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# GUIDE TO THE GEOLOGY AND TECTONIC EVOLUTION OF THE SOUTHERN COAST MOUNTAINS

James W.H. Monger and J. Murray Journeay

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Available from
Geological Survey of Canada
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A small number of copies of this guidebook were prepared for a field trip that was part of the Geological Society of America's Penrose Conference on the "Tectonic evolution of the Coast Mountains orogen," held on Bowen Island near Vancouver, on 17–22 May 1992. Since then, the guidebook has been in demand because it contains a field trip that can be traversed on good roads in two days, starting from and returning to Vancouver, and also because the introduction reviews current ideas on the tectonic evolution of the southern Coast Mountains. Therefore, it has been revised and is re-issued with modifications as GSC Open File 2490.

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**Cover description**

Folded pegmatites along the east side of the Scuzzy Pluton, east of Kwoiek Needle, southwestern British Columbia.

Photo: Murray Journeay

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# INTRODUCTION

This two-day field trip focuses on plutonic, stratigraphic and structural elements of the southern Coast Mountains that bear on its tectonic evolution. The trip was run initially as part of the Geological Society of America Penrose Conference on the "Tectonic evolution of the Coast Mountains orogen," held from 17–22 May 1992 (reported in GSA Today, October 1992). This guidebook is a revised and updated version of the one used for the conference. It does not claim to give an even-handed overview of all aspects of the geology, but emphasizes structure and tectonics; granitic rocks are **not** the focus of the guidebook even though they constitute 80% of the Coast Mountains!

The guidebook uses material from earlier mapping and field trips (e.g., Roddick, 1965; Roddick et al., 1977; Woodsworth and Roddick, 1977; Roddick and Woodsworth, 1979; Price et al., 1985; Monger, 1989a; Monger et al., 1990), and unpublished theses from the Universities of British Columbia, Washington and Western Washington. New data and ideas come from recent mapping in the region by the Geological Survey of Canada and British Columbia Geological Survey Branch (e.g., Schiarizza et al., 1990a,b, Journeay et al., 1992; Journeay, 1993, Monger, 1993; Journeay and Mahoney, 1994) and from university research conducted under the auspices of the Lithoprobe southern Canadian Cordillera transect (Friedman and Armstrong, 1990, in press; Coleman and Parrish, 1991; Varsek et al., 1993; Zelt et al., 1993).

The guidebook contains an introduction with summary overviews of the geology and tectonic evolution of the entire Coast Belt and some details of the geology of the southern Coast Mountains and its relationship with that of the Cascade Ranges, and field trip logs for Days 1 and 2 with "comments" whose purpose is to enlarge on the regional significance of the geology of that part of the route. The introduction and comments are intended to provide enough background so that the field trip can be taken without a leader. The route of Day 1 (Fig. 1; in pocket) follows the Fraser River east from Vancouver to Hope and then north to Lillooet, along the physiographic boundary between Coast and North Cascade mountains. Day 1 emphasizes similarities and differences between the two orogens. The route of Day 2 crosses the entire Coast Mountains, and runs southwest from Lillooet via Pemberton to Whistler and then south through Squamish to Vancouver. During Day 2, emphasis is placed on the stratigraphic, structural and tectonic evolution of the southeastern Coast Mountains and its bearing on interpretations of the tectonic evolution of the geology of the Coast Mountains as a whole.

In the guidebook, the geologic time scale of Harland et al. (1989) is used to relate ages in years, derived from isotopic determinations, to biostratigraphic stages. This aspect is important because of the nature of the geology of the Coast Mountains. In order to understand its tectonic evolution, it is necessary to establish the time and space relationships between the results of deeper lithospheric processes such as metamorphism and plutonism which dominate Coast Mountains geology, and surface processes such as basin sedimentation and volcanic effusion.

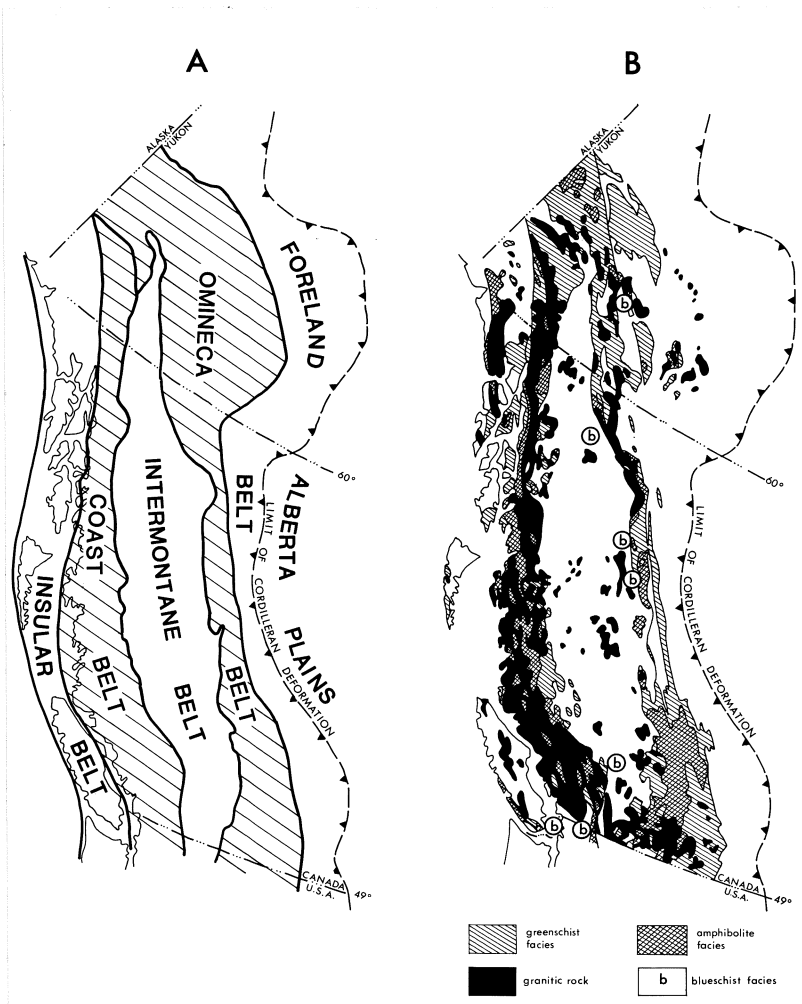



**Figure 2A.** Morphogeological belts of the Canadian Cordillera.
**Figure 2B.** Simplified metamorphic map of the Canadian Cordillera; this map displays the geological character of the belts in Figure 2A; two belts of granitic and high-grade metamorphic rocks separate three tracts of little metamorphosed and subgreenschist facies rocks.

## GENERAL GEOLOGY OF THE COAST MOUNTAINS

### *Terminology and general description*

The ''Coast Belt'' is one of the five **morphogeological** belts of the Canadian segment of the North American Cordillera, and includes both the **physiographic** Coast Mountains and Cascade Ranges (Fig. 2A; Wheeler and McFeely, 1991; Mathews, 1986). Because of the need on this field trip to discuss the geology of both Coast and Cascade mountains, generally the term ''Coast Mountains'' is used herein. Older names such as ''Coast Range batholith'' (Le Roy, 1908), ''Coast Crystalline Belt'' (Roddick, 1966; Roddick et al., 1967) and ''Coast Plutonic Complex'' (Wheeler and Gabrielse, 1972) reflect the predominance of plutonic rocks within the orogen. Note that the meaning of Coast Plutonic Complex, widely used synonymously with the now current term Coast Belt, has changed subtly in the most recent general description of Cordilleran plutonic rocks, and now refers exclusively to plutonic rocks within the Coast Belt (Woodsworth et al., 1991).

The Coast Mountains extend for 1700 km (between lats. 49° and 62°N), are between 100–200 km wide, and reach elevations of over 4000 m, although most summits are 2000–3000 m high. The Coast Mountains are composed mainly (80–85%; Roddick, 1983) of Late Jurassic to early Tertiary granitic rocks and form one of the largest Phanerozoic plutonic complexes in the world. In general, the plutonic rocks become younger from west to east. The Late Jurassic-Early Cretaceous plutonic rocks that characterize the western part of the Coast Mountains south of Prince Rupert (lat. 54°N) are absent from it north of this, so that the plutons which give integrity to the entire belt are of mid-Cretaceous through early Tertiary ages. Subordinate, variably metamorphosed septa and fault-bound slivers of stratified rocks, and rare stratigraphic caps of volcanic and sedimentary strata are distributed throughout the Coast Mountains (Fig. 2B), and range in protolith age from Proterozoic(?) to Holocene. The Coast Mountains lie mostly within the domain of the accreted terranes (Fig. 3; Table 1), which feature juvenile crust with low initial strontium ratios (Armstrong, 1988), and positive $\varepsilon_{Nd}$ values (Samson and Patchett, 1991).

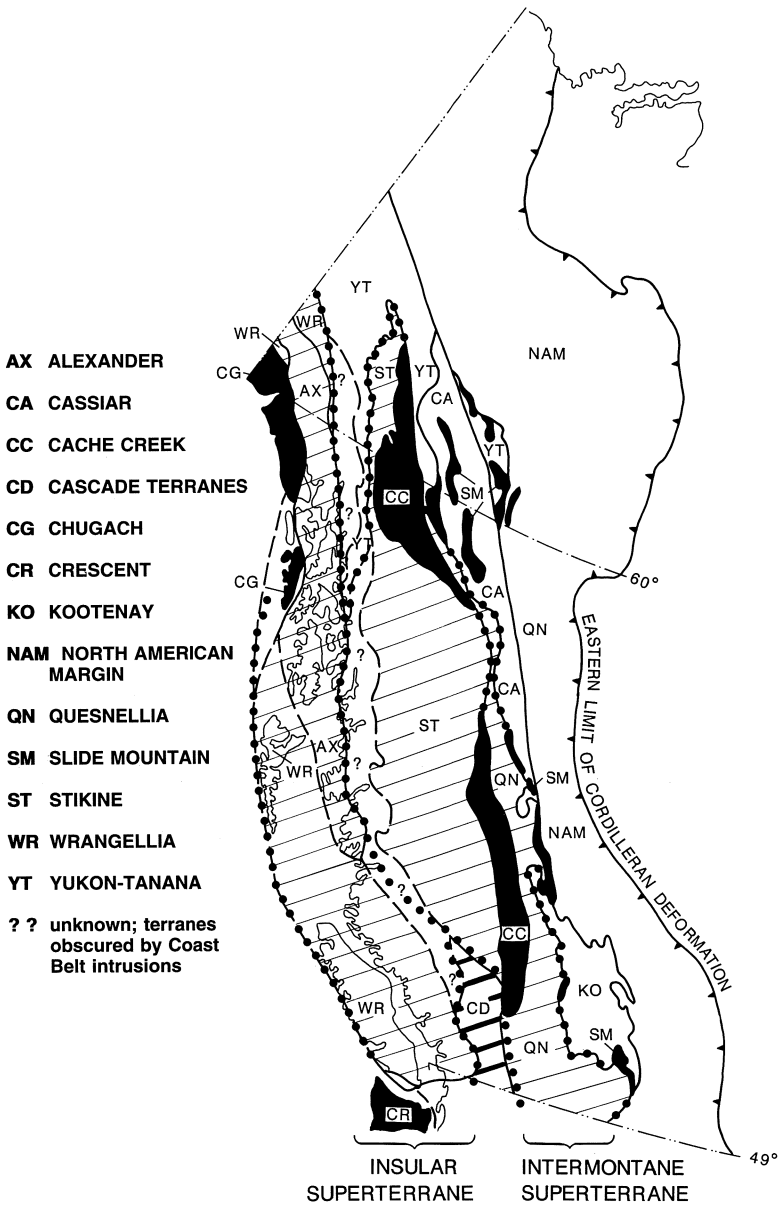
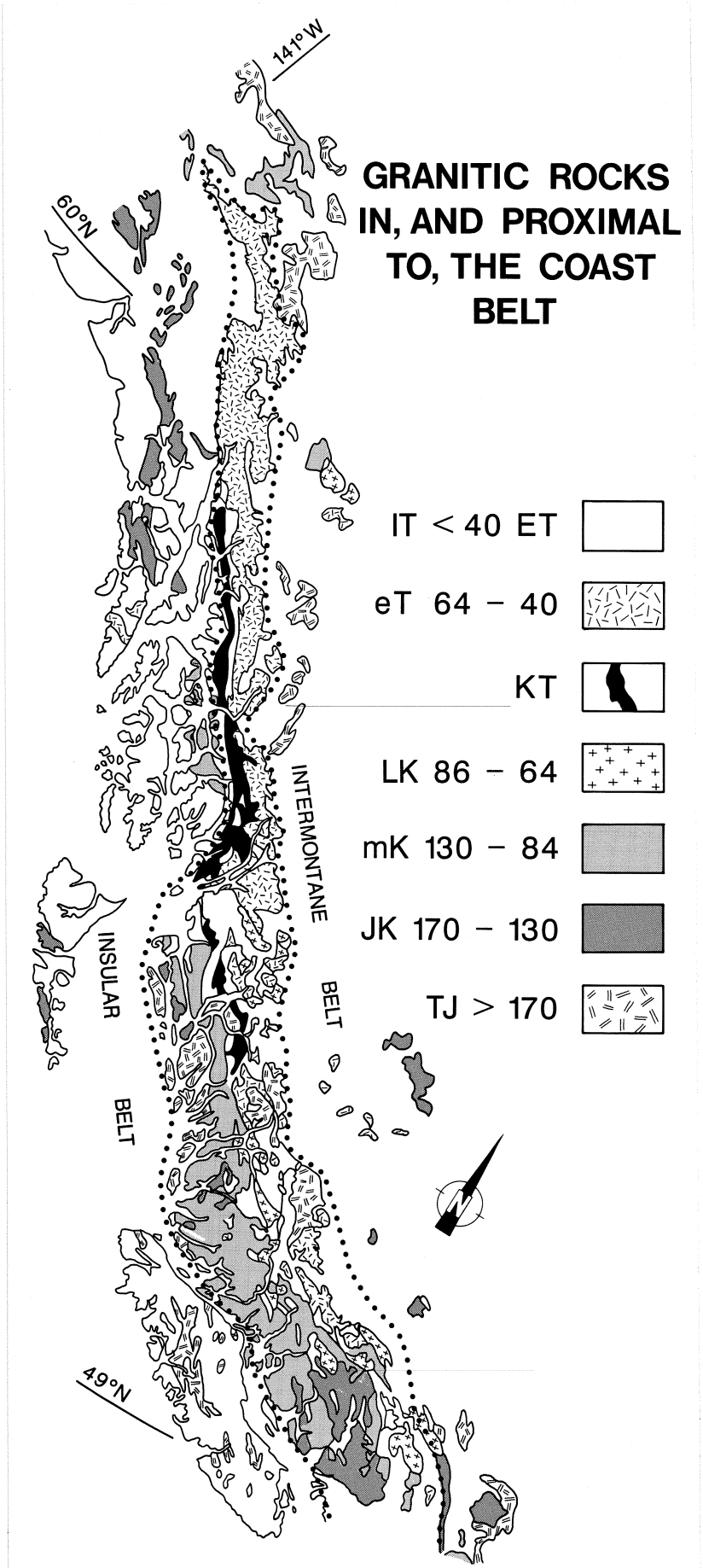



