BLOOM

The role of the Shuswap Metamorphic Complex in Cordilleran tectonism: a review

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The Shuswap Metamorphic Complex consists of three parts, each with unique stratigraphy and orogenic evolution, separated by major faults of diverse nature and having in common only a post-late Mesozoic tectonic history. The first part, the Monashee Complex, is a possible extension of the Precambrian Shield that contains limited evidence of Mesozoic orogenesis and that was rapidly uplifted during the Cretaceous to Paleogene. The Monashee Décollement, a warped mylonite zone interpreted as a regional thrust fault active through the Middle Jurassic, separates this complex from the second part, which contains rocks correlative with Hadrynian to late Paleozoic strata of the pericratonic prism. The third part, the Okanagan Complex, straddling the 49th Parallel from the Okanagan Valley to Kootenay Lake, contains the probable exhumed roots of a Mesozoic magmatic arc built upon possible North American continental and transitional crust and includes late Paleozoic and early Mesozoic suspect terranes.

The Columbian Orogen formed during westward drift of the craton into a continent of accreted elements. Response of the craton, attenuated during at least two episodes of Proterozoic rifting, and its overlying sedimentary prism to underthrusting from the east and simultaneous collision with an accreting collage from the west took place in two stages. First, attenuated crust was telescoped and thickened to its approximate original configuration while westernmost parts of the bordering prism were deformed and metamorphosed (the Jura-Cretaceous Columbian Orogeny that affected the Okanagan Complex and strata above the Monashee Décollement). Second, the thickened crust, the deformed prism, and platformal strata were thrust eastward (the Late Cretaceous – Paleocene Laramide Orogeny that formed the Rocky Mountains Thrust Belt). Waning convergent tectonism led to ascendancy of crustal extension (primarily in the Okanagan Complex) and final uplift of cratonic massifs.

Le complexe métamorphique de Shuswap comprend trois parties dont chacune se distingue par sa propre stratigraphie et évolution orogénique, elles sont séparées par des failles majeures de nature différente et elles n'ont en commun que leur histoire tectonique du post-Mésozoïque supérieur. La première partie, le complexe de Monashee, semble être une extension du Bouclier précambrien montrant quelques indices d'une orogénèse mésozoïque, soulevée rapidement durant le laps de temps compris entre le Crétacé et le Paléocène. Le décollement de Monashee, une zone de mylonite déformée, interprétée comme une faille de chevauchement qui fut active durant tout le Jurassique moyen, sépare ce complexe de la deuxième partie constituée de roches corrélées avec les strates du prisme péricratonique de l'Hadrynien jusqu'au Paléozoïque supérieur. La troisième partie, le complexe d'Okanagan, est à cheval sur le 49ième parallèle de la vallée d'Okanagan jusqu'au lac Kootenay, elle est formée d'affleurements qui représentent possiblement les racines d'un arc magmatique mésozoïque construit sur ce qui pourrait être une croute continentale et transitionnelle et incluant des terranes suspectes du Paléozoïque supérieur et du Mésozoïque inférieur.

L'orogénèse du Columbien s'est développée lors de la dérive du craton vers l'ouest en un continent constitué par une accrétion d'éléments. La réponse du craton, atténuée durant au moins deux épisodes de rifting protérozoïque, et son prisme sédimentaire sus-jacent sous-charrié de l'est et la collision simultanée résultant en une accrétion par collage d'éléments de l'ouest, ont survenu en deux étapes. Premièrement, la croûte atténuée fut télescopée et l'épaisseur augmenta jusqu'à reconstituer approximativement sa puissance originale tandis que les parties les plus occidentales du prisme en bordure furent déformées et métamorphisées (l'orogénèse du Columbien au Jura-Crétacé a affecté le complexe d'Okanagan et les strates recouvrant le décollement de Monashee). Deuxièmement, la croûte épaissie, le prisme déformé ainsi que les strates de la plate-forme ont été charriés vers l'est (l'orogénèse de Laramide au Crétacé supérieur – Paléocène qui a formé la zone de charriage des Montagnes Rocheuses). La réduction du tectonisme convergent a favorisé l'ascension de l'extension crustale (surtout dans le complexe d'Okanagan) et le soulèvement final des massifs cratoniques.

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Introduction

The Shuswap Metamorphic Complex of southeastern British Columbia has been studied for a century. During this time it has not been clearly or consistently defined, metamorphic rocks in the region have been variously included within it or excluded from it, and numerous overly generalized and conflicting hypotheses have been proposed to explain its origin and evolution. This paper attempts to resolve conflicting interpretations and to place the complex within the tectonostratigraphic framework of this part of the Cordillera. A brief historical review shows how earlier studies have influenced present thinking and how the implications of new data have not been adequately assessed. This review is followed by a description and definition of the complex; then recent hypotheses regarding its role in Cordilleran tectonism are examined.

Review of geological investigations

In British Columbia, metamorphic rocks in the vicinity of Shuswap Lake (Fig. 1) were first mapped by Dawson (1879) who in 1898 named them the Shuswap Series and suggested that they were correlatives of Archean rocks of the Canadian Shield. Adjacent rocks of lower metamorphic grade were assigned to the Nisconlith and Adams Lake series of possible Cambrian age. Other workers (Daly 1915; Brock 1934) and, in other areas, Dawson himself, added granitoid gneiss and metasediments near Okanagan, Kootenay, and Upper and Lower Arrow lakes to the Shuswap Series. Because no discrete age could be attributed to these rocks, Daly (1915) and Brock (1934) adopted the term Shuswap Terrane, which embraced Dawson's Nisconlith and Adams Lake series, and suggested that most of the strata of the terrane were "Beltian" (Pro-

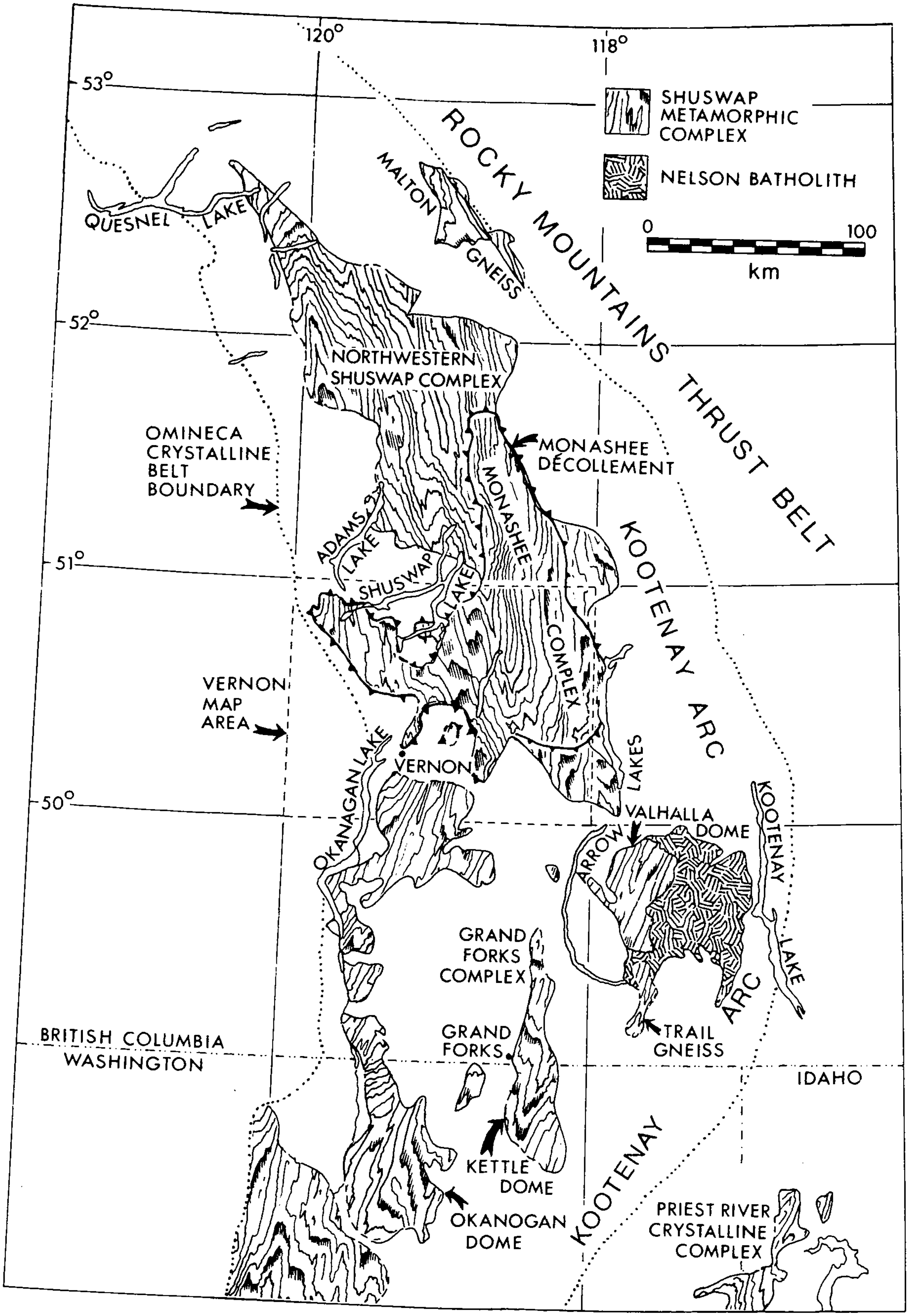

FIG. 1. The Shuswap Metamorphic Complex: geographic and major geologic features. Patterns do not reflect structural trends.

terozoic) and possibly Paleozoic in age.

Although Daly (1915) favoured a Precambrian age for their metamorphism, Brock (1934) and Cairnes (1939) stressed the close association of metasediments and Mesozoic granitic rocks. Brock (1903) first suggested correlation of the metamorphic rocks between Okanagan and Kootenay lakes with the Shuswap "Series" but also left open the possibility that they were metamorphosed equivalents of late Paleozoic and Mesozoic strata common to the region. This possibility was substantiated by Daly (1912) and Cairnes (1939) who, after observing apparent continuity of these rocks with those of the Shuswap Terrane in the region east of Vernon (Fig. 1) (Cairnes 1932), combined both of Brock's (1903) alternatives and applied this evidence of Mesozoic contact metamorphism to the whole terrane.

Jones' (1959) detailed observations of lithology, structure, and metamorphism in the Vernon map area led him to divide the terrane into three groups (Monashee, Mount Ida, and Chapperon) and return to Dawson's (1898) original proposals for a probable pre-Beltian or Archean age for strata and orogenesis of the terrane.

During the next 10 years several detailed studies (Reesor 1965; Wheeler 1965; Hyndman 1968; Ross 1968; Fyles 1970; McMillan 1970; Reesor and Moore 1971—summarized and discussed by Wheeler 1970) led to an informal and inconsistently applied definition of the terrane as a metamorphic complex bounded for the most part by the sillimanite isograd and containing rocks ranging in age from Proterozoic to Triassic. These were divided into granitoid core gneiss exposed in structural culminations (Frenchman Cap, Thor–Odin and Valhalla domes) and mantling paragneiss (Reesor 1970). Evidence obtained from areas adjacent to the complex indicated that metamorphism and deformation were of Jurassic age. The marked disparities in structural style between the Shuswap Complex and nearby strata were attributed to varying regional metamorphism within infra- and suprastructure (*cf.* Fyson 1970; Reesor 1970) rather than to episodes of folding widely separated in time (e.g., Jones 1959).

In southern parts of the complex, Little (1957, 1960, 1961) correlated paragneiss near and within the Jurassic Nelson batholith with Mesozoic strata on the basis of mappable transitions in grade but included paragneiss and granitoid gneiss between Okanagan and Lower Arrow lakes in the Precambrian Monashee Group (Jones 1959) because they lay near rocks so designated to the north. Western gneissic parts of the Nelson batholith (Little 1960) were remapped by Reesor (1965) who included them in the Shuswap Complex (Valhalla Dome). Some granitoid gneiss units were suggested by him to be related to the Nelson Batholith, implying a Mesozoic age of metamorphism. Preto (1969) investigated metamorphic rocks near Grand Forks and suggested that the paragneissic units may be correlative with Proterozoic and Paleozoic strata to the east.

Metamorphic rocks northwest of the main mass of the complex were examined near Shuswap Lake by Dawson (1898), Daly (1915), and Jones (1959). Reconnaissance mapping (Campbell 1963, 1967, 1978) revealed the extent of rocks within the sillimanite isograd and led to documentation of a metamorphic and structural transition from infrastructure to suprastructure within strata of Proterozoic to Triassic age (Campbell 1970, 1973) east of Quesnel Lake.

In keeping with geologists' favourite parable, that of the six blind men examining an elephant, the preceding apparently incompatible ideas seem to be *all* correct. Not one, however, can be applied to the whole of the complex. A sort of Peter Principle seems to apply to these and likely all geological hypotheses: they should not be promoted beyond areas where they are competent to explain the facts.

