The Cordilleran foreland thrust and fold belt in the southern Canadian Rocky Mountains

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SUMMARY: The thick (~40 km) slab of Hudsonian (>1750 Ma) continental crust that extends under western Canada from the Canadian Shield can be followed westward, on the basis of its distinctive magnetic anomalies, its influence on the Bouguer gravity values, the results of deep seismic refraction experiments, and the results of geomagnetic depth sounding of the deep electrical conductivity structure, to the Kootenay Arc. The Kootenay Arc is basically a W-facing monocline of crustal dimensions, across which the change in structural level involves an aggregate stratigraphic thickness of up to 20 km. It marks the western edge of the continental craton over which the displaced supracrustal rocks have been draped.

Balanced structure sections of the thrust and fold belt, which take into consideration the deep crustal structure, as constrained by the geophysical data, show that: (i) in early Campanian time the continental crust that now lies beneath the western Rocky Mountains and the Purcell anticlinorium was covered with the platformal Palaeozoic to Upper Jurassic rocks and the exogeoclinal Mesozoic rocks that now form the northeasterly verging imbricate thrust slices of the eastern Rocky Mountains; (ii) the Cordilleran miogeocline developed outboard from the edge of the continental craton, on tectonically attenuated continental crust, or oceanic crust; and (iii) tectonic shortening of about 200 km in the supracrustal rocks in the Rocky Mountains must be balanced at a deeper level, W of the Kootenay Arc, by the shortening of the oceanic or attenuated continental crust.

The net convergence between the Cordilleran magmatic arc and the autochthonous cover on the continental craton is a type of intra-plate subduction that was antithetic to the SW verging subduction zone marking the boundary of the North American Plate. The basement of the back-arc or marginal basin, in which the miogeocline formed, was consumed; but the adjoining continental margin was not. The foreland thrust and fold belt is a shallow subduction complex that was tectonically prograded over the margin of the continental craton, as the supracrustal cover scraped off the down-going slab was piled up against the overriding mass, and spread laterally eastward under its own weight.

The foreland thrust and fold belt of the North American Cordillera is a zone of easterly verging shallow thrust faulting and décollement folding, up to about 300 km wide, that follows the boundary between the Cordilleran miogeocline and the North American craton from the Yukon Territory of northern Canada to southeastern California (King 1969; Wheeler *et al.* 1974; Burchfiel & Davis 1972), from whence it may extend across southern Arizona and New Mexico and southwestern Texas into the Sierra Madre Oriental of Mexico (Corbitt & Woodward 1973), and finally into Guatemala and Honduras (de Cserna 1971). Within this zone of 'thin-skinned' deformation, an easterly tapering wedge of supracrustal rocks, comprising parts of the miogeocline, the cover of the cratonic platform, and the overlying exogeoclinal wedge of synorogenic clastic wedge deposits, was horizontally compressed and tectonically thickened, as it was displaced eastward relative to the underlying undeformed craton. The distinctive attributes of structural style and tectonic setting of this belt are characteristic of many other foreland thrust and fold belts, including: the Helvetic and Jura Mountains of the northern Alps (Bernoulli *et al.* 1974), the Valley and Ridge Province in the western Appalachian Mountains of the United States (Rodgers 1970), the Labrador 'trough' along the boundary between the Churchill Province (Proterozoic–Aphebian) and Superior Provinces (Archean) in the eastern Canadian Shield (Dimroth 1970), and the Asiak fold and thrust belt along the boundary between the Bear Province (Proterozoic–Aphebian) and the Slave (Archean) Province in the northwestern

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Canadian Shield (Hoffman 1973: Hoffman *et al.* 1978). Foreland thrust and fold belts appear to represent a geotectonic phenomenon that is widely distributed in space and time, but for which there is, as yet, no simple global explanation.

Foreland thrust and fold belts display obvious similarites in structural geometry to the zones of imbricate thrust faulting and associated folding that occur above subduction zones along the inner slope of trenches in which there is a thick sedimentary fill (Hamilton 1977: Seeley 1977); and some foreland thrust and fold belts have been interpreted as the result of a 'collision' between a passive continental margin and island arc or another continental margin, during which part of the supracrustal cover of the underriding passive margin was scraped off against the overriding plate (Dewey & Bird, 1970). However, many foreland thrust and fold belts cannot be formed in this way because they represent relatively small ($d \times 10^2$ km) displacements distributed within a single physically continuous lithospheric plate (intra-plate displacements), rather than relatively large ($d \times 10^3$ km) displacements across a suture along which one discrete plate has overridden another quite different one (inter-plate displacements).

The foreland thrust and fold belt of the North American Cordillera is a good example of an intra-plate thrust and fold belt. Although there are conspicuous changes in the details of tectonic setting, structural style and history of deformation from one segment to another of the foreland thrust and fold belt of the North American Cordillera, the segment containing the southern Canadian Rocky Mountains is as typical as any. It is a particularly appropriate place for a more detailed analysis of the nature and tectonic significance of this foreland thrust and fold belt, because its essential features are not obscured by younger sedimentary and volcanic deposits, or superimposed deformation of the Basin and Range type, and because it has been relatively thoroughly studied by deep drilling and seismic reflection exploration for hydrocarbon traps (Fox 1959: Dahlstrom 1970; Shaw 1963; Keating 1966; Bally *et al.* 1966) as well as by systematic mapping of surface exposures. The main objective of this paper is to elucidate the geological structure, tectonic evolution, and geotectonic significance of the segment of the Rocky Mountains thrust and fold belt between latitude 48°N and latitude 52°N (Fig. 1), and of the Cordilleran foreland thrust and fold belt in general.

Structure of the southern Canadian Rocky Mountains

The structure of the southern Canadian Rocky Mountains is dominated by thrust faults, almost all of which are easterly verging. Flexural-slip folds have developed in conjunction with the displacements along the thrust faults; and many of the thrust faults die out upward in the cores of anticlines or downward in the cores of synclines, where they mark the centres of curvature for strata that are concentrically folded (Price 1964*b*, 1965). The mechanical anisotropy due to the stratigraphic layering has exerted a profound influence on the style of the deformation. The thrust faults commonly follow bedding glide zones that are linked by ramps along which the faults step across the stratigraphic layering; and the thrust

FIG. 1. Geological map of the foreland thrust and fold belt of the North American Cordillera between 49° and 52°N latitude. The lines labelled W–E and SW–NE mark the locations of the structure sections presented in Fig. 2. Symbols identifying the more important faults are as follows: Bo-Bourgeau; BR-Bull River; Ca–Cnatter Creek: Co–Columbia River: Fl–Flathead: Ha–Hall Lake: Ho–Hope: Jo–Johnston Creek: Le–Lewis: Li–Livingstone: Mc–McConnell: Moy–Moyie: Mo–Mons: Ne–Newport: Pi–Pipestone Pass: Pu–Purcell: RMT-Rocky Mountain Trench: Ru–Rundle: Sa–Standfast Creek: Si–Simpson Pass: Sm–Sulphur Mountain: StM–St. Mary.

Symbols identifying batholiths are as follows: Ba–Battle Mountain: Bu–Bugaboo: By–Bayonne: F–Fry Creek: H–Horsethief Creek: Ka–Kaniksu: Ku–Kuskanax: N–Nelson: W–White Creek. The Valhalla gneiss complex is identified by the symbol V. The information which is compiled and interpreted in this map is taken from maps by: Balkwill, *et al.* (in press *a, b*): Benvenuto (1978): Bielenstein, *et al.*, (1971): Campbell & Loofbourow (1957): Douglas (1951, 1952, 1958): Fyles (1964, 1967): Fyles & Hewlett (1959): Gilman (1972): Glover (1978): Harrison & Jobin (1963, 1965): Harrison & Schmidt (1971): Henderson (1954): Höy (1977, 1978): Huntting *et al.* (1961): Johns (1970): Leech (1959, 1960): Little (1960): Miller (1974 a–d): Miller *et al.* (1961): Johns (1970): Ollerenshaw (1975, 1978): Park & Cannon (1943): Price (1962): Price & Mountjoy (1978 *a–d*, 1979 *a–b*, in press): Price *et al.* (in press): Price *et al.* (in press *a, b*): Balkwill *et al.* (in press *a* & *b*): Read & Wheeler (1976): Reesor (1973): Ross (1959): Sears (1979): Simony & Wind (1970): Thompson (1972): Wheeler (1963): Williams (1949): Yates (1964, 1971): Zwanzig (1973).

sheets have been folded (and unfolded) as they moved along the stepped fault surfaces (Douglas 1950). Flexural-slip folding and thrust faulting have proceeded concurrently as two different manifestations of the same basic process of horizontal compression and relative northeasterly upward translation of the supracrustal rocks. Many of the thrust faults have been folded together with the beds they cut, because of displacements on other curved faults which developed later, at lower structural levels (Douglas 1950, 1952, 1958; Dahlstrom 1970). Displacements on the thrust faults range up to a maximum of at least 50 km in the case of the Lewis Thrust (Price 1965; Dahlstrom *et al.* 1962); but nevertheless all the faults die out along the strike, and the rock mass is physically continuous around the ends of each fault. On a regional scale, the faults comprise a penetrative array of discrete, discontinuous overlapping and interfingering slip surfaces; and at this scale, the deformation can be viewed as a kind of compressive plastic flow (Nye 1952) involving both large distortion (E–W compression and vertical thickening) and large easterly relative translation, without any overall disruption and loss of cohesion (Price 1973).

The tectonically foreshortened easterly tapering wedge of supracrustal rocks is firmly attached in the E to the undeformed supracrustal rocks comprising the platformal and exogeoclinal deposits that cover the western flank of the Canadian Shield. The western part is penetrated by granitic intrusions and intercalated with and overlain by calc-alkaline volcanic rocks comprising a regional calc-alkaline magmatic belt that has been attributed to the easterly subduction of oceanic lithosphere beneath the Cordilleran orogenic belt (Monger *et al.* 1972; Monger & Price 1979). There are some structures in the deformed supracrustal wedge that reflect an easterly horizontal stretching; but, as will be shown below, the magnitude, location, and timing of this horizontal stretching leave no doubt that the development of the foreland fold and thrust belt involved a large net convergence between the magmatic belt and the North American craton. Therefore, the relative easterly translation of

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FIG 2. Structure Sections and palinspastically restored sections through the Cordilleran foreland thrust and fold belt in southern Canada. The locations of section are given in Fig. 1. Symbols identifying equivalent points on the hanging-wall side of the more important faults in each structure section and its palinspastically restored counterpart are listed: Bo–Bourgeau; BR–Bull River; BT–Burnt Timber; Bz–Brazeau; Co–Coleman; HL–Hall Lake; In–Inglismaldie; La–Lac des Arcs; Le–Lewis; Li–Livingstone; Mc–McConnell; Mi–Mill Creek; OB–Old Baldy; RMT–Rocky Mountain Trench; Ru–Rundle; Sm–Sulphur Mountain; Wa–Watson; Wh–Whaleback.

The horizontal sea-level datum for the restored sections is the boundary between marine and non-marine Upper Jurassic rocks. In the overlying synorogenic clastic wedge deposits of the exogeocline the Upper Jurassic and Lower Cretaceous Kootenay Formation is designated–JKk; the Lower Cretaceous Blairmore Group in designated–Kbl. and the Upper Cretaceous marine Alberta Group is designated Kag. The shallow structure shown in the eastern part of Section W–E is adapted, with modifications, from sections by Douglas (1950). Price (1962); that between the Bull River (BR) and Rocky Mountain Trench (RMT) faults is modified after Leech (1958) and Höy (1978); and that in the Kootenay Arc follows Fyles (1967) and Höy (1977). The eastern part of Section SW–NE, above −2400 m, is adapted with modifications from sections at a scale of 1:50 000 by Price & Mountjoy (1970*a*. 1970*b*. 1973*a*. 1973*b*) and Ollerenshaw (1972*a*, 1972*b*) and follows the basic pattern of a section by Bally *et al.* (1966, plate 1B and plate 3) that is based on seismic reflection data and deep drilling. The shallow structure of the Purcell anticlinorium is adapted with modifications from a section at 1:250 000 by Reesor (1973); and that in the Kootenay Arc follows a section by Fyles (1964).

Locations of wells drilled for oil and gas, from which logs were used in the preparation of the structure sections are shown by heavy lines and identified by number as follows: 1–Spring Point 2-4-10-29W4 (projected 460 m S); 2–Calstan C&E Cow Ck. 76-30-8-1W5 (projected 8400 m S); 3–Marjon Lundbeck No 1. (projected 4100 m S) (log by Douglas 1950); 4–Texaco Wilmont Todd Ck A-1 3-6-10-2W5 (Projected 4100 m S); 5–Union Quaich 10-3 10-3-9-3W5 (Projected 4500 m N); 6–Imperial Quaich 3-3-10-3W5 (Projected 4100 m); 7–Triad Union Quaich 10-21-9-3W5 (Projected 100 m S); 8–Texaco Livingstone East 3-20-9-3W5 (Projected 500 m N); 9–Coseka *et al.* Coleman 4-23-9-4W5 (Projected 500 mN); 10–Gulf PCP Coleman 7-33-10-4W5 (projected 460 m S); 11–California Standard-Crowsnest 6-14-8-5W5 (Projected 11500 mN); 12–Sinclair *et al.* Racehouse 15-29-9-5W5 (Projected 2100 m S); 13–Consumer Shell Cremanc 10-21-29-5W5 (projected 1200 m SW); 14–Dome Winchell 10-18-24-5W5 (Projected 1500 m S; 15–Western Imperial wildcat 6-16-18-6W5 (Projected 2400 m SW); 16–Imperial Fina Pacifac Benjamin 3-14-38-W75 (Projected 800 m SW).