**Figure 3.** Simplified terrane map of the Canadian Cordillera; summary descriptions of terranes in and near Coast Mountains in Table 1; others in Gabrielse et al., 1991. Pericratonic terranes (CA, KO, NAM, YT) that formed along the margin of the craton have no ornamentation; accreted terranes include (1) magmatic arc terranes (AX, QN, ST, WR) shown hachured; (2) oceanic terranes (CC, CR, SM) incorporated as subduction complexes shown black; CD (Cascade terranes) shown as a mixture of oceanic and arc terranes, are in detail on Figure 6 (in back pocket). Heavy dots delineate Intermontane, Insular superterrane boundaries.

### *Plutonic rocks*

The plutonic rocks of the Coast Mountains are mainly quartz diorite to granodiorite, with an average composition of tholeiitic andesite in the Irvine and Baragar (1971) classification (Roddick, 1983; Armstrong, 1988). The general distribution of ages of granitic rocks in and near the Coast Mountains is shown in Figure 4; it should be recognized that far more isotopic dating is needed even on a reconnaissance level, particularly between latitudes 51°N and 53°N. South of Prince Rupert (lat. 54°N), most granitic rocks are late Middle Jurassic to mid-Cretaceous (ca. 165–91 Ma) in western parts, and mid-Cretaceous to early Tertiary (110–46 Ma) in eastern parts, with plutons ranging in age from about 110–90 Ma common to both parts (Fig. 4; Armstrong, 1988; Friedman and Armstrong, 1990, in press; Wheeler and McFeely, 1991; Monger and McNicoll, 1993). North of Prince Rupert, the plutonic rocks of the Coast Mountains are Late Cretaceous and early Tertiary (84–46 Ma; Gehrels et al., 1991); Late Jurassic-Early Cretaceous plutons lie to the west, in the Insular Belt.




A distinctive suite of high-pressure (ca. 8 kb) mid-Cretaceous (ca. 95 Ma) quartz diorite plutons is present just west of the western boundary of the Coast Mountains in southern southeast Alaska, and coeval, high-pressure plutons occur in the southeastern Coast and Cascade mountains, and on the west side of the Idaho Batholith (e.g., Zen, 1985). On the Tectonic Assemblage Map of the Canadian Cordillera, Wheeler and McFeely (1991, and Fig. 4) include these high-pressure plutons within a broader suite of mid-Cretaceous igneous rocks (130–87 Ma) that also contains granodioritic plutons emplaced at high crustal levels in the southwestern Coast Mountains.

In places, Paleozoic, Triassic, and Early Jurassic plutons and orthogneisses are present in the Coast Mountains, but are a minor component of the whole. As these rocks are coeval with plutons that are integral parts of flanking terranes (Fig. 3; mainly Stikinia and Yukon-Tanana (or Nisling) north of lat. 52°N, Wrangellia and Alexander terrane south of lat. 54°N), they bear no direct relationship to the Late Cretaceous and younger plutonic suites that give integrity to the Coast Belt. Under debate is whether Middle Jurassic to Early Cretaceous plutons and associated arc-related volcanic and sedimentary strata (Gambier Group; Gravina-Nutzotin strata), emplaced in and overlying Wrangellia, Alexander and Harrison terranes, are specific to these terranes, or are part of a broader "ancestral Coast Belt" (van der Heyden, 1992; Monger et al., 1994).

### *Structures*

Major structures affecting rocks of the entire Coast Mountains are mid-Cretaceous to early Tertiary in age and include systems of both west and east-vergent contractional faults, right-lateral strike-slip faults and Eocene extensional faults (Fig. 5, Woodsworth et al., 1983; Monger, 1986; Crawford et al., 1987; Rubin et al., 1990; Journeay, 1990a,b; Rusmore and Woodsworth, 1991; Coleman and Parrish, 1991; Journeay and Friedman, 1993; Varsek et al., 1993).

West-vergent thrust faults in southeast Alaska, near Prince Rupert, and in northwestern Washington are interpreted to be segments of a regionally extensive contractional belt that formed in mid- to Late Cretaceous time (Fig. 5; Rubin et al., 1990). In the southern Coast Mountains, the Coast Belt Thrust System (CBTS; Journeay and Friedman, 1993) is part of this

**Figure 4.** Distribution of ages of Mesozoic and Tertiary granitic rocks in the Coast Belt (simplified from Wheeler and McFeely, 1991).

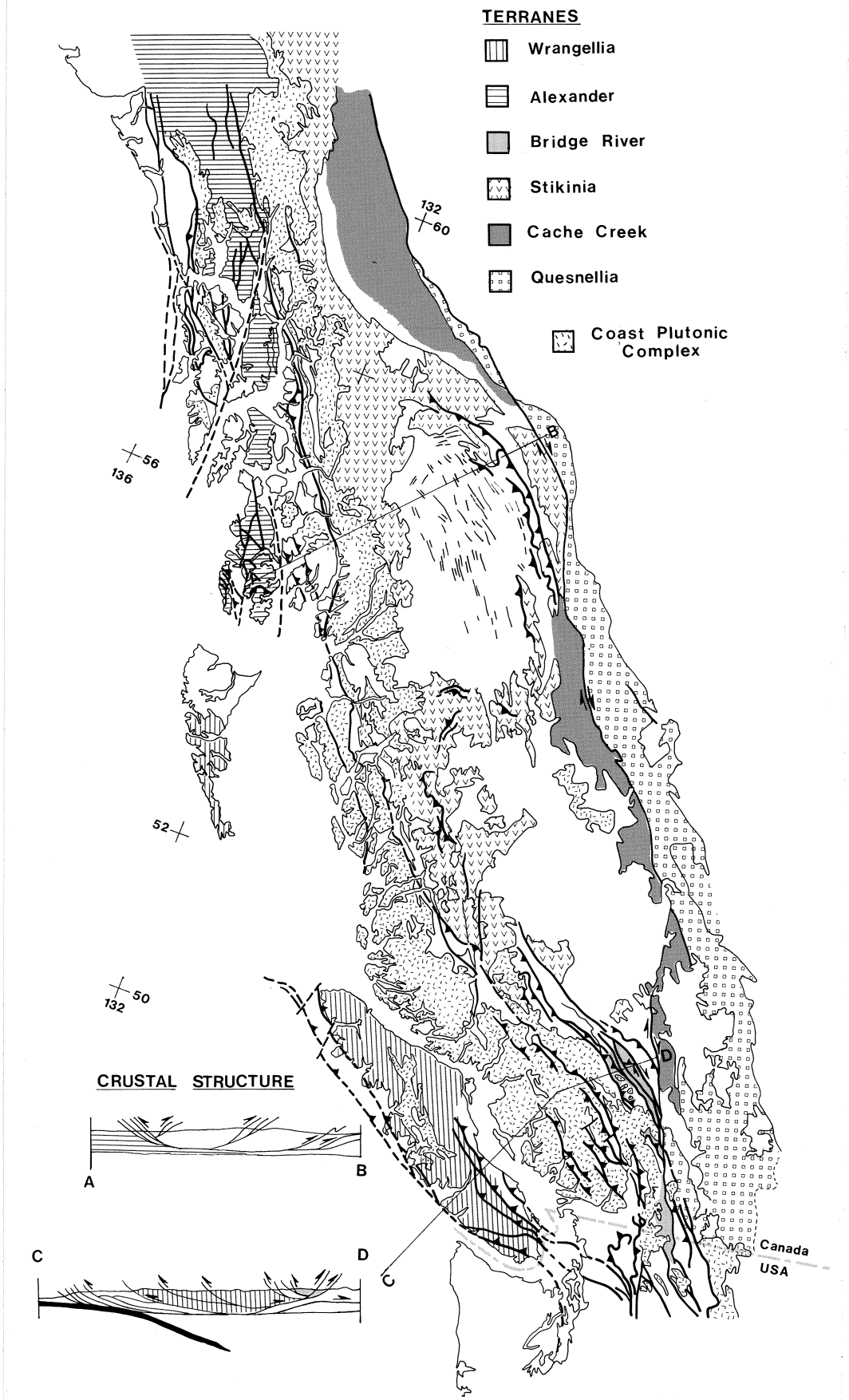



**Figure 5.** Major Cretaceous and early Tertiary structures within and flanking the Coast Belt. Modified from Rubin et al. (1990), Rusmore and Woodsworth (1991), Evenchick (1991), and Wheeler and McFeely (1991).

contractional belt and is an important structural link between thrust systems of the Cascade Ranges and the northern Coast Mountains. In the CBTS, crosscutting plutonic suites constrain timing of west-vergent thrusting to between 97–91 Ma. The thrust sheets and bounding faults of the CBTS extend south into the North Cascade ranges (Crickmay, 1930; Misch, 1966; Lowes, 1972; Davis et al., 1978; Monger, 1986). Northward, they have been traced within the Coast Mountains to lat. 51°N; west-vergent faults are not well-documented between lats. 51° and 53°N, possibly due to the abundance of post-kinematic Late Cretaceous and Tertiary intrusions in this segment of the Coast Mountains. North of lat. 54°N, thrust faults of the Prince Rupert shear zone carry metasedimentary rocks of the Central Gneiss Complex westward over lower grade rocks of the Alexander terrane and overlapping strata of the Jura-Cretaceous Gravina belt (Crawford et al., 1987). Crosscutting plutonic suites constrain the timing of west-vergent thrusting between 100–84 Ma in the Prince Rupert region (Crawford et al., 1987) and between 95 and 90 Ma for equivalent structures in southern Southeast Alaska (Rubin and Saleeby, 1987; Arth et al., 1988; McClelland et al., 1990). The similarities in geometry, timing and regional tectonic setting support correlation of segments of this west-directed thrust system. The west-vergent contractional fault system in both northern and southern parts of the Coast Mountains coincides with the inboard, eastern margin of the Insular superterrane (Fig. 3). This consists of Wrangellia, Alexander and Harrison terranes, linked by Middle Jurassic time (ca. 165 Ma), if not before, by the suite of magmatic and sedimentary rocks (Gravina-Nutzotin-Gambier volcanic-plutonic-sedimentary belt of Nokleberg et al., in press) in the western part of the southern half of the Coast Mountains and the northern half of the Insular Belt. The margin crosses the axis of the Coast Mountains at an acute angle, from the Harrison Lake-Lillooet River Valley in the southeast to the coastal fiord region on the west side near Prince Rupert and farther north. The west-vergent thrust system is consistent with the hypothesis that shortening was driven by oblique convergence and northeastward underplating of the Insular superterrane beneath Intermontane superterrane and intervening "Cascade" terranes (Bridge River, Cadwallader and Methow) in early Late Cretaceous time (Davis et al., 1978; Monger, 1986; Crawford et al., 1987; Journeay, 1990a; Rubin et al., 1990; McGroder, 1991).

