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Kinematics of compressional and extensional ductile shearing deformation in a metamorphic core complex of the northeastern Basin and Range

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Abstract—Analysis of shear criteria enables the kinematics of two main ductile-shearing events (D_1 and D_2) to be established in the Raft River, Grouse Creek and Albion 'metamorphic core complex'. The first event (D_1) is a NNE-thrusting and corresponds to Mesozoic shortening. A well developed non-coaxial ductile deformation (D_2), of Cenozoic age, is marked by the occurrence of opposing eastward (in Raft River) and westward shear criteria (in Albion–Grouse Creek). These characterize an arch structure where the shear strain increases outwards. In the axial zone of the complex, D_2 seems coaxial. Cenozoic extension is considered to be related to gravitational instability induced by mesozoic overthickening of the crust (involving uplift, erosion and abnormal heating). Brittle extension occurs in the upper part of the uplifted domain. It is transformed laterally above undeformed basement towards stretched domains of the middle and lower crust through the ductile shear zones localized at the Precambrian–Paleozoic interface. This extension of the middle and lower crust occurred in the vicinity of the root zones of the former Mesozoic thrusts, which may thus have been reactivated as ductile normal faults during Cenozoic extension.

INTRODUCTION

SEVERAL 'metamorphic core complexes' (Crittenden *et al.* 1980, Armstrong 1982) occur from Canada to Mexico along an average N–S-striking belt. They expose strongly deformed and metamorphosed rocks. The tectonic significance of such ranges in the general framework of the North American Cordillera is under debate. During the last two decades, numerous interpretative models have been published (Armstrong 1968, Mudge 1970, Price & Mountjoy 1970, Campbell 1973, Hose & Danes 1973, Price 1973, Roberts & Crittenden 1973, Scholten 1973, Fox *et al.* 1977, Armstrong 1978, Brown 1978, Drewes 1978, Davis & Coney 1979, Coney 1980, Crittenden *et al.* 1980, Allmendinger & Jordan 1981, Wernicke 1981, Armstrong 1982, Davis 1983, Miller 1983a, Coney & Harms 1984, Smith & Bruhn 1984, Davis *et al.* 1986) and discussed (Thorman 1977, Crittenden 1979, Davis *et al.* 1980, De Witt 1980, Wernicke 1982, Brown & Read 1983, Mattauer *et al.* 1983).

Some models attempt to relate metamorphism and ductile deformation to the Mesozoic compressional orogeny, rather than Cenozoic extensional tectonics. Mesozoic compressional deformation and metamorphism were first documented by Misch and collaborators (Misch 1960, Misch & Hazzard 1962, Dover 1969, Nelson 1969, Thorman 1970), who also established the main characteristics of the 'décollement' faulting, of younger-on-older style. Armstrong introduced the 'mobile infrastructure' concept, and proposed both Mesozoic and Cenozoic metamorphic ages on the basis of geochronological data (Armstrong 1964, Armstrong & Hansen 1966, Armstrong 1972). Thus, he considered that the infrastructure of the metamorphic core complexes represents the Sevier Orogenic Belt compressional hinterland, and that the décollements are the results of Tertiary gravitational tectonism and metamorphism. Further south, in Arizona, Davis & Coney (1979) propose Cenozoic exensional tectonics in their boudinaged crust model to account for the different deformation structures observed. Wernicke (1981) and Davis *et al.* (1986) attributed the ductile deformation and metamorphism of the metamorphic core complexes to a crust-penetrating low-angle shear zone involving simple shear as the main deformational mechanism. Several authors (Proffett 1977, Eaton 1979, Rehrig & Reynolds 1980, Miller *et al.* 1983, Smith & Bruhn 1984, Gans *et al.* 1985) explain the stretching of the crust by a pure-shear model which involves a subhorizontal, mid-crustal decoupling horizon. Hamilton (1983) and Kligfield *et al.* (1984) proposed a model in which extension in the lower crust is accommodated by lenses of anastomosing ductile shear zones. Different models associate ductile deformation and metamorphism with doming and granite emplacement, of mostly Jurassic age in the Ruby Mountains (Howard 1980) and Tertiary age in Grouse Creek and the Raft River Mountains (Compton 1980, Todd 1980). Structural and geochronological data reveal a polyphased tectono-metamorphic history. The Albion, Grouse Creek, and Raft River metamorphic core complexes were studied to gain further understanding of the structure and kinematics of both Mesozoic compressional and Cenozoic extensional tectonics. The approach was to define the geometry of the ductile deformation using microtectonic techniques, particularly strain analysis and shear-criteria analysis. These results and their bearing on the kinematic history are discussed and placed in the general framework of the Cordilleran thrust belt and extensional Basin and Range.

GEOLOGIC FRAMEWORK

The Albion–Grouse Creek–Raft River Mountains are located along the western part of the Idaho–Utah border (Fig. 1a). Prior to Mesozoic and Cenozoic deformations, the Archean basement of the North American margin was overlain by a thick sequence of continental shelf sediments made of upper Proterozoic to lower Cambrian clastic rocks (Miller 1983b) and Paleozoic carbonate rocks. Previous geological studies and mapping (Armstrong 1968, 1970, Compton 1972, 1975, 1980, Compton *et al.* 1977, Miller 1980, Todd 1980, Jordan 1983) show the structural resemblance of the three ranges (Fig. 1b). Two main structural units are usually defined: a metamorphic basement (unit 1, Fig. 1) consists of 2500 Ma old Archean granite and metasediments unconformably overlain by Proterozoic Z and lower Paleozoic schists and quartzites (see Compton *et al.* 1977). Above this basement are several allochthonous sheet that are usually divided into two main groups; the lower allochthonous sheets (unit 2, Fig. 1) consist of upper Proterozoic to Ordovician metamorphosed sedimentary terranes; the upper allochthonous sheets (unit 4, Fig. 1) include slightly metamorphosed or non-metamorphosed upper Paleozoic and lower Mesozoic rocks. The thick, allochthonous sequence (unit 3, Fig. 1) of Proterozoic Z to Cambrian quartzites and schists exposed on Mount Harrison (Miller 1983b) in the Albion Mountains belongs to the lower sheets. These rocks are part of an overturned sequence of metasedimentary rocks overthrusted on inverted metamorphosed Paleozoic series.

All these units are subparallel and bounded (and often cut) by low-angle faults. These low-angle faults are commonly subparallel to the bedding and/or to the main foliation, carry younger rocks over older ones (or unmetamorphosed over metamorphosed) and locally eliminate stratigraphic units. Repetitions of units are extremely uncommon (e.g. Mount Harrison Nappe), and are the result of early thrusting. It seems that many of the major tectonic contacts originated as polyphase ductile faults and were later reactivated as brittle shear zones, as evidenced by tectonic breccias and gouges. Compton *et al.* (1977) have shown that major brittle eastward transport (several kilometres) on low-angle faults occurred after metamorphism (or late during metamorphism). More recent gravity-induced tectonic denudation emplaced large slices of Paleozoic material into the Tertiary basins (Compton 1983, Covington 1983, Todd 1983).