The logs of the wells are given in the Schedule of Wells Drilled for Oil and Gas, Province of Alberta, published annually by the Energy Resources Conservation Board, Calgary, Canada.

the magmatic belt, and the attendant horizontal compression of the supracrustal wedge, must be balanced by an equivalent large horizontal displacement and/or compression of a basement of continental or oceanic lithosphere at a deeper crustal level. This raises a number of important questions concerning the foreland thrust and fold belt:

1. What is the minimum amount of net horizontal shortening across it?
2. How is this shortening accommodated at a deep crustal level?
3. What is the significance of the foreland fold and thrust belt in terms of the geotectonic evolution of the whole of the Cordilleran Orogen?

In attempting to answer these questions it is necessary, first, to consider the nature of the northeasterly tapering wedge of supracrustal rocks, in order to provide a framework for interpreting the structures that have developed within it.

Tectonostratigraphic framework

The northeasterly tapering wedge of supracrustal rocks, within which the structures of the foreland fold and thrust belt have developed, consists of a series of distinctive tectonostratigraphic assemblages, each comprising a suite of rock units, the overall characteristics of which imply deposition in a particular tectonic setting. These tectonostratigraphic assemblages record the existence of contrasting eugeoclinal, miogeoclinal and platformal tectonic domains, and of distinct preorogenic and synorogenic stages of tectonic evolution (Fig. 1). The preorogenic stage spans the interval from late Proterozic to Middle Jurassic, and involves a western eugeoclinal domain characterized by the widespread occurrence of basic to intermediate and acidic volcanic rocks and of immature, wacke-type clastic rocks: and eastern platformal and miogeoclinal domains characterized by mature, shallow-water clastic and carbonate rocks that become thicker and more shaly westward, from the platform to the miogeocline. The synorogenic stage spans the interval from late Jurassic to Palaeogene and is characterized by clastic deposits that are of western or else local provenance. These accumulated mainly in a fore-deep or exogeocline that migrated northeastward from late Jurassic to Palaeocene time (Bally *et al.* 1966), as the continental lithosphere subsided isostatically in reponse to the load imposed on it by tectonic thickening and relative northeasterly displacement of the overlying supracrustal rocks (Price 1973).

The Belt-Purcell Supergroup, the oldest assemblage in the supracrustal wedge, has a maximum exposed thickness of about 11 km in the Purcell anticlinorium in southern Canada (Reesor 1973), and about 20–25 km in NW Montana (Harrison 1972). The lower part, which is up to at least 6 km thick in southern Canada, consists mainly of fine-grained quartz-wacke turbidites and dark pelites that grade upward through light grey, green, purple and red shallow-water argillites and sandstones, into the carbonate rocks of the middle part. The great volume of relatively homogeneous fine-grained sediment comprising the very thick lower part of the Belt-Purcell assemblage appears to have been supplied by a very large river system of the size and character of the modern Mississippi, and to have been prograded outboard from the margin of the continent (Price 1964*a*). From the thickness and volume of the Belt-Purcell sediments, and the fact that they lie athwart the structural grain of the buried crystalline basement complex, one may infer that the basin in which they accumulated formed during the initial stages in the rifting of a former Precambrian continental mass which was eventually severed to produce the Pacific margin of the North American continent (Sears & Price 1978). The middle and upper parts of the Belt-Purcell assemblage consist of shallow-water carbonate and terrigenous clastic, tidal-flat and flood-plain deposits that accumulated over the prograded delta plain. Eastward thinning and convergence of all the units in the Belt-Purcell assemblage is accompanied by the disappearance of the turbidite facies, and by the appearance of coarse feldspathic detritus that apparently came from the nearby margin of the basin (Price 1964*a*).

The Windermere Supergroup, of late Proterozoic age, provides a sharp contrast with the Belt-Purcell assemblage, which it overlies unconformably. Coarse-grained feldspathic wacke sandstone and quartz-feldspar pebble conglomerates are interbedded with green and grey pelites, and locally with carbonate rocks. Conglomeratic mudstones, some of which have been interpreted as glacial deposits (Aalto 1971), occur on the unconformity at the base of the assemblage, and locally also at various other levels, including the top. The clasts in these conglomerates consist of distinctive rock types that occur in the underlying Belt-Purcell Supergroup. They are of local provenance, and imply high structural relief that exposed Belt-Purcell rocks, whilst Windermere strata accumulated nearby to thicknesses of up to 9 km (Lis & Price 1976). In contrast, the coarse feldspathic wackes had an external provenance and were transported southwestward (Young *et al.* 1973) over the Belt-Purcell rocks to their site of deposition. The Eo-Cambrian quartz sandstones, which form the base of both the overlying Lower Palaeozoic miogeoclinal and platformal assemblage and the eugeoclinal assemblage, unconformably overlap the thickest part of the Windermere Supergroup, and also the Upper and Middle Belt-Purcell strata in the Purcell anticlinorium in southern Canada. Relationships beneath the unconformity show that the St Mary Fault (Fig. 1. 49°30'N. 116°15'N), which is now an important right-hand reverse fault (Rice 1941; Leech 1958. 1962) and is linked to thrust faults in the western Rocky Mountains, follows the locus of an older NE trending structure across which there were up to 13 km of stratigraphic separation with the northwest side down, during the deposition of the Windermere Supergroup (Lis & Price 1976).

The Lower Palaeozoic platformal sequence consists of about 1 km or less of shallow-water carbonate rock and interbedded shale. It thickens westward in the miogeocline, to about 5 km adjacent to a carbonate bank margin that is now located in the centre of the Rocky Mountains (Fig. 1, 49°40'N, 115°10'W to 52°0'N, 117°20'W). At this carbonate bank margin all the Cambrian and Ordovician formations, except one Upper Cambrian limestone unit and an Upper Ordovician and Lower Silurian dolomite unit undergo a relatively abrupt change in facies to light-coloured shale and argillaceous limestone (Aitken 1971). Farther W. there is a profound change across the Purcell anticlinorium between the light-coloured miogeoclinal shale facies and the coeval eugeoclinal facies. The latter consists mainly of dark graphitic pelites, but contains significant intercalations of basic volcanic rocks and feldspathic grits and wacke-type sandstones (Fyles 1964) that must have had either a western or a local provenance. The Lower Palaeozoic platformal and miogeoclinal strata appear to have accumulated as shelf and slope deposits along a continental margin (Monger *et al.* 1972) that was prograding westward into a back-arc or marginal basin. Repeated basic volcanism punctuates the record of deposition in this basin; and it includes intermittent incursions of igneous clastic detritus, that presumably were derived from a magmatic arc situated further west, (Monger & Price 1979). Abrupt thickness and facies changes, which occur in the miogeoclinal rocks of the Purcell and western Rocky Mountains, record the tilting of large blocks, perhaps caused by displacements on deep down-to-the-basin listric normal faults.

Important unconformities separate the Lower Palaeozoic and Upper Palaeozoic assemblages. Mississippian and younger eugeoclinal rocks, consisting of dark pelites with intercalated sandstones, limestones and basic volcanic rocks, overlie the Lower Palaeozoic assemblage with angular unconformity. A basal conglomerate, containing clasts among which there is random orientation of a low-grade metamorphic tectonite foliation, shows that the Lower Palaeozoic rocks in the eugeoclinal domain were deformed and metamorphosed prior to late Mississippian time (Read & Wheeler 1976). In the platformal and miogeoclinal domain, Upper Devonian shallow-water carbonate rocks unconformably overlap Silurian, Ordovician and Upper Cambrian formations toward the E (Price & Mountjoy 1970, Figs. 2, 3) across the flank of a broad epeirogenic structure, the Alberta arch (Douglas *et al.* 1970, Fig. VIII–15). However, in the Purcell Mountains and Rocky Mountains near Cranbrook (Fig. 1, 49°30', 113°30') relationships beneath the unconformity show that the Moyie-Dibble Creek Fault, another right-hand reverse fault like the St Mary Fault, follows the locus of an older northeasterly trending structure across which there was stratigraphic separation of more than 10 km, with the NW side down, during the deposition of the Lower Palaeozoic assemblage (Leech 1958. 1962; Norris & Price 1966; Price 1972).

The Triassic–Middle Jurassic assemblage is represented on the cratonic platform by about 100 m of Jurassic marine shale; but it is more than 1000 m thick in the miogeocline, where it includes about 600 m of Triassic shallow-water marine shale, siltstone, quartz sandstone, carbonate rock and evaporites. The coeval eugeoclinal assemblage, which is only preserved W of the Purcell anticlinorium and the Kootenay arc (Fig. 2) appears to be more than 10 km thick (Little 1960; Read & Wheeler 1976) and consists of dark pelites, volcanogenic sandstones, and mainly andesitic volcanic rocks. Some of the mid to late Jurassic granitic plutons have been deformed along with the rocks they have intruded, and they may be partially coeval with the Jurassic volcanic rocks; whereas the Cretaceous granitic plutons cut across previously developed folds and faults, and have no obvious extrusive counterparts (Gabrielse & Reesor 1974).

The dramatic changes in thickness and lithofacies that occur along, as well as across, the strike of the northeasterly tapering wedge of Proterozoic–Jurassic supracrustal rocks involve important changes in mechanical properties, and have exerted a profound influence on the nature and orientation of the structures that have developed within it. They account for the development of a series of distinct linear tectonic subprovinces in the Rocky Mountains, including the Foothills (platformal Palaeozoic and exogeoclinal Upper Cretaceous), Front Ranges (platformal to miogeoclinal Palaeozoic and exogeoclinal Jurassic–Lower Cretaceous), Eastern Main Ranges (miogeoclinal carbonate facies) Western Main Ranges (miogeoclinal Cambrian shale facies), and Western Ranges (miogeoclinal Cambro-Ordovician shale facies with Ordovician–Silurian carbonate rocks). Each of these displays distinctive characteristics of structural style and physiography, as well as stratigraphy (North & Henderson 1954; Price & Mountjoy 1970). Most change abruptly along the strike at about 49°20′N latitude because of changes in sedimentary facies and thicknesses resulting from the influence of the transverse, northeasterly trending structures that were active during Proterozoic and Palaeozoic time. For example, SE of this zone, where the thick and relatively homogenous Belt-Purcell assemblage above the Lewis thrust fault is overlain by very thin Lower Palaeozoic and Upper Palaeozoic assemblages that are in the platformal facies, broad open folds in relatively flat-lying beds are carved into castellated mountain massifs, and a structural and physiographic style that is characteristic of the eastern Main Ranges further N, extends out to the edge of the Foothills. The steeply-dipping, moderately thick thrust slices of Palaeozoic platformal to miogeoclinal carbonate rocks, that form the characteristic linear mountain ranges of the Front Ranges subprovince further N, are missing completely. The conspicuous changes in structural trend that occur across this transverse zone can be related, as will be shown below, to changes in the configuration and orientation of the margins of the basin in which the various tectonostratigraphic assemblages accumulated.

Balanced structure sections

Reliable estimates for the minimum net tectonic shortening or convergence across the thrust and fold belt can be obtained from balanced structure sections (Dahlstrom 1969*b*), in which the deep crustal structure is also considered, and is constrained by the relevant geophysical data. A balanced section is one in which there is geometric compatibility from one level to another. In a thrust faulted terrain the simplest application of this principle is the requirement that the rock units which occur on one side of a thrust fault must have matching counterparts on the other side. Provided that the deformation accompanying the thrust faulting has been accomplished mainly by flexure and interstratal slip, and that the penetrative strain within the beds is so small in comparison with the size of the folds and thrust faults that it can be ignored, a balanced section is one in which there is consistency of stratigraphic markers from one level to another, and in which the lengths of the stratigraphic markers in the deformed state are the same as they were in the undeformed state, only the shape having changed during the deformation. Under these circumstances one test of a balanced section is that it leads to a logical and reasonable palinspastic restoration (Dahlstrom 1969*b*).

