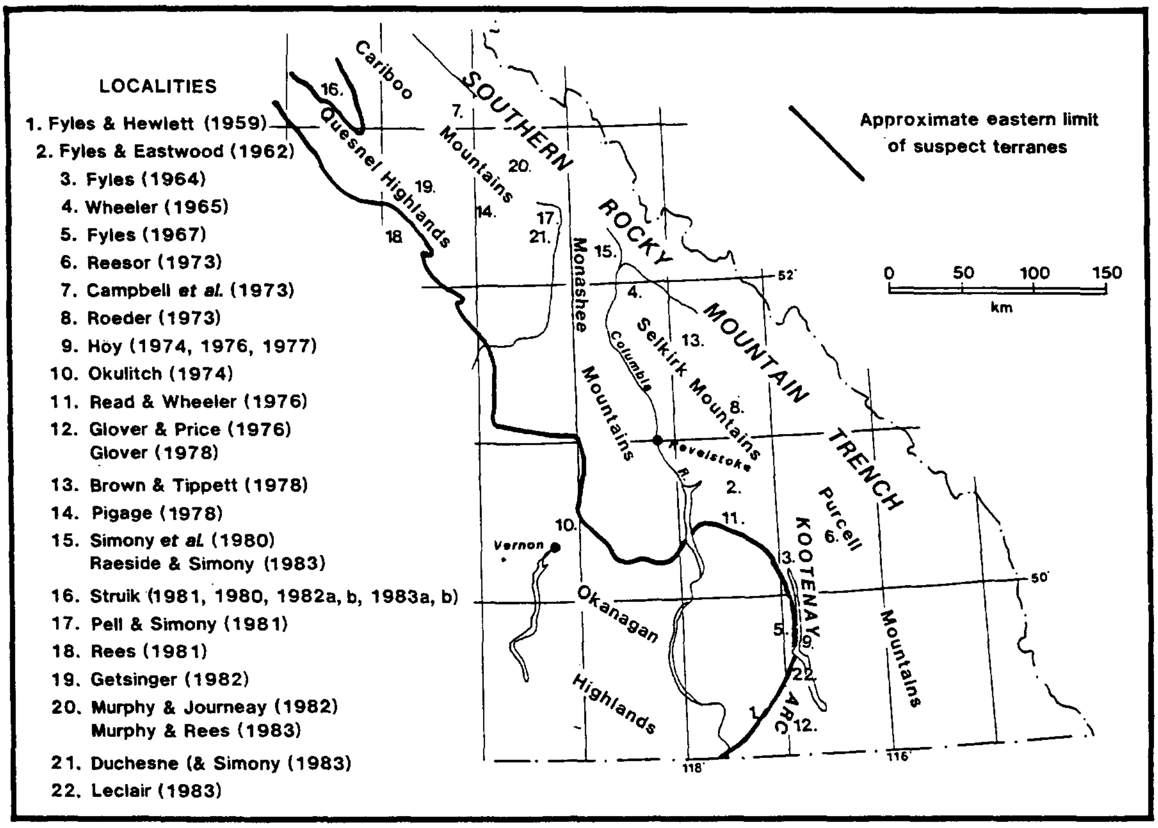

Fig. 13. Areas in the southeastern Canadian Cordillera in which regional-scale (south)westwardly verging structures have been described.