Effects of low-angle faulting on the metamorphic zonation can be summarized as follows (Armstrong 1968, Compton *et al.* 1977, Todd 1980). In some places the upper allochthonous sheets are not metamorphosed whereas they generally show a downward-increasing low-grade metamorphism. At a regional scale, metamorphism in the basement increases towards the west. Metamorphic grades also increase downward within lower sheets and basement rocks. In Raft River Mountain, the basement is metamorphosed in green-

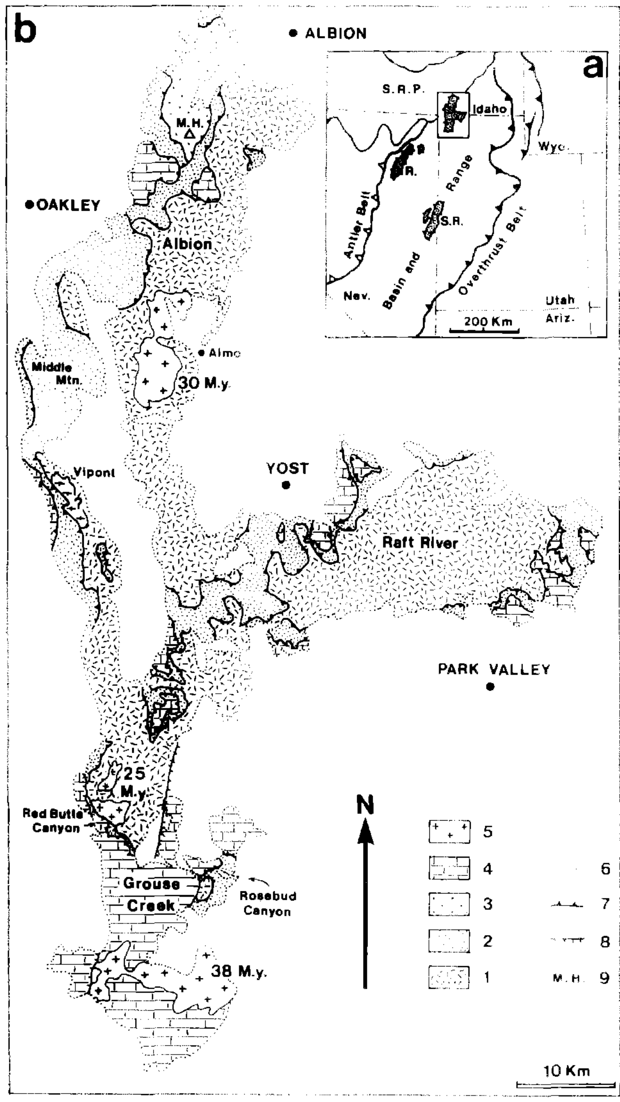

Fig. 1(a). Tectonic setting of metamorphic core complexes in the Basin and Range Province. R., Ruby Mountain; S.R., Snake Range; S.R.P., Snake River Plain. (b) Structural sketch map. 1, Basement; 2, Lower allochthonous sheets; 3, Inverted Mount Harrison allochthonous sheet; 4, Upper allochthonous sheets; 5, Granitic rocks; 6, Cenozoic series boundaries; 7, Main low-angle faults; 8, Recent normal faults; 9, Mount Harrison (M.H.).

schist facies and, locally, to kyanite–chloritoid–muscovite assemblages (Compton *et al.* 1977). In the Grouse Creek Mountains, the metamorphic grade reaches the amphibolite facies, and the Albion Mountains show greenschist facies assemblages in the eastern part of the range and amphibolite facies in the western part (Armstrong 1968). A temperature increase in the vicinity of Almo Pluton (south-west of Albion) is marked by cordierite–andalusite–sillimanite assemblages. The highest grade (sillimanite and diopside) is reached in the westernmost part of the ranges, at Middle Mountain (Miller *et al.* 1983). Lateral variations in metamorphism in the allochthonous sheets indicate major post metamorphic displacements of the different units relative to the underlying basement (Compton *et al.* 1977).

SUPERIMPOSED DUCTILE DEFORMATIONS

After allowing for the effects of late doming, penetrative foliation was approximately horizontal and subparallel to bedding during the main tectonic events in the three ranges. Ductile deformation affects an important thickness of rocks (close to 1 km) and strain intensity decreases downwards in the basement rocks. For example, the Archean granite of the core of Raft River Mountain is undeformed a few hundred meters below the allochthonous sheets. In the allochthonous sheets, finite strain is more heterogeneous. Rocks of the lower sheets are often strongly deformed, whereas ductile deformation is mainly restricted to narrow fault zones in the upper sheets. The most outstanding microstructures throughout the ranges are two prominent sets of stretching lineations (Fig. 2). The first trends N–NE and is related to the first phase of deformation (D_1). The second trends E–W and is related to the second phase D_2 (Compton *et al.* 1977).

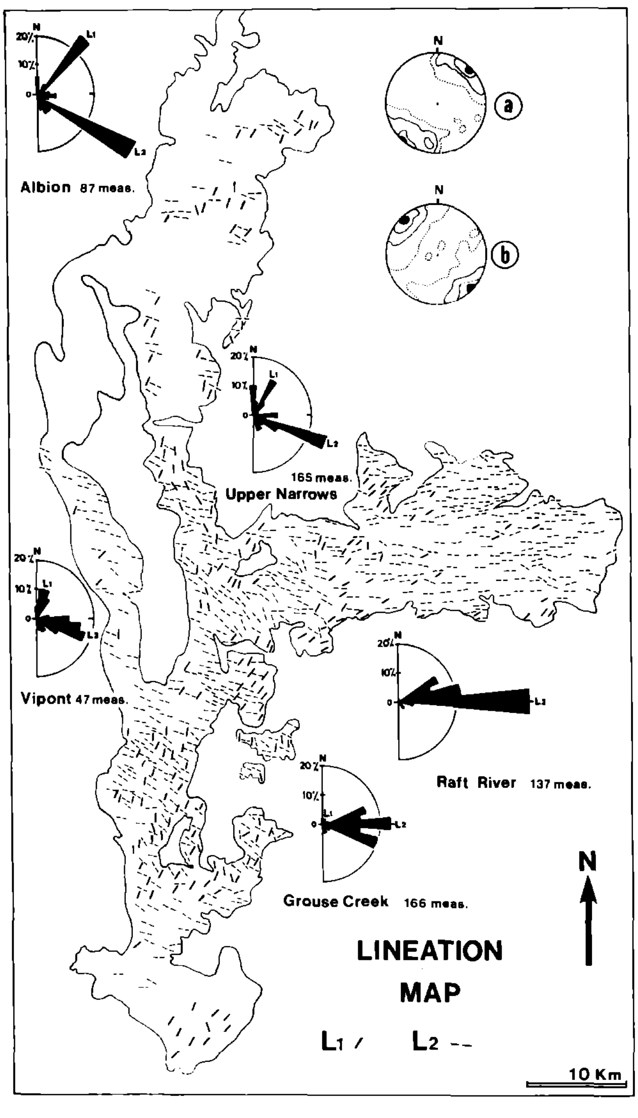

Fig. 2. Lineation map; data from Compton (1972), Compton *et al.* (1977), Miller (1980), Jordan (1983) and personal observations. (a) L_1 lineations (208 measurements), and (b) L_2 lineations (650 measurements), plotted on lower hemisphere Schmidt stereograms (Albion Mountain; after Miller 1980). Rose diagrams for various localities from personal observations.

D_1 *structures*

Most D_1 structures are prevalent in the basement rocks. They are well expressed along the N–S-trend from Grouse Creek to the Albion Mountains, their abundance decreases rapidly towards the east and they disappear completely in the eastern part of the Raft River Mountains (Fig. 2). The foliation S_1 is subhorizontal and roughly parallel to the initial bedding in the sedimentary rocks. It bears a prominent stretching lineation L_1 whose direction varies from NNE–SSW in Albion to N–S in Grouse Creek (Fig. 2). It is outlined in orthogneissic rocks by the elongation of feldspar, quartz ribbons, mica aggregates, stretching of porphyroclasts and elongation of recrystallized tails; and in quartzitic rocks by the elongation of quartz grains and pebbles. Although L_1 is barely preserved in the calcareous Paleozoic sediments of the allochthonous sheets, it is outlined in the fossiliferous limestones of the southeast part of Grouse Creek by highly stretched crinoid ossicles (Fig. 3a) and well-developed pressure shadows around pyrite (Fig. 3b). Boudinage of competent layers also indicates stretching in a N–S direction.

In the basement, scarce folds related to the D_1 phase have axes parallel to the N–S stretching lineation. Generally of slight amplitude, these isoclinal or tight folds are overturned or recumbent to the west. No major D_1 fold closures have been described in these ranges, but the thick overturned rock sequence of Mount Harrison suggests that large-scale recumbent folds were present.