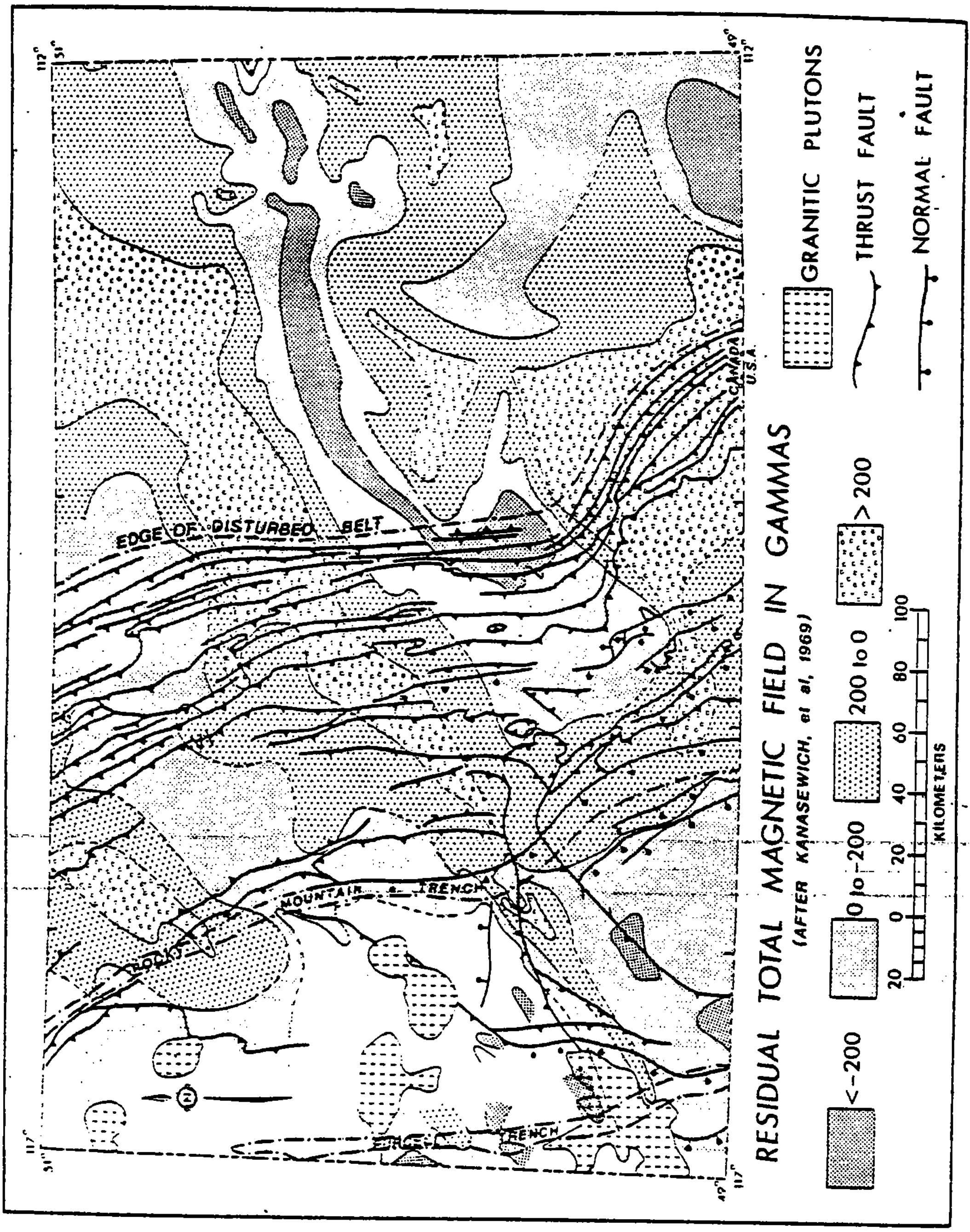
Systematic regional geological mapping and many detailed studies of the stratigraphy and sedimentary petrography of the rocks of the exogeoclinal assemblage and the eastern carbonate facies of the platformal and miogeoclinal assemblages show that stratigraphic units do not change significantly in thickness between the limbs and hinge zones of folds, and also that even the finest details of original sedimentary and organic structures have survived the deformation without significant distortion. This leaves no doubt that intrastratal penetrative strain within these rocks is very low and can be neglected for purposes of constructing balanced sections. Accordingly, the structure sections through the carbonate facies of the platformal and miogeoclinal assemblages (Fig. 2) have been balanced, on the premise that the lengths of the stratigraphic contacts have not been altered significantly in the course of the folding and faulting; and the palinspastic sections have been constructed on the premise that the lengths of the beds in the deformed state are essentially the same as in the undeformed state. In the shale facies of the Lower Palaeozoic miogeoclinal assemblage, widespread development of penetrative cleavage and schistosity, and conspicuous distortion of primary structures (Balkwill 1972; Gardner 1977) preclude the application of this technique except at the highest stratigraphic levels (Middle Ordovician quartz sandstones and

Upper Ordovician and Lower Silurian dolomites), where interstratal penetrative strains are low, and shortening has been accomplished mainly by flexural-slip folding and associated thrust faulting (central parts of Sections NE–SW and E–W, Fig. 2).

In the Proterozoic rocks of the Windermere and Belt-Purcell assemblages. Penetrative intrastratal strain is variable, increasing from relatively low levels in the E to higher levels with progressively higher metamorphic grade toward the W. However, the general structure of the Purcell anticlinorium can be outlined satisfactorily by projecting what occurs at the surface along the northerly and northwesterly plunge into the plane of the sections. The Kootenay Arc, which forms the SW flank of the Purcell anticlinorium, is characterized by high, penetrative strains and complex tectonic overprinting associated with greenschist and amphibolite facies regional metamorphism and coaxial refolding about N and NW trending axes (Fyles 1964, 1967); and the general character of the structures within it can also be outlined on the basis of projections along the plunge (Fyles 1967; Höy 1977). There is a very large change in structural level across the Kootenay Arc, from the lowest part of the Belt-Purcell Supergroup in the core of the Purcell anticlinorium to an extensive tract of exposures of the Triassic–Jurassic eugeoclinal assemblage, which occurs adjacent to the Nelson batholith and the Shuswap Metamorphic Complex. The aggregate thickness of the succession of rock stratigraphic units which occur in superposition between these two horizons is more than 20 km. Lateral variations, involving a westerly decrease in the thickness of older units and an easterly decrease in the thickness of younger units, may account for some of this large change in stratigraphic level, but most of it is an expression of change in structural level. Accordingly, the Kootenay Arc, in spite of the complexity of the detailed structure within it, is basically a simple structure—a westerly facing monocline of crustal dimension. Geophysical investigations of deep crustal structure indicate that this monoclinal structural step in the supracrustal rocks marks the western edge of the thick (~40 km) slab of Hudsonian (>1750 Ma) continental crust that forms the basement of the North American craton in western Canada.

Deep drilling and seismic exploration for hydrocarbons show that the thrust faults flatten with depth within the supracrustal succession, above the Hudsonian basement, at least as far W as the central part of the Rocky Mountains, and probably beyond the Rocky Mountain Trench beneath the eastern flank of the Purcell anticlinorium (Bally *et al.* 1966). The distinctive pattern of broad, high-amplitude, NE trending magnetic anomalies that is characteristic of the Hudsonian basement, which extends under the cratonic platform from the Churchill Province of the Canadian Shield, can be followed across the whole of the Rocky Mountains into the eastern part of the Purcell anticlinorium, and perhaps as far W as the Kootenay arc (Fig. 3). These relationships corroborate the interpretation from the seismic reflection data. The Hudsonian basement does, indeed, extend under the thrust and fold belt and the Purcell anticlinorium without any apparent disruption of its characteristic NE trending tectonic fabric.

The results of seismic refraction experiments, using explosions in lakes in the interior of the Cordillera (Chandra & Cumming 1972; Berry & Forsyth 1975; Mereu *et al.* 1977), mine blasts in the coalfields of the southern Rocky Mountains and the porphyry copper mines of the interior of the Cordillera (Bennett *et al.* 1975; Spence *et al.* 1977; Cumming *et al.* 1979), show that the change in structural level that occurs in the supracrustal rocks across the Kootenay Arc coincides with a westerly decrease in crustal thicknesses, from a maximum thickness of 50–55 km in the vicinity of the western Rocky Mountains and the Purcell anticlinorium to about 30–40 km beneath the interior of the Cordillera (Monger & Price 1979). The monoclinal step at the Kootenay Arc also coincides with a westerly increase in crustal and upper mantle electrical conductivity as outlined by geomagnetic depth sounding experiments (Caner 1971; Dragert & Clarke 1976); and with a westerly change in the Bouguer gravity anomaly from about −200 mgal, where the crust is thickest, to about −140 mgal in the interior of the Cordillera (Fig. 4). It is worthy of note, that the lowest Bouguer gravity anomaly values occur within the widest part of the foreland fold and thrust belt, where the amount of tectonic thickening of the supracrustal wedge would be expected to be greatest. Morever, the amount of crustal thickening in this area, as indicated by the analysis of the Bouguer anomalies (Stacey 1973) and by the results of the seismic refraction experiments that were cited above, is about that which is required to accommodate the tectonic thickening of the supracrustal wedge that was calculated by Price (1973) from a comparison of structure sections and corresponding palinspastically restored sections by Bally *et al.* (1966) and Price &


RESIDUAL TOTAL MAGNETIC FIELD IN GAMMAS
(AFTER KANASEWICH, et al, 1969)
<-200
0 to -200
200 to 0
>200
GRANITIC PLUTONS
THRUST FAULT
NORMAL FAULT
KILOMETERS
20 0 20 40 60 80 100
EDGE OF DISTURBED BELT
ROCKY MOUNTAIN TRENCH
PURCELL TRENCH
CANADA
U.S.A.
112°
51°
49°
117°
②

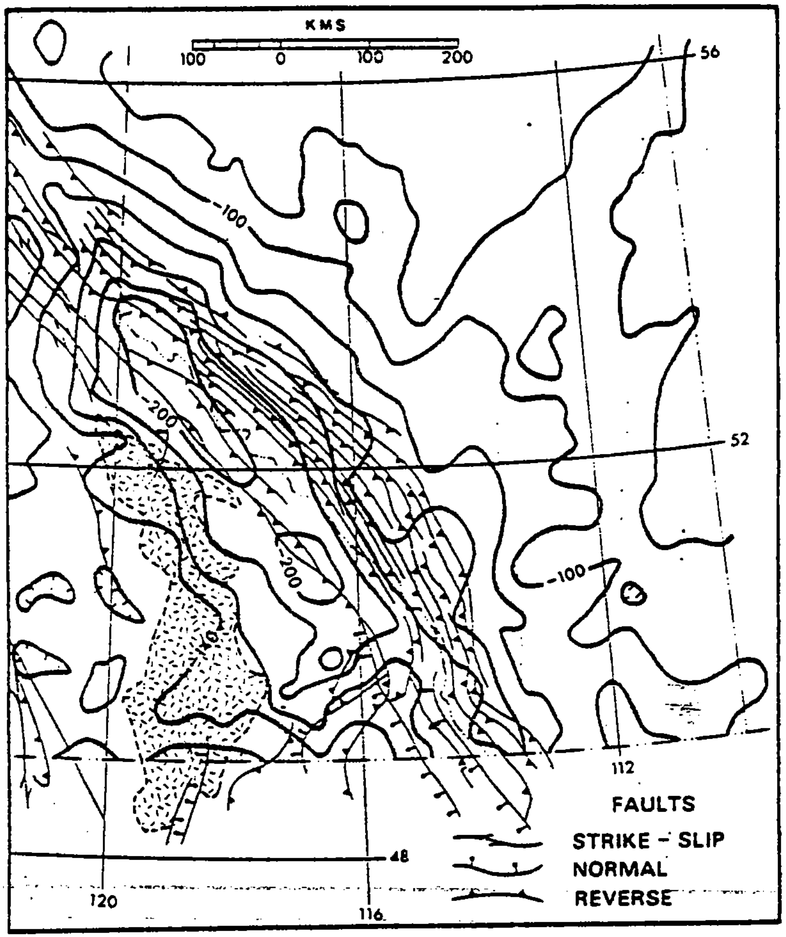


FIG. 4. Relationships between the structures in the foreland thrust and fold belt of the southern Canadian Rocky Mountains and the Bouguer gravity anomalies. The Bouguer gravity anomaly values are after the Earth Physics Branch of Canada (1974). The faults and the limits of the Shuswap Metamorphic Complex are from Fig. 1 and from King (1969).

Mountjoy (1970)—10 km of increase in crustal thickness, of which 2 km stand above sea level. However, it is not enough to balance, at a deep crustal level below the basement surface, the amount of tectonic shortening that occurs in even the eastern part of the thrust and fold belt, where deep drilling and seismic reflection data leave little scope for imagination in interpretations of tectonic shortening in the supracrustal wedge. Thus, a consideration of the deep crustal structure, based on the seismic refraction, magnetic and gravity data leads to the conclusion that the slab of continental crust, about 40 km thick, that underlies the cratonic platform in front of the thrust and fold belt, extends under the deformed wedge of supracrustal rocks, beneath the Rocky Mountains and the Purcell anticlinorium, to the Kootenay Arc, where the supracrustal rocks wrap around the edge of it (Fig. 5). This is contrary to the interpretation of Eisbacher *et al.* (1974) who suggested that the structure in the western Rocky Mountains and beyond is the result of uplift involving both the supracrustal wedge and its Precambrian crystalline basement, and that the flat thrust faults and undeformed basement are limited to the eastern Rocky Mountains. Their interpretation was based on structure sections which cannot be balanced between the supracrustal rocks and the basement because they involve relatively little shortening at the level of the basement,

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FIG. 3. Relationships between the structures of the thrust and fold belt of the southern Canadian Rocky Mountains and the Hudsonian structural fabric of the Precambrian basement that extends under it. The magnetic anomaly map is after Kanasewich *et al.* (1969). The thrust faults and normal faults are a simplified portrayal of those in Fig. 1.