East-directed thrust faults are recognized in various places along the east side of the Coast and Cascade mountains, and are antithetic to the west-directed structures (Fig. 5; Wheeler and Gabrielse, 1972; Schiarizza et al., 1990b; Rusmore and Woodsworth, 1991; McGroder, 1991; Journeay et al., 1992). Although physical continuity has not been established, they may correlate with the extensive Cretaceous, northeast-directed thrust system in the Bowser Basin of the Intermontane Belt of north-central British Columbia (Fig. 5; Evenchick, 1991). In the eastern Coast Mountains near of lat. 52°N, these structures root westward beneath high-grade metamorphic rocks, and crosscutting plutons indicate that the faults were active between 86 Ma and 68 Ma (Rusmore and Woodsworth, 1991). In the south, east-directed thrust faults on the east side of the Cascade Ranges root westward beneath the Cascade metamorphic core and appear to have been active between 100–88 Ma (McGroder, 1991).

At the latitude of the field trip, the pattern of west- and east-vergent folds is obscured at the surface by offset of about 130 km on the Eocene dextral strike-slip Fraser-Straight Creek Fault. However, the pattern at depth of seismic reflectors in the Lithoprobe transect, some of which align with thrust faults at surface, suggests that the southeastern Coast Mountains are a two-sided contractional orogen whose crust is composed of a stack of west- and east-vergent rootless tectonic flakes (cross-section on Fig. 6). East-directed thrust faults intersect and displace west-directed thrusts at depths of between 9–18 km (3–6 s two-way travel time) below the southeastern Coast Mountains, and apparently root westward into a decollement zone at the base of the crust at about 12 s; (section on Fig. 6; Journeay, in Monger et al., 1990; Varsek et al., 1993). The fabric and scale of the seismic images in the southeastern Coast Mountains generally resemble those across the Alps and Pyrenees respectively reported by Pfiffner et al. (1991) and Blois et al. (1991). Note that the Alps and Pyrenees are considered to be "collisional orogens." The main difference between them and the Coast Mountains is that there is no deep crustal root beneath the southeastern Coast Mountains; the Moho there is relatively flat, and the locus of active subduction is west of Vancouver Island.

Orogen-parallel dextral strike-slip faults of Late Cretaceous and earliest Tertiary age contribute to the general transpressional fabric of the southern Coast Mountains (Monger, 1986; Brown and Talbot, 1989; Schiarizza et al., 1990a,b), but have not been recognized in the northern Coast Mountains. Suggestions for displacements on these structures range from a few hundred kilometres deduced from offsets of sedimentary basins (Kleinspehn, 1985), to displacements relative to the southwestern Intermontane Belt of nearly 2000 km, derived from interpretations of paleomagnetic data from the eastern margin of the Coast Mountains and the Cascade metamorphic core (Ague and Brandon, 1992; Wynne et al., 1993).

Eocene (ca. 47 Ma) extensional structures are recognized in the eastern margin of the southern Coast and Cascade mountains (e.g., Coleman and Parrish, 1991), and in regions just east of the Coast Mountains (Friedman and Armstrong, 1988). Such structures appear to be restricted to the eastern side of the southern Coast Mountains, but are widespread farther east.

A fragmentary record of pre-mid-Cretaceous deformation is recognized at scattered locatities within the Coast Mountains and flanking terranes; much has been obliterated by the widespread younger intrusions. There is evidence for Jurassic contraction in southwestern Coast Mountains, along the east side of the central Coast Mountains, and in the flanking southwestern Intermontane Belt rocks (Monger, 1991a; Greig, 1992; Greig et al., 1992; van der Heyden, 1992). Some rocks of the Bridge River terrane on the east side of the southern Coast Mountains were deformed and metamorphosed in blueschist facies, probably in Triassic time (Archibald et al., 1991; Monger, 1991a). A rare record of probable Jura-Cretaceous extension is preserved in parts of

southwestern Coast Mountains, west of the region of intense Late Cretaceous deformation (Monger, 1991a).

### *Metamorphism*

High-grade regional metamorphism in the Coast Belt is closely associated with the Late Cretaceous and early Tertiary plutons and structures. Where most completely studied (near lats. 54° and 50°N), high pressure/medium temperature (7–8 kb; 600–650°C) regional metamorphism can be directly related to west-directed thrust faults and to the mid-Cretaceous, locally epidote-bearing (=high pressure?) plutons noted above (Zen, 1985). The high-pressure metamorphism is succeeded by lower pressure/high temperature (≤ 4 kb) metamorphism, again associated with emplacement of Late Cretaceous-early Tertiary plutons (Woodsworth et al., 1983; Crawford et al., 1987; Journeay, 1990a). Widespread greenschist facies metamorphism occurs both with Late Cretaceous-early Tertiary deformation and with earlier episodes of Jurassic deformation and plutonism (van der Heyden, 1992). In the central, southern Coast Mountains, near Harrison Lake, zeolite and lawsonite-albite metamorphism is developed in, respectively, Jurassic and Early Cretaceous strata (Beaty, 1974; Hettinga, 1989). Triassic blueschist is present locally in rocks of the Bridge River terrane, in southeastern Coast Belt (Archibald et al., 1991).

### *Concepts of the tectonic evolution of the Coast Mountains*

Before the 1960s, Cordilleran tectonics were viewed within the geosynclinal framework. In 1881, Dawson surmised that the Coast Range (*sic*) had undergone great uplift, metamorphism and "Appalachian-type folding" in addition to introduction of great masses of hornblendic granite. From their work in the northern Canadian Cordillera, Gabrielse and Wheeler (1961) similarly recognized the Coast Mountains as the locus of long-lived plutonism, metamorphism and uplift, and interpreted it as the geanticlinal core of their Pacific orogen (Wheeler and Gabrielse, 1972), flanked by divergent thrust systems, with east-vergent thrusts antithetic to the dominant west-vergent thrusts. Thus, from the earliest days of geological mapping, the close association of plutonic rocks, metamorphism and structure, were recognized as characteristics of the Coast Mountains.

Following acceptance of the plate tectonic hypothesis, the Coast Plutonic Complex (*sic*) was identified as a Jurassic to Tertiary magmatic arc formed above an east-dipping subduction zone (Fig. 7A; Monger et al., 1972), a view that persists (Armstrong, 1988; van der Heyden, 1992). Within such a framework, the Late Cretaceous-early Tertiary metamorphism and deformation can be attributed to concentration of strain in the hot, weak, crust of a continental margin arc during strong plate convergence, as in the modern central Andes (Isacks, 1988).

In 1975, Godwin explained the distribution of porphyry copper deposits of the central Coast Mountains (ca. lat. 54°N), and crustal thickening in that region deduced from the high metamorphic grades, as the result of imbrication along two Late Cretaceous-early Tertiary east-dipping subduction zones, one west of the Queen Charlotte Islands and the other along the western margin of the central Coast Mountains.

The distribution of terranes in the Coast Mountains and their time-space relations, and especially the presence of the oceanic Bridge River terrane (Fig. 6) in the southeastern Coast Mountains (containing then recently identified Permian to Jurassic radiolarians; D.L. Jones, pers. comm. to Monger, 1977), and the Late Cretaceous metamorphism, structures and uplift, led Monger et al. (1982) to view the structure and metamorphism of the entire Coast Mountains as subsequent to mid-Cretaceous accretion of amalgamated Wrangellia and Alexander terranes and overlapping Jura-Cretaceous strata (=Composite terrane II or Insular superterrane) to the North American Plate (Fig. 7B). The suture in this model, within the Coast Mountains outlined by the terrane boundaries deduced from scattered variably metamorphosed septa, more-or-less follows the early Late Cretaceous west-directed thrusts of the later named Coast Belt Thrust System (cf. Fig. 3, 5).

However, north of lat. 54°N, Gehrels and Saleeby (1985) found no evidence for oceanic strata between (mainly) Stikine and Alexander terranes (Fig. 3). Instead, they suggested that Insular Superterrane moved northwards within the western edge of the North American plate in Jura-Cretaceous time (Fig. 7C) and was finally accreted in the Late Cretaceous. In this model, Jura-Cretaceous flysch basins along the western Cordilleran margin from southern Alaska to California, formed as pull-apart basins associated with northward, dex-

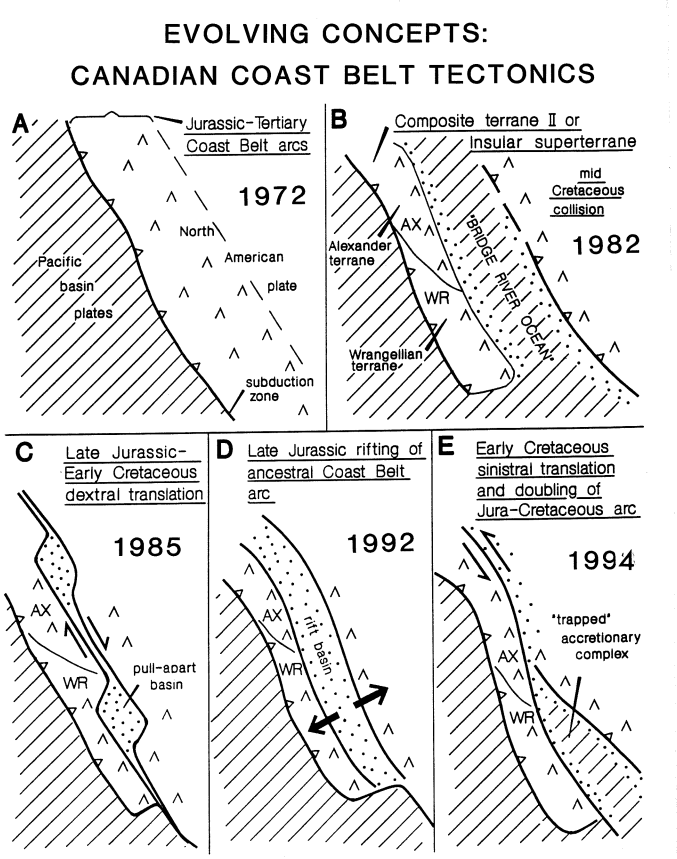



**Figure 7.** Evolving concepts of Coast Belt tectonics.

tral, translation of the Insular Superterrane, and were collapsed during its final accretion.

Plafker (1990) also proposed that the Insular Superterrane was translated northwards from Californian latitudes, but in Late Cretaceous-Paleocene time. He drew on reconstructions of (Kula) plate motions (Engebretson et al., 1985) in which sinistral and orthogonal convergence was succeeded after 85 Ma by dextral plate motions. Although this model agrees with plate reconstructions, and is supported by the structural evidence for Late Cretaceous-early Tertiary dextral transpression in the southern Coast Mountains and by paleomagnetic data, it does not explain the origin of the Jura-Cretaceous flysch basins.

In 1992, van der Heyden suggested on the basis of isotopic dating in the central segment of the Coast Mountains that the orogen is a long-lived (Middle Jurassic to Tertiary; ca. 165–47 Ma) Andean-style magmatic arc. It was built across Stikine, Wrangellia and Alexander terranes which were accreted in the Middle Jurassic (ca. 170 Ma) along a Cache Creek suture in the Intermontane Belt (Fig. 3). In this model, the Jura-Cretaceous flysch basins formed during intra-arc rifting, and closed in mid-Cretaceous time (Fig. 7D). This interpretation is based on the presence of Middle and Late Jurassic granites on both sides of the Coast Mountains near lat. 53°N, the absence of any record of oceanic crust within the Coast Mountains north of lat. 51°N, and the general eastward younging of granitic rocks across the Insular and Coast belts.