D_2 *structures in Albion–Grouse Creek Mountains*

In the Albion and Grouse Creek Mountains, D_2 is heterogeneous, with local development of mylonitic zones. Stretching and mineral lineations L_2 directed N080°E to N120°E lie in subhorizontal composite S_{1-2} foliations. L_2 is marked by stretched minerals within mylonitic zones which developed parallel to S_{1-2}. Away from mylonitic zones, the L_2 stretching is outlined by the alignment of muscovites or biotites superimposed on the still visible L_1. In the Grouse Creek basement, metre-thick mylonitic zones bound large gneissic slices within which D_1 structures are preserved. The D_2 deformation intensifies towards the west. For instance, the granites of the Almo Pluton (southeast of the Albion Mountains) are undeformed, while those of the Middle Mountains (Armstrong 1976) and Vipont Mountains in the west are mylonitized (Fig. 4f) (Compton *et al.* 1977).

In the lower allochthonous sheets, D_2 is locally intense. For example, an intense folding affects the Mount Harrison Ordovician limestones. In the vicinity of stop 8 of the Utah geological and mineral survey field trip (Miller *et al.* 1983, p. 13), the authors noted the scattered orientations of fold axes. My measurements of fold axes (Fig. 5) show a clear maximum close to the regional direction of the stretching lineation. Observation on the cliff in sections normal to L_2 show numerous 'eyelike' and 'noselike' structures (Fig. 6) at different scales, which are characteric of sheath folds (Quinquis *et*

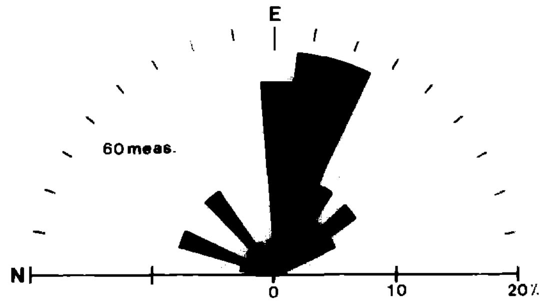

Fig. 5. Rose diagram of fold axes directions (Ordovician limestones of Mount Harrison).

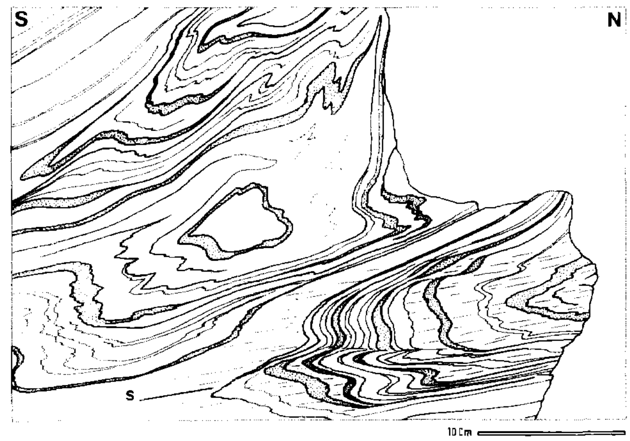

Fig. 6. Sheath folds in metamorphic limestones of Mount Harrison (section perpendicular to stretching lineation L_2).

al. 1978). In these strongly deformed limestones, previous deformations have been totally overprinted.

Outside localized zones of strong deformation, D_2 is much less developed. F_2 folds with N–S axes are gently recumbent and overturned towards the west. Curved fold axes are locally observed. Intermediate zones of D_1–D_2 superimposition show superimposed structures. The previous N–S linear fabric of rocks is curved (Fig. 3d) in the later composite $S_{1\text{-}2}$ foliation, and a less penetrative E–W mineral lineation L_2 develops, often outlined by muscovites or biotites. Some F_1 folds parallel to L_1 are also deformed and curved locally by more than 100° (Fig. 3e). The superimposed L_2 lineation direction is always E–W.

D$_2$ *structures in the Raft River Mountains*

The D_2 deformation is characterized by a strong stretching lineation with a rather constant N80°E trend throughout the range (Fig. 2), and a subhorizontal S_2 foliation generally parallel to S_1 where present. The lineation L_2 lies in the S_2 foliation plane or in composite $S_{1\text{-}2}$ foliation. It is defined by the elongation of pebbles (Fig. 3c) in the conglomeratic levels of the Elba quartzite, by the stretching of porphyroclasts (feldspar, kyanite, tourmaline, etc.), by mineral elongation and boudinage of micas and of quartz grains in quartzites, and by elongation of mineral-filled pressure shadows. F_2 folds are more often observed than F_1 folds, which are very scarce. The most spectacular expression of F_2 folds in the basement is a set of large kilometre-scale recumbent folds, which extend for more than 20 km over the southern flank of the range. These major folds and associated smaller scale folds have axes parallel to the E–W stretching lineation. Most F_2 folds are overturned to the north. The change in style and increase in amplitude of folds indicate an eastward increase of strain magnitude (Malavieille, 1987). Decimetre-scale sheath folds are observed within the largest strain zones. In the allochthonous sheets, folds are of flexural type and are gently overturned SE. They clearly refold S_1 and L_1. A schistosity (S_2) and a roughly E–W stretching lineation (L_2) are developed in fold hinges (e.g. at Rosebud Canyon, easternmost part of Grouse Creek Mountain).

Deformation regime

In the area studied, D_1 and D_2 microstructures developed in a non-coaxial deformation regime. However, in the central transitional zone, D_2 is mostly of coaxial type. The shear sense has been determined in the field and on oriented samples in the laboratory. More than 200 samples were collected from the three mountain ranges and studied in thin section. Observations of different criteria, in different rocks and in numerous localities, demonstrate the regional consistency of the determined shear senses.

The observed shear criteria are as follows.

(a) Sigmoidal foliation (Fig. 3f) across shear zones at different scales (Ramsay & Graham 1970).
(b) Rotated porphyroclasts (Passchier & Simpson 1986, Van Den Driessche & Brun 1987).
(c) Asymmetric recrystallized tails (Figs. 4 & b) around feldspars (Simpson & Schmid 1983).
(d) Crystallization of asymmetric pressure shadows around minerals (Fig. 3b) (Etchecopar & Malavieille 1987) or fossils.
(e) Asymmetric C'-planes (Berthé *et al.* 1979) (Figs. 4d and 7).
(f) Asymmetric boudinage (Fig. 8) (Hanmer 1986).
(g) Sigmoidal shape of deformed minerals (i.e. 'mica fish', Fig. 4a & c) (Eisbacher 1970, Etchecopar 1974, Lister & Snoke 1984).
(h) Oblique foliation in dynamically recrystallized quartz aggregates (Fig. 4e) (Brunel 1980, Etchecopar & Vasseur 1987).
(i) Asymmetry of the quartz $\langle c \rangle$ axis preferred orientations (Fig. 9) (Bouchez & Pecher 1976, Etchecopar 1977, Lister & Williams 1979, Behrman & Platt 1982, Bouchez *et al.* 1983, Brunel 1983, 1986, Etchecopar 1984, Etchecopar & Vasseur 1987).