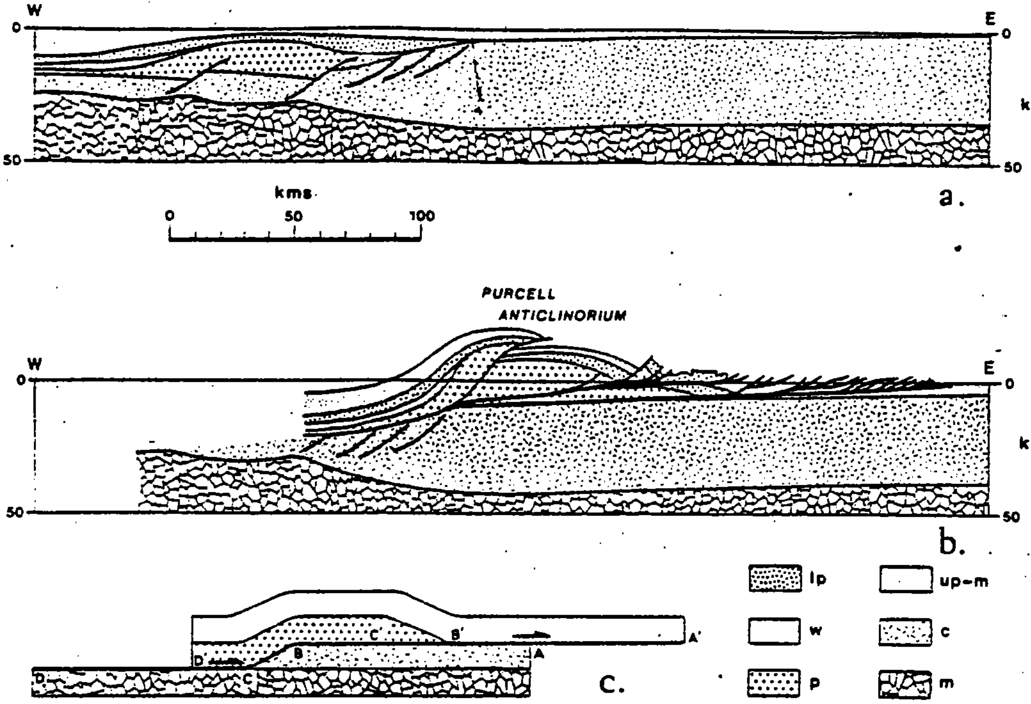


FIG. 5. Evolution of the Purcell anticlinorium a. Restored Section through the southeastern Canadian Cordillera along 49°45'N Latitude (as in Fig. 7). b. Generalized structure of the Cordilleran fold and thrust belt and the Purcell anticlinorium along 49°45'N Latitude (based on Section W–E of Fig. 2: but drawn to eliminate the effects of the erosion of supracrustal rocks that occurred during and after the thrust faulting). c. Schematic representation of the development of the Purcell anticlinorium (Points on the footwall and hanging wall of the zone of detachment, that were initially contiguous, are labelled with the same letters) m—mantle; c—continental crust; p—Belt-Purcell assemblage; w—Windermere assemblage; lp—Lower Palaeozoic assemblages; up–m—Upper Palaeozoic and Triassic–Jurassic assemblages.

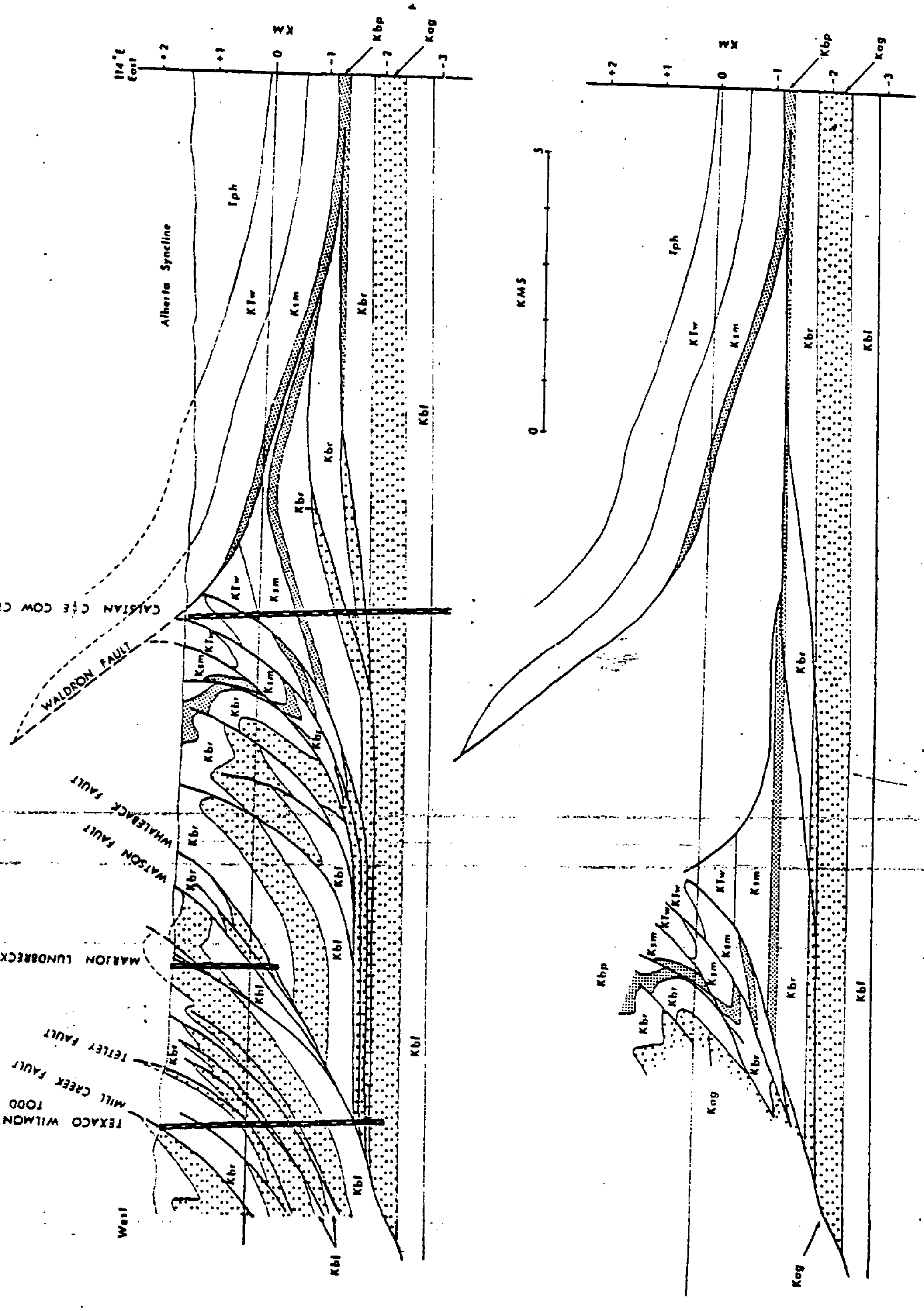
whereas tectonic shortening in the supracrustal rocks to the E amounts to 65 km across the Foothills alone (Bally *et al.* 1966, plate 12), and all of this would have to be accommodated within the basement E of the boundary of their postulated basement uplift.

A conspicuous increase in the Bouguer gravity values occurs southward along the Purcell anticlinorium where the transverse northeasterly trending St Mary and Moyie Faults cut across it. Further S. in Montana, at about 48°15'N. Wynn *et al.* (1977), on the basis of their analysis of a gravity and audio-frequency magnetotelluric traverse across the Purcell anticlinorium. concluded that crystalline basement rocks occur in the core of the anticlinorium there. Transported basement rocks could be expected within the deformed wedge of supracrustal rocks at this structural position because suspected basement rocks occur in about the same position above the Purcell thrust fault. to the N, at about 52°30'N (Campbell 1968; Price & Mountjoy 1970).

Structure sections W–E and SW–NE (Fig. 2) have been drawn perpendicular to the local structural trend, in order to facilitate the projection of surface and well data, along the plunge of the structures, into the plane of the section. Errors in estimates of tectonic displacements, introduced by the fact that the structure sections may not be parallel with the direction of net tectonic displacement, vary with the cosine of the angle between the structure section and the direction of net tectonic displacement. The error introduced in this way is normally less than 15%, if the angle between the line of section and the average direction of net tectonic displacement is less than 30°. Both sections are based on the premise that the Hudsonian crystalline basement complex extends under the thrust and fold belt to the centre of the Purcell anticlinorium. and that all the thrust faults lie above it, in the supracrustal wedge, or at the basement surface (Bally *et al.* 1966). In both sections, the depth to the basement, as shown, is the minimum depth that is required to accommodate the total thickness of the supracrustal succession, as it is known from one place to another, and to meet the geometric requirements of a balanced section. The eastern part of Section W–E is relatively tightly constrained by information from deep drilling, and it illustrates several important principles, some of which are portrayed in more detail in Fig. 6. It is a well established empirical rule that the thrust faults cut up through the stratigraphic succession in the direction of relative displacement of the overthrust side, and that they carry older rocks over younger and produce repetitions of the normal stratigraphic succession. Thus, westerly verging thrust faults, such as the Waldron fault, cut up through the stratigraphic succession toward the W; and easterly verging thrust faults, even if folded and E-dipping, cut up through the stratigraphic succession toward the E and carry older rocks over younger. Certain stratigraphic zones are a preferred locus of décollement or bedding glide, and other thrust faults branch away from them as concave upward imbricate splays. The condition of geometric compatibility for a balanced section requires that an extensive bedding glide zone in the hanging-wall of a thrust fault must have its matching counterpart in the footwall. The bedding glide zones in the upper part of the Alberta Group (Kag) in the eastern part of Fig. 6, require matching footwall counterparts to permit the palinspastic reconstruction shown in the lower part of the figure. The thrust faults are commonly stepped, comprising extensive bedding glide zones that are linked by relatively narrow ramps. The external rotation and folding which is inherent in movement along such stepped fault surfaces affects all the overlying rocks, including any thrust faults which occur within them. Thrust faults may merge upwards as well as downwards, as shown by the tectonically thickened wedge of Alberta Group (Kag), Belly River Formation (Kbr) and Bearpaw Formation (Kbp) which is responsible for the easterly dip of the Alberta Syncline, and separates it from the relatively flat-lying strata below (Fig. 6). Elsewhere, as for example in the Coleman Thrust sheet, beneath where it is overlapped at the surface by the Lewis Thrust sheet (Fig. 2, Section W–E), sigmoidal imbricate thrust faults that merge upward with one fault and downward with another result in substantial tectonic thickening of an isolated lenticular mass of rock, as well as a significant transfer of displacement from the lower fault to the upper.

In the Foothills, along and beneath the Livingstone Thrust fault (Li in Section W–E, Fig. 2), extensive bedding glide zones occur in Jurassic marine shales (Fernie Group) and in or at the base of a thick unit of Upper Devonian micritic limestone (Palliser Formation). The condition of geometric compatibility for a balanced section requires that offset counterparts of these bedding glide zones be represented along the opposite side of the same thrust surfaces to the W; and this leads to the conclusion that autochthonous Upper Devonian and Cambrian strata lie on undeformed Hudsonian crystalline basement under the western Rocky Mountains, beneath the Bourgeau (Bo) and Bull River (BR) thrust sheets. It also implies that along this line of section the Lewis Thrust sheet (Le) cannot contain a thick succession of Belt-Purcell strata, as it does further S (Fig. 1), unless there is an abrupt change in the depth to the basement surface to provide the necessary space. Seismic reflection profiling shows no such change; instead, the basement in this area slopes gently W without disruption' (P. L. Gordy pers. comm. 1979). Similar arguments, involving the Coleman thrust sheet (Co), the Lewis Thrust sheet (Le), and the Bourgeau Thrust sheet (Bo), indicate that the thick Lower Palaeozoic miogeoclinal shale facies and the underlying thick succession of Belt-Purcell strata that occurs between the Bull River Thrust fault and the Rocky Mountain Trench (RMT) first appears in the Bougeau Thrust sheet. The abrupt westerly increase in the thickness and facies of the supracrustal succession within the Bourgeau thrust sheet, which marks the change from platformal to miogeoclinal supracrustal rocks, must have had a matching counterpart in the 'autochthonous' cover succession on the basement below the Bourgeau thrust fault. Palinspastic reconstruction of only that part of the balanced section W–E (Fig. 2) which lies E of and includes, the Bourgeau Thrust sheet (eastern two-thirds of the Rocky Mountains) indicates that the locus of this matching 'autochthonous' counterpart is along the Kootenay Arc. This corroborates what was concluded from a consideration of the geophysical data—that the western edge of the thick slab of continental cratonic crust lies beneath the Kootenay Arc. It implies that all the thick miogeoclinal strata, including the Lower Palaeozoic shale facies and the underlying Windermere and Belt-Purcell assemblages, accumulated outboard from the edge of the craton on attenuated continental crust or oceanic crust.