The models of Gehrels and Saleeby (1985) and van der Heyden (1992) require formation of Jurassic pull-apart and/or rift-related basins within an existing arc. However, in the southeastern Coast Mountains this appears not to be the case. There, Mississippian to Middle and possibly Late Jurassic strata of pelagic facies (radiolarian chert) of the oceanic Bridge River terrane (Cordey and Schiarizza, 1993; F. Cordey, pers. comm., 1994), grade upwards into clastic rocks of the Jura-Cretaceous Cayoosh assemblage (Journeay and Northcote, 1992; Mahoney and Journeay, 1993). Thus the basin containing the Jura-Cretaceous clastics appears to have been in existence since at least the Mississippian, and is the youngest stage of a long-lived, and therefore probably originally large, ocean basin. The style of structures and older metamorphism suggests that the Bridge River-Cayoosh rocks represent, at least in part, a Mesozoic accretionary complex, analogous to the Chugach terrane in southern Alaska (lat. 60°N), or western Klamath terranes and oldest parts of the Franciscan Complex in California-Oregon (lat. 40°N). To accomodate both the accretionary complex in southeastern Coast Mountains and the intra- or back-arc basin model of Gehrels and Saleeby (1985) and van der Heyden (1992), Monger et al. (1994; Fig. 7E) suggested that Bridge River and Cayoosh rocks were an accretionary complex trapped in Early Cretaceous time (120–100 Ma) east of a fragment of the Jura-Cretaceous North American plate margin arc (carried on Wrangellia and Alexander terranes), which was displaced southward (sinistrally) in the Early Cretaceous. Such movement is in accord with local structural evidence for sinistral displacements, with reconstructions of Early Cretaceous relative motions between North American and various Pacific plates (Engebretson et al., 1985; D. Engebretson and K. Kelley, pers. comm., 1993), and permitted by paleomagnetic interpretations (Irving and Wynne, 1990). In Late Cretaceous time, the entire Coast Belt underwent contraction and dextral translation, collapsing the different kinds of basins lying between Insular Superterrane and North America.

Figure 7, in its variety, illustrates some of the explanations that have been offered to explain the tectonic evolution of the Coast Mountains. The explanations change as we acquire data. More data are needed to satisfactorily resolve the problem, and some fundamental questions require answers. For example, are the Late Cretaceous-early Tertiary plutonic, structural and metamorphic features that give the Coast Mountains integrity due to Andean-style arc collapse during strong plate convergence, or to accretion and underplating of Insular Superterrane, or to both? The available evidence suggests to us that accretion is important. Are the Late Cretaceous-early Tertiary granitic rocks arc-related and formed above a subducting oceanic slab, or were they generated by anatectic melting of juvenile crust thickened during contraction, or both? Do the Coast Mountains include relics of Jura-Cretaceous arc systems carried by Wrangellia and Alexander terranes, that migrated across an ocean basin, or are they parts of a single North American plate margin arc fragmented and duplicated by transcurrent faulting along the margin of the Pacific Ocean basin and its ancestors? We favour the latter. To resolve these questions requires not only wider knowledge of the composition and ages of the plutonic rocks, the kinematics of their structures, and the plate tectonic framework, but also of the stratigraphy of hosting terranes, and paleogeographies determined from paleomagnetism and paleontology.

## GEOLOGY OF THE SOUTHERN COAST MOUNTAINS

New information on the southern Coast Mountains includes geological map coverage at 1:50 000 to 1:500 000 scales (e.g., Journeay, 1990a; Schiarizza et al., 1990a,b; Journeay et al., 1992; Monger, 1989a, 1991b, 1993) (Fig. 1, in pocket), and theses mainly at the universities of British Columbia, Western Washington and Washington. In addition to ongoing conodont and radiolarian studies (e.g., Cordey and Schiarizza, 1993), about 50 new U-Pb dates of granitic rocks have been added to the Coast Mountains geological database south of lat. 51°N, as part of the Lithoprobe supporting geoscience program and on-going GSC mapping (Friedman and Armstrong, 1990, in press; Parrish and Monger, 1992; Journeay and Friedman, 1993; Monger and McNicoll, 1993). Integration of Lithoprobe seismic reflection data with structural profiles drawn from geological mapping (Journeay in Monger et al., 1990 and Fig. 6; Cook et al., 1992; Varsek et al., 1993) provides a window into the deep crustal structure of the southern Coast Mountains, and the mechanisms of crustal deformation.

That part of the Coast Mountains between lats. 49° and 51°N is divided into **southeastern** and **southwestern** parts based on the distribution of plutonic rocks, terranes and structures (Fig. 1, 6, 8; Table 1; Monger and Journeay, 1994).

The two-fold division of the southern Coast Mountains was first recognized by Crickmay (1930) who saw the southwestern part as a region of plutonic rocks and "Jurasside structures" which were distinct from the region of "Laramide structures" in the southeastern Coast Mountains and Cascade Range. Misch (1966; his Fig. 7-17), showed a similar division. Mid-Cretaceous (ca. 110–90) plutonic rocks are emplaced in both parts of the southern Coast Mountains, linking them at a time slightly earlier than that at which the earliest structural linkages can be demonstrated.

The **southwestern Coast Mountains** feature mainly Middle Jurassic to mid-Cretaceous plutons (ca. 165–91 Ma) which intrude supracrustal sequences ranging in age from Middle Triassic to Middle Jurassic of Wrangellia and Harrison Lake terranes, and overlapping Jura-Cretaceous volcanic and sedimentary rocks (Fig. 8, 9). The western boundary is the western limit of Middle Jurassic intrusions that possibly were localized along pre-and syn-plutonic faults (Nelson, 1979; Monger and McNicoll, 1993). Its eastern boundary is delineated by the high-grade, internal, metamorphic thrust nappes of the Coast Belt Thrust System (CBTS) that are derived in large part from basinal strata (Bridge River terrane) characteristic of southeastern Coast Mountains (Fig. 1, 8, 9; Journeay and Friedman, 1993). Note that rocks (Harrison terrane and Gambier Group) characteristic of the eastern part of southwestern Coast Mountains also are internally imbricated along west-directed thrust faults of the external part of the CBTS (Fig. 1, 9), below nappes featuring high grade metamorphism to the east. Thus, the southwestern Coast Mountains is a plutonic-dominated crustal block that acted as a foreland buttress during early Late Cretaceous (91–97 Ma) west-directed thrusting centred in southeastern Coast Mountains, essentially as first recognized by Crickmay (1930).

The **southeastern Coast Mountains** feature mid-Cretaceous through early Tertiary (103–47 Ma) plutonic rocks, emplaced within (mainly) Bridge River, Cadwallader and Methow terranes (Fig. 1, 6, 8; Table 1). This part of the Coast Mountains was the site of the most intense deformation and highest grade metamorphism (Fig. 1, 3) in Late Cretaceous-early Tertiary time. All three terranes in the southeastern Coast Mountains appear to be founded on oceanic crust (Fig. 6, 8; Table 1). Of these, the Bridge River terrane consists of lithologies characteristic of oceanic crust, such as pillow basalts associated with radiolarian chert and alpine-type ultramafic rocks, whereas the other terranes are predominantly of arc affinity. Bridge River terrane contains fossils in chert facies ranging in age from Mississippian to late Middle Jurassic (350–160 Ma; Cordey and Schiarizza, 1993) and possibly Late Jurassic (Oxfordian, 155 Ma; F. Cordey, pers. comm., 1994). The long time spanned by this terrane, at least 190 million years, suggests that it represents remnants of a former large ocean basin. In addition, Journeay and Northcote (1992) found that locally chert grades into clastic rocks of the Cayoosh assemblage of Jurassic(?) and Early Cretaceous age. The Bridge River and overlying strata occur both as highly disrupted tectonic melange and relatively coherent units. Rare Triassic blueschist in the Bridge River terrane indicates that part underwent early Mesozoic subduction; other $^{40}Ar/^{39}Ar$ data show that the terrane underwent no other thermal perturbation until the Cretaceous (Archibald et al., 1991). Monger (1991a) argued that the Darrington Phyllite and Shuksan Greenschist of the North Cascades, which contain blueschist yielding Jura-Cretaceous ages (peaking at 140–120 Ma), can be traced northwards into Settler Schist which is correlated with the Cayoosh assemblage. The Bridge River terrane is disposed in thrust slices structurally interleaved with Cadwallader and(?) Methow terranes, variably metamorphosed and intruded by Late Cretaceous and early Tertiary plutons.

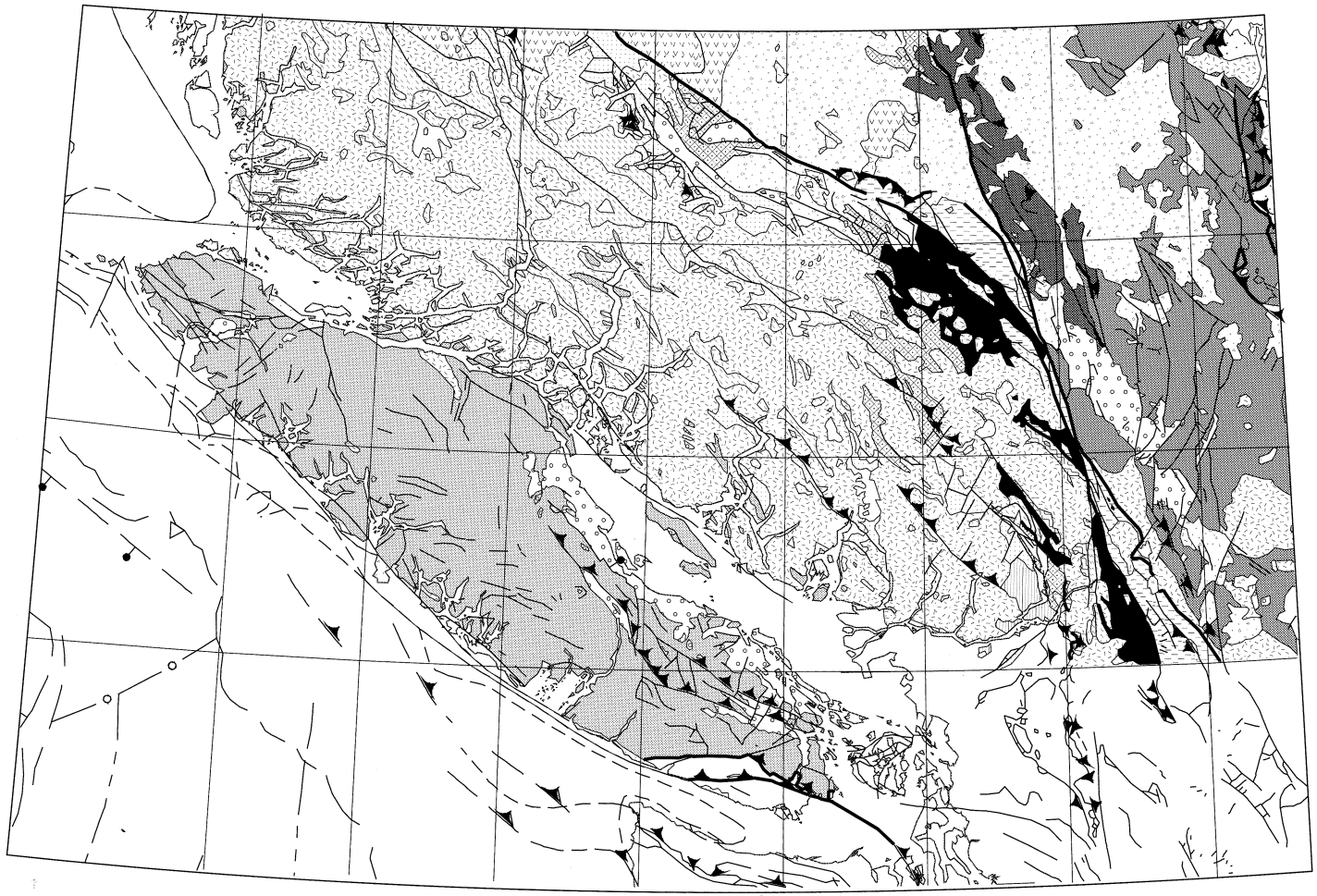
On the east, the Coast Mountains are flanked by the **southwestern Intermontane Belt** which contains Quesnel, Cache Creek and (north of lat. 52°N) Stikine terranes (Fig. 6, 8, 9; Table 1) and plutonic rocks ranging in age from Late Triassic to earliest Tertiary (212–62 Ma). Between lat. 51–52°N, Late Cretaceous-early Tertiary plutons characteristic of the Coast Mountains are emplaced in possible Stikine strata (Fig. 3) and in Early Cretaceous strata that may be correlative with the Gambier Group in southwestern Coast Belt.

The along-strike extension of the Coast Mountains to the south across Fraser River is the **Cascade Range.** This comprises a core of Late Cretaceous-early Tertiary plutons and high-grade metamorphic rocks, equivalent to rocks in southeastern Coast Mountains, but offset dextrally by ca. 130 km on the early Tertiary Fraser-Straight Creek Fault (Fig. 1). The core is flanked by little metamorphosed strata; on the east are Methow and Bridge River terranes, and on the west rocks of Bridge River (?), Chilliwack and Harrison terranes. Not recognized in the Cascades are Wrangellian strata (Dickinson, 1976; Davis et al., 1978), and the voluminous Middle Jurassic to Early Cretaceous plutonic rocks so abundant in the southwestern Coast Mountains.