Quartz $\langle c \rangle$ axis fabrics are highly asymmetric with respect to the structural framework defined by the foliation plane and the stretching lineation. The girdle which contains the maxima is oblique to the foliation plane (Fig. 9) and its normal is expected to track the bulk

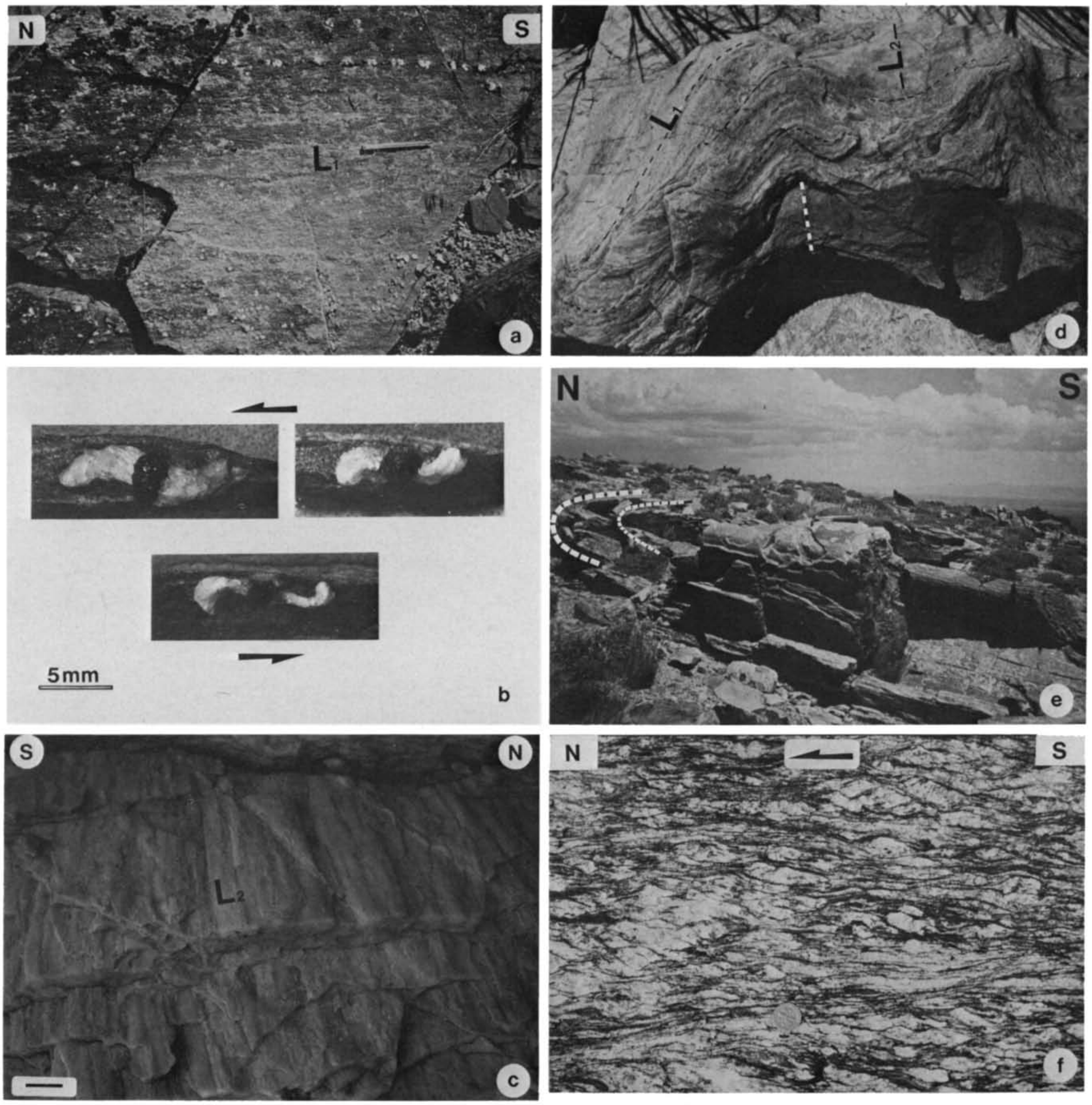

Fig. 3(a). Lineation L_1 outlined by stretched crinoids in metamorphic limestones of Rosebud Canyon. The pen provides scale. (b) Asymmetrical pressure shadows around euhedral pyrites in metamorphic limestones of Rosebud Canyon; northward shearing. (c) L_2 marked by stretched pebbles in conglomeratic Precambrian quartzites of Raft River; scale bar, 3 cm. (d) 'Horseshoe-like' structures due to superimposition of D_2 on D_1 in Precambrian quartzite of Raft River West. The N–S L_1 lineation is curved on the composite foliation planes; the weak E–W L_2 lineation is outlined by muscovites. (e) Curved fold axis resulting from superimposition of D_2 westward shearing on a first phase 'a' type fold (Grouse Creek); N–S fold axis (F_1) is rotated about 100° (N040°E–N140°E). (f) Shear bands in the mylonitic Archean gneiss of Grouse Creek (northward shearing).

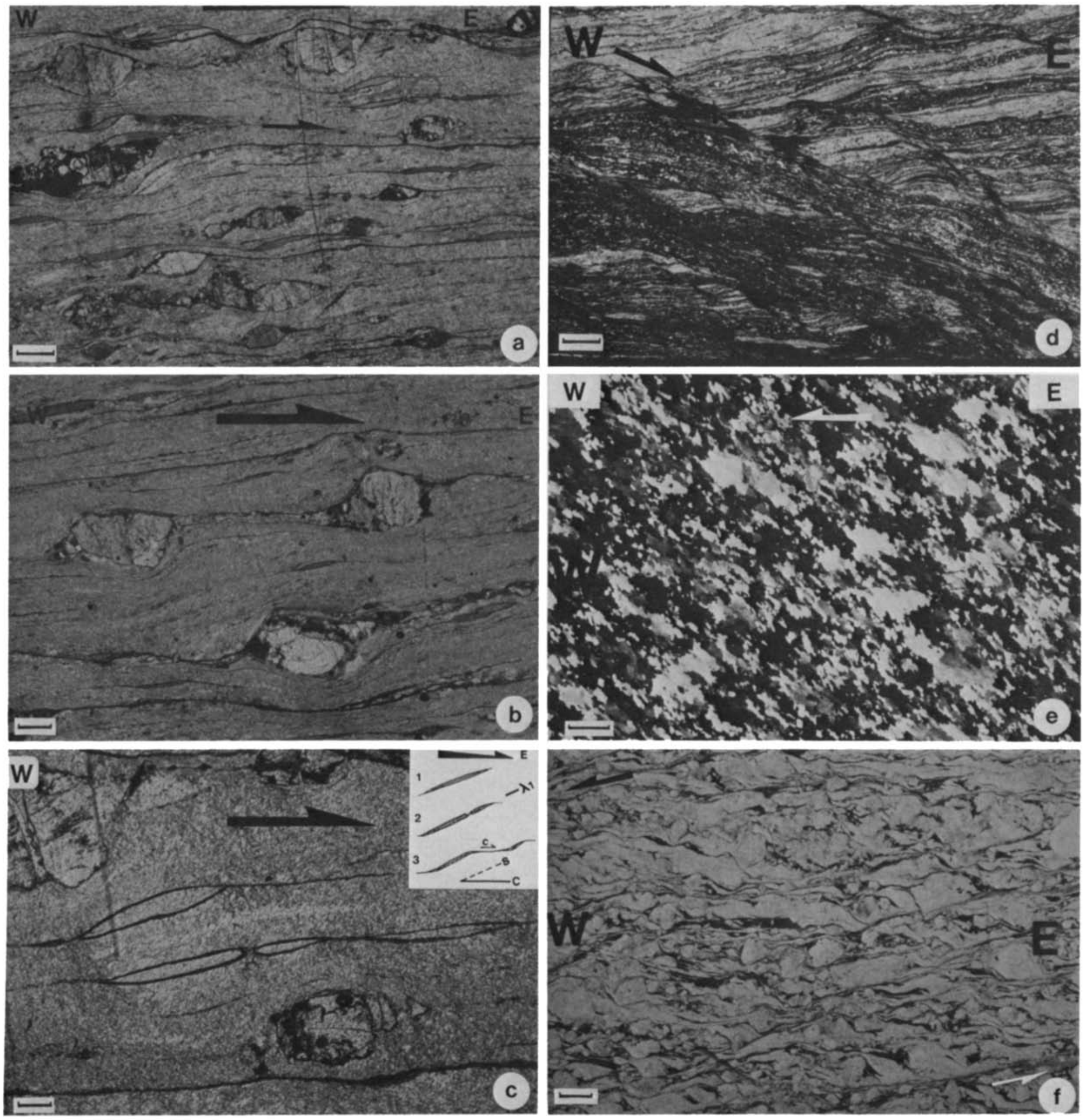

Fig. 4(a). Mylonitic impure quartzite showing asymmetrical mica fish, stretched feldspar porphyroclast with asymmetrical tails (Precambrian quartzite of Raft River East). (b) Detail of asymmetrical tails around broken feldspars; two feldspars in the centre are one broken clast. (c) Asymmetrical mica 'fish' structures within C–S mylonite. C-planes develop on the stretched part of mica. (d) Micro-scale shear bands in phyllitic quartzites (Raft River East). (e) Oblique foliation in dynamically recrystallized quartz aggregates (Grouse Creek). (f) mylonitic granite of Vipont Mountain; shear bands indicate a westward sense of shear. The scale bars are: (a) 500 μm; (b) 300 μm; (c) 160 μm; (d) 1 mm; (e) 260 μm; (f) 2 mm.