The structure in section SW–NE differs in


114° E
East
West
KM
+2
+1
0
-1
-2
-3
Kbp
Kag
Kbl
Alberta Syncline
Tph
Klw
Ksm
Kbr
KMS
0
5
CALSTAN C&E COW CK
WALDRON FAULT
WHALEBACK FAULT
WATSON FAULT
MARION LUNDBRECK No. 1
TETLEY FAULT
MILL CREEK FAULT
TEXACO WILMONT
TODD CK. A-1

detail from that in Section W–E because of changes in the character of the supracrustal succession, but also because of an increase in the amount of overall tectonic shortening or convergence across the fold and thrust belt. The Lower Palaeozoic platformal and miogeoclinal carbonate facies is substantially thicker and more complete beneath the sub-Devonian unconformity, and extends further W than in Section W–E. It forms a series of five moderately thick, relatively steeply dipping thrust sheets (Mc, La, In, Ru, and Sm in Section SW–NE of Fig. 2) beneath the Bourgeau Thrust fault (Bo); whereas in Section W–E there are only two in the same interval. As in Section W–E, the Bourgeau Thrust fault marks the approximate position of the relatively abrupt increase in thickness of the Lower and Upper Palaeozoic assemblages that defines the boundary between the platform and the miogeocline. It also marks the first appearance of Precambrian rocks, but these belong to the Windermere assemblage rather than the Belt-Purcell assemblage.

A broad, northerly plunging anticlinorium occurs W of the Bourgeau Thrust fault, in the centre of the Rocky Mountains near the Continental Divide. The Simpson Pass Thrust fault follows its crest, and several large thrust faults that develop as northeasterly diverging splays from the Simpson Pass Thrust fault are folded across the anticlinorium and adjacent syncline, and extend far to the NW (Price & Mountjoy 1970).

There are about 2 km of Windermere strata exposed above the Simpson Pass Thrust fault, and these, together with the thicknesses of the overlying formation, define a lower limit on the total thickness of supracrustal rocks that must be accommodated between the crystalline basement and the rocks that are exposed at the surface to the W of the Simpson Pass Thrust fault. This minimum estimate of the depth to the basement is what has been used to construct the section. Because the section is based on the premise that the Hudsonian basement is undeformed and slopes beneath the deformed supracrustal wedge to beyond the Rocky Mountain Trench, the space between the basement surface and the core of this northerly plunging anticlinorium is considered to be occupied by deformed supracrustal rocks, the specific structure of which is completely hypothetical. The particular interpretation shown in the section follows the style of the structures which occur in a similar setting beneath the Coleman fault and the Livingstone Thrust in Section W–E, where their existence is documented by deep drilling; and beneath the Lewis thrust sheet further S, where they are exposed (Fermor & Price 1976). The choice of the number of thrust faults that occur beneath the Bourgeau Thrust fault and merge upward with it may be arbitrary, but it does have implications regarding the rate at which the stratigraphic succession in the supracrustal wedge thickens westward between the Sulphur Mountain (Sm) and Bourgeau thrust faults. The interpretation shown implies that the thickening is gradual and relatively uniform. If the thickening is abrupt, fewer thrust faults and less shortening between the Sulphur Mountain and Bourgeau thrust sheets would be involved.

The carbonate bank margin, marking the boundary between the Lower Palaeozoic carbonate and shale facies, occurs W of the Simpson Pass fault and is, in part, offset by late listric normal faults (Cook 1975). W of this boundary, there is a profound change in structural style marked by the widespread development of slaty cleavage and high penetrative strain. A conspicuous fan structure is outlined by the orientation of the cleavage and the axial surfaces of the folds, which are northeasterly verging on the NE side and southwesterly verging on the SW side. Locally, it is situated over an anticlinal culmination in underlying Eo-Cambrian quartz sandstones (Balkwill 1972), but elsewhere (Section SW–NE) it coincides with a monoclinal flexure (see also Price & Mountjoy 1979*a* and *b*; Price *et al.* in press). The fan structure can be attributed to décollement above the Eo-Cambrian quartz sandstone succession, in Cambrian slates of the Lower Palaeozoic shale facies, in conjunction with displacement on the Purcell Thrust fault (as shown in Section SW–NE). Cross-cutting relationships amongst northeasterly and southwesterly verging thrust faults on the W flank of the fan structure, under the Purcell Thrust fault, outline two phases of tectonic overprinting. Older northeasterly verging thrust faults were rotated during continued folding, until overturned, together with the beds in which

FIG. 6. Structure of the eastern Rocky Mountain Foothills at 49°45′N Latitude. The line of section is the same as that of Section W–E in Fig. 2, and the sources of information on wells drilled for oil and gas are those given in Fig. 2. The interpretation in the upper part of the section is adapted, with some modifications, from Douglas (1950). The symbols identifying lithostratigraphic units are as follows: Kbl—Blairmore Group; Kag—Alberta Group; Kbr—Belly River Formation; Kbp—Bearpaw Formation; Ksm—St. Mary River Formation; KTw—Willow Creek Formation; Tph—Porcupine Hills Formation.

they occur (Balkwill 1972). They are cut by younger southwesterly verging thrust faults that involve displacements out of the core of the fan structure. These, in turn, are cut by still younger northeasterly verging thrust faults that are spatially related to the Purcell thrust fault which has overridden the W flank of the fan structure. The Lower Palaeozoic shale facies appears to have been 'scraped' off the underlying quartz sandstones, and piled up in front of the Purcell Thrust sheet. The same fan structure extends southward to Section W–E, beyond which it terminates against the northeasterly trending Moyie-Dibble Creek Fault (Moy) that follows the southern margin of the basin in which the Lower Palaeozoic shale facies accumulated. This asymmetric basin, which formed on top of the late Precambrian miogeoclinal prism, was probably controlled by deep-seated listric normal faulting. The steep, E flank forms a natural, local boundary between the miogeocline and platform.

In the Purcell anticlinorium, W of the Purcell fault, the Lower Palaeozoic miogeoclinal succession, which is about 8 km thick in the western Rocky Moutains, occurs as a condensed section that is overlain by Upper Devonian strata, and locally is less than 500 m thick (Reesor 1973). This contrast can be attributed to tectonic foreshortening, across the Purcell thrust fault, of the less steep W flank of the Lower Palaeozoic basin. The structure shown in the upper levels of the Purcell anticlinorium in Section SW–NE is based on projections along the plunge of structures exposed nearby; but the structure at depth is based on the hypothesis that the edge of the undeformed cratonic basement extends to the W flank of the Purcell anticlinorium, and that the Purcell Thrust fault and the décollement above the Eo-Cambrian quartz sandstone succession extends just about as far.

Palinspastic restorations

The tectonic displacements involved in the evolution of the thrust and fold belt can be analyzed most conveniently in terms of a horizontal datum defined by the transition from late Jurassic marine to non-marine deposits, which marks the onset of synorogenic sedimentation; and a reference frame fixed relative to the 'autochthonous' supracrustal rocks that are attached to the Hudsonian basement E of the thrust and fold belt (Price & Mountjoy 1970). The relative horizontal (and vertical) displacements of points in the thrust and fold belt, and the amount of tectonic shortening or convergence between them and the autochthon can be estimated by comparing their position in the restored section with their present position, provided that the angle between the section and the actual direction of net tectonic displacement is not more than about 30°. The minimum total convergence between the leading edge of the Bourgeau thrust sheet and the autochthon estimated in this way is about 170 km in Section SW–NE, and about 105 km in Section W–E. The estimated minimum total shortening up to and including the McConnell Thrust sheet, which stands above the Foothills in Section SW–NE, is about 75 km, whereas that up to and including the Lewis Thrust sheet, which stands above the Foothills in Section W–E, is about 90 km. However, the Lewis thrust fault dies out along the strike to the N within the Rundle Thrust sheet, and the estimated minimum net shortening up to and including the Rundle Thrust sheet in Section SW-NE is about 105 km. Although the total convergence across the fold and thrust belt probably does decrease southward, some of the difference in estimated tectonic convergence between the two sections can be attributed to errors arising from sections that are not parallel with the direction of net tectonic displacement.

The palinspastic restoration based on Section SW–NE, indicates that the Mesozoic exogeoclinal rocks occurring under and E of the McConnell Thrust fault have been horizontally compressed by at least 10 km more than the underlying Upper Palaeozoic platformal rocks from which they are separated by a regionally important bedding glide or décollement zone (Dahlstrom 1969*a*). A similar relationship in a nearby section has been described by Bally *et al.* (1966, p. 350), and attributed to cross-cutting relations involving a younger fault which developed in the Palaeozoic carbonate rocks, under the Mesozoic rocks, after the latter had been horizontally compressed above the décollement. In Section SW–NE, this younger fault is the McConnell Thrust fault, and the tectonic shortening that occurred above the décollement before it was offset by the McConnell Thrust must be related, at a deeper level, to other more westerly thrust faults.

Times of deformation

The record of deformation within the thrust and fold belt in this region spans an interval of almost 100 Ma from late Jurassic to Palaeocene time (Price & Mountjoy 1970).

The first stratigraphic record of deformation, uplift and erosion of the miogeoclinal assemblages occurs in the Upper Jurassic rocks (Kootenay Formation and upper part of Fernie Group). It is marked by an abrupt reversal in the provenance of clastic sediment, from the NE cratonic provenance, that characterized the miogeoclinal assemblages, to the SW Cordilleran provenance that characterized the exogeoclinal assemblage. Although the ages of individual thrust faults cannot be defined precisely, some broad limits can be established. Thrust faults in the western Rocky Mountains, particularly those along the W side of the fan structure that occurs in the Lower Palaeozoic shale facies, are linked spatially and kinematically to NE trending, right-hand reverse faults that cut across the Purcell anticlinorium (Fig. 1). Several of these NE trending, right-hand reverse faults are cut by early to mid-Cretaceous batholiths, and this establishes an upper limit for the times of last displacement on them and on the thrust faults that are linked to them. The Hall Lake fault, which is linked to faults in, and E of, the Rocky Mountain Trench that form part of the same thrust system as the Purcell fault, is cut by the White Creek batholith, for which Rb-Sr and K-Ar whole rock dates of 129 Ma and 111 Ma have been reported (Wanless *et al.* 1968). The St Mary fault, which is linked to faults in the western Rocky Mountains NE of Cranbrook, is cut by the Bayonne batholith, and K-Ar dates from the Bayonne batholith indicate that it was emplaced prior to 90 Ma (Archibald *et al.* 1977). Along the Kootenay Arc the late to mid-Cretaceous granitic plutons post-date most, or all, of the penetrative strain in the country rocks and have relatively low pressure thermal aureoles (Glover 1978; Sears 1979), in contrast with the late Jurassic plutons which are associated with intermediate pressure (Barrovian) regional metamorphism, and were deformed along with the enclosing strata (Gabrielse & Reesor 1974; Glover 1978). Thus, deformation in the eugeoclinal rocks W of the Kootenay Arc was underway in late Jurassic time, and thrust faults in the western Rocky Mountains had already developed by mid-early Cretaceous time.

The McConnell and Lewis thrust sheets both overlap Cenomanian to Santonian marine deposits and early Campanian non-marine deposits of the exogeoclinal assemblage (Alberta Group and Belly River and Brazeau Formations, respectively), and this establishes a lower limit for the time of displacement on these faults. An upper limit for the time of displacement on the Lewis Thrust is given by the late Eocene–early Oligocene age of the Kishenehn Formation, which unconformably overlaps the structures in the Lewis thrust sheet along the downthrown side of the Flathead fault (Price 1965), a SW dipping listric normal fault that merges at depth with the Lewis thrust fault (Bally *et al.* 1966). Thus, a minimum of about 100 km of the total horizontal convergence across the thrust and fold belt occurred between early Campanian and late Eocene time; and if the subsidence and sedimentation in the migrating foredeep was an isostatic response of the lithosphere to the loads imposed by the displacement of the thrust sheets (Price 1973), this amount of convergence must have occurred within less than 30 Ma, the time represented by the thick succession of early Campanian to Palaeocene fluvial sediments that makes up the youngest part of the exogeoclinal assemblage (McLean & Jerzykiewicz 1978). The implications of these conclusions warrant further consideration.