Studies during the past 10 years have concentrated on clarifying relationships of the Shuswap Complex with adjacent strata and establishing the ages of its constituents and the timing of episodes of deformation and metamorphism. These studies have made it possible to better describe this complex elephant.

Two key findings have been instrumental in enhancing our understanding of the nature of the complex. Firstly, radiometric dating (Wanless and Reesor 1975; Okulitch *et al.* 1981; Parrish and Armstrong 1983; Duncan 1984) indicated that the east-central part of the complex contains a core of paragneiss intruded by gneissic plutons 2000–2100 Ma old, mantled by unconformably overlying metasediments between 2000 and at least 770 Ma old. The mantling strata may be correlative with the Belt–Purcell Supergroup (1600–1350 Ma, McMechan and Price 1982) or older (Brown and Read 1983). Secondly, meticulous field mapping (Read and Brown 1981) delineated a tectonic bounding surface to this ancient part of the complex (Fig. 2). Called the Monashee Décollement, it is a zone of mylonite up to 1 km thick that formed in the Jurassic and extends along the eastern margin of the complex and passes westward into it. Read and Brown (1981) designated the ancient part of the complex below the décollement the Monashee Complex after the Monashee Group with which it largely coincides. Dawson's (1898) and Jones' (1959) suggestions of a "pre-Beltian" age to Shuswap rocks in the type area were proven at least partly correct.

Several detailed studies (e.g., Fletcher 1972; Pigage 1978; Simony *et al.* 1980; Pell and Simony 1982; Raeside and Simony 1983) in northwestern parts of the Shuswap Complex that lie structurally above the Monashee Décollement have confirmed the presence of a transition from infra- to suprastructure as described by Campbell (1970, 1973). Remapping of low- to medium-grade strata near Shuswap and Adams lakes, formerly of the Nisconlith and Adams Lake series (Dawson 1898) and the Mount Ida Group (Jones 1959), has indicated the likelihood of correlation of most units with Paleozoic successions lying east of the Monashee Complex in the Kootenay Arc and to the north in the Quesnel Lake region (Okulitch 1979*a*). These strata above the Monashee Décollement share a common hierarchy of structures (polyphase, predominantly coaxial, northwest-trending, recumbent and upright folds cut by thrust faults) that formed before, during, and after Jurassic metamorphism. The hypotheses summarized by Wheeler (1970) apply admirably to this part of the Shuswap Complex.

The region between Okanagan and Kootenay lakes contains a variety of metamorphic rocks, numerous plutons, and sedimentary and volcanic rocks predominantly of late Paleozoic, Triassic, Jurassic, and Paleogene ages. Ryan (1973) showed that late Paleozoic strata east of the Okanagan Valley near the 49th Parallel can be traced through a narrow aureole formed by Mesozoic plutons into high-grade equivalents previously included in the Monashee Group (Little 1961). Simony (1979) demonstrated that Late Carboniferous rocks can be followed across isograds into gneissic units of Valhalla Dome (Reesor 1965). Parrish (1981) suggested on the basis of lithologic correlation and Rb–Sr isotopic ages that the north flank of Valhalla Dome contains Paleozoic and Triassic strata. The core of the dome experienced intrusion and metamorphism during

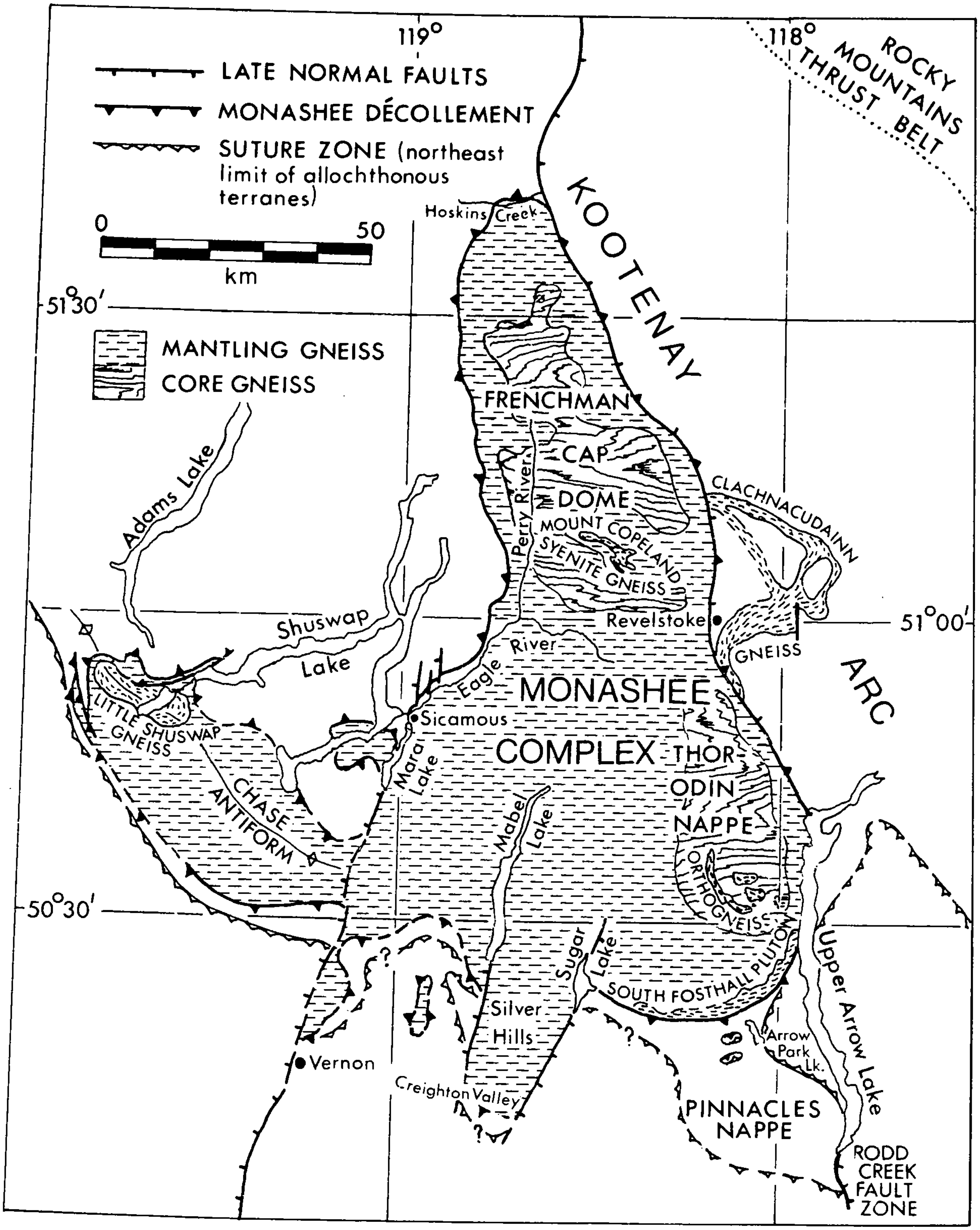

FIG. 2. The Monashee Complex: structural and plutonic components.

the Mesozoic and Paleogene (Parrish 1983). In the region south of the type area of the Shuswap Complex, therefore, ample evidence exists to support Cairnes' (1939) hypothesis of its development.

A definition

The Shuswap Complex is not readily definable; it includes rocks of plutonic, sedimentary, and volcanic parentage predominantly within the sillimanite isograd ranging in age from possibly Archean to earliest Jurassic. The term is synonymous with the core zone of the Columbian Orogen in the southern Canadian Cordillera, a part of the Omineca Crystalline Belt. The complex can be divided into three main parts, separated by two major fault zones, and numerous smaller features. The first part is the Monashee Complex (Read and Brown 1981). The second part of the Shuswap Complex lies above the Monashee

Décollement within the sillimanite isograd.

Metamorphic rocks of the southern region between Okanagan and Kootenay lakes pose a special problem. None have been established as part of the Monashee Complex and only some are correlatives of Proterozoic and Paleozoic strata to the east and north that overlie the Monashee Décollement. Many may be allochthonous relative to the North American craton and its bordering sedimentary prism. Foliated and gneissic granitoids, subordinate in northern parts of the Shuswap Complex, here are plentiful. Because of the heterogeneous nature of these units and because they cannot readily be affiliated with other parts of the Shuswap Complex at this time, it is provisionally proposed to group them in a third part, the Okanagan Plutonic and Metamorphic Complex (Okulitch 1979*a*). This term is considered useful only for general discussion and must in time be discarded as more specific information about these rocks becomes available. In the same way, the term Shuswap Complex should also pass out of use.

Part one: The Monashee Complex

The radiometric studies mentioned previously have established the existence of Archean(?) and Proterozoic plutonic and presumably depositional and (or) metamorphic events within the Monashee Complex (Fig. 2). Paragneiss of the core zone, 2800(?)–2200 Ma old, was intruded by granitoid plutons 2100–1960 Ma ago. Extensive migmatization of the core may have occurred at about 750 Ma. The core is unconformably overlain by a basal quartzite conglomerate unit followed by pelitic, psammitic, and calc-silicate gneiss, in turn overlain by thick psammitic gneiss (Brown and Psutka 1979; Höy and McMillan 1979; Brown 1980; Read 1980). These units, possibly correlative with either lower Paleozoic strata to the east (Höy and McMillan 1979; Duncan 1984), Hadrynian or Helikian successions (Read 1980), or perhaps as old as Aphebian (ca. 2000 Ma) (Brown and Read 1983), were intruded by the Mount Copeland syenite (Fig. 2), which may be as old as 770 Ma (Okulitch *et al.* 1981). Little is known about western parts of the complex. Nielsen (1982) has suggested that metasediments southwest of Shuswap Lake (Silver Creek Formation, Okulitch 1979*a*) may be lower grade equivalents of adjacent units of the complex.

Structural and stratigraphic mapping (see above and Read 1979; Journeay 1983) has allowed delineation of a hierarchy of major and minor structures. This consists of an earliest observed phase of recumbent isoclinal folds of variable but predominantly northerly trend that have been pervasively deformed during development of a prominent set of easterly trending nappes, thrust slices, and recumbent isoclines. Flattening and extension parallel and perpendicular to axes of both sets of folds accompanied folding (Jones 1959, pp. 107–116; Duncan 1984, pp. 121–122). Pronounced uniaxial stretching apparently did not occur, unless a prominent east–west stretching lineation, subparallel to phase-two fold axes, formed at this time. Second-phase folding accompanied by high-grade metamorphism did produce an east–west lineation, but analysis by Shore and Duncan (1983) indicated that this lineation is not associated with stretching, contrary to unsupported assertions by Mattauer *et al.* (1983). Structures in the complex described by Reesor and Moore (1971) near Thor–Odin Dome were interpreted as having formed during diapirism. Read (1979, 1980) and Duncan (1984) have shown that two episodes of isoclinal folding followed by broad warping produced domed nappes by interference. This major structure is therefore now called the "Thor–Odin Nappe." Brown (1978) demonstrated a similar evolution for Frenchman Cap Dome.

The timing of these deformational episodes remains uncertain. Originally they were considered to be Precambrian, but upon acceptance of correlations of Monashee Complex strata with Hadrynian and younger successions to the east it became necessary to include them in the Jura-Cretaceous Columbian Orogeny. The enlightening radiometric data that indicated the probable Precambrian ages of units and possible timing of thermal events in the Monashee Complex, and the discovery of the Monashee Décollement, which separates the complex from younger rocks, require reevaluation of the assumed Mesozoic age of deformation in the complex.

It is necessary to be very specific as to which episodes of folding are being discussed. The Monashee Décollement began to form before or during the onset of high-grade metamorphism in the hanging wall and heat from the hanging wall affected parts of the décollement mylonite and subjacent parts of the Monashee Complex (Journeay 1983). Movement continued in the interval 164–157 Ma, that is, after metamorphism and second-phase deformation in the Kootenay Arc east of the complex and before third-phase folding that affected the décollement, the complex, and Kootenay Arc strata (Read and Brown 1981; Brown and Read 1983). First- and second-phase structures and metamorphic isograds above and below the décollement are truncated. The recumbent isoclines of the complex cut by the décollement are those whose age is controversial.