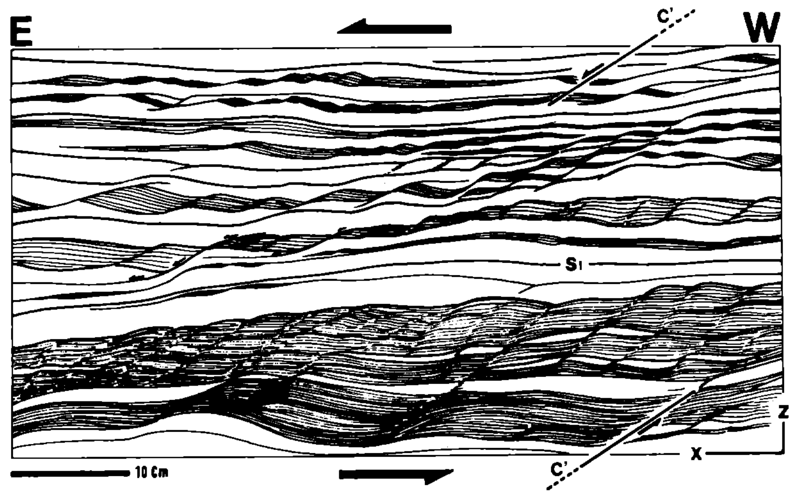

Fig. 7. Shear bands (C') in alternating beds of micaschist and quartzite (Raft River East).

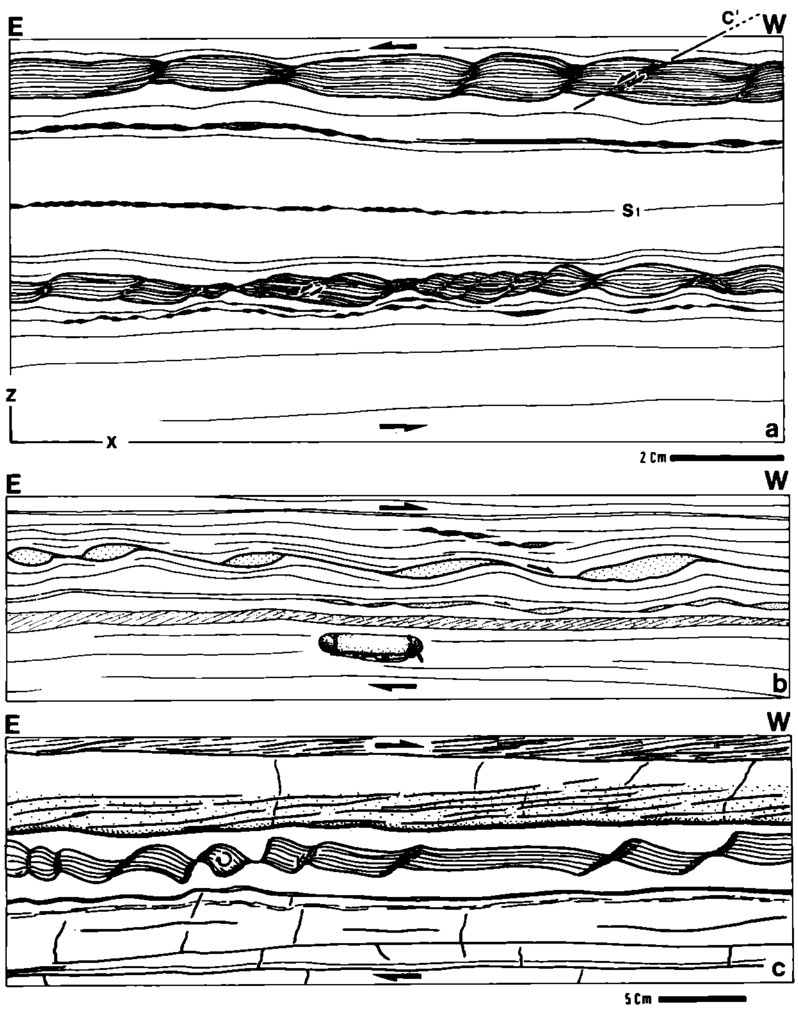

Fig. 8(a). Asymmetric boudinage of phyllitic layers in quartzite (Raft River East); note the back rotation of boudins related to synthetic movements on shear bands. (b) Asymmetric boudinage of a quartzitic layer inside marble (Mount Harrison); note the considerable stretching and the back-tilting of boudins. (c) Asymmetric boudinage of a quartzitic intercalation inside marble (Vipont Mountain); clockwise rotation of boudins indicates westward shear sense.