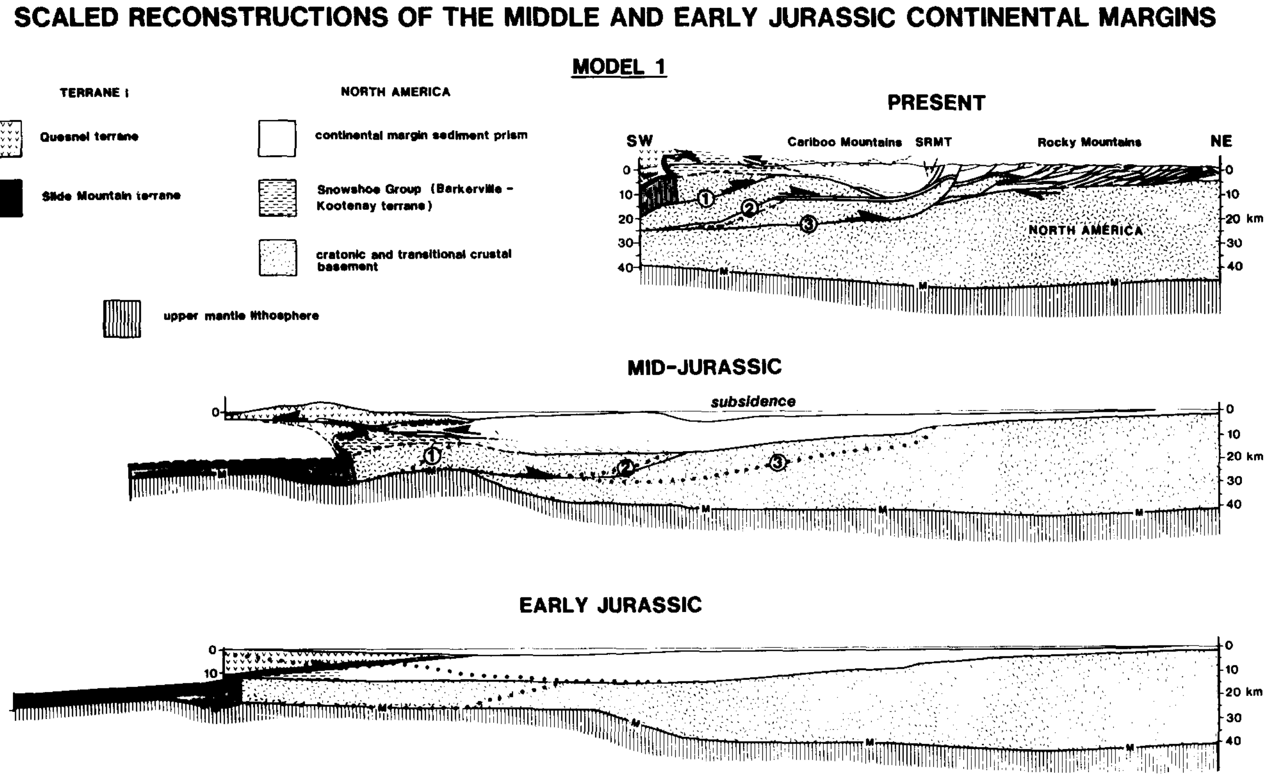

Fig. 14. Model 1 (see text for discussion). Vertical and horizontal scales are equal.