### *Coast Belt Thrust System*

The Coast Belt Thrust System (CBTS) is a major west-vergent contractional belt that formed along the inboard margin of the Insular superterrane in Late Cretaceous time (Fig. 1, 5, 6, and 9; Journeay and Friedman, 1993). The foreland of this contractional belt comprises rocks of southwestern Coast Mountains and Wrangellia, which are deformed by a system of east-dipping frontal thrusts that imbricate Jurassic and Lower Cretaceous (mainly) arc sequences and related plutons of the southwestern Coast Mountains. Local post- or syn-Albian folds and thrusts that predate the dominant east-dipping structures have southeast-directed vergence (Lynch, 1992). The dominant faults root eastward into an imbricate zone of folded thrusts and out-of-sequence reverse faults across which the metamorphic gradient is structurally inverted. To the east and structurally above the imbricate zone is a folded stack of thrust sheets mainly comprising metamorphosed island arc and oceanic rocks (mainly Cadwallader and Bridge

**Facing Page:**
**Figure 9.** Map derived from 1:1 000 000 scale Tectonic Assemblage map of the southern Coast Mountains (Vancouver, NM-9), modified from "The Tectonic Assemblage Map of the Canadian Cordillera, "by Wheeler and McFeely (1991; GSC Map 1712A) showing distribution of terranes, plutonic and metamorphic complexes, and overlap assemblages.



Pacific Rim/ Crescent

Gambier Assemblage

Cadwallader

Metamorphic Assemblages

Methow

Quesnellia

Coast Plutonic Complex

Wrangellia

Harrison

Bridge River

Shulaps Complex

Cache Creek

Stikinia

LK clastic and Tertiary volcanic overlap asssemblages

River terranes). These rocks represent the exhumed crustal root of the CBTS. Metamorphosed volcanic and volcaniclastic rocks that occur in westernmost, structurally lowest, thrust sheets of the imbricate zone yield U-Pb zircon dates which support their correlation with foreland/lower plate Jurassic and Early Cretaceous arc sequences of southwestern Coast Mountains.

In the southeastern Coast Mountains, southwest-directed structures are cut by northeast-directed thrust faults and by sinistral and dextral transcurrent faults of Late Cretaceous and early Tertiary age (Schiarizza et al., 1990b; Journeay et al., 1992; Parrish and Monger, 1992). These structures are in turn truncated by the dextral transcurrent Fraser Fault and offset about 130–140 km to the north from equivalent structures in the Cascade metamorphic core (Fig. 9).

In the southwestern Coast Mountains the effects of Late Cretaceous contraction are far less, and typically manifested by discrete shear zones spaced up to 10 km apart, that are steeply dipping and mainly southwest, locally northeast, vergent (Fig. 1; Monger, 1991b, 1993; Lynch, 1992). Some major early Late Cretaceous plutons were emplaced during the period of Coast Belt Thrust System contraction (Monger and McNicoll, 1993) and one (ca. 91 Ma), at least, postdates it. In rocks between the shear zones, sparse records of Jurassic contraction (between 185–155 Ma) and Jura-Cretaceous extension are preserved (Monger, 1991b,c).

### *Metamorphism*

The metamorphic grade varies considerably across the southern Coast Mountains. In the southwestern part, metamorphism in septa is mostly of greenschist facies, with higher metamorphic grades locally present at the contacts with large plutons. The grade locally drops to zeolite facies in Jurassic rocks just west of Harrison Lake (Beaty, 1974), and is of lawsonite-albite facies in Early Cretaceous rocks along west and east shores of Harrison Lake, and in thrust sheets in the lowest part of the Coast Belt Thrust System (CBTS; M.T. Brandon, pers. comm., 1986; Hettinga, 1989). Within 20 km to the east, across the imbricate zone of the CBTS, the grade rises to sillimanite-kyanite facies (Lowes, 1972; Pigage, 1976; Monger, 1986; Hettinga, 1989; Journeay, 1990a). A synthesis of available data from this area suggests that peak metamorphic conditions associated with 91–97 Ma thrust faulting are 7.0–9.6 kb and 600–750°C. The conditions dropped to 3.7–6.1 kb and 575–690°C by 84 Ma (Journeay, 1990a). Note that the metamorphic grade of the southeastern Coast Mountains decreases along strike towards the northwest, so that where traversed on Day 2, the southeastern Coast Mountains are entirely greenschist or lower grade.

In the easternmost Coast Mountains, metamorphic grade generally decreases from greenschist to prehnite-pumpellyite facies to the northeast. Crosscutting relationships with dated plutonic rocks indicate that metamorphism in this part of the eastern Coast Mountains is Late Cretaceous and early Tertiary. Near Lillooet, high greenschist grade rocks are exposed in slivers bounded by outward-dipping Middle Eocene normal faults (Coleman and Parrish, 1991). An older metamorphic history is present locally. $^{40}Ar/^{39}Ar$ dating in adjacent parts of the eastern Coast Mountains reveals local Late Triassic (221 Ma) blueschist metamorphism in rocks of the Bridge River terrane, followed by Early or mid-Cretaceous (ca. 130 Ma or 110–95 Ma) regional and/or contact metamorphism (Archibald et al., 1991).

### *The foreland basin of the Coast Mountains*

Clastic foreland basin deposits of the southern Coast Belt are represented by Late Cretaceous (Turonian to Maestrichtian; ca. 90–65 Ma) ''molassic'' marine and nonmarine clastic rocks of the Nanaimo Group (Fig. 1), which lap up on to the southwestern Coast Mountains, and are best-exposed on Vancouver Island and the Gulf Islands, where they overlie Wrangellia (Fig. 1; Mustard, 1991, 1994; Mustard and Rouse, 1991, 1992). The oldest Nanaimo deposits (early Turonian, ca. 90 Ma, Haggart, 1992) are slightly younger (ca. 2 Ma) than the 91-97 Ma thrust faults of the CBTS (Journeay and Friedman, 1993). The Nanaimo Group is involved in Eocene, thick-skinned southwest-directed thrust faulting (England and Calon, 1991), and appears to be an erosional remnant preserved in the late Tertiary(?)-Holocene Georgia Depression, centred along Georgia Strait between Vancouver Island and the mainland. England and Calon (1991) regard Nanaimo Group deposition as occurring in a **forearc** basin that lies oceanward of the coeval plutonism in southeastern Coast Mountains. Our view, which coincides with that of Brandon et al. (1988) for the northwest Cascades Ranges, is that the Nanaimo Group was deposited in a **foreland** basin built on the Wrangellian-southwestern Coast Mountains basement, and that clastic sedimentation resulted from tectonic uplift and erosion associated with the Coast Belt Thrust System, rather than erosion from a volcanic edifice built by extrusion. Separation of the Nanaimo foreland basin from the complementary thrust system is at least in part due to uplift of several kilometres and erosion of the southwestern Coast Mountains in late Tertiary to Recent time (Fig. 10; Parrish, 1983; Holdahl et al., 1989; Monger, 1991b; Monger and Journeay, 1994).

### *Terrane linkages in southern Coast Mountains and adjacent areas*

Terranes that were associated with one another prior to mid-Cretaceous (110 Ma) plutonism and Late Cretaceous (97 Ma) contraction and transpression fall into the following groups (Fig. 6, 8, 11; Table 1).

***Southwestern Coast Mountains and Insular Belt:*** Wrangellia is linked to Harrison terrane by Middle and Late Jurassic (165–147 Ma) plutonic rocks, which intrude Wrangellia on the west and extend more-or-less continuously across southwestern Coast Mountains to intrude Harrison Lake terrane west of southern Harrison Lake, and by widespread Early Cretaceous (Berriasian to Albian; ca. 140–100 Ma) calcalkaline volcanic and sedimentary rocks of the Gambier Group (Fig. 6: Friedman et al., 1990; Friedman and Armstrong, 1990, in press; Monger, 1991b). Wrangellian Triassic(?) and Early Jurassic strata in septa of southwestern Coast Mountains extend at least as far east as southern Howe Sound and middle Jervis Inlet, to within about 70 km of Harrison strata west of Harrison Lake (Friedman et al., 1990; Monger, 1993).

Jurassic strata of the two terranes are separated by Cretaceous intrusions and Cretaceous strata along Howe Sound, and by Jura-Cretaceous intrusions east of it.

***Southeastern Coast Mountains:*** Bridge River, Cadwallader and Methow terranes are founded on oceanic crust of, respectively, pre- or syn-Mississippian, Permian and Triassic ages (Fig. 6, 8, 11). On a non-palinspastic map these appear to underlie a trough that existed until the latest Early Cretaceous, between the southwestern Coast Mountains-Wrangellian block to the west, and the Intermontane Belt to the east (Fig. 12).

Until the Late Jurassic, the histories of these three terranes apparently differed. Bridge River terrane features abundant Upper Triassic to Middle and possibly Late Jurassic (Oxfordian; ca. 156 Ma) radiolarian chert, Cadwallader terrane contains Upper Triassic arc and Lower Jurassic clastic sedimentary facies, and Methow terrane contains a Jurassic clastic succession including Middle Jurassic arc volcanics (Fig. 6, 8). There is general agreement that Bridge River, Cadwallader and Methow terranes were linked by the mid-Cretaceous (ca. 100–110 Ma) as Albian chert-rich clastics derived from the Bridge River terrane occur in the uppermost parts of Cadwallader and Methow terranes (Taylor Creek and Virginian Ridge Formations, Garver, 1992). Prior to this time, linkages between the three terranes are tenuous, but possible, and Late Jurassic clastic basinal strata may overlap all three terranes (Fig. 11; Journeay and Mahoney, 1994).

The linkages between these terranes and flanking regions are also tenuous. The presence of radiolarian chert as young as Middle and(?) Late Jurassic (ca. 160-(?)156 Ma) in the Bridge River terrane suggests that pelagic conditions persisted until the later Jurassic, and that the arc to the west, represented by Middle Jurassic (167 Ma) and younger plutons, was

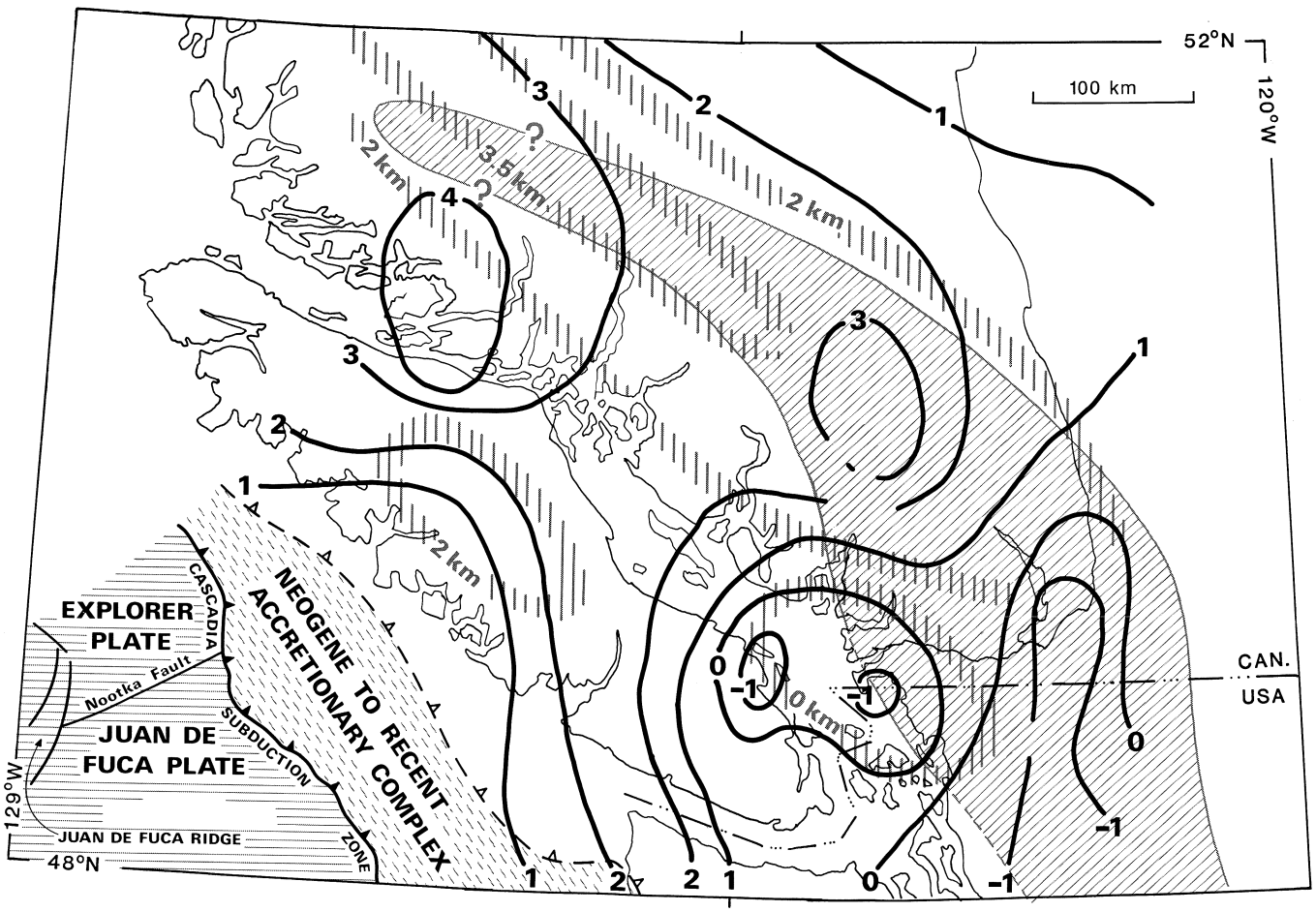



**Figure 10.** Amounts of uplift in southwestern British Columbia in the last 10 million years and present rates of uplift in millimetres per year.