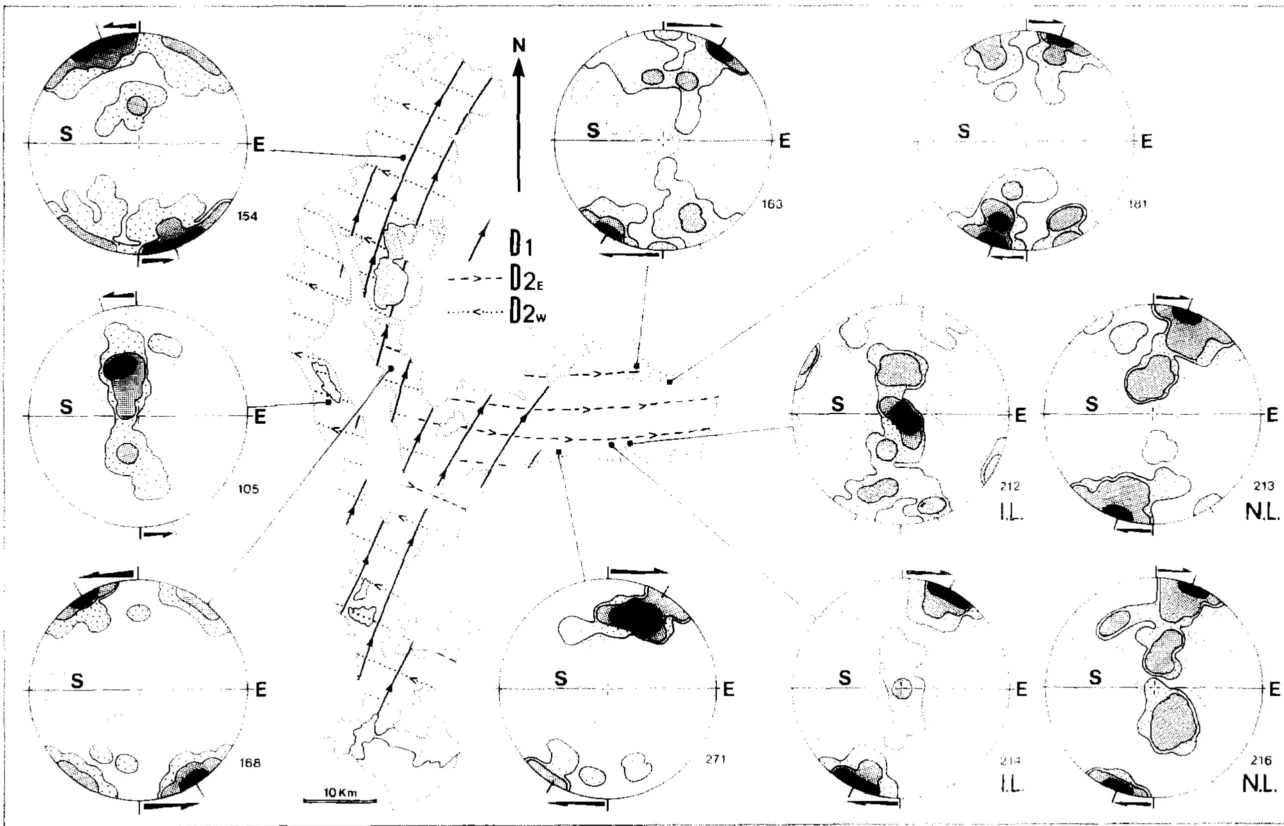

Fig. 9. Sketch map of lineation trajectories and shear senses criteria; D_1, northeastward shearing; D_{2E}, eastward shearing; D_{2W}, westward shearing. Quartz ⟨c⟩ axis fabrics and corresponding sample locations are shown; each diagram, 200 measurements. The foliation plane S ($\lambda_1\lambda_2$ plane) is related to D_2 deformation, L_2 horizontal. Contours (per 1% area): 0.5, 1, 3, 5, max 7% (sample 154); 0.5, 2, 4, 7, max 9% (sample 163); 0.5, 2, 4, 6, max 8% (sample 181); 0.5, 2, 6, 9, max 13% (sample 105); 0.5, 1, 3, 6, max 8% (sample 212 inverted limb); 0.5, 2, 3, 9, max 15% (sample 213 normal limb); 0.5, 2, 5, 9, max 15% (sample 168); 0.5, 2, 4, 6, max 8% (sample 271); 0.5, 2, 5, 8, max 16% (sample 214 inverted limb); 0.5, 2, 4, 10, max 12% (sample 216 normal limb).

shearing direction of the non-coaxial deformation (Etchecopar 1977, 1984). In all the samples studied, the fabric asymmetry is consistent with the overall sense of regional shear of D_2 deformation. Nevertheless, samples belonging to the transitional zone (samples 168 and 154, Fig. 9) show rather symmetric patterns of ⟨c⟩ axes girdles which would suggest a rather coaxial deformation history.

A sketch map of stretching lineation trajectories and associated shear senses is shown in Fig. 9. D_1 deformation is characterized by north to northeastward-vergent shear in the three ranges. D_2 deformation shows east-vergent shear in Raft River (Sabisky 1985, Malavieille & Cobb 1986) and west vergent shear in Albion–Grouse Creek (Malavieille & Cobb 1986). In the central transitional zone between Raft River and Grouse Creek–Albion (e.g. around Clarks Basin or Upper Narrows) no clear rotational component related to D_2 deformation has been observed, suggesting a coaxial deformation history.

Finite strain

Foliation and stretching lineation are parallel to the λ_1–λ_2 plane and the λ_1 direction of the strain ellipsoid, respectively ($\lambda_1 > \lambda_2 > \lambda_3$, Flinn 1965). The rock shape fabric is mostly of the L–S tectonite type (Turner & Weiss 1963) for both D_1 and D_2 phases. Strain measurements on quartz grains and pebbles of the Precambrian quartzites in Albion (Miller & Christie 1981, Miller 1983a), and in Raft River (Compton 1980), suggest plane strain. Strain magnitudes show vertical and lateral variations. In the basement, the strain decreases downwards to undeformed rocks. It roughly decreases upwards within the allochthonous sheets where highest strains are concentrated along particular tectonic or lithologic interfaces. D_1 is well developed in the Albion and Grouse Creek Mountains but strongly decreases in magnitude in the Raft River Mountain and disappears completely eastwards. D_2 occurs in all three ranges, but it appears to be very slight in the bordering zone between Raft River and Grouse Creek, where rock fabrics suggest a coaxial deformation. When moving eastward in Raft River, the strain and rotational features show an increasing progressive east-vergent shearing (Malavieille 1987). An equivalent but westward increase in west-vergent shear magnitude characterizes D_2 in Albion–Grouse Creek Mountains.

KINEMATIC INTERPRETATION OF DEFORMATION AND FOLDING

For intense and non-coaxial progressive deformation, stretching lineations track the shearing direction (e.g. Nicolas *et al.* 1972, Mattauer *et al.* 1981, Escher & Watterson 1984, Malavieille *et al.* 1984). Results presented above show that rocks of the allochthonous sheets were thrust N–NE relative to the basement rocks during D_1. After this early event, during D_2, they moved eastward in Raft River, and westward in Grouse Creek and Albion, away from a central transitional zone.

In the different areas previously studied, kinematic reconstitutions of tectonic phases were mainly deduced from fold overturning (vergence). Important scatters of fold axis directions, as well as superposed or curvilinear folds, were supposed to result from numerous distinct deformation phases. Thus, in this area, Armstrong (1970), Compton (1980), Miller (1980), Todd (1980) and Jordan (1983) dervied tectonic vergences completely different from those suggested by the present analysis of shear criteria. Recent studies on fold processes in ductile shear zones (Escher & Watterson 1974, Hobbs *et al.* 1976, Carreras *et al.* 1977, Rhodes & Gayer 1977, Bell 1978, Cobbold & Quinquis 1980, Hugon 1982), emphasize that progressive shearing yields variable fold geometries. Folds may be sheath folds (Quinquis *et al.* 1978) or straight 'a' type folds (Mattauer 1975) oblique or parallel to the regional stretching lineation (Escher & Watterson 1974, Nicolas & Boudier 1975, Hobbs *et al.* 1976). Both 'a' type and sheath folds are often closely associated, and geological examples have been documented in other areas of intense shearing (see for instance, Carreras *et al.* 1977, Quinquis *et al.* 1978, Minnighi 1980, Faure & Malavieille 1981, Henderson 1981, Lacassin & Mattauer 1985).

My approach, based on shear criteria rather than fold geometry, suggests that folding should be reinterpreted in the Raft River–Albion–Grouse Creek area. I suggest that straight folds subparallel to the stretching lineation formed oblique to the shear direction and rotated progressively with increasing shear strain (Bryant & Reed 1969, Sanderson 1973, Escher & Watterson 1974, Rhodes & Gayer 1977, Bell 1978, Quinquis *et al.* 1978). Such folds may be induced by shearing of layers oblique to the shear plane (Berthé & Brun 1980, Hugon 1982). Sheath folds or curvilinear folds can result from kinematic amplification of primary deflections in the folded layer (Cobbold & Quinquis 1980). Thus, fold geometries observed in the study area are consistent with a progressive shearing model. The set of large-scale D_2 folds of Raft River has been studied in detail and interpreted as 'a' type folds formed subparallel to the E–W shearing direction (Malavieille 1987). Curvilinear folds observed in zones of D_1–D_2 interference, in Grouse Creek, may result from kinematic curvature of previously N–S-trending 'a' type D_1 folds. In the same way, some of the D_2 sheath folds observed in Albion may be due to strong amplification of early folds or S_1 deflections. The systematic westward overturning of F_2 fold observed in the less deformed zones of Grouse Creek–Albion confirms the westward vergence of D_2 deformation.