In early Campanian time, prior to displacement on the McConnell, Lewis and Rundle Thrusts, the Precambrian crystalline basement that now extends as an undeformed autochthonous slab under the Rocky Mountains and the eastern part of the Purcell anticlinorium was still covered by a thin sequence of autochthonous, Palaeozoic platformal carbonate rocks with an overlying easterly tapering blanket of autochthonous late Jurassic–Campanian exogeoclinal deposits (Fig. 2). At almost the same time, but perhaps somewhat earlier, prior to the displacement on the Bourgeau thrust fault, the hinge zone between the platform and the miogeocline lay above the edge of this slab of continental crust, and the tectonically compressed miogeocline was situated outboard from it, and was underlain by tectonically attenuated continental crust, and perhaps, in part, oceanic crust. At the end of Middle Jurassic time, prior to the displacements on the thrust faults in the western Rocky Mountains and the the eastern Purcell anticlinorium, and on the NE trending right-hand reverse faults that cut across the Purcell anticlinorium, the miogeocline was essentially undeformed and comprised a series of overlapping and interfingering, unconformity-bounded, tectono-stratigraphic assemblages, for which the aggregate thickness at any one place was about 15–20 km.

Tectonic models

The only obvious actualistic model for this situation seems to be a rifted continental mar-

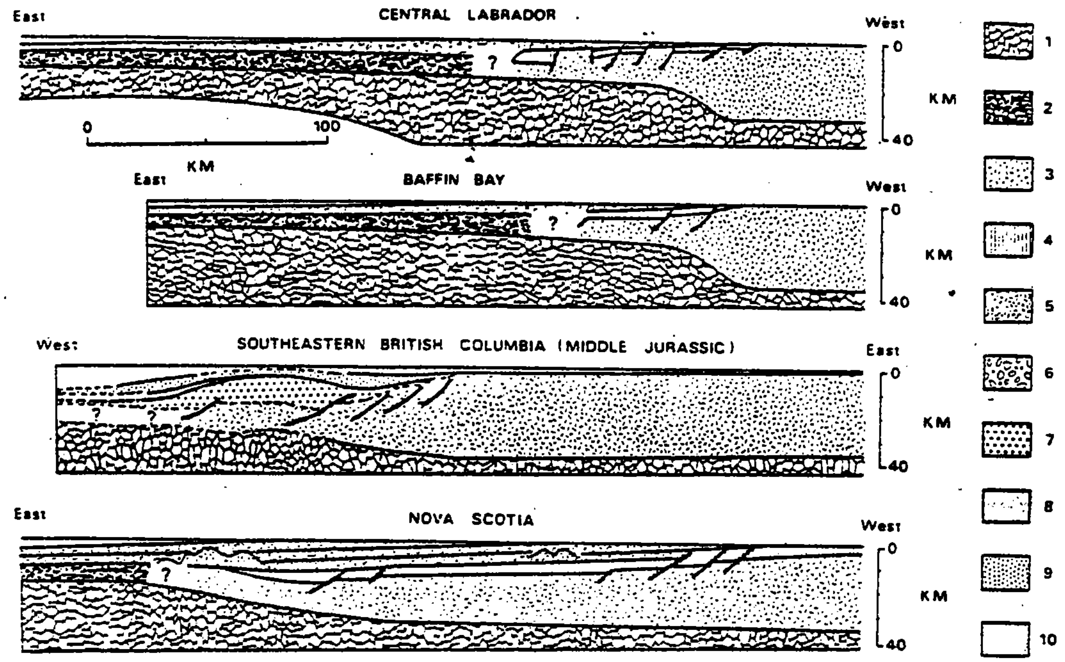


FIG. 7. Comparison of a palinspastically restored section through the foreland thrust and fold belt of southern Canada with crustal sections through the Atlantic margin of Canada. 1—Mantle; 2—oceanic crust; 3—continental crust: 4—continental basement rocks (pre-rifting); 5—early Jurassic and older sediments on the Nova Scotian shelf; 6—post-rifting sediments; 7—Belt-Purcell assemblage; 8—Windermere assemblage: 9—Lower Palaeozoic assemblages; 10—Upper Palaeozoic and Triassic–Jurassic assemblages.

The sections through the Atlantic continental margin are taken from Keen & Hyndman (1979). The palinspastically restored section for the Cordilleran margin in southern British Columbia and Alberta in Middle Jurassic time is based on the Section W–E of Fig. 2, and on the premise that the slab of continental crust approximately 40 km thick, which underlies the craton E of the foreland thrust and fold belt. extends under it to the vicinity of the Kootenay Arc, where it is tectonically attenuated as a result of rifting associated with the development of the western margin of North America in Proterozoic and early Palaeozoic time.

gin of the Atlantic-type. The results of recent investigations of the Atlantic continental margin of Canada (Keen & Hyndman 1979) provide some indication of the possible range in the configurations and relationships among continental and oceanic crust and an overlying, continental terrace wedge of supracrustal rocks (Fig. 7). The abrupt increase in thickness of the supracrustal wedge along section W–E can be compared with the abruptly attenuated continental margins that occur in Baffin Bay or Central Labrador, whereas the apparently more gradual increase in thickness along section SW–NE may be more comparable with gently tapering continental crust off Nova Scotia.

When the evolution of the thrust and fold belt is viewed from this perspective, the origin and significance of the Purcell anticlinorium and Kootenay Arc become obvious (Fig. 5). The Purcell anticlinorium formed in Campanian to early Palaeocene time, as a geometric consequence of the juxtaposition of the thick northeasterly tapering prism of sediment that had accumulated above the zone of abrupt crustal attenuation, outboard from the continental margin, with relatively flat, planar basement surface on the continental platform. The Kootenay Arc formed at the same time, as a W-facing monoclinal step beyond the crest of the Purcell anticlinorium where the sedimentary prism was draped over the edge of the continental platform. The zone of detachment, along which this juxtaposition occurred, reached the surface far to the E in the thrust and fold belt where it is expressed as a series of stacked imbricate thrust faults in the horizontally compressed supracrustal rocks. The overall structure is a crustal scale version of the stepped thrust faults which control the configuration of structures within the deformed supracrustal wedge (Fig. 5c).

The amount of displacement of the miogeoclinal prism relative to the cratonic basement can be visualized readily on even a regional geological map such as Fig. 1, provided that the map shows the relative distrubutions of the Palaeozoic shale and carbonate facies, and the location of the Kootenay Arc. The boundary between the Lower Palaeozoic shale and carbonate facies formed above, or close to, the edge of the continental craton, and the Kootenay Arc now follows the locus of the edge of this continental craton. Thus, the distance between the Kootenay Arc and the shale-carbonate facies boundary is a measure of how far this boundary has been displaced, and therefore, of the amount of convergence between the miogeocline and the autochthonous cover on the craton to the E. It is worth noting, that adjacent to the International Boundary, where there is a conspicuous right-hand deflection about 50 km long in the otherwise curvilinear N to NW trend of the Kootenay Arc, there is a matching right-hand deflection in the eastern and southern limit of the Lower Palaeozoic shale facies, which lies about 150 km NE of the Kootenay Arc. Moreover, S of 50°N Latitude, where the structural relief across the Kootenay Arc is greatest and the change in level across it is most abrupt, the change from a thin platformal facies of the Lower Palaeozoic assemblage to a very thick shale facies is also most abrupt. It seems clear that the miogeocline formed outboard from the continental craton, on tectonically attenuated continental crust, or on oceanic crust, and has subsequently been juxtaposed with the continental craton.

The idea that foreland thrust and fold belts may develop by lateral gravitational spreading is based on the premise that some other process operates to produce the topographic gradient, and thereby the gravitational potential, that drives the lateral spreading. Price & Mountjoy (1970) ascribed the gravitational potential in the southern Canadian Rockies to the buoyant upwelling of a hot, ductile infrastructure of high-grade metamorphic rocks, the exposed parts of which are represented by the intermediate-pressure (Barrovian) series metamorphic rocks of the Shuswap Metmorphic Complex. However, they left unspecified the processes that had carried the protolith of the metamorphic complex to the depths from which the subsequent upwelling occurred; and they also, mistakenly, placed too much emphasis on a conceptual model involving simple diapiric upwelling without any attendant convergence or crustal shortening. The dominantly flat foliation and mainly northeasterly oriented stretching lineation in the Shuswap Metamorphic Complex can be taken as evidence of large-scale vertical compression and horizontal northeasterly extension (Price & Mountjoy 1970). However, the fact that the less metamorphosed rocks in the suprastructure around the southern (Fig. 1) and northern (Campbell 1973) ends of the metamorphic culminations have a fabric indicative of horizontal compression, and moreover, show no indication of horizontal extension on the scale that is necessary to balance the more than 200 km of horizontal shortening that occurs between the metamorphic complex and the autochthonous cover on the craton, implies that there has been a large net horizontal shortening or convergence across the entire belt, and that the horizontal stretching in the deep infrastructure is a second-order effect associated with the special conditions which existed there. For example, tectonic slides are common around some of the margins of the metamorphic culminations. These are shallow-dipping extension faults (see Fig. 1) across which young rocks are juxtaposed on old rocks, and low-grade metamorphic rocks on high-grade metamorphic rocks. Faults of this type, marked by strong cataclasis, occur along the E side of the Shuswap Metamorphic Complex (Read 1977; Sears 1979), on the flanks of the Valhalla gneiss complex (Reesor 1965; Little 1960); on the W flank of a NNW-trending metamorphic culmination that extends along the Purcell Trench at about 116°30′ to 117°E Latitude (Fyles 1967); and around three sides of a local structural and metamorphic depression within a U-shaped metamorphic culmination at Newport in northeastern Washington State (Miller 1974 *b–d*). There is a contrast in structural style across the faults from recumbent folds below, to upright folds above. Some of these faults are cut by early to mid-Cretaceous plutons (Read 1977); others cut early (?) Tertiary rocks (Miller 1974*b–d*). The basic structural relationships are similar to those around many metamorphic culminations further S in the Cordilleran belt of the United States. The faults around these metamorphic culminations have been interpreted (Davis & Coney 1979) as ductile, normal, growth faults related to Neogene thermal upwelling with horizontal stretching, pinch-and-swell, and mega-boudinage of the crust. However, they occur in the basin-and-range province where evidence of Neogene–Recent crustal extension abounds; and moreover, the displacements on them appear to be much younger than on the comparable faults in S Canada and NE Washington, where the shallow extension faults appear to be related to local diapiric movement involving the buoyant rise of culminations in the metamorphic infrastructure and relative subsidence of parts of the suprastructure. Lateral continuity in nearby rocks indicates that they do not represent a major regional horizontal extension of the crust.

Brown & Tippet (1978) have interpreted the evolution of the Selkirk fan structure, along the eastern margin of the metamorphic complex NW of Golden, as a product of two distinct phases of deformation; and Brown (1978) has concluded that the evolution of the SE Canadian Cordillera involved earlier 'eastward underthrusting from the W', and then 'westward underthrusting of North America from the E'. However, the Selkirk fan structure has also been interpreted as a fan fold (Price & Mountjoy 1970; Zwanzig 1973; Price 1979), like that in the Lower Palaeozoic shale facies of the western Rocky Mountains (Fig. 2), involving a single progressive deformation above a zone of regional décollement, and comprising a local reversal in the vergence of the structures, like that which occurs along the W side of the Alberta syncline (Fig. 6).

The conclusion that there has been a large net horizontal shortening or convergence across the entire tract from the E edge of the thrust and fold belt to the W edge of the Shuswap Metamorphic Complex constrains interpretations of the geotectonic significance of the foreland thrust and fold belt. Convergence between the W flank of the Shuswap Metamorphic Complex and the autochthonous cover on the continental craton appears to be approximately equal to the amount of crustal shortening across the thrust and fold belt— about 200 km (Bally *et al*. 1966; Price & Mountjoy 1970). The shortening must have been balanced by an equivalent convergence or crustal shortening at a deeper crustal level, and this constitutes a type of subduction. Bally (1975) has called this 'Alpinotype or A-subduction', and has contrasted it with 'Benioff or B-subduction', in which a slab of oceanic lithosphere dips deep into the mantle with décollement or scraping off of supracrustal rocks and imbrication of oceanic crust and upper mantle. He has suggested that it is the shallower conjugate homologue of B-subduction and that it involves the disposal of large amounts of continental crust. In the southern Canadian Rocky Mountains, this A-subduction may have been the conjugate homologue of B-subduction which occurred along the Pacific Margin of the Cordillera from mid-Jurassic to Palaeogene time (Monger & Price 1979). However, it apparently did not involve the disposal of any significant amount of continental crust, because it occurred outboard from the edge of the continental craton, in an area floored by attenuated continental crust or oceanic crust.

The following scenario is suggested. As the attenuated continental crust and its supracrustal cover of eugeoclinal and miogeoclinal rocks were carried into the subduction zone, they were metamorphosed to relatively high grades in an intermediate pressure (Barrovian) metamorphic facies series, became ductile, rose bouyantly and spread laterally into the overlying shallower suprastructure of less metamorphosed rocks, partly along ductile zones of extension faulting that are now represented by tectonic slides. Easterly verging imbricate thrust faults and related folds developed concurrently, at a higher level, farther to the E in the subduction zone, in a subduction complex that was tectonically prograded eastward up to and over the edge of the continental craton, in much the same way that subduction complexes above sediment-filled oceanic trenches have been tectonically prograded over the adjacent oceanic crust (Hamilton 1977). The topographic slope produced in this way was primarily due to the compression of the supracrustal cover that was scraped off the down-going slab, and therefore, was a second-order effect, rather than the primary cause of the thrust and fold belt (Chapple 1978; cf. Elliot 1976). Thus, the evolution of the foreland thrust and fold belt in the southern Rocky Mountains can be viewed as an example of intra-plate convergence, involving the collapse of a back-arc or marginal basin, behind the main easterly dipping Cordilleran subduction zone. As the supracrustal cover of miogeoclincal rocks was scraped off the down-going slab, it was tectonically prograded over an adjacent continental craton to form a very wide and shallow subduction complex.