Direct evidence for *any* age for these recumbent structures is lacking. If they formed at the same time as early structures in the hanging wall and during early movement on the Monashee Décollement, as suggested by Brown and Read (1983) and Journeay (1983), they must be of Early to Middle Jurassic age. However, no radiometric data from the complex indicate the effects of a thermal event of Mesozoic age. Application of radiometric data from younger strata of the Kootenay Arc across the décollement into the complex is indefensible, as the relative positions of the complex and those strata at the time of the latter's deformation are unknown. The diapiric mechanisms proposed by Reesor and Moore (1971) that had been linked with compression in the Rocky Mountains Thrust Belt (Price and Mountjoy 1970) have been refuted by later work (Brown 1978; Read 1979; Simony *et al.* 1980; Price 1981; Duncan 1984). The strongest arguments linking structures above the décollement with those of the complex rest upon the apparent transition in structural style and orientation across the décollement. In the décollement mylonite zone, the predominantly northwesterly trending, polyphase, largely coaxial fabric in the hanging wall may have been progressively rotated into an easterly orientation approximately parallel with the pervasive easterly trending second-phase lineation of the complex. Rare sheath folds and dismembered metamorphic minerals in the hanging wall, mylonite, and footwall testify to intense stretching in this zone (Journeay 1983). This part of the Shuswap Complex is viewed as an infrastructural ductile shear zone that includes the polyphase isoclines of the Monashee Complex, the mylonite, and high-grade, polydeformed strata above the décollement. Above is a suprastructure only locally affected by underlying deformation. Post-metamorphic movement on the décollement and deeper faults is thought to have formed a duplex of crustal dimensions (R. L. Brown and M. Journeay, personal communication, 1984). This persuasive hypothesis, which finds support in the detailed studies (cited above) within

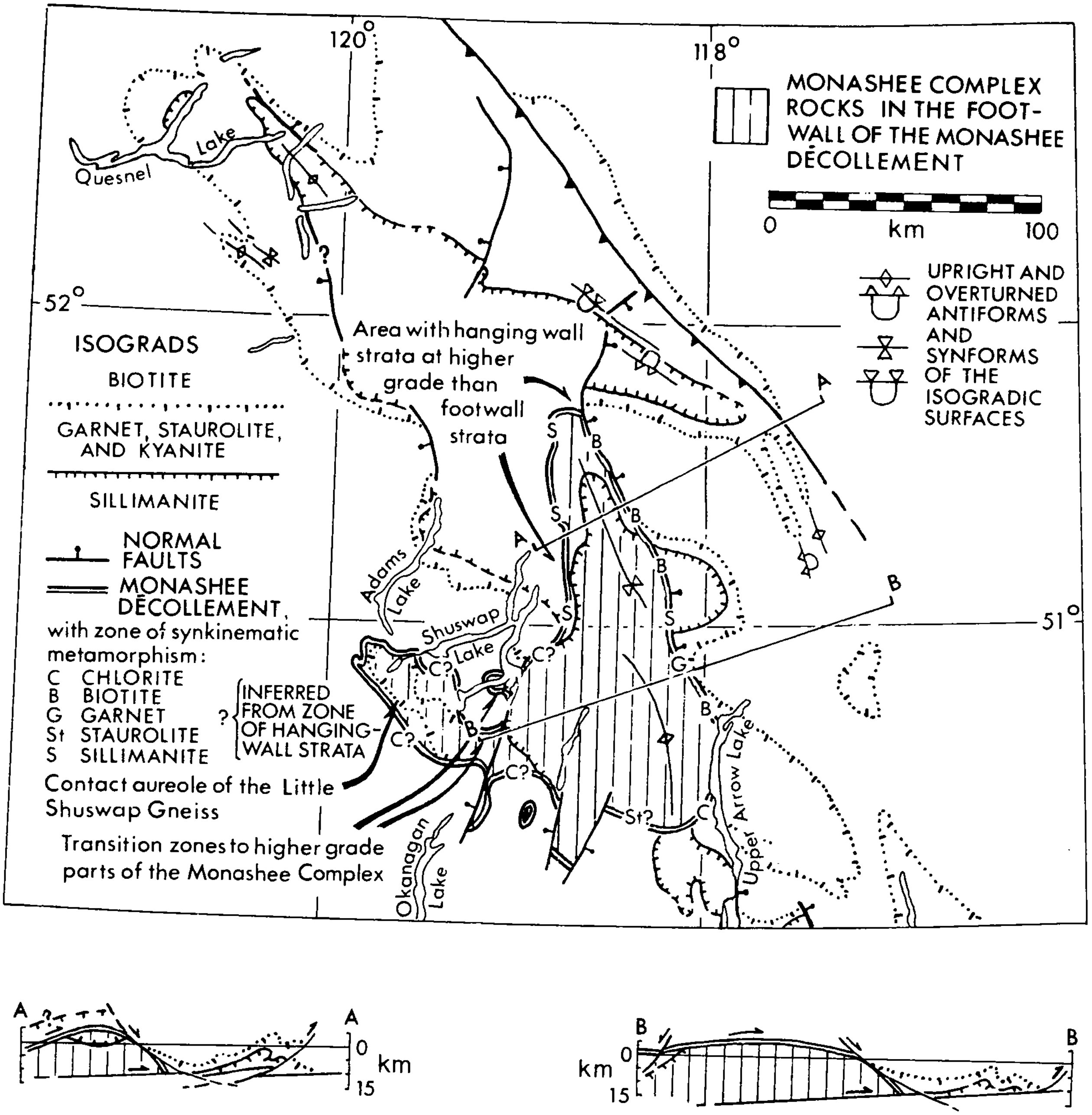

FIG. 3. Patterns of metamorphism in the Monashee Complex, the Monashee Décollement, and the hanging wall of the décollement. Information primarily from Wheeler *et al.* (1972) with additions from Leatherbarrow and Brown (1978), Okulitch (1979*a*), Read and Brown (1981), Ghent *et al.* (1981), Brown and Murphy (1982), Journeay (1983), and L. Lane (personal communication, 1984).

and near the décollement around Frenchman Cap Dome, has the virtue of providing an apparently unified kinematic model for the core zone and foreland of the Columbian Orogen. Examination of data from a broader perspective introduces some inconsistencies.

Integration of the differing hierarchies of structures above and below the Monashee Décollement requires the presence of different stress fields or differing metamorphic conditions or both in the two structural levels. Tectonic models invoking different stress fields remain speculative, albeit possible. Differing metamorphic conditions are present along much of both the east and west sides of the Monashee Complex (Fig. 3), but no zone of transition is preserved. In two areas, the west flank of Frenchman Cap Dome and north of Revelstoke (Fig. 2), younger hanging-wall strata were at higher temperatures than adjacent rocks in the complex (Brown 1980; Journeay 1983; L. Lane, personal communication, 1984). Such juxtaposition is likely the result of movement of hanging-wall strata from a region of higher temperature onto the complex. Metamorphic grade seems to vary independently within the complex and within younger hanging-wall strata without regard to the distinctly different hierarchies of structures found in each. No obvious reason derived from differences in metamorphic conditions seems to exist for the development of northerly verging folds below the décollement (Read and Klepacki 1981; Duncan 1984) and east-directed shear and compression above it, if all these features formed simultaneously.

The seductive parallelism of linear fabric in the complex and the décollement is of no independent value for their temporal correlation (Park 1969); however, control of structures by preexisting anisotropy is a well documented phenomenon. Moreover, metamorphic minerals that can be linked to extension and shearing within the décollement have been shown to overgrow an earlier subparallel mineral lineation that may be related synkinematically to either second-phase isoclines in the complex or to earliest movement on the décollement or to both. The inferred timing of this earlier mineral growth is dependent upon which relationship is favoured and bears significantly on metamorphic conditions in the décollement mylonite and the age of early structures in the Monashee Complex.

Rotation of earlier structures within a zone of ductile flow such as the décollement mylonite is also an expected phenomenon. The preservation of structures deeper within the complex at high angles to shear in the mylonite suggests that such shear was restricted to it and that the décollement is a discrete structure and not a zone of transition between the two hierarchies of structure above and below it.

Further difficulties may arise when palinspastic restoration of hanging-wall and footwall strata is attempted. Read and Brown (1981, Fig. 9) clearly illustrated the inconsistencies encountered if the décollement is interpreted solely as a normal fault with eastward and downward displacement of hanging-wall strata of 15–20 km (the minimum required to attain a match of isograds and stratigraphic successions) relative to the Monashee Complex. Mapping of the décollement around the north, northwest, and south ends of the complex indicates that it must be a warped, gently dipping surface. Read and Brown (1981) therefore suggested a minimum of 80 km eastward displacement (the exposed width of the Monashee Complex) in an attempt to explain juxtaposition of chlorite to sillimanite zone rocks against those bearing sillimanite–K-feldspar assemblages. Subsequent mylonite fabric studies confirmed the west over east thrust nature of the décollement (Brown and Murphy 1982; Journeay 1983).

As long as the décollement is viewed as a warped thrust fault, however, the restoration problem is not necessarily solved by increasing the relative displacement because, with the exception of areas within the sillimanite zone northeast of the north end of the complex and southeast of the complex at Kootenay Lake (Fig. 3), rocks of progressively lower grade lie farther east. Although it is possible to restore subsurface extensions of these metamorphic culminations over northern parts of the complex utilizing known or inferred directions of tectonic transport, substantial portions of the complex (particularly near Thor–Odin Nappe) are left without the requisite cover of high-grade strata.

The variations of Mesozoic metamorphism above and within the décollement (Fig. 3) are difficult to link consistently with the broad expanse of high-grade gneiss of the Monashee Complex. Nonetheless, it has been clearly demonstrated that in the Frenchman Cap Dome area inverted isograds in the complex were most likely produced by the tectonic superposition of hot hanging-wall strata (Journeay 1983). By this mechanism at least parts of the complex were affected by Mesozoic metamorphism. Apparent restriction of inverted isograds to areas bordered by mylonite whose mineral assemblages indicate movement at considerable depth and high temperature suggests that where adjacent mylonite and hanging-wall strata were relatively cool, Mesozoic metamorphism was not imposed upon the complex. The relation between the north-verging, phase-two nappes found throughout the complex to this metamorphism therefore remains to be demonstrated.

The assumption of pervasive Mesozoic metamorphism and deformation of the complex must therefore be questioned. A viable alternative is that early metamorphism and tectonism in the complex are unrelated to those in younger strata. Suggestive but not definitive data concerning the timing of orogenesis in the complex come from comparison of fabric in plutons and in adjacent gneiss in the complex.

The South Fosthall Pluton (Okulitch 1979*a*; Okulitch *et al.*, in preparation) on the southern flank of Thor–Odin Nappe (Fig. 2) is a lineated but weakly foliated granitoid of probable Ordovician age below and within the Monashee Décollement. Comparison of its fabric with that of adjacent metasediments based on limited field observations and reinterpretations of meticulous descriptions made by Reesor and Moore (1971) indicates that the pluton suffered deformation during movement on the Monashee Décollement but apparently does not contain structures possessed by its country rocks. Metasediments near the pluton are permeated with pegmatitic intrusions (not *in situ* melts), whereas the pluton has only minor quantities. This suggests that pegmatite genesis preceded or accompanied emplacement of the pluton. Xenoliths of paragneiss within the pegmatite contain a polyphase fabric that may therefore be pre-Ordovician. It is important to remember that the Shuswap Complex contains pegmatite of many ages and such relationships should not be expected throughout the complex.

In contrast to the South Fosthall Pluton, the Clachnacudainn Gneiss, which lies within hanging-wall strata east of Revelstoke (Fig. 2) and is also of Paleozoic age (Okulitch *et al.*, in preparation), shares the polyphase fabric of its country rocks and is intruded by pegmatite (Ross 1968; Thompson 1978). At high metamorphic grade, therefore, granitoids can be expected to deform together with their country rocks.

Possible westernmost parts of the Monashee Complex southwest of Shuswap Lake (Fig. 2) are also intruded by a pluton of probable Ordovician age (Okulitch *et al.*, in preparation). Garnets in the contact aureole (Fig. 3) contain inclusions delineating an early folded fabric that is presumed to be of pre-Ordovician age.

The absence of radiometric data indicating Mesozoic thermal events in the Monashee Complex and the difficulties in relating structures and patterns of metamorphism across the Monashee Décollement leave the assumption of a Mesozoic age for early structures in the complex with little support. If the early structures of the complex formed prior to the Ordovician, as suggested by relationships with the South Fosthall and Little Shuswap granitoids, it is possible that they may be related to post-Early Cambrian, pre-Carboniferous orogenesis in the Kootenay Arc (Read and Wheeler 1976), to Proterozoic events affecting the Belt–Purcell Supergroup about 1300–1350 and 800–900 Ma ago (McMechan and Price 1982), or to still older events such as the Hudsonian Orogeny (ca. 1700–1800 Ma) recorded in the Canadian Shield. Only the last exhibits structural trends similar to those of the Monashee Complex; others generally parallel the regional northwesterly trends formed during the Jura-Cretaceous. Although correlation of tectonic trends in the complex with those of the Canadian Shield is tenuous at best, few other similar trends are known.

Direct links between the Monashee Complex and the North American craton cannot be proven. K–Ar isotopic ages from the craton nearest the complex (presumably part of the Churchill Province) represent mainly cooling after the Hudsonian Orogeny, although some indicate events as old as 1820–2110 Ma (Burwash *et al.* 1962). In the complex, Rb–Sr and U−Pb ages overlap this older range and are similar to Rb−Sr and U−Pb ages from the Churchill Province (Stockwell 1982), but K−Ar isotopic ages record only Paleogene uplift, and effects of cooling after the Hudsonian Orogeny, if they were felt in the complex, are not preserved.