AGES OF DEFORMATIONS

Misch and co-workers (Misch 1960, Misch & Hazzard 1962, Dover 1969, Nelson 1969, Thorman 1970) showed that Mesozoic metamorphism and deformation have affected rocks of the Basin and Range province in Nevada, Utah and Idaho. Combining K–Ar geochronometry and geologic studies in the Eastern Great Basin, Armstrong pointed out that uplift and cooling of the metamorphic complexes were of Cenozoic age (Armstrong 1964, 1972, Armstrong & Hansen 1966). He further suggested that many of the low-angle younger-on-older faults, were probably Cenozoic extensional structures. The chronology of metamorphism and deformation in the Albion, Raft River and Grouse Creek complexes has been studied in detail (Armstrong & Hills 1967, Armstrong 1968, 1970, 1976, Compton *et al.* 1977, Compton & Todd 1979, Compton 1980, Miller 1980, 1982, Todd 1980). The first regional metamorphism reached the amphibolite facies in Jurassic time, more than 160 Ma ago. This was followed by a Cretaceous period of decreasing pressure and temperature. Granitic rocks were emplaced during early to mid-Cenozoic time. In the vicinity of the granitic injection complex, temperature increased up to sillimanite grade while the area was subject to regional extension. The late kinematic (with respect to D_2), Almo Pluton (Albion Mountain), Red Butte stocks and Immigrant Pass intrusion (Grouse Creek Mountains) (Fig. 1), yield Rb–Sr ages of 28, 25 and 38 Ma, respectively (Compton *et al.* 1977). They clearly postdate the N–S structures of the first deformation and most of the structures of the later E–W ductile deformation. However, the Red Butte pluton is locally deformed close to the contact with the lower allochthonous sheet (Todd 1980). In addition, the granitic rocks of the Middle Mountain injection complex and Vipont intrusion show a strong mylonitic foliation and an E–W-trending lineation associated with a westward sense of shear. Oligocene Rb–Sr ages have been obtained from the Middle Mountain injection complex (south of Oakley, Armstrong 1976), but the Vipont intrusion has not been successfully dated by whole rock Rb–Sr methods. It seems likely that the important heating produced by Cenozoic plutonism has had a strong influence upon D_2 deformation genesis. Thus structural and geochronological relations indicate that most of the E–W mylonitic fabric related to westward shearing is middle Cenozoic in age (i.e. pre- to syn-plutonism). $^{39}Ar/^{40}Ar$ ages from syntectonic muscovites give 19 ± 0.47 Ma (Fig. 10a) in Raft River (Clear Creek) and 20.67 ± 0.6 Ma in Grouse Creek (Dove Creek Pass) (Fig. 10b) (Maluski, unpublished data) suggesting that D_2 metamorphism ended at about lower Miocene. In my opinion, the NE-shearing (D_1) may be related to Jurassic compressional deformation. The subsequent D_2 progressive deformation, eastward shearing in Raft River

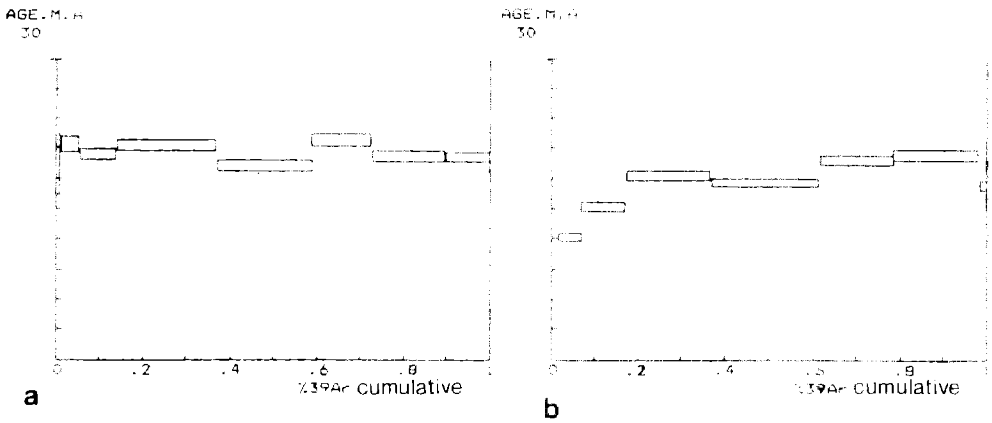

Fig. 10. Age diagrams of muscovites from (a) Grouse Creek (Dove Creek Pass), (b) Raft River (Clear Creek).

and westward shearing in Grouse Creek and Albion, probably began in early Cenozoic times and may have been active until lower Miocene ($^{39}Ar/^{40}Ar$ cooling ages). Brittle low-angle 'décollements' and gravity sliding developed later.

DISCUSSION AND MODEL

Several different hypotheses have been proposed recently to explain the ductile deformation in these metamorphic core complexes. Compton (1980) put forward the idea of 'gravity-driven deformation in a broad heated dome'. In his study of the Raft River Mountains, he considered that deformation was mainly pure shear rather than simple shear, and that the stretching lineation did not indicate the direction of displacement of the allochthonous sheets. Todd (1980) proposed a similar interpretation for the Grouse Creek Mountains, suggesting that ductile deformation occurred in response to gravitational instability created by uplift and controlled by a 'hot spot' within a broad region of high-grade metamorphism. For the Albion Mountains, Miller (1980) discussed in detail three different models in the general framework of the Cordilleran fold and thrust belt: gravity gliding (Mudge 1970, Hose & Danes 1973), gravity spreading (Price & Mountjoy 1970, Price 1971, 1973) and hinterland overthrusting (Armstrong 1978). This author clearly prefers the third model which also provides a straightforward cause for tectonism in the foreland thrust belt. Following the theoretical model of Cady (1980), Armstrong (1982) relates the core complexes to compressive crustal thickening during a Mesozoic orogeny, followed by deep erosion and Cenozoic ductile to brittle deformation due to isostatic readjustment. More recently, different models have focused on Tertiary extensional tectonics and attempted to relate ductile deformation in the core complexes to a unique episode of stretching of the continental crust, involving a lithospheric-scale normal shear zone (see Wernicke 1981, 1985, 'Raft River Stage' fig. 3, Davis *et al.* 1986); or mid-crustal brittle–ductile transition during Tertiary extension, as proposed by Miller *et al.* (1983) in the Snake Range (Nevada).

The proposed palinspastic reconstitution from the Jurassic (Fig. 11a) to the Quaternary (Fig. 11e) attempts to reconcile the geological structures, kinematics and ages observed in the study area. It involves the following deformation history.

The Mesozoic Cordillera in Nevada–Utah–Wyoming is generally accepted to be related to an E–W-directed shortening and eastwards thrusting. The N–S to N–E directions of transport observed all along the narrow zone trending N–S from Grouse Creek to Albion Mountains are difficult to integrate in this general frame work. In this area I propose that N–S transport direction to be the result of a combination between dextral wrenching and thrusting movements along a main cordilleran fault zone (Fig. 11b). Most of the allochthonous older-over-younger thrust slices (such as the Mount Harrison overturned sequence) were emplaced during this event.