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References

AALTO, K. R. 1971. Glacial marine sedimentation and stratigraphy of the Toby conglomerate (Upper Proterozoic) southeastern British Columbia, northwestern Idaho and northeastern Washington. *Can. J. Earth Sci.* **8**, 753–87.

AITKEN, J. D. 1971. Control of Lower Paleozoic sedimentary facies by the Kicking Horse rim, southern Rocky Mountains, Canada. *Bull. Can. Pet. Geol.* **19**, 557–69.

ARCHIBALD, D. A., GLOVER, J. K. & FARRAR, E. 1977. K-Ar ages from the Bayonne batholith and some nearby plutons in SE British Columbia and their geological implications. *Geol. Assoc. Can. Progs with Abstr.* **2**, 5.

BALKWILL, H. R. 1972. Structural geology, lower Kicking Horse river region, Rocky Mountains, British Columbia. *Bull. Can. Pet. Geol.* **20**, 608–33.

——, PRICE, R. A. & MOUNTJOY, E. W. (in press *a*). Geology, Golden (east half), British Columbia. *Geol. Surv. Can. Map* **1496A**.

——, —— & —— (in press *b*). Geology, Golden (west-half), British Columbia. *Geol. Surv. Can. Map* **1497a**.

BALLY, A. W. 1975. A geodynamic scenario for hydrocarbon occurrences. *9th World Petrol. Cong. (Tokyo)*, **2**, 23–44.

——, GORDY, P. L. & STEWART, G. A. 1966. Structure, seismic data and orogenic evolution of southern Canadian Rockies. *Bull. Can. Pet. Geol.* **14**, 337–81.

BENNETT, G. T., CLOWES, R. M. & ELLIS, R. M. 1975. A seismic refraction survey along the southern Rocky Mountain Trench, Canada. *Bull. seismol. Soc. Am.* **65**, 37–54.

BENVENUTO, G. L. 1978. *Structural Evolution of the Hosmer Thrust Sheet, South-Eastern British Columbia.* Queen's Univ. Ph.D. thesis (Unpubl.).

BERNOULLI, D., LAUBSCHER, H. P., TRÜMPY, R. & WENK, E. 1974. Central Alps and Jura Mountains. *In:* SPENCER, A. M. (ed). *Mesozoic-Cenozoic Orogenic Belts.* Spec. Publ. geol. Soc. London, **4**, 83–108.

BERRY, M. J. & FORSYTH, D. A. 1975. Structure of the Canadian Cordillera from seismic refraction and other data. *Can. J. Earth Sci.* **12**, 182–208.

BIELENSTEIN, H. U., PRICE, R. A. & JONES, P. B. 1971. Geology of the Seebe-Kananaskis area, Map. *In:* HALLADAY, I. A. R. & MATHEWSON, D. H. (ed). *A Guide to the Geology of the Eastern Cordillera along the Trans-Canada Highway between Calgary, Alberta and Revelstoke, British Columbia.* Alberta Soc. Petrol. Geol., Calgary.

BROWN, R. L. 1978. Structural evolution of the southeast Canadian Cordillera: A new hypothesis. *Tectonophysics.* **48**, 133–51.

—— & TIPPET, C. R. 1978. The Selkirk fan structure of the southeastern Canadian Cordillera. *Bull. geol. Soc. Am.* **89**, 548–58.

BURCHFIEL, B. C. & DAVIS, G. A. 1972. Structural framework and evolution of the southern part of the Cordilleran orogen, Western United States. *Am. J. Sci.* **272**, 97–118.

CAMPBELL, R. B. 1968. Canoe River, British Columbia. *Geol. Surv. Can. Map* **15–1967**.

—— 1973. Structural Cross-Section and Tectonic Model of the S.E. Canadian Cordillera. *Can. J. Earth Sci.* **10**, 1607–20.

CAMPBELL, I. & LOOFBOUROW, J. S., Jr. Preliminary geologic map and sections of the magnesite belt, Stevens county, Washington. *U.S. Geol. Surv. Map* **MF-117**.

CANER, B. 1971. Quantitive interpretation of geomagnetic depth-sounding data in western Canada. *J. geophys. Res.* **76**, 7202–16.

CHANDRA, N. N. & CUMMING, G. L. 1972. Seismic refraction studies in western Canada. *Can. J. Earth Sci.* **9**, 1099–109.

CHAPPLE, W. M. 1978. Mechanics of thin-skinned fold and thrust belts. *Bull. geol. Soc. Am.* **89**, 1189–98.

COOK, D. C. 1975. Structural style influenced by lithofacies, Rocky Mountain Main Ranges, Alberta-British Columbia. *Bull. geol. Surv. Can.* **233**.

CORBITT, L. L. & WOODWARD, L. A. 1973. Tectonic framework of cordilleran foldbelt in southwestern New Mexico. *Bull. Am. Assoc. Petrol. Geol.* **57**, 2207–16.

CUMMING, W. B., CLOWES, R. M. & ELLIS, R. M. 1979. Crustal structure from a seismic refraction profile across southern British Columbia. *Can. J. Earth Sci.* **16**, 1024–40.

DAHLSTROM, C. D. A. 1969*a*. The upper detachment in concentric folding. *Bull. Can. Pet. Geol.* **17**, 326–46.

—— 1969*b*. Balanced cross sections. *Can. J. Earth Sci.* **6**, 743–57.

—— 1970. Structural geology in the eastern margin of the Canadian Rocky mountains. *Bull. Can. Pet. Geol.* **18**, 332–406.

——, DANIEL, R. E. & HENDERSON, G. G. L. 1962. The Lewis thrust at Fording mountain, British Columbia. *J. Alberta Soc. petrol. Geol.* **10**, 373–95.

DAVIS, G. & CONEY, P. J. 1979. Geological development of the Cordilleran metamorphic core complexes. *Geology*, **7**, 120–4.

DE CSERNA, Z. 1971. Development and structure of the Sierra Madre Oriental of Mexico. *Geol. Soc. Am. (Abstracts with programs)*, **3**, *Rocky mountain section*, 377–78.

DEWEY, J. A. & BIRD, J. M. 1970. Mountain belts and the new global tectonics. *J. geophys. Res.* **75**, 2625–47.

DIMROTH, E. 1970. Evolution of the Labrador geosyncline. *Bull. geol. Soc. Am.* **81**, 2717–42.

DOUGLAS, R. J. W. 1950. Callum creek, Langford creek and Gap map-areas, Alberta. *Mem. geol. Surv. Can.* **255**.

—— 1951. Pincher creek, alberta; preliminary map. *Pap. geol. Surv. Can.* **51–22**.

—— 1952. Waterton Alberta; preliminary map. *Pap. geol. Surv. Can.* **52–10**.

—— 1958. Mount Head map-area, Alberta. *Mem. geol. Surv. Can.* **291**.

—— 1970. in Douglas, R. J. W., Gabrielse, H., Wheeler, J. O., Stott, D. F. & Belyea, H. R. Geology of Western Canada, Chap. VIII of Geology and Economic Minerals of Canada. *In:* Douglas, R. J. W. (ed), *Geol. Surv. Can. Econ. Geology report,* **1,** 396.

Dragert, H. & Clarke, G. K. C. 1976. A detailed investigation of the Canadian cordillera geomagnetic transition anomaly. *J. geophys.* **42,** 373–90.

Earth Physics Branch. 1974. Bouguer anomaly map of Canada. *Dept. Energy Mines & Res., Can. Gravity map series* **74–1.**

Eisabacher, G. H., Carrigy, M. A. & Campbell, R. B. 1974. Paleodrainage pattern and late-orogenic basins of the Canadian Cordillera. *In:* Dickinson, W. R. (ed). *Tectonics and Sedimentation.* Spec. Publ. Econ. Paleon. & Min. **22,** 143–66.

Elliot, D. 1976. The motion of thrust sheets. *J. geophys. Res.* **81,** 949–63.

Fermor, P. R. & Price, R. A. 1976. Imbricate structures in the Lewis thrust sheet around the Cate Creek and Haig Brook windows. *Pap. geol. Surv. Can.* **76-1B,** 7–10.

Fox, F. G. 1959. Structure and accumulation of hydrocarbons in southern Foothills, Alberta, Canada. *Bull. Am. Assoc. Petrol. Geol.* **43,** 992–1025.

Fyles, J. T. 1964. Geology of the Duncan lake area, Lardeau district, British Columbia. *Bull. Br. Columbia Dep. Mines Petrol. Res.* **49.**

—— 1967. Geology of the Ainsworth-Kaslo area, British Columbia. *Bull. Br. Columbia Dep. Mines Petrol. Res.* **53.**

—— & Hewlett, C. G. 1959. Stratigraphy and structure of the Salmo lead-zinc area. *Bull. Br. Columbia Dep. Mines.* **41.**

Gabrielse, H. & Reesor, J. E. 1974. The nature and setting of granitic plutons in the central and eastern parts of the Canadian Cordillera. *Pacific Geology.* **8,** 109–38.

Gardner, D. A. C. 1977. *Structural Geology and Metamorphism of Calcareous Lower Paleozoic Slates. Blaeberry River-Redburn Creek Area, near Golden, British Columbia.* Queen's Univ., Ph.D. thesis (Unpubl.).

Gilman, R. A. 1972. Geology of the Clachnacudainn salient near Albert canyon. British Columbia. *Can. J. Earth Sci.* 9, 1447–54.

Glover, J. K. 1978. *Geology of the Summit Creek Map Area, Southern Kootenay Arc, British Columbia.* Queen's Univ. Ph.D. thesis (Unpubl.).

Hamilton, W. 1977. Subduction in the Indonesian region. *In:* Talwani, M. & Pitman, W. C. (ed). *Island Arcs, Deep Sea Trenches and Back-Arc Basins.* Am. geophys. Union. Maurice Ewing series **1.** 15–31.

Harrison, J. E. 1972. Precambrian Belt basins of northwestern United States: its geometry, sedimentation, and copper occurrences. *Bull. geol. Soc. Am.* **83,** 1215–40.

—— & Jobin, D. A. 1963. Geology and ore deposits of the Clark Fork quadrangle, Idaho-Montana. *Bull. U.S. geol. Surv.* **1141-K.**

—— & —— 1965. Geologic map of the Packsaddle mountain quadrangle, Idaho. *U.S. geol. Surv. Map* **GQ-375.**

—— & Schmidt, P. W. 1971. Geologic map of the Elmira quadrangle, Bonner county, Idaho. *U.S. geol. Surv. Map* **I-603.**

Henderson, G. G. L. 1954. Geology of the Stanford range of the Rocky mountains. *Bull. Br. Columbia Dep. Mines* **35.**

Hoffman, P. 1973. Evolution of an early Proterozoic margin: the Coronation geosyncline and associated aulacogens of the northwestern Canadian Shield. *Philos. Trans. R. Soc. London* **A273,** 547–81.

Hoffman, P. F., St. Onge, M., Carmichael, D. M. & De Bie, I. 1978. Geology of the Coronation geosyncline (Aphebian) Hepburn lake sheet (86J). Bear province, district of Mackenzie; current research. *Pap. geol. Surv. Can.* **78-1A,** 147–51.

Höy, T. 1977. Stratigraphy and structure of the Kootenay arc in the Riondel area, southeastern British Columbia. *Can. J. Earth Sci.* **14,** 2301–15.

—— 1978. Geology of the Estella-Kootenay King area, southeastern British Columbia. *Br. Columbia Dep. Mines prelim. Map* **28.**

Huntting, M. T., Bennett, W. A. G., Livingston, V. E., Jr., & Moens, W. S. (Compilers). 1961. Geologic map of Washington. Washington Div. Mines and Geology.

Johns, W. M. 1970. Geology and mineral deposits of Lincoln and Flathead counties, Montana. *Bull. Montana Bur. Mines Geol.* **79.**

Kanasewich, E. R., Clowes, R. M. & McCloughan, C. H. 1969. A buried Precambrian rift in western Canada. *Tectonophysics,* **8,** 513–27.

Keating, L. F. 1966. Exploration in the Canadian Rockies and Foothills. *Can. J. Earth Sci.* **3,** 713–23.

Keen, C. E. & Hyndman, R. D. 1979. Geophysical review of the continental margins of eastern and western Canada. *Can. J. Earth Sci.* **16,** 712–47.

King, P. B. (Compiler) 1969. Tectonic map of North America. *U.S. geol. Surv.* Washington, D.C.