At present the complex is surrounded by rocks ranging in age from Helikian to Jurassic. The details of relationships of these rocks to the craton are beyond the scope of this paper. To summarize, however, they can be divided into three successions. The oldest (Helikian to early Paleozoic) is linked to the craton through numerous closely constrained interpretations of structures and facies relationships (cf. Monger and Price 1979, pp. 786−787). Above this succession are units largely of Paleozoic age whose stratigraphic associations with the craton are suspect but on the basis of structural relationships were likely attached to the pericratonic prism since the Carboniferous. Above these units are possibly allochthonous strata of Carboniferous to Jurassic age (cf. Monger and Price 1979, pp. 780−786), which will be discussed later.

If the complex is suggested to be allochthonous relative to the craton, the time of its arrival is constrained by the foregoing relationships. If it arrived any time after initiation of deposition of Hadrynian pericratonic strata, a mechanism must be envisaged that moved it eastward against the craton beneath those strata. Structures that support such a model of relative movement are found south of the Malton Gneiss along the northern margin of the Shuswap Complex (Fig. 1). A large first-phase nappe verges southwestward (Raeside and Simony 1983) in this area, but its relation to the Monashee Complex is uncertain. Available data noted above indicate that movement of Hadrynian strata was east relative to the Monashee Complex during the Jurassic. In areas adjacent to the Monashee Complex there is no evidence for earlier motion of opposite sense. Arrival of the complex prior to the Hadrynian could have occurred during the East Kootenay Orogeny (ca. 1350 Ma; McMechan and Price 1982), as the Helikian Belt−Purcell Supergroup affected by this compressive tectonism lies east of the complex. Arrival prior to the Helikian is also possible. These latter two possibilities are presently in the realm of unconstrained speculation.

Brown (1981) suggested that the complex was a continental fragment separated from the craton by a marginal basin. Helikian and younger sediments spread from the craton into the basin and over the complex. No sedimentological data are available to support or refute such a relationship.

In the model to be presented below, lacking evidence or consistent models to the contrary, the Monashee Complex is considered to be a western continuation of the North American craton. Early deformation in the complex is assumed to be Aphebian on the basis of the 2.8−2.0 Ma isotopic ages. Subsequently the complex was extended but not separated from the craton during Helikian rifting. The complex possibly was compressed during the East Kootenay Orogeny and extended again during Hadrynian rifting. The only manifestation of Paleozoic orogenesis appears to be Ordovician magmatism. Finally the complex was thrust against the craton during Jura-Cretaceous convergent tectonism. Its intimate relations with pericratonic strata and the nature of structures separating it from these rocks militate against origins for the complex distant from North America.

The Monashee Décollement

The eastern parts of this structure have been described and defined by Read and Brown (1981), wherein they specified criteria that enable delineation of its probable western extent through areas where it has not yet been mapped. These criteria are metamorphic and structural: rocks of differing grade are juxtaposed and the décollement is affected only by late open folds.

The décollement extends from the north end of Frenchman Cap Dome (Brown 1980; Read and Brown 1981) southward near the Perry River (Journeay 1983) and may join a gently west-dipping fault along the north side of the Eagle River valley and extend as far west as Sicamous (Fig. 2). Here, Paleozoic rocks in chlorite, biotite, and, at low structural levels, garnet zones are exposed within 100 m of sillimanite gneiss of the Monashee Complex (Okulitch 1979*a*). Nielsen (1982) correlated paragneiss, amphibolite, and calc-silicate gneiss of the complex through a slightly faulted transition in grade into quartz−mica schist, amphibolite, and calcareous schist in the Mara Lake area south of Sicamous. The criteria of Read and Brown (1981) require separation of these units from overlying greenstone and calcareous phyllite whose grade does not exceed that of the chlorite zone. Both low- and medium-grade units have been tightly folded several times; however, the contact between them is a gently warped surface whose development presumably postdates most of the internal structures of units above and below it. This contact, herein interpreted as a continuation of the Monashee Décollement, can be traced throughout the Shuswap Lake region, around the crest of the Chase Antiform, and southeast to the vicinity of Vernon (Fig. 2).

The southern and southwestern extensions of the décollement remain uncertain and open to several interpretations. Read and Brown (1981, Fig. 2) chose to link the décollement with the Rodd Creek Fault Zone (Hyndman 1968). This choice seems unsuitable for several reasons. Near the South Fosthall Pluton and to the north the décollement is a 300−1000 m thick mylonite zone. South of the pluton, east of Arrow Park Lake, a fault correlated with it is ". . . a thin but poorly exposed fault zone" (Read and Brown 1981, p. 1132). Movement on the Rodd Creek Fault Zone is constrained by faulted and cross-cutting plutons to the interval 140−110 Ma (Hyndman 1968; Read and Brown 1981), well after that of the décollement, which ended 157 Ma ago (Brown and Read 1983). Hyndman (1968) and Parrish (1981) correlated Paleozoic and Mesozoic strata in the hanging wall of the Rodd Creek Fault Zone with those in the footwall. Reesor (1970, and personal communication, 1982) and Read (1979) extended the correlation to strata of the Pinnacles Nappe. As the Monashee Complex contains, as far as is known, primarily Precambrian rocks, an interpretation consistent with these correlations but one presently without supporting data would extend the décollement along the southern margin of the South Fosthall Pluton through an unmapped and poorly exposed area to Sugar Lake. The Rodd Creek Fault Zone may be analogous to the Columbia River Fault Zone north of Hoskins Creek (see Read and Brown 1981, Fig. 2, p. 1132), that is, a younger normal fault that reactivated but diverged from the décollement. The klippen on Pinnacles Nappe and the faults southeast of Arrow Park Lake, thought by Read and Brown (1981, Fig. 4) to be part of the décollement, are discussed later.

West of Sugar Lake (Fig. 2) the probable trace of the décollement is displaced to the south by a horst that forms the Silver Hills and presumably lies south of Creighton Valley, perhaps beneath Eocene lava flows. West of the horst high- and low-grade rocks lie in close proximity in a region of poor exposure. Outcrops at various elevations suggest that the décollement may be represented by several tectonic interdigitations of sillimanite zone gneiss and chlorite zone pelitic and metavolcanic rocks (Okulitch 1979*a*).

A further complication along the southern margin of the Monashee Complex is the probable presence of a folded suture between strata of the pericratonic prism and terranes that may be allochthonous or at least in fault contact with respect to the North American craton. These features are described later.

The metamorphic grade of the décollement mylonite zone is variable (Fig. 3). Study of this variation is incomplete, yet it bears significantly on the inferred initial crustal level of the décollement, the timing of movement, the changing depths of the décollement as it evolved, and, as has been mentioned above, the timing of metamorphism and deformation in the Monashee Complex. Estimates of initial crustal level of the décollement depend upon which mineral assemblages are interpreted to have formed during mylonitization and which might predate movement. Journeay (1983) has shown that in parts of the mylonite on the west side of Frenchman Cap Dome sillimanite has grown within dismembered kyanite crystals, and he interpreted the preservation of kyanite to mean that it too grew during earliest stages of mylonitization. The crustal level implied by this interpretation is 20–25 km and movement was synmetamorphic. Similar relationships have been discovered on the east side of the dome north of Revelstoke (L. Lane, personal communication, 1984). To the north and south of Lane's study area, however (Brown and Murphy 1982; Read and Brown 1981), and likely throughout much of the rest of the southern and western parts of the décollement (inferred from the grade of hanging-wall strata) mylonitic fabric developed after regional metamorphism under low-grade conditions. No evidence of overprinted synmetamorphic mylonitization has been described from these areas.

Several models are currently under development to describe the evolution of the Monashee Décollement, but such variations remain difficult to account for. A tentative working hypothesis, described further in a concluding section of this paper, incorporates formation of narrow metamorphic belts during the Early Jurassic (such as can be presently seen northeast of the Monashee Complex; Fig. 3) within polydeformed Hadrynian and younger strata during tectonic burial by allochthonous terranes arriving from the west. Variations in depths of burial or local zones of heat flow may have produced the narrow metamorphic belts bounded by steeply dipping isograds (see Ghent *et al.* 1981 for examples). These narrow belts moved during and after their formation across the complex and variably metamorphosed it and the mylonite.

Part two: The northwestern Shuswap Complex

The extensive region northwest of the Monashee Complex above the Monashee Décollement and bounded by the sillimanite isograd has not been fully mapped in detail. Where such studies have been carried out, primarily along the region's northeast margin (Pigage 1978; Campbell 1967, 1978; Simony *et al.* 1980; Pell 1981; Raeside and Simony 1983), it has been demonstrated that Hadrynian strata pass through a normal metamorphic transition into the sillimanite zone. This part of the complex could be considered as an extension of the Selkirk Allochthon of the Kootenay Arc (Read and Brown 1981). The nature of the western margin is less well understood. Near Quesnel Lake, Hadrynian strata have been traced through a narrow metamorphic transition with vertical isograds (Fletcher 1972; Fletcher and Greenwood 1979). North of Adams Lake and between Adams and Shuswap lakes the contact between gneissic rocks of the complex and low-grade strata largely of Paleozoic age is partly intruded by Devonian granitoids and may also be faulted (Okulitch 1979*a*). Rocks within the sillimanite isograd throughout the northwestern part of the complex are undated and may include gneiss correlative with the Monashee Complex and Hadrynian to Paleozoic sedimentary and volcanic units.

Earliest structures are large recumbent nappes that have been deformed by smaller recumbent isoclinal folds and refolded by more upright, open structures. All observed folds trend northwest. Earliest nappes may have formed in the Paleozoic, as did earliest structures in the Kootenay Arc (Read and Wheeler 1976), or may be the result of protracted Jurassic deformation (Raeside and Simony 1983). Second-phase folds formed during the mid-Jurassic peak of metamorphism, and third-phase folds are post-metamorphic and deform isograds.

This part of the Shuswap Complex can be considered as a high-grade extension of the fold belts of the Kootenay Arc and Cariboo Mountains. The transitions in metamorphic grade and structural style have been summarized by Campbell (1973) who postulated the involvement of cratonic basement in Jurassic tectonism, below Hadrynian strata.

Part three: The Okanagan Plutonic and Metamorphic Complex

The Okanagan Complex is distinguished from the rest of the Shuswap Complex on the basis of its consituent strata, intrusions, and orogenic history. Most mapping of this complex is only of a reconnaissance nature, and the necessarily generalized description that follows is based on extrapolation (Okulitch and Peatfield 1977) from widely separated detailed studies.

Basement of the complex is exposed in several domal culminations (Fig. 1). Eastern culminations, the Spokane Dome (Cheney 1980; Reynolds *et al.* 1981) and the Trail Gneiss (Simony 1979; R. L. Armstrong, personal communication, 1982), contain gneiss, possibly of Proterozoic age and perhaps a part of cratonic North America. Extensions of the craton may also occur in Valhalla and Kettle domes (Reesor 1965; Cheney 1980). The oldest rocks known in western parts of the complex are Paleozoic, and their affinities to North America are suspect. The basement of the complex may therefore include a transition from the craton in the east, through attenuated craton, to possibly oceanic crust in the west, although the last has not been observed.

Successions above the basement similarly change character westward. Pericratonic Proterozoic strata occur near Spokane Dome, and Paleozoic units of the Kootenay Arc are found on the north flank of Valhalla Dome (Parrish 1981) and the south flank of Kettle Dome (Cheney 1980). The remainder of the complex contains Carboniferous to Jurassic strata in fault contact with basement units. These strata are part of allochthonous or suspect terranes (Quesnel and Slide Mountain terranes; J. W. H. Monger, personal communication, 1984; formerly the Quesnellia and Eastern terranes of Tipper 1981 and Monger *et al.* 1982). Although no detailed documentation of the lithologic distinctions or structures that separate these terranes from coeval strata on the craton has been assembled. available data presented in highly condensed form (Monger *et al.* 1982) suggest that rocks of these terranes bear little relation to those bordering the craton. Volcanic arc and oceanic assemblages of Devonian to Permian, Triassic, and Early Jurassic ages constitute most of these terranes, and faunal zonation suggests that Jurassic strata of Quesnel Terrane were deposited maybe hundreds of kilometres south of their present location (Tipper 1981). Triassic and Paleozoic rocks underlying allochthonous Jurassic units are also likely allochthonous, given disparities between their faunal assemblages and stratigraphy and those of pericratonic units (Tozer 1970, 1982; Monger 1977).