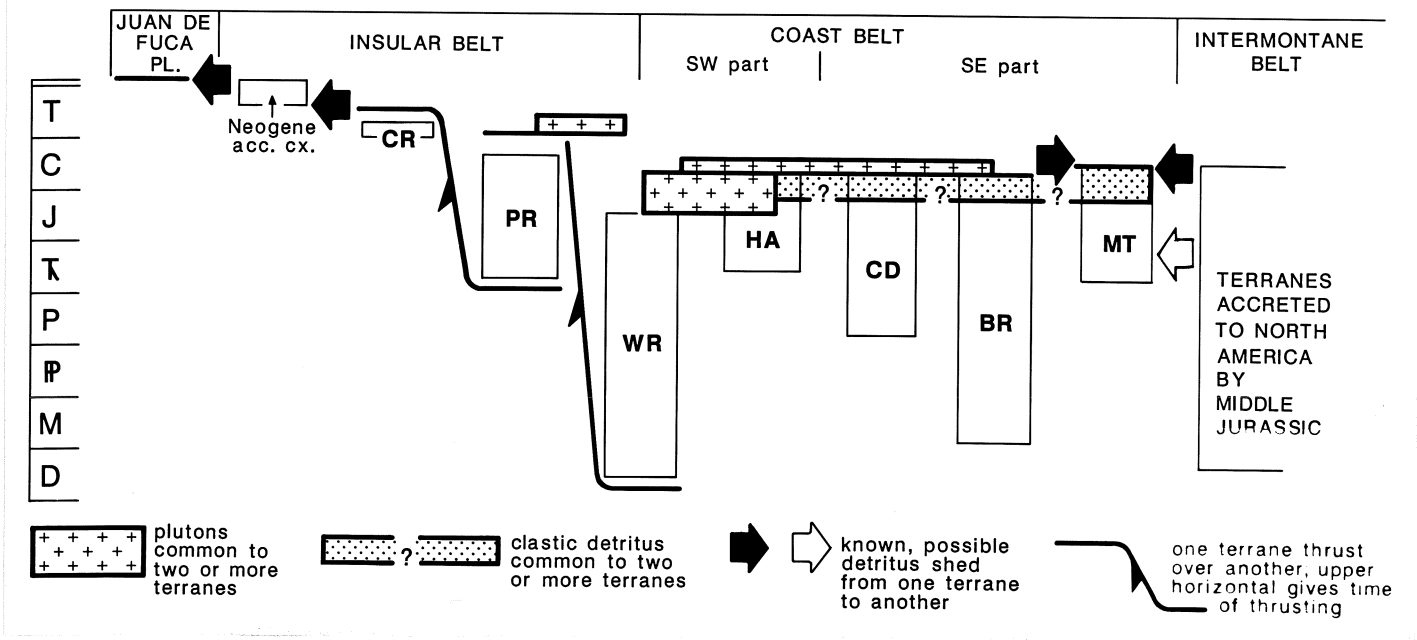



**Figure 11.** Times and natures of terrane linkages, southwestern British Columbia; BR Bridge River; CD Cadwallader; CR Crescent; HA Harrison; MT Methow; WR Wrangellia.

either not present at this latitude or else was separated by a wide basin. To the west, it has long been recognized that some clastic rocks in the late Callovian to Hauterivian (160–135 Ma) Relay Mountain Group, which may lie on Cadwallader terrane, have a southwestern provenance; these have been linked to uplift in the southwestern Coast Mountains (Jeletzky and Tipper, 1968). Tuff beds within the Relay Mountain Group may record Gambier Group volcanism (P. Schiarizza, pers. comm., 1990). To the east, late Early Cretaceous clastic rocks of Methow terrane long have been considered as having been derived from the uplifted western margin of Quesnel terrane and/or forearc deposits of the continental Spences Bridge Group, which is founded on Quesnel and Cache Creek terranes (e.g., Coates, 1970; Tennyson and Cole, 1978; Kleinspehn, 1985; Thorkelson and Smith, 1989). However, recent interpretations of paleomagnetic data (Wynne et al., 1993) from rocks that appear to be correlative with those in uppermost Methow terrane, and comparisons with paleomagnetic data from Spences Bridge volcanics, cast doubt upon these linkages, and suggest Quesnel and Methow terranes were separated in the early Late Cretaceous by ca. 1800 km (*see* discussion by Cowan, 1994).

The complex structure of the southern Coast Mountains, including the ''tectonic flaking'' recorded by surface structures and interpreted from seismic reflection data (Fig. 6), makes it difficult to restore these terranes to their relative pre-mid-Cretaceous paleogeographies. Based on paleocurrent data from chert-rich clastics derived from the Bridge River, Garver (1992) inferred that the probable order from west to east in mid-Cretaceous (Albian to Cenomanian) time was Cadwallader, Bridge River and Methow terranes (as in Fig. 8, 12). McGroder (1991) suggested a minimum of 500 km of shortening across the Cascade orogen along strike to the south. In addition, as noted above, displacement on dextral, orogen-parallel faults (e.g., Brown and Talbot, 1989) may be considerable; estimates range from a few hundred kilometres based on geology (e.g., Kleinspehn, 1985) to nearly 2000 km based on paleomagnetic interpretations.

***Intermontane Belt:*** The Intermontane Belt comprises Quesnel, Cache Creek, and (farther north) Stikine terranes (Fig. 3, 6, 8, 9, Table 1). It is separated from Methow terrane by the Cretaceous Pasayten Fault and the Tertiary Fraser Fault. On the field trip, we visit only the westernmost part of Quesnel terrane at one stop. Quesnel terrane is characterized by a west-facing, Late Triassic-Early Jurassic arc (Nicola-Rossland volcanics; Mortimer, 1987), built upon older oceanic and arc rocks (Fig. 8; Monger, 1989a). The arc is unconformably overlain by non-volcanic clastic rocks of the Lower to Middle Jurassic (late Pliensbachian to earliest Callovian; ca. 190–160 Ma) Ashcroft Formation. The early Mesozoic Quesnellian arc probably formed on the western edge of the North American plate, but was not emplaced on pericratonic and continental margin rocks to the east until late Early Jurassic time (ca. 185 Ma). Cache Creek terrane, east of the Fraser River in southern British Columbia, comprises Middle Pennsylvanian to Early Jurassic chert, pelite and volcaniclastic rocks, melange with blocks of chert, gabbro, ultramafics and limestone, and a large body of Late Permian to Late Triassic limestone. (Cordey et al., 1987; Monger, 1989a; Beyers and Orchard, 1991). Its lithologies, structural character and position relative to the early Mesozoic Quesnel arc, suggest that the Cache Creek represents the accretionary complex complementary to the Quesnellian arc.

In southern British Columbia, the Cache Creek terrane was thrust eastwards over Quesnellia and the Ashcroft Formation in Jurassic time (Travers, 1978). Quesnel and Cache Creek terranes are overlapped by mid-Cretaceous (ca. 105 Ma) continental volcanics of the Spences Bridge Group.

### *Coast Mountains-North Cascade correlations*

Early workers used both Coast and Cascade ranges as interchangeable names, and both are included within the Coast Belt (*see* Dawson, 1881; Daly, 1912; Gabrielse et al., 1991). The Cascade orogen is subdivided into three distinct tectonic domains, each of which contains some rocks correlative with those in the Coast Mountains (Fig. 1, 6; Misch, 1966; Tabor et al., 1989). (1) An eastern tract of low-grade metamorphic rocks intruded by scattered Late Cretaceous and early Tertiary granitic rocks is the southeastern extension (across the Fraser-Straight Creek Fault) of Bridge River and Methow terranes. (2) The Cascade core features abundant Late Cretaceous and early Tertiary plutons and high-grade metamorphic rocks. Some of which (Napeequa and Chiwaukum schists) resemble high-grade metamorphic rocks (Cogburn Group, Settler Schist) in the southeasternmost Coast Belt (Evans and Berti, 1986; Tabor et al., 1987; Monger, 1991a), that are derived from Bridge River and Cayoosh rocks. (3) A western tract, called the Northwest Cascades System (Misch, 1966; Brown, 1987), consists of mainly subgreenschist strata belonging to one of four thrust fault-bound terranes (Monger, 1991a). The terranes from base to top, are: (a) Harrison terrane (Wells Creek volcanics; Nooksack Group); (b) Chilliwack terrane; (c) Elbow Lake Formation (= Bridge River terrane?) and (d) Easton terrane (comprising Shuksan Greenschist and Darrington Phyllite = ? Cayoosh assemblage). The small exposure of rocks correlated with Harrison terrane in the Mount Baker structural window, and possibly the Chilliwack terrane, are the sole representatives of rocks in southwestern Coast Mountains recognized in the Cascades.

The physiographic boundary between the Coast Mountains and the Cascade Ranges is the Fraser River, which follows the north-trending early Tertiary (47–35 Ma) Fraser-Straight Creek Fault north of Hope, and Neogene (24–14 Ma?) northeast-trending faults southwest of Hope (Fig. 1; Monger, 1989a). These faults are superimposed on the fundamental Late Cretaceous-earliest Tertiary structural elements common to the Coast Mountains and North Cascades. The northwestern Cascade and Cascade core structures are congruent with the contractional faults of the Coast Belt Thrust System (Journeay and Friedman, 1993). However, in the Cascades, because the high-grade metamorphic rocks in the core are separated from the low-grade rocks to the west by the Fraser-Straight Creek Fault, historically the two have been treated separately. Restoration of dextral movement on the Fraser-Straight Creek Fault aligns rocks in the southeastern Coast Mountains with those in the Cascade metamorphic core (e.g., Evans and Berti, 1986). Note that different workers in the Cascades emphasize either thrust faulting or strike-slip faulting in interpretations of Cascades structure (cf. Brandon et al. (1988) and McGroder (1991)); with Brown (1987) and Brown and Talbot (1989). Both structural modes are present there, as they are in the southeastern Coast Mountains.