Leech, G. B. 1958. Fernie map-area, west half, British Columbia. *Pap. geol. Surv. Can.* **58-10.**

—— 1959. Canal Flats, British Columbia. *Geol. Surv. Can. Map* **24-1958.**

—— 1960. Fernie, west half, British Columbia. *Geol. Surv. Can. Map* **11-1960.**

—— 1962. Structure of the Bull river valley near latitude 49°35'. *J. Alberta Soc. petrol. Geol.* **10,** 396–407.

Lis, M. G. & Price, R. A. 1976. Large-scale block faulting during deposition of the Windermere supergroup (Hadrynian) in southeastern British Columbia. *Pap. geol. Surv. Can.,* **76-1A,** 135–6.

Little, H. W. 1960. Nelson map-area, west half, British Columbia. *Mem. geol. Surv. Can.* **308.**

McLean, J. R. & Jerzykiewicz, T. 1978. Cyclicity, tectonics and coal: Some aspects of fluvial sedimentology in the Brazeau-Paskapoo formations. Coal valley area. Alberta. Canada. *Mem. Can. Soc. petrol. geol.* **5**, *Fluvial Sedimentology*. 441–68.

Mereu. R. F.. Majumdar. S. C. & White. R. E. 1977. The structure of the crust and upper mantle under the highest ranges of the Canadian Rockies from a seismic refraction survey. *Can. J. Earth Sci.* **14**. 196–208.

Miller. F. K. 1974*a*. Preliminary geologic map of the Newport number 1 quadrangle. Pend Oreille county. Washington and Bonner county. Idaho. *Washington Dept. Natural Res. geol. Map* **GM-7.**

—— 1974*b*. Preliminary geologic map of the Newport number 2 quandrangle, Pend Oreille and Stevens counties. Washington. *Washington Dept. Natural Res. geol. Map* **GM-8.**

—— 1974*c*. Preliminary geologic map of the Newport number 3 quadrangle. Pend Oreille. Stevens and Spokane counties. Washington. Washington *Ddpt. Natural Res. geol. Map* **GM-9.**

—— 1974*d*. Preliminary geologic map of the Newport number 4 quadrangle. Spokane and Pend Oreille counties. Washington and Bonner county, Idaho. *Washington Dept. Natural Res. geol. Map* **GM-10.**

——, Clark. L. D. & Engels. J. C. D. 1975. Geology of the Chewelah-Loon Lake area. Stevens and Spokane countries. Washington. *Prof. Pap. U.S. geol. Surv.* **806.**

Monger. J. W. H. & Price. R. A. 1979. Geodynamic evolution of the Canadian Cordillera—progress and problems. *Can. J. Earth Sci.* **16**, 770–91.

——. Souther. J. G. & Gabrielse. H. 1972. Evolution of the Canadian Cordillera: a plate-tectonic model. *Am. J. Sci.* **272**, 577–602.

Nguyen. E. K., Sinclair. A. J. & Libby. W. G. 1968. Age of the northern part of the Nelson batholith. *Can. J. Earth Sci.* **5**, 955–7.

North. F. K. & Henderson. G. G. L. 1954. Summary of the geology of the southern Rocky Mountains of Canada. *Alberta Soc. petrol. Geol., Calgary, Guidebook, Fourth Annual Field Conference.* 15–81.

Norris. D. K. 1958*a*. Beehive mountain. Alberta and British Columbia. *Pap. geol. Surv. Can.* **58-5.**

—— 1958*b*. Livingstone river. Alberta. *Geol. Surv. Can. Map* **5–1958.**

—— & Price. R. A. 1966. Middle Cambrian lithostratigraphy of southeastern Canadian Cordillera. *Bull. Can. Pet. Geol.* **14**, 385–404.

Nye. J. F. 1952. The mechanics of glacier flow. *J. Glaciol. London.* **2**, 82–93.

Ollerenshaw. N. C. 1972*a*. Geology. lake Minnewanka (east half). Alberta *Geol. Surv. Can. Map* **1347A.**

—— 1972*b*. Geology. Wildcat Hills (west half). Alberta. *Geol. Surv. Can. Map* **1351A.**

—— (Compiler). 1975. Rocky mountain Foothills and Front ranges: Calgary region. In guidebook, Savanna creek to Panther river, exploration update W5. *Can. Soc. Pet. Geol.*, Calgary.

—— (Compiler), 1978. Geology. Calgary. Alberta-British columbia. *Geol. Surv. Can. Map* **1457A.**

Park. C. F.. Jr & Cannon. R. S.. Jr. 1943. Geology and ore deposits of the Metaline quadrangle, Washington. *Prof. Pap. U.S. geol. Surv.*. **202.**

Price. R. A. 1962. Fernie map-area, east half Alberta and British Columbia. *Pap. geol. Surv. Can.* **61-24.** (includes *Map* **35-1961).**

—— 1964*a*. The Precambrian Purcell System in the Rocky mountains of southern Alberta and British Columbia. *Bull. Can. Pet. Geol.* **12**, 399–426.

—— 1964*b*. Flexural-slip folds in the Rocky Mountains, southern Alberta and British Columbia. Seminars on Tectonics IV, Dept. Geol. Sci., Queen's Univ., Kingston, Ont. (Also as Geol. Surv. Can. reprint 78).

—— 1965. Flathead map-area. British Columbia and Alberta. *Mem. geol. Surv. Can.* **336.**

—— 1970. Geology, Canmore (east half) Alberta. *Geol Surv. Can. Map* **1265A.**

—— 1972. Rocky Mountain Trench—Golden to Cranbrook. *In:* Price. R. A.. Balkwill. H. R., Charlesworth.. H. A. K., Cook. D. G. & Simony. P. S. *The Canadian Rockies and Tectonic Evolution of the Southeastern Cordillera.* Guidebook, field excursion AC15, XXIV Inter. Geol. Congress. Montreal, 77–85.

—— 1973. Large-scale gravitational flow of supracrustal rocks. southern Canadian Rockies. *In:* de Jong. K. A. & Scholten. R. (ed). *Gravity and tectonics.* Wiley, New York. 491–502.

—— 1979. The Selkirk fan structure of the southeastern Canadian Cordillera: Discussion. *Bull. geol. Soc. Am.* **90**, 695–8.

——, Balkwill H. R. & Mountjoy, E. W. in press. Geology, McMurdo (east half), British Columbia. *Geol. Surv. Can. Map* **1501A.**

——, Cook, D. G., Aitken, J. D. & Mountjoy, E. W. in press *a*. Geology, Lake Louise, (west half), Alberta-British Columbia. *Geol. Surv. Can. Map* **1483A.**

——, ——, —— & —— in press *b*. Geology, Lake Louise, (east half), Alberta-British Columbia. *Geol. surv. Can. Map* **1482A.**

—— & Mountjoy. E. W. 1970. Geologic structure of the Canadian Rocky Mountains between Bow and Athabasca revers—a progress report. *In:* Wheeler. J. O. (ed), *Structure of the Southern Canadain Cordillera.* Spec. Pap. geol. Assoc. Can. **6**, 7–25.

—— & —— 1972*a*. Geology. Banff. (east half). Alberta-British Columbia. *Geol. Surv. Can. Map* **1294A.**

—— & —— 1972*b*. Geology, Banff, (west half), Alberta-British Columbia. *Geol. Surv. Can. Map* **1295A.**

—— & —— 1978*a*. Geology, Hector Lake. (east half), Alberta-British Columbia. *Geol. Surv. Can. Map* **1463A.**

—— & —— 1978*b*. Geology, Hector Lake, (west half), Alberta-British Columbia. *Geol. Surv.*

Can. Map **1464A.**

—— & —— 1978*c*. Geology Siffleur River, (east half), Alberta. *Geol. Surv. Can. Map* **1465A.**

—— & —— 1978*d*. Geology, Siffleur River, (west half), Alberta. *Geol. Surv. Can. Map* **1466A.**

—— & —— 1979*a*. Geology, Mount Goodsir, (east half), Alberta-British Columbia. *Geol. Surv. Can. Map* **1476A.**

—— & —— 1979*b*. Geology, Mount Goodsir, (west half), Alberta-British Columbia. *Geol. Surv. Can. Map* **1477A.**

—— & in press. Geology, McMurdo, (west half). British Columbia. *Geol. Surv. Can. Map* **1502A.**

READ, P. B. 1977. Relationship of the Kootenay arc to the Shuswap Metamorphic Complex, Southern British Columbia. *Geol. Assoc. Can. Progs with Abstr.* **2,** 43.

—— & WHEELER, J. O. 1976. Geology of Lardeau west-half map area. *Geol. Surv. Can. open-file Report* **288.**

REESOR, J. E. 1965. Valhalla gneiss complex, British Columbia. *Bull. geol. Surv. Can.* **129.**

—— 1973. Geology of the Lardeau map-area, east half, British Columbia. *Mem. geol. Surv. Can.* **369.**

RICE, H. M. A. 1941. Nelson map-area, east half, British Columbia. *Mem. geol. Surv. Can.* **228.**

RODGERS, J. 1970. *The Tectonics of the Appalachians.* Wiley-Interscience, New York.

ROSS, C. P. 1959. Geology of Glacier National Park and the Flathead region, northwestern Montana. *Prof. Pap. U.S. geol. Surv.* **296.**

SEARS, J. W. 1979. *Tectonic Contrasts between the Infrastructure and Suprastructure of the Columbian Orogen, Albert Peak Area, Western Selkirk Mountains, British Columbia.* Queen's Univ., Ph.D. thesis (Unpubl.).

—— & PRICE, R. A. 1978. The Siberian connection: A case for Precambrian separation of the North American and Siberian cratons. *Geology,* **6,** 267–70.

SEELEY, D. R. 1977. The significance of landward vergence and oblique structural trends on trench inner slopes. *In:* TALWANI, M. & PITMAN, W. C. (eds), *Island Arcs, Deep Sea Trenches and Back Arc Basins.* Am. geophys. Union. Maurice Ewing Series **1,** 187–98.

SHAW, E. W. 1963. Canadian Rockies—orientation in time and space. *In:* CHILDS, O.E. (ed), *Backbone of the Americas.* Mem. Am. Assoc. Petrol. Geol. **2,** 231–42.

SIMONY, P. S. & WIND, G. 1970. Structure of the Dogtooth Range and adjacent portions of the Rocky Mountain Trench. *In:* WHEELER, J. O. (ed), *Structure of the Southern Canadian Cordillera.* Spec. Pap. geol. Assoc. Can. **6,** 41–53.

SPENCE, G. D., CLOWES, R. M. & ELLIS, R. M. 1977. Depth limits of the M discontinuity in the southern Rocky Mountain Trench, Canada. *Bull. seismol. Soc. Am.* **67,** 543–6.

STACEY, R. A. 1973. Gravity anomalies, crustal structure, and plate tectonics in the Canadian Cordillera. *Can. J. Earth Sci.* **10,** 615–28.

THOMPSON, R. I. 1972. *Geology of the Akolkolex River Area near Revelstoke, British Columbia,* Queen's Univ. Ph.D. thesis (Unpubl.).

WANLESS, R. K., LOVERIDGE, W. D. & MURSKY, G. 1968. A geochronological study of the White Creek batholith, southeastern British Columbia. *Can. J. Earth Sci.* **5,** 375–86.

WHEELER, J. O. 1963. Rogers pass map-area, British Columbia and Alberta. *Pap. geol. Surv. Can.,* **62–32.**

—— & GABRIELSE, H. 1972. The Cordilleran Structural Province. *In:* PRICE R. A. & DOUGLAS, R. J. W. (eds), *Variations in Tectonic Styles in Canada*; Spec. Pap. Geol. Assoc. Can., **11,** 1–81.

——, CHARLESWORTH, H. A. K., MONGER, J. W. H., MULLER, J. A., PRICE, R. A., REESOR, J. E., RODDICK, J. A. & SIMONY, P. S. 1974. Western Canada. *In:* SPENCER, A. M. (ed), *Mesozoic-Cenozoic orogenic belts.* Spec. Publ. geol. Soc. London. **4,** 592–623.

WILLIAMS, E. P. 1949. Preliminary map, Cardston, Alberta. *Pap. geol. Surv. Can.* **49–3.**

WYNN, J. C., KLEINKOPF, M. F. & HARRISON, J. E. 1977. Audio-frequency magnetotelluric and gravity traverse across the crest of the Purcell anticlinorium, northwestern Montana. *Geology,* **5,** 309–12.

YATES, R. G. 1964. Geologic map and sections of the Deep Creek area, Stevens and Pend Oreille counties, Washington. *U.S. geol. Surv. Map* **I-412.**

—— 1971. Geologic map of the Northport quadrangle, Washington. *U.S. geol. Surv. Map* **1-603.**

YOUNG, F. G., CAMPBELL, R. B. & POULTON, T. P. 1973. The Windermere Supergroup of the southeastern Canadian Rockies. *Belt Symposium* **1,** 181–203.

ZWANZIG, H. V. 1973. *Structural Transitions between the Foreland Zone and the Core Zone of the Columbian Orogen, Selkirk Mountains, British Columbia.* Queen's Univ. Ph.D. thesis (Unpubl.).

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