"Docking" of these terranes is post-Pliensbachian, likely post-Toarcian (Tipper 1981) and before emplacement of the Kuskanax Batholith (Read and Wheeler 1976). Docking initiated the Columbian Orogeny but preceded most deformation and metamorphism, according to relationships described by Montgomery (1978), Read (1979), Parrish (1981), Struik (1981), Rees (1981), and Schiarizza (1983). It therefore occurred from about 188 ± 18 and 182 ± 6 to 173 ± 5 Ma (Parrish and Wheeler 1983) and may have coincided with earliest motion on the Monashee Décollement.[1]

Some of the plutons in the Okanagan Complex are known to be Triassic to Paleogene in age, but most are undated. The older plutons may be genetically allied with Triassic and Jurassic volcanic arc rocks. Many, such as the Kuskanax and Nelson batholiths, intruded the arc and its Carboniferous to Triassic basement after attachment of the terranes to North America. Their intrusion accompanied and followed extensive metamorphism and deformation during the Columbian Orogeny. The younger plutons were emplaced during later tectonism that persisted into the Tertiary.

Most deformation and metamorphism in the Okanagan Complex occurred from the Permo-Triassic (Read and Okulitch 1977) to the Paleogene. Episodes of Proterozoic and Paleozoic orogenesis likely affected the craton and its autochthonous cover, but evidence for these events is sparse (e.g., Simony 1979; Okulitch 1979*b*). Recumbent, polyphase isoclines are prevalent in high-grade parts of the complex (Preto 1969; Christie 1973; Ryan 1973; Medford 1975; Ross 1981). Less intense polyphase deformation affected lower grade strata (e.g., Okulitch 1969, 1973; Simony 1979). Much of this tectonism likely occurred during the Columbian Orogeny. Deformation as young as the Late Cretaceous – Paleocene is recorded in gneissic granitoids that form large sheets within Valhalla Dome (Little 1960; Reesor 1965) and yield 89–97 and 64 Ma U–Pb isotopic ages (Parrish 1983) and a foliated pluton in the Okanagan Valley with a U–Pb isotopic age of 66 Ma (R. D. Loveridge, personal communication, 1981). Gently dipping mylonite zones border major domal culminations (Figs. 1, 4) and formed during these tectonic events (Reesor 1965; Snook 1965; Preto 1969; Fox *et al.* 1976; Cheney 1980; Ross 1981; Parrish 1984). Overprinting of mylonite along cataclastic and brittle zones took place during extensional tectonism, likely of Paleogene age, whose effects increase southward.

Although much remains to be learned of the evolution of the Okanagan Complex, a general model (and one that is quite likely inaccurate in detail) includes the following elements. A continental to transitional to oceanic basement formed about 1600 Ma during rifting and received sediments from the craton up to the late Paleozoic. It is likely that several tectonic episodes (McMechan and Price 1982; Okulitch 1979*b*) interrupted this deposition. Quesnel and Slide Mountain terranes began as a late Paleozoic volcanic arc and marginal basin, presumably with an oceanic basement possibly as old as Devonian, an unknown but not necessarily great distance south and west of North America. Deformation, minor plutonism, and low-grade metamorphism of this arc and its adjacent reefs and basins were followed by erosion and then by deposition of Triassic strata (Read and Okulitch 1977). During the Triassic the marginal basin filled, and in the Early Jurassic a second magmatic arc formed on the older one as the terranes approached North America. During and after collision, which will be described further below, extensive plutonism permeated the continental margin and the obducted allochthon. Extensive crustal shortening and attendant uplift have exposed the roots of these two magmatic arcs, septa of their volcanic carapace, the overthrust pericratonic wedge, and perhaps parts of the craton itself. During the Tertiary, extensional tectonism and volcanic extrusion predominated and gave rise to exposure of still deeper crustal levels.

The Quesnel – North America suture zone

Parts of the suture zone between suspect or allochthonous terranes and North America form an important component of the Shuswap Complex. The zone (Figs. 4, 5) lies west of the complex in the north, may pass between the Monashee and Okanagan complexes south of Shuswap Lake, and may lie east of the complex near Kootenay Lake.

North of Quesnel Lake, Struik (1981, 1982) has described structures of the suture zone near the base of a thick pile of basic volcanic rocks and chert of the late Paleozoic Slide Mountain Group. Southward to Quesnel Lake, the suture zone may follow the contact of ultramafic and tholeiitic volcanic rocks against Hadrynian(?) and (or) Paleozoic metasediments of the Snowshoe Formation (Campbell 1978; Rees 1981). A small ophiolitic klippe has been mapped north of Clearwater (Montgomery 1978). All these volcanic units are interpreted to be part of Slide Mountain Terrane.

Northwest of Adams Lake (Fig. 4, left inset) the Fennell Formation (Campbell and Tipper 1971; Okulitch 1979*a*; Preto *et al.* 1980) of pillow basalt, greenstone, chert, and associated intrusive and sedimentary rocks likely lies above and west of the suture zone. Eastern parts of the formation consist of greenstone intruded by mid-Devonian (387 ± 4 Ma; R. L. Armstrong, personal communication, 1982) feldspar porphyry dykes. A pre-metamorphic fault zone separates these rocks from Early Carboniferous clastic units (Preto 1981; Schiarizza 1983), which lie disconformably on pre-Late Devonian metasediments correlated with Cambro-Ordovician strata in the Kootenay Arc (Okulitch 1979*a*). This part of the Fennell Formation is therefore of pre-mid-Devonian age and faulted against strata believed to be part of the North American succession. Western parts of the formation contain basic volcanic rocks intercalated with late Paleozoic chert (M. Orchard, personal communication, 1982). The whole of the Fennell Formation is considered part of Slide Mountain Terrane.

West and south of Shuswap Lake, structures that are part of the suture zone lie in the North Thompson River region and to the southeast near Vernon (Okulitch 1979*a*, Map B).

Between Vernon and Upper Arrow Lake the suture zone is difficult to trace but can be tentatively positioned between units of Quesnel and Slide Mountain terranes and those of either the Monashee Complex or the Kootenay Arc. Given that the patterns and ages of metamorphic isograds differ in each of these

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[1] The time scale compiled by Palmer (1983) is used to estimate isotopic ages of paleontological stages.

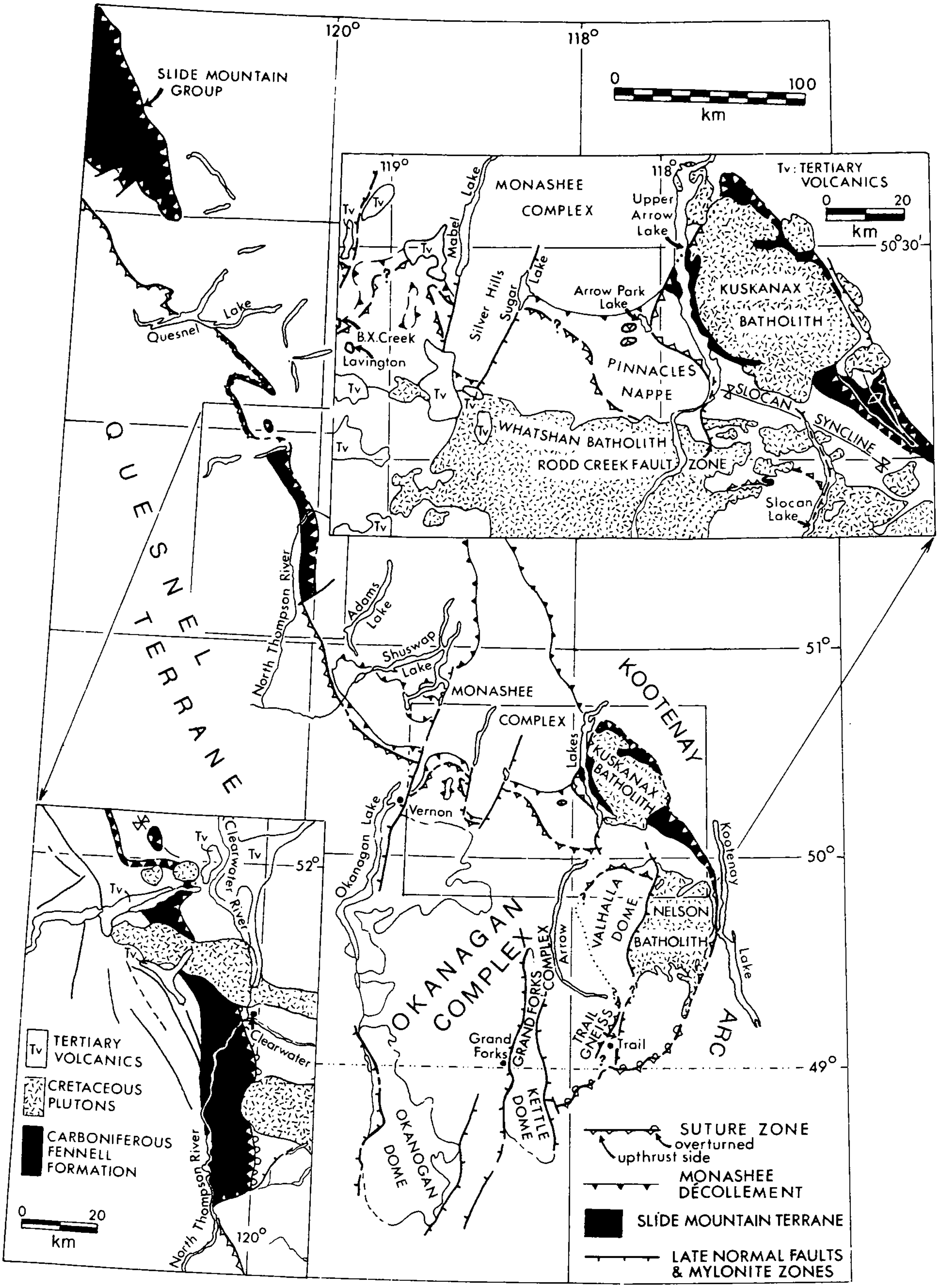


FIG. 4. The Okanagan Plutonic and Metamorphic Complex: major structural components and the Quesnel – North America suture zone.

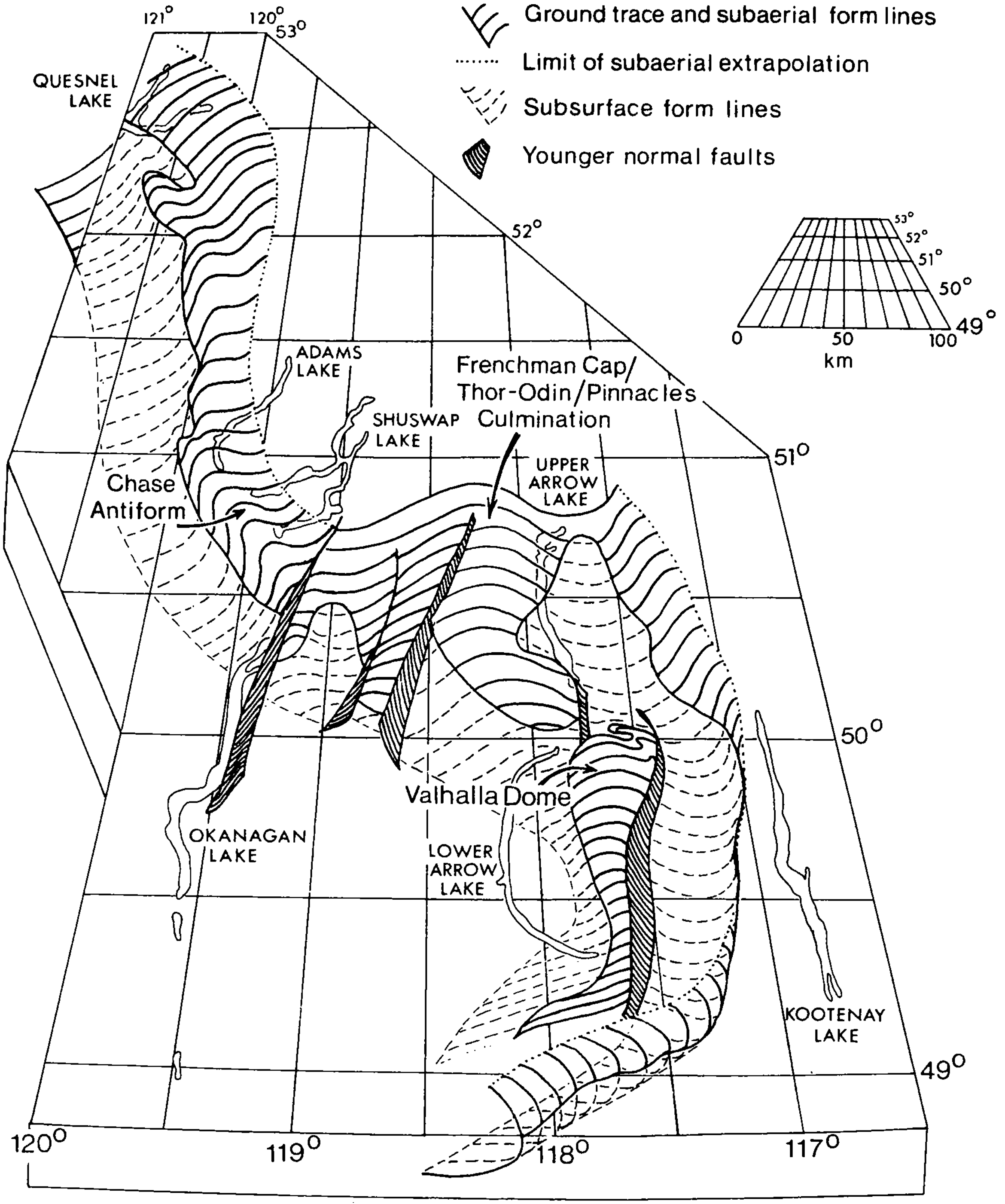


FIG. 5. Perspective sketch looking north, portraying the general configuration of the Quesnel – North America suture zone, represented by a folded and faulted surface whose form lines are generated by the intersection of the surface with east–west vertical planes.

elements, it is to be expected that juxtaposition of rocks of diverse grade would be observed in some areas. These criteria lead to the suggestion that the klippen west of Arrow Park Lake, interpreted by Read and Brown (1981, Fig. 4) to rest on Monashee Décollement, are parts of Quesnel Terrane resting on Paleozoic pericratonic strata. The fault west of Pinnacles Nappe (Jones 1959; Read and Brown 1981, Fig. 2) may also be part of the suture zone, but there and between the Silver Hills Horst and Vernon relationships remain ambiguous because, in the absence of paleontological information, correlations of pelitic units may be made with either those of Quesnel Terrane (Slocan Group) or pericratonic strata (Lardeau and Milford groups). Specific localities such as Lavington and BX Creek, where "unconformities" were postulated (Read and Okulitch 1977), might be reinterpreted as klippen akin to those near Arrow Park Lake.