East–west shortening continued during the Cretaceous resulting in the Foreland overthrust belt (Fig. 11c). A shortening of more than 100 km is observed in supracrustal rocks east of Raft River Mountain (Royse *et al.* 1975). This suggests that an equivalent shortening of the whole basement must have occurred. According to thin-skinned tectonic models, the pre-orogenic layered structure of the crust (made of thick sedimentary sequences overlying the Precambrian basement) accounts for different locations of shortening deformation within basement and upper crustal rocks. If so, Cordilleran foreland thrust belts would be a superficial expression of basement shortening further west, beneath the future metamorphic core complexes (Royse *et al.* 1975, Price *et al.* 1981). Armstrong (1982) previously outlined this peculiar geometry and, following Kehle (1970), Price (1973), Elliott (1976) and Chapple (1978), suggested that the area of thickened crust provided sufficient slope and push to drive supracrustal rocks towards the fold and thrust belt. In my opinion, overthrusting in the foreland thrust belt cannot be directly related to a gravity-driven phenomenon due to thickening in the west; rather, it may be the direct consequence of underthrusting of basement slices below the present site of metamorphic core complexes during the compressional orogeny. Thus, the thickness of the lower crust may have

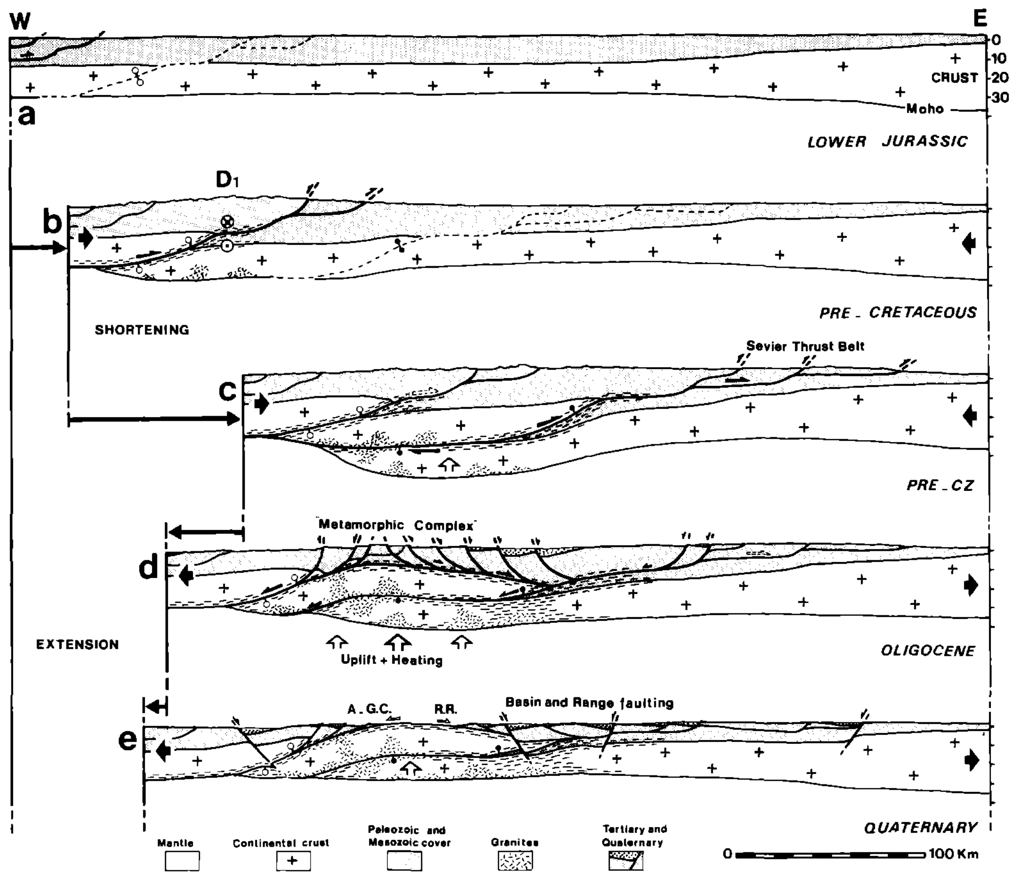

Fig. 11. Interpretative crustal-scale cross-sections from the Albion Raft River and Grouse Creek Mountains to Sevier Thrust Belt and palinspastic restorations since early Mesozoic to Quaternary (see further details in the text). (a) Initial stage, (b) 'dextral wrench-thrust' Cordilleran fault zone associated with D_1, (c) thickening of the crust by underthrusting of crust slices under the future domain of metamorphic core complex, (d) 'metamorphic core complex stage', growing of the ductile 'décollements' and (e) 'Basin and Range' faulting and brittle low-angle faulting.

been doubled by such a mechanism (Fig. 11c). COCORP seismic reflection profiles across the Sevier Desert in west-central Utah (Allmendinger *et al.* 1983) confirm such a geometry. Indeed, W-dipping reflectors rooted in the lower crust beneath the Confusion Range may represent the trace of previous underthrusting of basement slices below the Snake Range metamorphic complex. In this case, too, the foreland thrust belt shortening (about 100 km; Allmendinger *et al.* 1986) could be directly related to this deep underthrusting; as a consequence, an important crustal thickening must have occurred below the Snake Range, suggesting a similar tectonic evolution for this range.

Tertiary extension began after doubling of the lower crust under the metamorphic core complex (Fig. 11d). This was probably favoured by changes in plate boundaries (Coney 1980, Sbar 1982, Engebretson *et al.* 1984). The former overthickening of the crust (Coney & Harms 1984) caused important uplift, abnormal heating (Glazner & Bartley 1985, Gaudemer 1986), local melting of rocks in the basal crust (Farmer & De Paolo 1984), and subsequent rapid rising of S-type granites (Chappel & White 1974, Armstrong 1983) through the crust. Consequent relief and gravity induced instabilities and an important decoupling zone formed at the Precambrian basement–Paleozoic series lithological interface. This interface lies at about 10–15 km depth (Armstrong 1968), which is in the field of ductile deformation. The ductile shear zones simply transformed the shallow brittle extension of the upper-plate rocks towards the zones of deep crustal extension. D_2 deformation developed during east–west extension and emplacement of Tertiary granites. Opposite senses of shearing (eastward in Raft River, westward in Grouse Creek and Albion Mountains) occurred on each side of the arch structure with respect to the transitional zone of coaxial deformation. The domains of important extension in the middle and lower crust located east and west of the metamorphic complexes correspond, in this model, to the root zones of former Mesozoic thrusts, and may have been reactivated as deep low-angle normal faults (Fig. 11d) during the Tertiary. Indeed, the mylonitic deformation disappears downward in the basement rocks of the core complex domain, suggesting that these crustal terranes have remained intact since Precambrian times.

After Oligocene times, uplift, doming and important erosion continued until the original crustal thickness was reconstituted. Brittle deformation occurred along the deep formerly ductile interface 'basement upper-plate' and 'cataclastic décollements' were superimposed on former ductile shear zones. Large displacements of upper-plate rocks, and gravity sliding in the adjacent basins occurred. This extensional model shows a complete progressive evolution from ductile deformation of rocks in large-scale shear zones to brittle deformation on low-angle normal faults and gravity sliding on shallow-dipping planes. Lastly (Fig. 11e) high-angle normal faults crosscut the basement and the present Basin and Range structure developed (Gilbert & Reynolds 1973, Stewart 1978, Zoback *et al.* 1981).

CONCLUSIONS

The results of this study help resolve several important questions about the origin and kinematic history of ductile deformation in metamorphic core complexes. The observations show that the most prominent structures of metamorphic core complexes are the stretching lineations. These lineations are good indicators of the shear direction. Analyses of various shear criteria reveal the kinematics of compressional as well as extensional deformation and tectonic settings. Similar structures and microstructures may form in both compressional and extensional low-angle shear zones. The only difference seems to be the important development of boudinage and intense stretching within extensional shear zones, as opposed to large-scale folding and shortening in compressional zones. Nevertheless, in extensional shear zones, bedding-parallel shearing can be accompanied by folding, as suggested by the occurrence of 'a' type folds and sheath folds.

The overall structure of metamorphic core complexes results from superimposed compressional and extensional tectonics. The Mesozoic thickening of the crust in the Cordillera hinterland has probably controlled the location of the core complexes. Most of the mylonitic fabrics, décollement zones and complex structures in the upper-plate rocks are the results of the reequilibration of the thickened crust during Tertiary extensional tectonics. The ductile extensional deformation may have been favoured by the input of heat by intrusion of the granitic rocks at depth. Some of the early extensional features probably developed while thrusting in the foreland belt was still active. The late ductile deformations ceased at about Lower Miocene time, as suggested by superimposition of late deformational features on plutons of middle Tertiary age, and geochronological data from D_2 syntectonic minerals. Brittle displacements along low-angle décollements and superficial attenuation started later and could be related to gravity sliding of upper-plate rocks into adjacent basins.

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