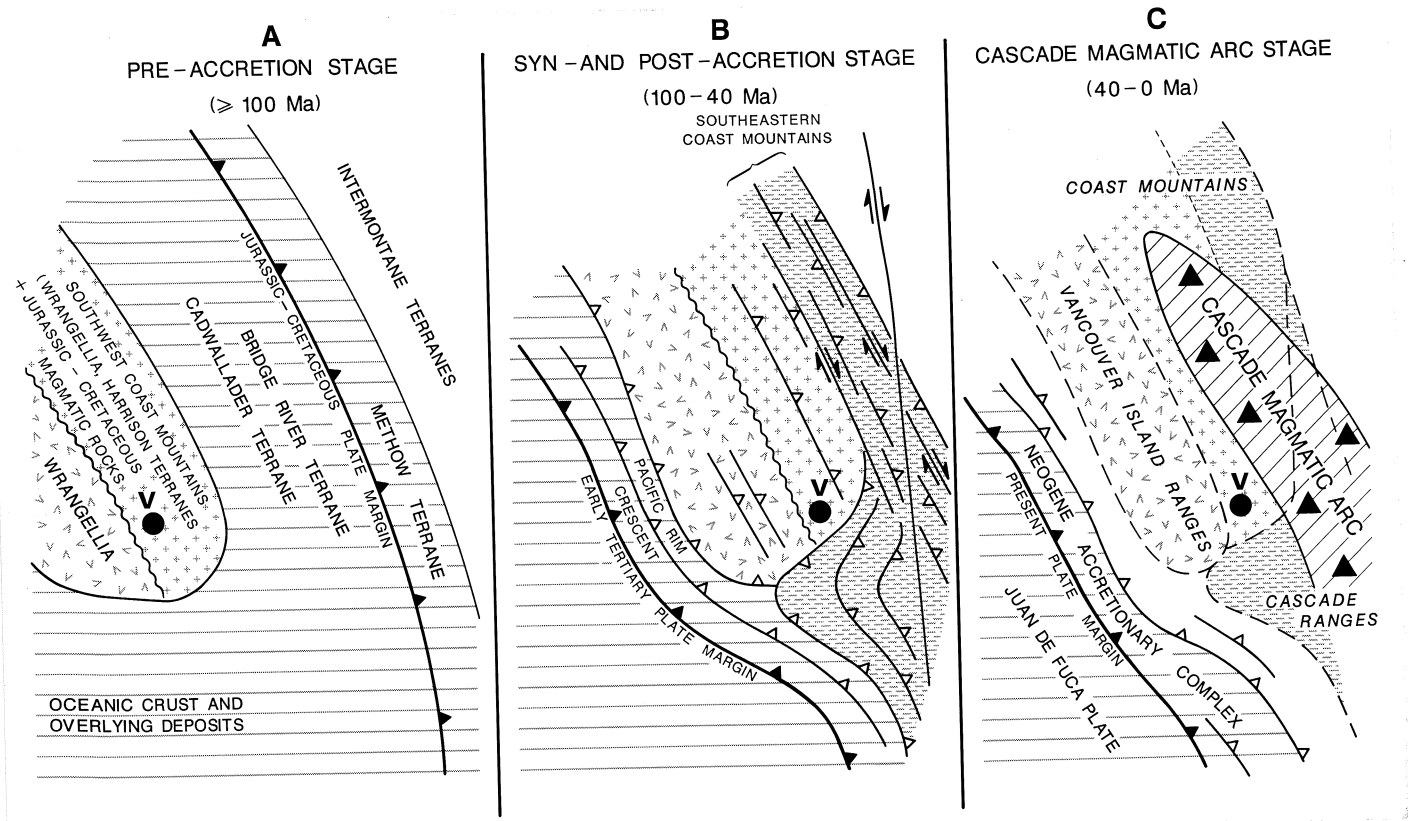



**Figure 12.** Summary tectonic evolution of southwestern British Columbia. Location of city of Vancouver shown by V. No attempt is made at palinspastic restoration, other than the schematic gap east of the southwestern Coast Mountains-Wrangellian block (=Insular Superterrane) in A; orogen parallel displacements may have been large, and sinistral before 100 Ma, and dextral after that time (Monger et al., 1994; Monger and Journeay, 1994).

The major differences between the geology of the Coast and Cascade mountains include: (1) absence of Wrangellia in the Cascade orogen; (2) near absence of the Jura-Cretaceous plutonic rocks common in southwestern Coast Mountains in the Cascades; and (3) presence of a Cretaceous-Tertiary ''Western Melange Belt'' in the western foothills of the Cascades (Frizzell et al., 1987; Tabor et al., 1989). The latter lithologically and structurally resembles the Leech River Schist and Pacific Rim Complex exposed along the southern and western coast of Vancouver Island, which is thrust beneath Wrangellia (Fig. 6, Yorath et al., 1985; Wheeler and McFeely, 1991). This relationship suggests that the Wrangellian-southwestern Coast Mountains block (=? Insular Superterrane) can be considered as an enormous tectonic flake, mostly not present south of lat. 49°N, incorporated in a long-lived accretionary complex (Fig. 7B, 7E).

## *Crustal evolution of the southern Coast Mountains region.*

In summary, we consider the crust of the southern Coast Mountains to have evolved through three stages (Fig. 12).

***(1) Pre-accretion stage (100 Ma):*** Before the mid-Cretaceous, the North American plate margin probably lay within the southeastern Coast Mountains and Cascade Ranges, which are located on the site of a long-lived accretionary and forearc complex represented by Bridge River and Methow terranes. The paleogeographic relationship between Insular Superterrane (Wrangellia and rocks of southwestern Coast Mountains) and the Early Cretaceous plate margin is uncertain, and only shown schematically on Figure 12A.

***(2) Syn- and post-accretion stage*** (ca. ***100–40 Ma):*** This is the major crust building stage, and appears to have been initiated by mid-Cretaceous (ca. 100 Ma) accretion of Insular Superterrane to the Cretaceous North American plate margin. Following accretion, the margin stepped westwards to a position near the west coast of Vancouver Island. Intense Late Cretaceous-early Tertiary (96–47 Ma) intra-plate contraction and dextral transpression, accompanied by metamorphism and granitic intrusion, were focused along the former plate boundary in southeastern Coast Mountains and Cascade Ranges (Fig. 12B). There, the crust was thickened to >30 km, so that the region of the former accretionary and forearc complex in southeastern Coast Mountains was inverted and became the major source of detritus for the Late Cretaceous foreland basin to the west.

***(3) Cascade magmatic arc stage (40–0 Ma):*** The tectonics of the region today, and probably of the last 40 million years, bear no direct relationship to those which formed most of the crust of the Coast Mountains, although they are dominated by plate convergence. Subduction of the oceanic Juan de Fuca Plate beneath the North American Plate was accompanied by generation of the Cascade Magmatic Arc, which extends from southwestern British Columbia to northern California (Fig. 12C). The physiography mainly formed in Neogene to Quaternary time (10 Ma), with uplift of the Coast Mountains and(?) western Vancouver Island, and subsidence of the intervening area, which today is covered by the water of Georgia Strait (Fig. 10).

# TABLE 1

## Summary descriptions of terranes in and near the Coast Mountains (Fig. 3, 6, 8; in part from Nokleberg et al., in press).

**AX** **ALEXANDER TERRANE** ***(Island Arc)***
**Rock units: Wales, Descon, Donjek, Kaskawulsh, Alexander, Cedar Cove, Karheen, Cannery, Iyokeen, Halleck, Pybus, Hyd** Upper Proterozoic to Triassic volcanic and sedimentary rocks in a variety of depositional settings (ocean arc, back arc, platform, rift, trough, offshelf) and comagmatic intrusions (Gehrels and Saleeby, 1987; Wheeler et al., 1991).

**BR** **BRIDGE RIVER TERRANE** (in Cascade terranes of Fig. 3; distribution on Fig. 6) ***(Accretionary Wedge/Subduction Complex)***
**Rock units: Bridge River and Hozameen complexes, correlative units? Cogburn Group, Elbow Lake Formation, Napeequa Schist, Orcas chert.** Where little metamorphosed, typically highly disrupted radiolarian (ribbon) chert, argillite and basalt, subordinate siltstone and greywacke, mafic intrusive rocks, minor Late Triassic olistostromal carbonate and associated alpine-type ultramafics (Potter, 1983; Monger, 1989b; Schiarizza et al., 1990a). The terrane ranges in age from Mississippian to late Middle Jurassic (Callovian) and possibly younger, but Late Triassic and Early Jurassic fossils are most common (F. Cordey, pers. comm., 1992). Metamorphic grade is mainly sub-greenschist to greenschist, but locally is of amphibolite grade. Triassic blueschist occurs in one locality (Archibald et al., 1991). It may include gradationally overlying Upper Jurassic and/or Lower Cretaceous clastic rocks with minor volcanics (**Brew Group, Cayoosh assemblage**; Journeay and Northcote, 1992); alternatively, these clastic rocks may be an overlap assemblage. Bridge River, Cadwallader and (?) Methow terranes were linked by mid-Cretaceous time (Garver, 1990).

**CC** **CACHE CREEK TERRANE** ***(Accretionary Wedge/Subduction Complex)***
**Rock unit: Cache Creek Complex.** In its type-area of southern British Columbia, the Cache Creek Complex is in part disrupted, in part coherent, and comprises Permian to Lower and(?) Middle Jurassic chert and pelite, with melange blocks of basalt, alpine type ultramafic rocks, and carbonates; the latter of which are Middle Pennsylvanian and Permian in age. A distinctive large body of shallow water Late Permian and Triassic reefoidal carbonate is called **Marble Canyon Formation** (Beyers and Orchard, 1991). On the west side, Late Triassic volcaniclastic rocks are of possible arc origin (Cordey et al., 1987; Monger, 1989a). The terrane is mainly of sub-greenschist grade, but blueschists of Late Triassic age (210–217 Ma) occur in central British Columbia (ca. lat. 55°N) (Paterson and Harakal, 1974).

The large carbonates contain Late Permian fusulinid, coral and conodont faunas similar to those in regions bordering the western Pacific and the Permian Tethys, and are the only clearly "exotic" faunas recognized in the Canadian Cordillera (Ross and Ross, 1983). The Marble Canyon carbonate contains both latest Permian (late Dzhulfian to Chanxingian) and earliest Triassic (Griesbachian) conodonts (Beyers and Orchard, 1991). Although a continuous section is not proven, recognition of these two faunas in one rock body is unique to western North America. Coeval faunas only occur together in south China, Tibet and Armenia (Tozer, 1988), regions with "Permian Tethyan" faunas.

**CD** **CADWALLADER TERRANE** (in Cascade terranes of Fig. 3) ***(Island Arc)***
**Rock units: Pioneer greenstone, Hurley Formation (Cadwallader Group), Tyaughton Group and Last Creek Formation.** Pioneer greenstone, diorite, trondhjemite, gabbro, and alpine-type ultramafic rocks of Permian age, are overlain by Upper Triassic arc volcanics, carbonates, and distinctive carbonate-volcanic-plutonic clast conglomerates of the Hurley Formation (Cadwallader Group). In fault contact are interlayered Upper Triassic conglomerates and siltstones of the Tyaughton Group, and Lower and Middle Jurassic fine grained, non-volcanic clastics of the Last Creek Formation (Rusmore, 1987; Leitch et al., 1991; P. Schiarizza, pers. comm., 1991). In turn, these rocks may be overlain by Jura-Cretaceous clastic rocks (**Relay Mountain Group** and(?) **Cayoosh assemblage**), and structurally interleaved with the Bridge River Complex along early Late Cretaceous thrust faults (Journeay and Northcote, 1992).

**CK** **CHILLIWACK TERRANE** (in Cascade terranes of Fig. 3) ***(Island Arc)***
**Rock units: Chilliwack Group, Cultus Formation.** Devonian to Permian arc volcanics, clastics and carbonates of the Chilliwack Group are stratigraphically overlain by Upper Triassic and Lower Jurassic arc-related clastics of the Cultus Formation and (?) Upper Jurassic pelites, of mainly subgreenschist (lawsonite-albite) grade; folded and internally imbricated on early Late Cretaceous thrust faults of the Northwest Cascades System (Monger, 1989a). Early Permian fossils are most like those of Quesnel and Stikine terranes.

**HA** **HARRISON TERRANE** (in Cascade terranes of Fig. 3) ***(Island Arc)***

**Rock units: Camp Cove, Harrison Lake formations.** Middle Triassic cherty argillites and mafic volcanics of the Camp Cove Formation are overlain by thick Lower Jurassic (Toarcian; ca. 185 Ma) andesitic and dacitic volcanics (Harrison Lake Formation). Overlying Middle and Late Jurassic (early Callovian and early Oxfordian; ca. 160–155 Ma) sedimentary and andesitic volcanic rocks (**Mysterious Creek, Billhook Creek formations;** Arthur, 1986) are coeval with plutonic rocks (≤ 165 Ma) coextensive with, and herein included in the Gravina-Gambier arc. The plutonic rocks link Harrison terrane with Wrangellia; the strata are the only known surfical representatives of older, Jurassic, parts of the Gravina-Gambier arc at these latitudes.

**MT** **METHOW TERRANE** (in Cascade terranes on Fig. 3) ***(Dominantly Arc/Turbidite)***

**Rock units: Spider Peak Formation, Ladner Group (Dewdney Creek and Boston Bar formations), Thunder Lake sequence, Jackass Mountain and Pasayten groups.** Methow terrane comprises rocks representing three different tectonic settings: (1) Basal Triassic MORB basalt (Spider Peak Formation) associated with gabbro and serpentinized ultramafic rocks of the Coquihalla serpentinite belt (Ray, 1986, 1990); (2) overlying Jurassic (Sinemurian(?), Pliensbachian to Bajocian; and (?) Oxfordian) pelites, siltstones and tuffs (Boston Bar Formation, Ladner Group) and laterally interfingering arc-related flows and volcaniclastics of Aalenian-Bajocian age (ca. 170–178 Ma; Dewdney Creek Formation, Ladner Group), and Upper Jurassic (Oxfordian to Tithonian; ca. 155–145 Ma) volcanic sandstone and argillite (O'Brien, 1987; Ray, 1990; O'Brien et al., 1992). The last is (mostly) lithologically inseparable from the Boston Bar Formation (Ray, 1986); (3) uppermost are thick (4500 m) Lower to Upper Cretaceous (Hauterivian to Turonian; ca. 135–90 Ma) argillite, sandstone and conglomerate of the Jackass Mountain and Pasayten groups (Monger, 1989a; McGroder, 1991).