Probable but unmapped extensions of the suture presumably lie on the southwest, south, and east flanks of Pinnacles Nappe, pass northward to cross Upper Arrow Lake, and wrap around the Kuskanax Batholith. South of the batholith, the suture zone seems to be represented by thrust faults of possibly limited (tens of kilometres) displacement (Klepacki 1983) that telescope Permo-Triassic marginal basin floor (Kaslo Group), but major displacement, if such occurred in this region, may have been on faults in the Slocan Syncline and near Valhalla Dome. South of Klepacki's study area, the suture zone presumably parallels the Kootenay Arc until it abuts the mylonitic shear and cataclastic zone on the east side of Kettle Dome (Rhodes and Cheney 1981).

The highly folded part of the probable suture zone described by Parrish (1981) west of Slocan Lake may delineate the northern extremities of a window (Valhalla Dome) in the obducted sheet. Eastward, the suture zone is cut off by the Slocan Lake Fault Zone (Parrish 1981, 1984), a mylonitic shear and cataclastic zone along the east side of Valhalla Dome (Reesor 1965). To the west and south the suture vanishes within a sea of Jurassic to Tertiary plutons to perhaps reappear near Trail as the mylonitic contact between Upper Carboniferous rocks of Quesnel Terrane (Mount Roberts Formation) and the Trail Gneiss (Simony 1979). Its configuration west and south of Valhalla Dome is highly conjectural.

The suture zone formed during initial phases of the Columbian Orogeny and has been affected to varying degrees by it and later tectonic events. Its present approximate form, based on the data presented, is illustrated in Fig. 5.

Related metamorphic complexes

The Malton Gneiss Complex (Fig. 1)

The Malton Gneiss (Campbell 1967) contains granitoid gneiss and paragneiss of probable Archean and Proterozoic ages (Chamberlain *et al.* 1979). Southeastern parts of the gneiss are in stratigraphic and fault contact with basal strata of the Hadrynian Miette and Horsethief Creek groups (Oke and Simony 1981). Southern parts of the gneiss are interdigitated with presumed Hadrynian strata by four phases of folding and mylonitization (Morrison 1982). Simony *et al.* (1980) and Morrison (1982) have described the Malton Gneiss as thrust slices of the craton intimately involved with its Hadrynian cover during Mesozoic deformation and metamorphism. Lowermost gneissic units are less pervasively deformed by Mesozoic tectonism and may preserve older structures of presumed Precambrian age.

The Malton Gneiss may lie within the footwall of the Monashee Décollement as an extension of the Monashee Complex carried upwards on the post-metamorphic Purcell Thrust. Alternatively, the gneiss may occupy a position above the Monashee Complex within the hanging wall of the décollement. The hierarchy of structures affecting cover strata and upper parts of the gneiss is identical to that within the hanging wall of the décollement throughout the Kootenay Arc and Cariboo Mountains fold belts and the second (northwestern) part of the Shuswap Complex. The Purcell Thrust may join the Monashee Décollement below the Columbia River valley (Read and Brown 1981; Brown and Read 1983); however, movement on the décollement ended at 157 Ma, whereas the Purcell Thrust was apparently active in the interval 140–100 Ma (Simony *et al.* 1980). Alternatively, the thrust may root east of and below the Monashee Complex as a later splay from a basal detachment below the Monashee Décollement. Both latter alternatives are favoured and illustrated in Fig. 6, cross section A–A.

Priest River Crystalline Complex (Fig. 1)

To the south of Kootenay Lake in Idaho, Proterozoic gneiss (R. L. Armstrong, personal communication, 1982) cut by numerous Mesozoic and Tertiary plutons forms the Priest River Complex, exposed in the Spokane Dome (Cheney 1980) (Fig. 6, inset, cross section C–C). Little is known of the internal evolution of this complex. Like others to the northwest and west, it is largely bounded by mylonite zones (Harms 1982) and has been compared to metamorphic core complexes of the American southwest (Reynolds *et al.* 1981) that formed during Neogene crustal extension and denudation (Davis and Coney 1979; Coney 1980). The Priest River Complex may be an uplifted part of the North American craton and, as such, similar to the Monashee Complex, but appears to have been modified by later extension not evident to the northwest.

The Shuswap Complex and Cordilleran tectonism

Key problems regarding the role of the Shuswap Complex in Cordilleran tectonism have been the following.

(1) Is the basement, i.e., the North American craton, involved in deformation associated with the Rocky Mountains Thrust Belt and, if so, how?

(2) How is at least 200 km of shortening of the cover accommodated at basement levels?

(3) What are the driving forces for the thrust belt?

(4) What effects have extensional tectonics had in the evolution of the complex?

These four problems will be discussed with reference to a favoured model and to other recently proposed hypotheses. It should be noted that what follows is not, therefore, a fully balanced review and other papers (e.g., Armstrong 1982; Monger *et al.* 1982; Brown and Read 1983) should be consulted for alternative views. The favoured model is illustrated by three cross sections (Fig. 6) and an evolutionary sequence (Fig. 7).

The question of involvement of western parts of the North American craton in Jurassic to Paleocene development of part of the Rocky Mountains Thrust Belt has been one of extent and location. Borehole and seismic data have clearly established that cratonic basement was undisturbed as far west as the Rocky Mountain Trench south of 50°N (Bally *et al.* 1966; P. Gordy, personal communication, 1981). To the north, no information on basement involvement is available until exposures of the Malton Gneiss and Monashee Complex are encountered. On the basis of the assumption of intense Mesozoic deformation in the Shuswap Complex, a sophisticated hypothesis was proposed describing buoyant upwelling and extending flow in an infrastructure that drove a suprastructure eastward to form the thrust belt (Price and Mountjoy 1970). Immediately east of such dramatic tectonism, the basement was interpreted to be passive (Price and Mountjoy 1970, Fig. 2-1). Campbell (1973) proposed a transition from ductile flow in the Shuswap Complex through high-angle reverse faulting over passive basement in the east. Clarification of the nature of the Malton Gneiss (Morrison 1982) and the discovery of a large, uplifted, presumably thrust-faulted block of basement under the core of the Purcell Anticlinorium in Montana (Wynn *et al.* 1977; Harrison *et al.* 1980) indicated that west of the Rocky Mountain Trench basement was involved in thrusting (Fig. 6, sections A–A and C–C). The infrastructure–suprastructure model (Price and Mountjoy 1970) was thereby modified to one of intraplate convergence with associated subduction and upwelling of a core zone (Price 1981) (see also Fig. 8*b*).

The favoured model describes involvement of basement beginning in the Jurassic (Fig. 7*a*). Middle Jurassic tectonism (Fig. 7*b*) has affected the Okanagan Complex, the Monashee Décollement, and northwestern parts of the Shuswap Complex (including the Malton Gneiss). The Monashee Complex was apparently little affected except by stretching associated with

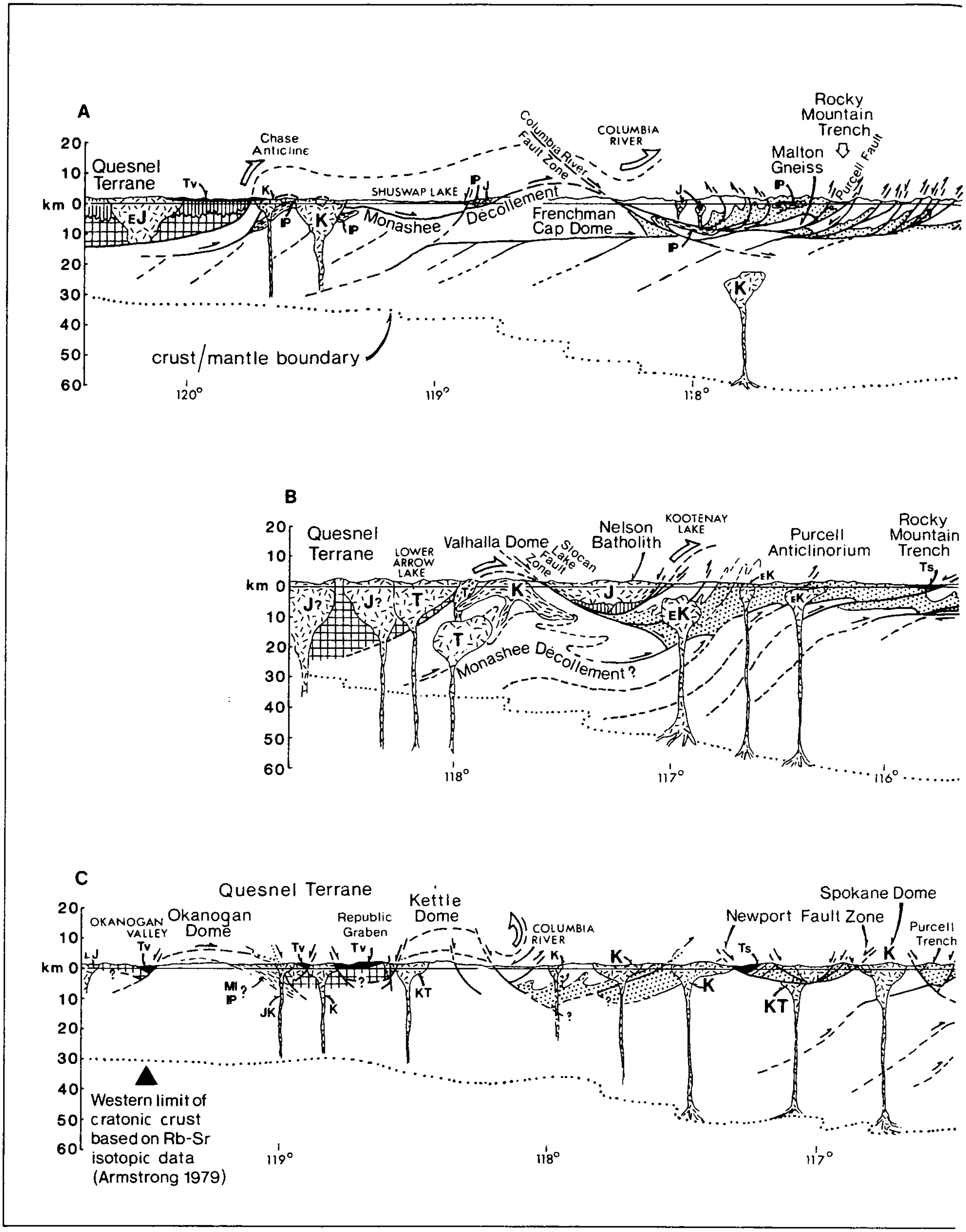
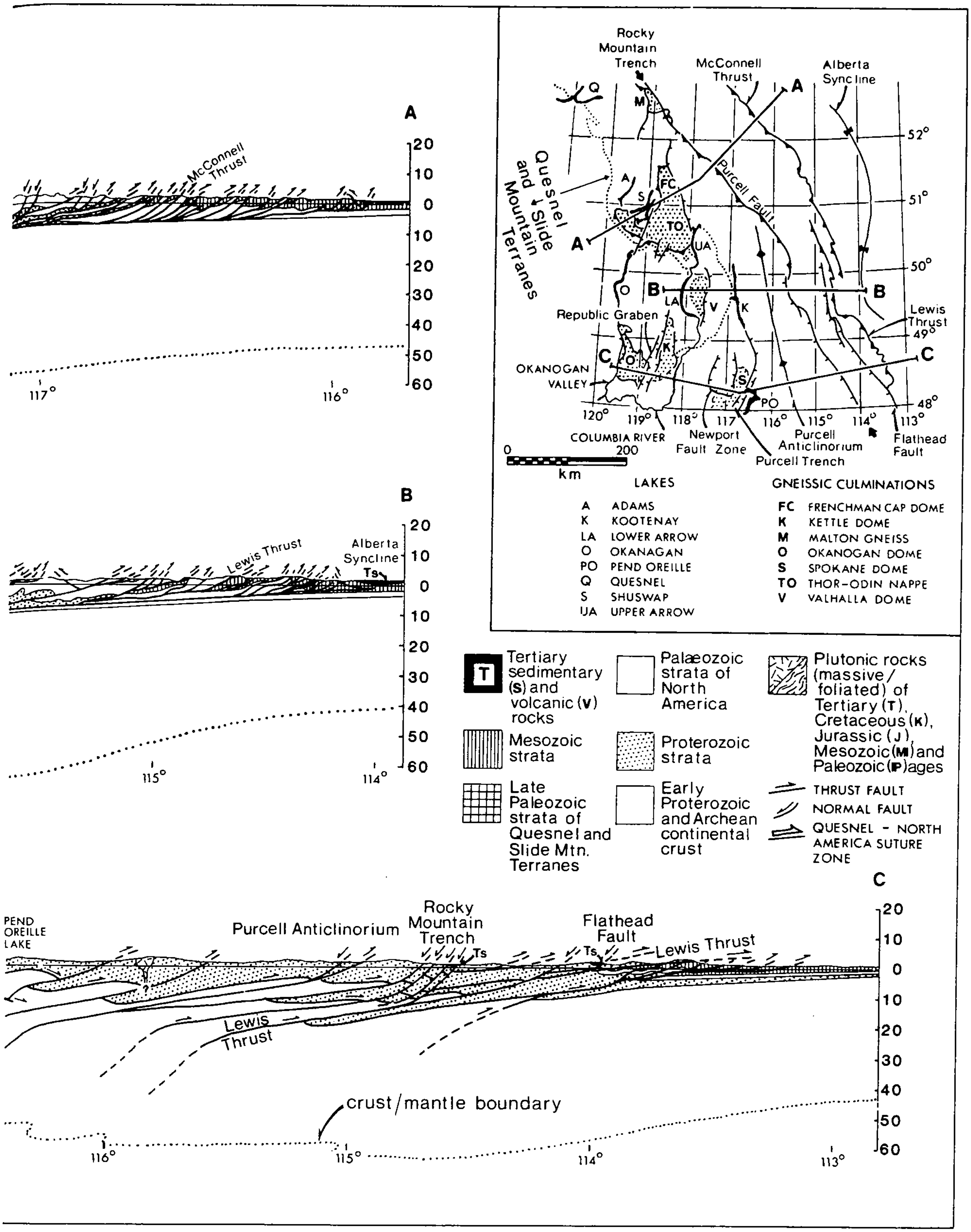