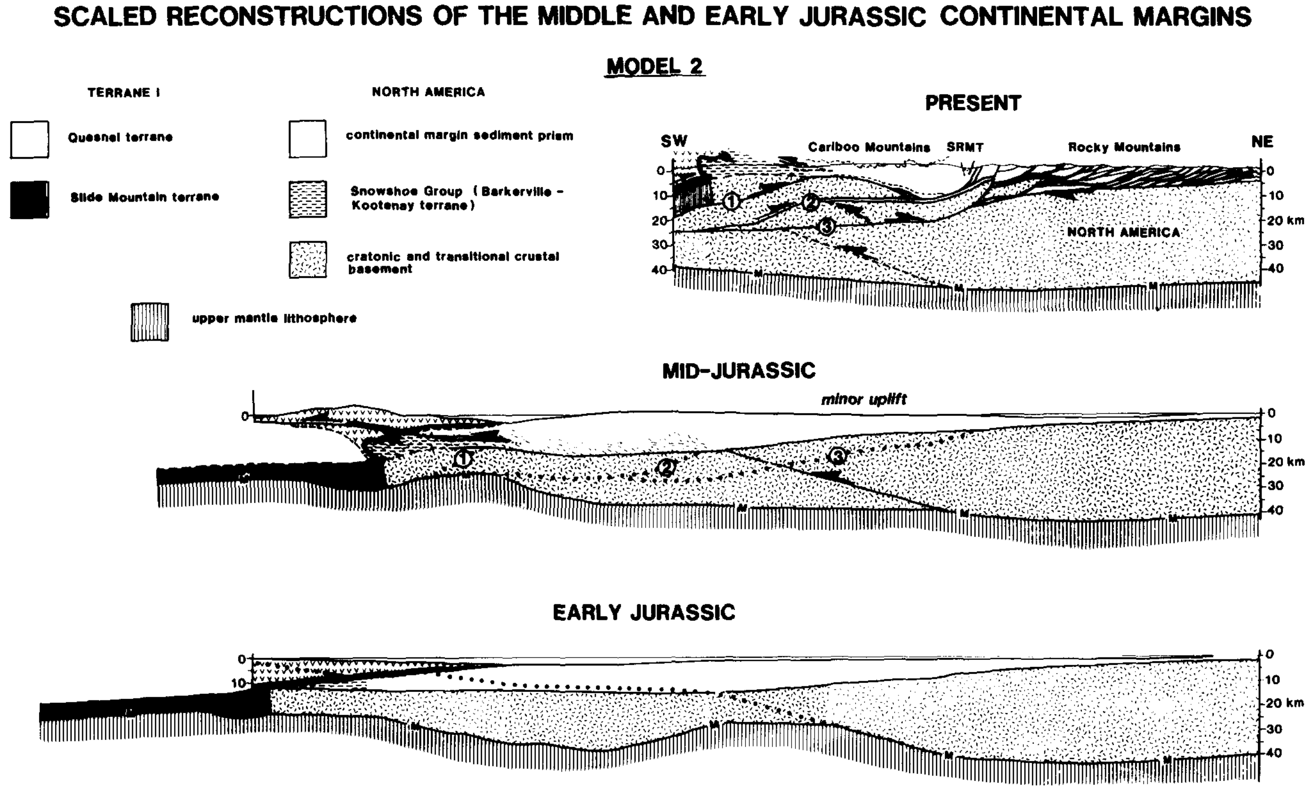

Fig. 15. Model 2 (see text for discussion). Vertical and horizontal scales are equal.

basal shear zone resulted in folding of inactive W- to SW-verging structures and initiation of shortening in the Rocky Mountain fold and thrust belt.

Model 2 (Fig. 15)

The early Jurassic obduction of Terrane I was followed by an eastward transfer of the active boundary between Terrane I and North America into the North American continental margin prism and its underlying crustal basement (Monger 1977, Brown 1981, Brown *et al.* 1986). The new boundary dipped E and subsequent convergence was accommodated in the basement by underplating the transitional crust that originally underlay the continental margin prism beneath the transitional crust still attached to cratonic North America. After crustal thickening and metamorphism, E to NE vergence was re-established with the formation of a new, W-dipping boundary. E to NE displacement on this boundary and associated shear strain resulted in folding of inactive W- to SW-verging structures and the initiation of shortening in the Rocky Mountain foreland.

Surface geometric and kinematic data, as presently understood, are not sufficient to choose between the two models. They differ primarily in how, and where, basement is shortened; as such, an unequivocal choice can only be made on the geometry of basement deformation, which cannot be observed directly. However, the different geometries and locations of basement deformation imply different initial geometries and geodynamic responses which may have different signatures in the pre- and syn-kinematic geological records. To examine the implications of each model, the Cordilleran continental margin was restored to its configurations just after D_2 deformation, but before shortening of the Rocky Mountain foreland, and just before D_2 deformation (Figs. 14 and 15).

The reconstructions shown in Figs. 14 and 15 are based on the balanced restoration of displacements associated with shortening in the Rocky Mountain foreland and with D_2 deformation of the outer part of the North American continental margin prism. D_2 displacement trajectories in the basement are determined by assuming that the eastern limit of D_2 deformation (essentially the southern Rocky Mountain Trench) marks the farthest advance of the 'hanging wall' of each model. In Model 1, this eastern limit is interpreted as the location of the leading edge of the delaminating wedge, where the oppositely verging shear zones bounding the wedge intersect. In Model 2, the eastern limit is inferred to be where the trajectory of W- to SW-verging shear strain leaves the crystalline basement and enters the supracrustal sediments. These 'hanging wall cut-offs' are combined with a displacement of 50 km to construct internally consistent displacement trajectories and to locate corresponding footwall cut-offs. Displacement trajectories in the basement associated with shortening of the Rocky Mountain foreland were constructed by combining the inferred geometry of basement thrust slices in the Rocky Mountain Main Ranges with displacement of 125 km (Brown *et al.* 1986). Four preliminary conclusions can be drawn from the geometries shown in Figs. 14 and 15.

(1) The different styles and locations of basement

thickening in the two models would be expected to have different implications for the geological record of the Rocky Mountain foreland. The loading of basement and thickened metasediments onto a footwall still attached to the North American craton (Model 1) would be expected to depress the footwall, inducing subsidence and marine transgression in the Rocky Mountain foreland. Alternatively, underthrusting of transitional crust beneath a hanging wall of transitional crust still attached to North America (Model 2) would be expected to induce uplift in the Rocky Mountain foreland. This uplift would be somewhat counterbalanced by the effects of loading the deformed supracrustal metasediments onto the footwall.