FIG. 6. Cross sections of the southeastern Canadian Cordillera and adjacent parts of the American Cordillera illustrating structures of the thrust belt, core zone, and allochthonous terranes. Horizontal and vertical scales are equal. Section A–A from Price and Mountjoy (1970), Okulitch



(1979*a*), and Read and Brown (1981). Section B–B from Reesor (1965) and Price (1981). Section C–C from Fox *et al.* (1976), Cheney (1980), Harrison *et al.* (1980), and Harms (1982).

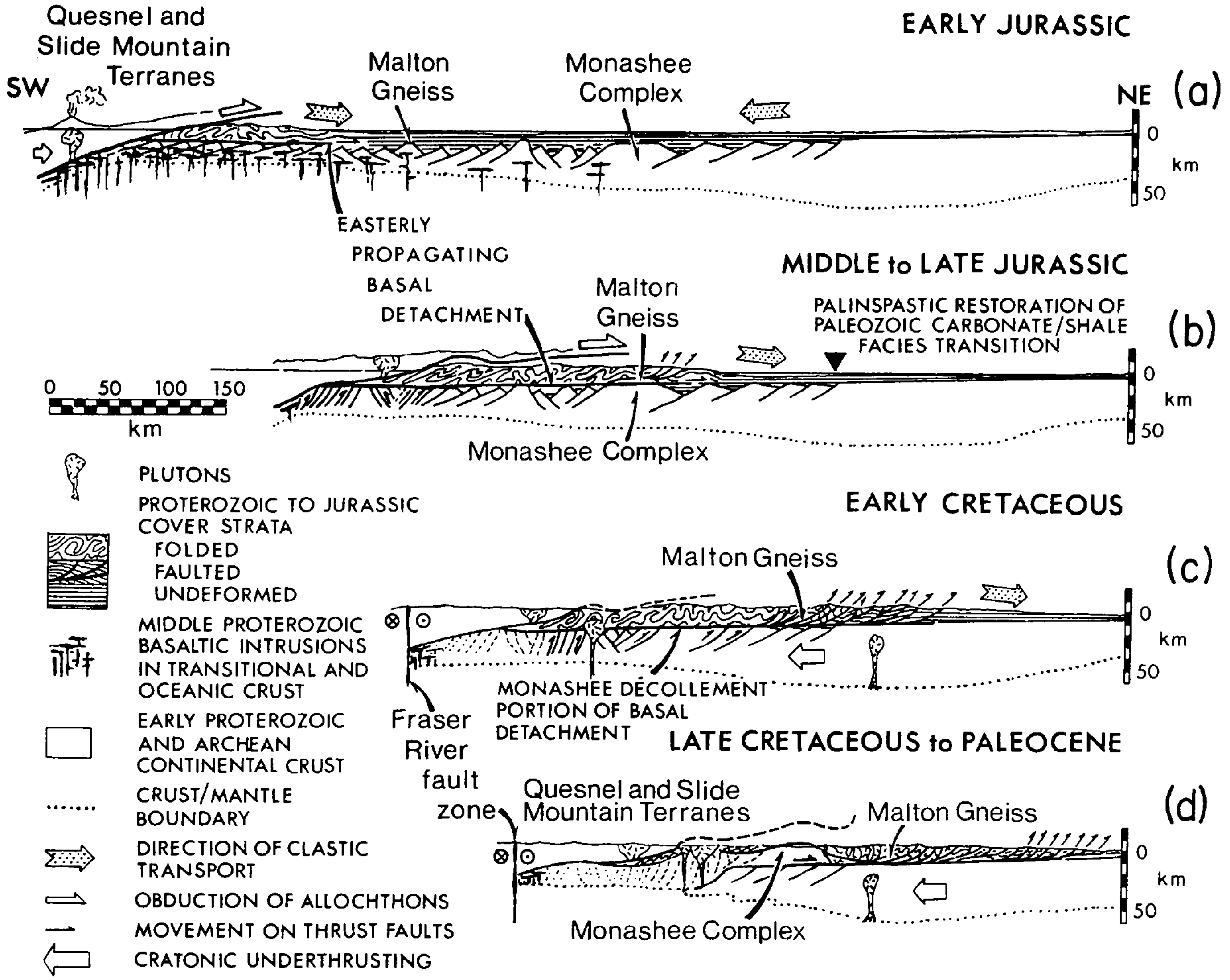


FIG. 7. Evolution of the Columbian Orogen along section A–A, Fig. 6. Structures formed during episodes of compressional and extensional tectonism during the Proterozoic (McMechan and Price 1982) and Paleozoic (Okulitch 1979*b*; Read and Wheeler 1976; Bond and Kominz 1984) are not shown. The Malton Gneiss provides an indicator of convergence of cover and craton (possibly as much as 400 km). Horizontal and vertical scales are equal.

eastward propagation of the Monashee Décollement. The intimate association of the Malton Gneiss with its cover suggests that both originated west of the Monashee Complex prior to Jurassic orogenesis. Continuations of the Malton Gneiss in the footwall of the décollement presumably lie below Quesnel Terrane (Figs. 6, 7).

In response to westward drift of North America during opening of the Atlantic Ocean, continued convergence in the Cretaceous and Paleocene with allochthonous terranes west of Quesnel Terrane (Monger 1977; Monger and Irving 1979) further compressed Quesnel and Slide Mountain terranes, the previously deformed pericratonic prism, and the outer edge of the craton (Fig. 7 *c* and *d*). Subduction zones and their associated belts of heat flow are presumed to have lain to the west of the accreted terranes (Monger and Price 1979, Fig. 10) and apparently did not play a significant part in tectonism of the Shuswap Complex. At this time the crystalline slices beneath the Purcell Anticlinorium, the Monashee Complex, and possibly parts of the Okanagan Complex acted as crystalline massifs and moved along zones of weakness such as Proterozoic normal faults that converged upward with the Monashee Décollement. As shortening proceeded the décollement was itself overthrust (Fig. 6, section A–A, below Frenchman Cap Dome; Fig. 7*d*). At about the same time, the Purcell Thrust carried the Malton Gneiss eastward and upward. The crystalline massifs and their deformed and metamorphosed cover, which had cooled by the Late Jurassic, were rigid constituents of the thrust belt. The essence of this model was first proposed by Bally *et al.* (1966) and it has proved difficult to improve upon their suggestions.

The second question concerns the imbalance in shortening between cover and basement. Price and Mountjoy (1970) showed a minimum of 160 km of shortening of cover over passive basement between the plains and the Rocky Mountain Trench. Still more shortening of cover must have occurred between the trench and Quesnel Terrane. This amount is indeterminable, but total shortening is suggested to be about 400 km in Fig. 7. West of the trench, geophysical evidence (Stacey 1973) indicated the absence of a deep crustal root beneath the Shuswap Complex. Commensurate shortening of the basement to provide mass balance above the Mohorovičić discontinuity

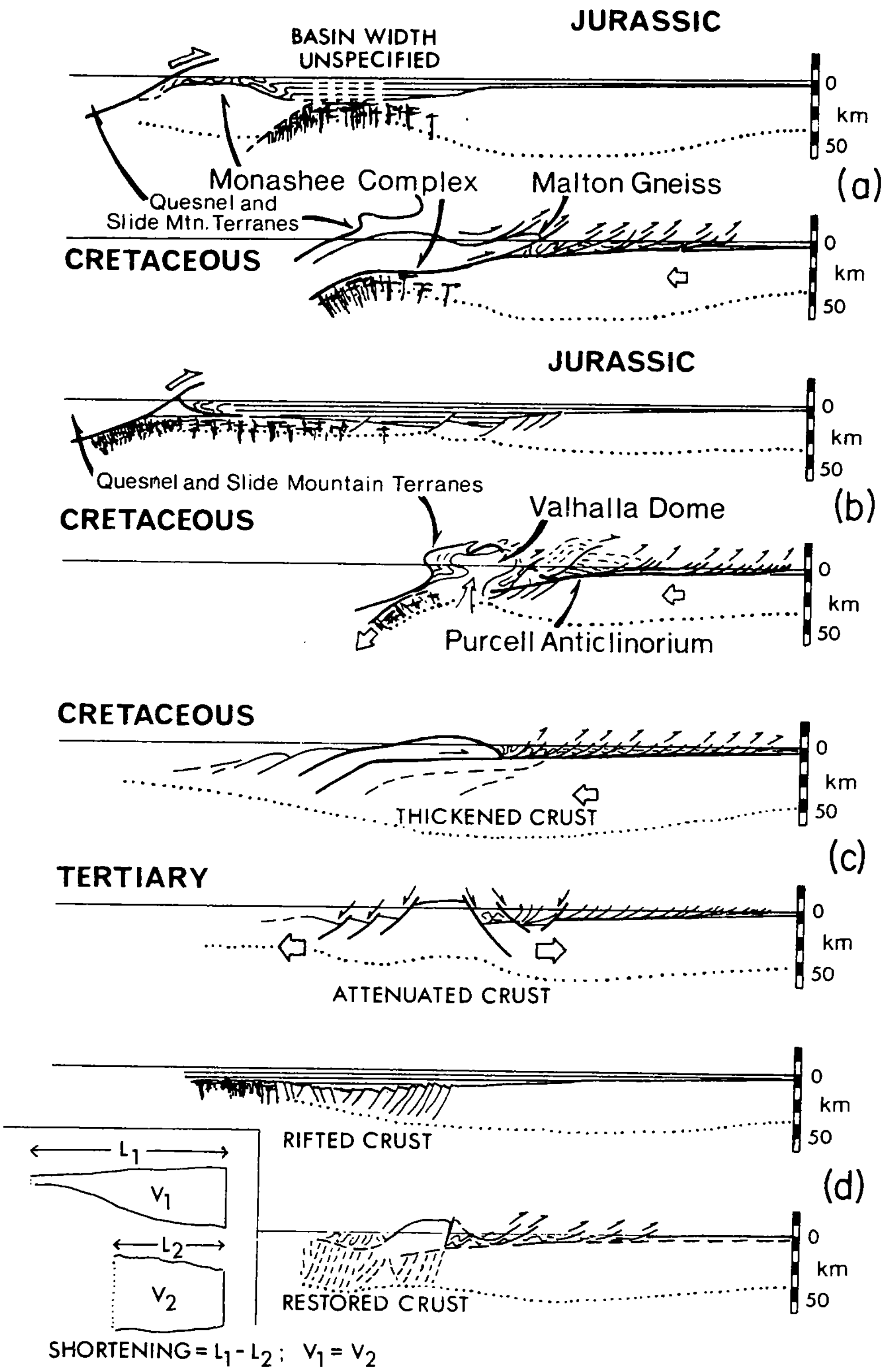


FIG. 8. The balance of cover and basement as achieved by different models. (*a*) From Brown (1981) and Brown and Read (1983). (*b*) From Fig. 5 and text descriptions in Price (1981). (*c*) From oral descriptions by Coney (1983). (*d*) From Helwig (1976). All figures modified to have consistent format and scale. Legend as for Fig. 7. Horizontal and vertical scales are equal.