During the Sinemurian to Callovian interval in which D_2 deformation was taking place, the Rocky Mountain foreland was characterized by alternating periods of deposition of shallow marine sediments and non-deposition in which little or no erosion occurred (Frebold & Tipper 1970). Thus, the Rocky Mountain foreland ranged from weakly positive to weakly negative, apparently uninfluenced by the deformation that was occurring to the west. Such a geological history is not conclusive, but is mildly supportive of Model 2, which would be expected to produce minor uplift in the Rocky Mountain foreland.

(2) The footwall cut-off of the delaminating wedge (Model 1) is constrained to lie in the basement to the North American continental margin prism. Thus, the leading edge of the rigid indenter would be composed of transitional continental margin basement, not the oceanic basement to Terrane I, as suggested in Model 1. Displacements on the order of 150 km would be necessary to transport rocks of Terrane I from their early Jurassic position outboard of the North American continental margin prism to beneath rocks presently at the southern Rocky Mountain Trench.

(3) Model 2 predicts that the basement of cratonic thickness that presently underlies the Cariboo Mountains consists partly of a slice of allochthonous transitional crust that was underplated beneath the autochthonous transitional crust still attached to the craton. If Model 2 is correct, the westward transition from 50–55 km thick crust to 30–35 km thick crust that occurs beneath the Cariboo Mountains and Quesnel Highlands may be related to the geometry of D_2 deformation rather than to Paleozoic continental rifting, as has been suggested by numerous workers (Monger & Price 1979, Price 1981, Price & Fermor 1981, Chamberlain & Lambert 1985).

(4) Model 1 predicts that the westward transition from cratonic to attenuated transitional (or oceanic?) crustal basement produced by crustal extension during the formation of the Cordilleran miogeocline lay 100–150 km west of autochthonous crust that presently underlies the southern Rocky Mountain Trench (Fig. 1). The lower Paleozoic rocks of the Cordilleran miogeocline undergo a transition from platformal facies to basinal facies in the vicinity of the southern Rocky Mountain Trench. This 'hinge line' is displaced approximately 125 km to the west when the shortening in the Rocky Mountain fold and thrust belt is restored. Its palinspastic location coincides approximately with the transition from attenuated transitional to cratonic crust predicted by Model 1. This coincidence is not conclusive, but is indirect support for Model 1.

None of the transitions from cratonic to transitional or oceanic crust indicated in any of the deformed or restored sections of Model 1 or 2 correspond with the transition from 50–55 km crust of low electrical conductivity to 30–45 km higher conductivity crust that is observed at the southern Rocky Mountain Trench (Price 1981, Chamberlain & Lambert 1985). The deformed and restored sections in Figs. 14 and 15 imply that this transition is due to Mesozoic tectonism, rather than extension during the formation of the Cordilleran continental margin.

The data and interpretations presented in this paper are useful in constraining the two models that have been proposed for the deformation of the Jurassic continental margin, but they cannot be used as an unequivocal basis to choose between them. The two models must be refined by more detailed field work and tested mechanically and experimentally before firm conclusions can be drawn.

CONCLUSIONS

(1) The transition from upright to near-recumbent structures in the Cariboo Mountains involves syn-metamorphic, but pre-metamorphic-peak, SW-verging D_2 structures.

(2) The transition is gradual and continuous and results from the non-coaxial component of strain produced during northeastward underthrusting of the North Americal continental margin prism.

(3) Structural thickening at infrastructural levels is much greater than at suprastructural levels, requiring that shallow and deep structural levels be detached southwest of this portion of the Cariboo region. A SW-verging syn-metamorphic fold and thrust belt near Quesnel Lake is likely to be the shallow-level manifestation of the deep-level shortening evident in the east-central Cariboo Mountains.

(4) The structural geometry and distribution of structural thickening of the Kaza Group in the east-central Cariboo Mountains can be accounted for by approximately 50 km of northeastward underthrusting of deep structural levels with respect to a fixed suprastructure.

(5) This amount of displacement has been combined with the eastern limit of SW-verging deformation to specifically define the structural geometries implicit in the two different tectonic models that have been proposed to account for backfolds and backthrusts in the southeastern Canadian Cordillera. Neither model can be disproved on the basis of geological or geophysical observations, or on the basis of the predicted geological record of neighbouring regions determined from scaled reconstructions of the Jurassic continental margin.

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