is not evident. Several solutions have been proposed to account for this missing crust (Fig. 8). Brown (1981) suggested that westward subduction of oceanic or transitional basement that originally separated the craton from the Monashee Complex could explain the absent mass. Shortening of the cover that mantled the craton, the marginal basin, and the Monashee Complex was thereby accomplished, the Monashee Complex joined, or rejoined, the craton, and the missing crust vanished in the depths (Fig. 8*a*). At the latitude of the Monashee Complex, however, little evidence for a magmatic arc associated with this subduction exists, nor is there any evidence for an oceanic or transitional basement for Proterozoic sediments of the pericratonic prism. Hadrynian sediments apparently lie on possibly attenuated continental crust (the Malton Gneiss) to the north (Oke and Simony 1981). Amphibolitic sills and dykes in Hadrynian units, cited as possible evidence for subjacent oceanic or transitional crust (Brown 1981), also occur throughout Paleozoic platformal to miogeoclinal successions over the length of the Cordillera (Gabrielse 1976) and are not likely to be genetically related to rifting that formed such a crust at least 200 and possibly 1000 Ma before these intrusions were emplaced.

Price (1981) proposed similar mechanisms to account for the imbalance but located the subduction zone west of the Shuswap Complex (Fig. 8*b*). Attenuated continental crust followed oceanic crust into the subduction zone, was heated, and then rose diapirically to form the Shuswap Complex. However, structures in the complex are not the result of diapirism (Brown 1978; Read 1979, 1980; Duncan 1984). Bally (1981) also appealed to such "A-subduction" to create the necessary balance. The buoyancy of continental crust may make such a process too limited to fully balance the observed and inferred shortening (McKenzie 1969) unless attenuated crust was involved or other, deeper seated (and entirely hypothetical) mechanisms such as crustal delamination are invoked (Bird 1979).

Coney (1983) suggested that the crust was thickened but that it later thinned during Tertiary extension (Fig. 8*c*). The absence of evidence for extension of the necessary magnitude is discussed below.

Helwig (1976) provided a geometrically elegant and geologically consistent solution for the Alps by accepting that an apparent imbalance between cover and basement was a demonstrated fact and a necessary consequence of the development of a passive margin that is later compressed. His hypothesis is adopted for the Shuswap Complex (Fig. 8*d*). Rifting of the western margin of North America formed the basin of deposition for Proterozoic and younger sediments. These were therefore laid down upon attenuated crust. Shortening of this crust and the overlying strata during collision with Quesnel and Slide Mountain terranes initially restored the crust to its approximate original (pre-Belt–Purcell) configuration while extensively shearing and compressing cover strata (Fig. 7 *a–c*). A large imbalance was thereby created. Clearly, a major décollement that propagated eastward as compression progressed must have simultaneously formed between basement and cover. Mylonite zones likely formed at various levels. In places, cover may have remained attached to basement. The Monashee Décollement and the mylonite zones bounding other gneissic culminations are envisaged as tectonically exhumed parts of such a structure. Some of the variation in metamorphic grade within the décollement mylonite zone might be explained by the juxtaposition of tectonic slivers that originated at various levels. Continued shortening during the Cretaceous involved the basement as crystalline massifs (Fig. 7*d*). Overthrusting of the eastern part of the Monashee Décollement occurred during this shortening, possibly along reactivated Proterozoic normal faults, and movement up basement ramps provided the uplift necessary for exposure of the Monashee Complex and other gneissic culminations. Final uplift during the Tertiary (not incorporated in Fig. 8*d*) thinned the crust to that shown in Fig. 6, section A–A.

In this model, convergence in the orogen can be viewed as occurring in three episodes. During the first (Early Jurassic) episode (Fig. 7*a*), collision and obduction of Quesnel and Slide Mountain terranes buried and metamorphosed parts of the pericratonic prism wherein predominantly west over east shearing and east-directed compression (Spang *et al.* 1980) occurred. Near the allochthons, early folds verge eastward (Brown and Read 1983). To the east, attenuated basement began to shorten, producing west-verging nappes (Raeside and Simony 1983), possibly as their lower limbs moved east with the basement. Some early west-verging folds may be of Paleozoic age (Read and Wheeler 1976; Höy 1977; Brown 1981).

During the second (Middle to Late Jurassic) episode (Fig. 7*b*), substantial shortening and eastward displacement of cover strata took place above the Monashee Décollement while the décollement was an integral part of an easterly propagating basal detachment. Tectonic burial by obducted allochthons may have depressed parts of the décollement to depths of 20–25 km (Journeay 1983). In the southern Kootenay Arc, Proterozoic and Paleozoic strata were depressed to such depths between 166 and 156 Ma (Archibald *et al.* 1983). Synmetamorphic folds in the cover verge both west (Brown and Read 1983) and east (Simony *et al.* 1980; Raeside and Simony 1983). Some of this variation may have resulted from later regional warping. The Monashee Complex appears to have been largely unaffected except for imposition of metamorphism from hot cover strata and formation of a stretching lineation that may (Journeay 1983) or may not occur throughout the complex, depending upon interpretations of the genesis of metamorphic fabrics. Thermal relaxation in cover strata and erosion of obducted terranes may have allowed final movement on the Monashee Décollement (up to 157 Ma, Read and Brown 1981) to proceed at low-grade conditions. During this second episode, basement shortening occurred to the west as the attenuated part of the craton was restored.

Only some of this Jurassic tectonism appears to be spatially and temporally related to the thrust belt. In the foreland clastic wedge, earliest west-derived detritus is of Callovian–Bathonian age (Poulton 1984), possibly the time of deformation in western parts of the thrust belt. Thrust faulting and folding migrated eastward and culminated in the Cretaceous, as indicated by major clastic influx into the foreland basin (Price and Mountjoy 1970).

At some point restoration of basement was completed and further shortening required overthrusting of basement. During this third (Cretaceous–Paleocene) episode (Fig. 7*c*), the Monashee Complex moved eastward on a lower detachment, up a basement ramp, and along the eastern (autochthonous) part of the Monashee Décollement. The western (now allochthonous) part of the décollement became warped during movement up the ramp. Thickening of the craton below but largely to the east of the Monashee Complex and other rising massifs may have resulted in local anatexis and generation of the zone of mid-Cretaceous plutons that parallels exposure of the massifs (Figs. 6, 7 *c* and *d*). Armstrong (1983) has supported this last hypothesis with isotopic studies.

Final stages of this uplift, recorded by numerous K–Ar isotopic ages from the complex, occurred primarily 60–50 Ma ago. Eastward movement of the Monashee Complex must be balanced by the shortening of the cover that took place during this time. Price (1981) reviewed evidence for a minimum of 100 km shortening in the thrust belt between 80 and 50 Ma ago. It therefore seems likely that convergence between the complex and the craton was about 100 km or less.

The third question, that of driving forces for the thrust belt, cannot be answered. However, some of the major features of the orogen might be explained by considering the probable relative movements of major crustal elements. Structures of the thrust belt must be related to convergence between the craton and its bordering sedimentary wedge and a collage of allochthonous terranes of continental dimensions (Monger *et al.* 1982). Obducted and underthrust allochthons (e.g., Archibald *et al.* 1983), together with the thickened sedimentary prism, moved eastward. The craton was telescoped; this can be viewed as easterly overthrusting or westerly underthrusting. Differences in structural styles between the core zone (Shuswap Complex) and the thrust belt may have resulted from the differing interactions among the allochthons, the pericratonic prism, and the craton. The heterogeneous nature of the Shuswap Complex, particularly the Okanagan Complex, may be the result of the meeting of a collage of terranes with the craton and its sediments along a highly irregular suture zone. In contrast, the thrust belt is a coherent structural entity over 2000 km long. Interaction of crustal elements responsible for its formation presumably could not have varied significantly over that distance. Westward underthrusting of the whole of the western margin of the craton and the resulting off-scraping of its bordering sediments seems necessary to form the thrust belt.

The final question concerning the evolution of the Shuswap Complex arises from proposed parallels between it and "metamorphic core complexes" studied in the American southwest (Coney 1980; see Armstrong 1982 for an excellent review). The late history of the complex does indeed embody some extension and both tectonic and erosional denudation. The amount of extension that can be documented at present, however, is an order of magnitude less than that proposed for the Basin and Range Province (see Wernicke *et al.* 1982 for a summary of estimates).

The Monashee Complex is least like a denuded core complex. The characteristics of the latter, clearly summarized by Coney (1980), should be compared with what follows. Pervasive easterly trending lineations within the complex can be related to early (Precambrian?) north-verging nappes and to stretching related to formation of the Monashee Décollement. As has been noted previously, various interpretations of the timing of these structures have been made. Metamorphic grade in the décollement mylonite, which presumably reflects quenching during final movement, is variable and suggests the imposition of a complicated distribution of transported metamorphism within overthrust allochthonous terranes and cover strata. The extensional fabric of the décollement can be kinematically linked to this eastward overthrusting (Brown and Murphy 1982; Journeay 1983), and is part of shear and compression produced by intraplate convergence, not extension.

Cretaceous to Eocene normal faults, the Columbia River and Rodd Creek fault zones on the east, and minor faults in and north of Mara Lake on the west have displacements of less than 1 km, usually only a few 100 m (Read and Brown 1981; Nielsen 1982; Brown and Read 1983). Rotation of hanging-wall strata and structures is minimal. The complex yields evidence of final uplift and cooling at about 60–50 Ma. It was soon after covered with volcanic flows (45–35 Ma) after which uplift continued, as indicated by erosional stripping of the flows from most parts of the complex that contain their feeder dykes. Post-Eocene uplift did not occur by renewed faulting along the Monashee Décollement or younger faults, as flows lie undisturbed across these structures in some places (Okulitch 1979*a*; Mathews 1981).

In the Okanagan Complex, the Valhalla, Kettle, and Okanagan domes are bounded on at least one side by major mylonite zones. Much of this southern region has been postulated to have suffered extension and normal faulting (Price *et al.* 1981) during Eocene to Oligocene times. In places, high-grade metamorphic rocks are exposed, but the faulting, except where demonstrably older mylonite zones are reactivated, commonly bears no relationship to these exposures. The most obvious, and possibly unique, extensional feature is the Newport Fault Zone (Miller 1971; Harms 1982). Extension on this structure has been estimated as between 10 and 30 km at its south end and zero to the north (Fig. 6, inset). Extension on other obvious structures such as the Purcell Trench (Harrison *et al.* 1972; Rehrig and Reynolds 1981) and the Republic Graben (Fig. 6, inset) may not exceed 10 km. Dip slip on the Slocan Lake – Champion Lakes Fault is limited to several kilometres (Parrish 1984). Throughout the 500 km wide region between the Okanagan Valley and the Flathead Valley Graben only a few tens of kilometres of extension can be demonstrated with presently available data (about 10 and not more than 20%). The amount of extension may increase southward in a gradual transition to tectonism of the Basin and Range type.

Movement of high-grade metamorphic rocks into upper crustal levels was therefore apparently accomplished during Cretaceous–Paleocene compression rather than by rotation of crustal boudins during extension. Once within upper levels, final cooling, uplift, and limited extensional denudation led to further exposure of high-grade rocks. Tensional forces might be a regional phenomenon, as suggested by Price (1979) and Harms (1982). In short, the special conditions attributed to evolution of metamorphic core complexes in the southwestern states do not appear to have been present in the region of the Shuswap Complex except to a limited degree. However, it is very likely that both the southwestern core complexes and the Shuswap Complex participated in Mesozoic compressional tectonics.

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