Dated clasts (of 235 Ma) from a granitic clast in the basal Ladner can be most closely correlated in the region with ages from the Mount Lytton Complex, which is part of Quesnel terrane; a Late Jurassic (156 Ma) clast in east-derived Albian Jackass Mountain conglomerate is age-equivalent to granitic rocks in the western edge of Quesnel terrane; westerly-derived chert-rich Albian-Cenomanian clastics are probably derived from the Bridge River terrane (Tennyson and Cole, 1978; O'Brien et al., 1992).

**QN** **QUESNEL TERRANE** ***(Island Arc)***

**Rock units: Harper Ranch Group and Apex Mountain Complex, Nicola and Rossland groups, (?)Ashcroft Formation.** Integrity is given to Quesnel terrane (or Quesnellia) by the west-facing (Mortimer, 1987) Late Triassic to Early Jurassic (Karnian to Toarcian; ca. 230–195 Ma) arc volcanics, comagmatic granitic and alkaline intrusions and laterally equivalent clastic rocks (Nicola and Rossland formations; Monger, 1989a; Andrew et al., 1990; Parrish and Monger, 1992). These are unconformably overlain by clastic rocks of the Pliensbachian to Callovian (ca. 190–160 Ma) Ashcroft Formation. The early Mesozoic arc overlaps two basements: Devonian through Permian oceanic Apex Mountain Complex and the roughly coeval island arc-related Harper Ranch Group, which were deformed and juxtaposed with one another and possibly with parts of Kootenay terrane still farther east, prior to Late Triassic time (probably ca. 185 Ma; Read and Okulitch, 1977; Klepacki, 1983; Orchard and Danner, 1991).

Permian faunas are mixture of western Cordilleran forms similar to those in Stikinia, Chilliwack terrane and the eastern Klamaths, with some Tethyan elements (Orchard and Danner, 1991). Pliensbachian faunas of Quesnellia appear to be latitudinally comparable to those of Stikine (Smith and Tipper, 1986), that is south of their present position with respect to cratonic North America. However, May and Butler (1986) conclude from their analysis of paleomagnetic data that there is no evidence for northward displacement in the Triassic-Jurassic rocks.

Plutonic rocks emplaced in southwestern Quesnellia include early Mesozoic arc-related magmatism (212–195 Ma), post-accretionary Late Jurassic (168–155 Ma) granitic rocks, and mid-Cretaceous (ca. 105–100 Ma) plutonics, probably related to the Spences Bridge arc (Monger, 1989a; Thorkelson and Rouse, 1989; Parrish and Monger, 1992).

**ST** **STIKINE TERRANE** ***(Island Arc)***

**Rock units: "Stikine assemblage," Asitka Group, Takla, Stuhini, Hazelton and Spatsizi groups** in the Intermontane and eastern Coast belts, mainly north of lat. 52°N. Includes Lower Devonian, Mississippian, Upper Pennsylvanian and Permian arc volcanics, interbedded with carbonates and fine grained clastics (Brown et al., 1991), overlain unconformably by Upper Triassic arc volcanics and interfingering clastic rocks (Stuhini and Takla groups), granitic rocks that are coeval with the volcanics, and unconformably overlying Lower Jurassic mainly andesitic arc volcanics and intercalated sedimentary rocks of the Hazelton and Spatsizi groups (Tipper and Richards, 1976; Monger, 1977; Anderson, 1989; Gabrielse, 1991). The basement of Stikine is not exposed; it may be underlain by rocks correlative with continentally-derived Nisling or Yukon-Tanana terranes (Gehrels et al., 1990; McClelland, 1992). However, Nd-isotopes derived from Stikinian rocks (Samson et al., 1991) indicate a primitive source.

Stikine strata on the east side of the Coast Belt (north of lat. 52°N) appear to be overlain by rocks correlated with Gambier Group (Wheeler and McFeely, 1991). Van der Heyden (1992) argued that Middle Jurassic granitic rocks on both sides of the central Coast Belt are related, and that the amalgamated Stikine-Alexander-Wrangellia terranes were emplaced against rocks to the east in Lower to Middle Jurassic time along a Cache Creek suture.

**WR** **WRANGELLIA** ***(Island Arc)***

**Rock units: Sicker and Buttle Lake groups, Karmutsen, Quatsino, Parsons Bay and Harbledown and Bonanza formations** and comagmatic granitic rocks in the Insular Belt and southwestern Coast Belt. Oldest rocks of Wrangellia on Vancouver Island comprise Upper Devonian (367 Ma) island arc-related volcanic and sedimentary rocks and intrusions comagmatic with the volcanics (Sicker Group), and overlying mainly sedimentary strata of Late Carboniferous-Early Permian ages (Buttle Lake Group; Muller, 1980; Massey and Friday, 1989; Parrish and McNicoll, 1992). Locally overlying these rocks on Vancouver Island are Ladinian black argillites. The most distinctive elements of Wrangellia, are Middle(?) to Late Triassic tholeiitic basalt (Karmutsen Formation) and associated mafic intrusions and overlying shallow water carbonates (Quatsino limestone) grading upwards into deeper water clastics; these rocks may represent the remnants of a large oceanic volcanic plateau and associated overlying deposits (Jones et al., 1977). In turn this sequence is succeeded on Vancouver Island by Lower and Middle Jurassic arc volcanics (Bonanza Formation) and comagmatic granitic intrusions (ca. 200–180 Ma).

Permian fossils in Wrangellia of Vancouver Island are high-latitude Boreal or Uralian cool water mainly brachiopod and bryozoa faunas (Yole, 1969). Pliensbachian faunas indicate that northern Wrangellia was at similar latitudes to Stikinia (northern California-Nevada) in the Early Jurassic (Smith and Tipper, 1986). Paleomagnetic data suggest that the Triassic Nikolai Greenstone in southern Alaska has shifted northward by about 2500 km relative to the craton, whereas the correlative Karmutsen Formation on Vancouver Island shows little displacement (May and Butler, 1986). Assuming a northern hemisphere location, early Mesozoic Wrangellia was located at northern Oregon-southern British Columbia latitudes.

Stitching plutons indicate that parts of Wrangellia and Alexander terranes were together in southern Alaska by Pennsylvanian time (Gardner et al., 1988), although the Alexander terrane lacks the distinctive early Mesozoic stratigraphy characteristic of all parts of Wrangellia. In Alaska (lat. 63°N) the broad Jura-Cretaceous Kahiltna flysch basin overlaps Wrangellia, separates it from Yukon-Tanana terrane to the north, and would seem to be an analogue of the coeval clastic basin in the southern Coast Belt. Wrangellia, Alexander and Harrison terranes are clearly overlapped and intruded by the Jura-Cretaceous Gravina-Gambier arc. Van der Heyden (1992) argued that this arc links these terranes with Stikinia terrane. In the southern Coast Belt, the Gravina-Gambier arc, together with Wrangellia and Harrison terranes, forms the foreland of an early Late Cretaceous west-directed thrust system, whose foreland basin is represented by Late Cretaceous (Turonian)-earliest Tertiary Georgia Basin strata (Journeay, 1990a; Monger, 1991c).

**NS** **YUKON-TANANA** ***(Cratonal Margin Terrane; Affinity Uncertain; rocks probably part of Yukon-Tanana terrane of central Yukon and Alaska, and called Nisling terrane on some maps)***

**Rock Units: Nisling and Nasina formations.** Metamorphosed Proterozoic to lower Paleozoic (?) passive continental margin assemblage and partly metamorphosed carbonaceous and siliceous offshore sediments (Gehrels et al., 1990; Wheeler et al., 1991).

# FIELD TRIP ROAD LOG AND COMMENTS

# DAY 1: VANCOUVER TO LILLOOET: COAST-CASCADE BOUNDARY

James W.H. Monger and J. Murray Journeay

The field trip for the Geological Society of America Penrose Conference on the "Tectonic evolution of the Coast Mountains orogen," 17-18 May 1992, started at the Delta River Inn, Richmond, 2.6 km east of Vancouver International Airport. To avoid Vancouver traffic, the trip was routed south on Highway 99 and then east through Cloverdale. Distances are given from the Delta River Inn to the point (51.7 km) where the route intersects the **Trans-Canada Highway at 232nd Street (Exit 66) south of Fort Langley. For guidebook users who wish to drive the ca. 45 km directly from downtown Vancouver, this exit is used as 0 km for the remainder of Day 1.** The route of Day 1 is given on Figure 1 (in back pocket).

The route of Day 1 mainly follows the physiographic boundary between the Coast Mountains and Cascade Ranges, which is delineated by the Fraser River (Mathews, 1986). However, note that the eastern part of the southern Coast and the Cascade mountains contain correlative rock units and a common structural fabric now known to be of Late Cretaceous to Eocene age (97-47 Ma), as recognized by Crickmay (1930) and Misch (1966). South and east of Vancouver, the bedrock is largely concealed beneath fluvio-glacial deposits of the Lower Mainland (or physiographic Fraser Lowland) region, which lies between Vancouver and Hope (Fig. 1). Towards the eastern end of this region the valley narrows, and there its orientation reflects the northeast-trending Neogene (ca. 25-14 Ma?) Sumas and Vedder faults. At Hope, the river turns abruptly northwards and runs along the Fraser Canyon, which follows the trace of the Fraser-Straight Creek Fault, a Paleogene (ca. 47-35 Ma) dextral strike-slip fault system. Thus, the boundary between Coast and Cascade mountains is delineated by late structures which overprint and disrupt an originally continuous orogen. The main difference between the two systems is the presence in the southwestern Coast Mountains of abundant Jura-Cretaceous granitic rocks, that on their western margin intrude Wrangellian strata and on their eastern margin intrude Harrison Lake terrane. These rocks are almost entirely absent from the Cascade Ranges. However, exposed in the small Mount Baker structural window, low in the structural stack of thrust faults of the northwest Cascades in northern Washington, are rocks probably correlative with Harrison terrane and overlying strata.

**0 km** Leave Delta River Inn, located on Sea Island in the Fraser River delta.

**0.3 km** Cross Middle Arm, a distributary of the Fraser River.

**2.0 km** Enter Highway 99; head southwards, crossing beneath the Main Channel of the Fraser River by the George Massey Tunnel at 10.4 km, and continue until turnoff on to Highway 10 at 20.4 km.

**COMMENT** The route crosses the Fraser River delta, a postglacial (≤ ca. 13 000 years in this area) feature in which much of the sediment south of the Main Channel was deposited between ca. 7500-5000 BP (Clague et al., 1991). The delta is built on glacial sediments that overlie mostly nonmarine clastic rocks. Where penetrated in boreholes, the latter form a discontinuous stratigraphic succession that ranges downwards in age from Miocene to Albian (Mustard and Rouse, 1991). In North and West Vancouver, Jura-Cretaceous granitic rocks of the southwestern Coast Mountains are overlain non-conformably by Eocene and Upper Cretaceous (Santonian-Campanian) nonmarine clastic rocks of the Burrard and Kitsilano formations, that dip southwards at 10-15° and are crossed at the end of Day 2. These rocks in part are correlative with the Cretaceous Nanaimo Group on Vancouver Island, and in part with the Tertiary Chuckanut Group near Bellingham, Washington. The southward dip of these strata is due partly(?) to late Tertiary and Recent uplift of the southern Coast Mountains (Fig. 10; Roddick, 1966; Parrish, 1983; Holdahl et al., 1989). Thickness of the strata near Vancouver is nearly 1400 m; in a borehole drilled near the International Boundary at Point Roberts they are 4400 m thick, with Albian strata at the bottom of the hole. The basement at Point Roberts is unknown. Twenty-five kilometres south of the International Boundary, near Bellingham, Washington, the early Tertiary Chuckanut Formation unconformably overlies rocks of the Northwest Cascades System (Johnson, 1984).

**20.4 km** Drive east on Highway 10, leaving the delta area and climbing over till, through Cloverdale (38.5 km), and skirting Langley on its northwest side (43.0 km). Route 10 (local name here is Glover Road) turns northeast and joins the Trans-Canada Highway at 51.7 km.

**0 km** Exit 66 (intersection of Trans-Canada Highway with 232nd Street, ca. 45 km from downtown Vancouver). Go southeast on Highway. Views of the southernmost part of the southwestern Coast Mountains to north.

**7.4 km** View ahead (to southeast) of Mount Baker in Washington State. This is a late Tertiary to Recent volcano, approximately 3300 m high, that last erupted in 1872. It is the northernmost of the volcanoes of the Cascade Magmatic Arc in the United States. The continuation of the chain in Canada is seen on Day 2 in the southwestern Coast Mountains.

**COMMENT** The young volcanics of Mount Baker lie on older rocks of the North Cascade Range, whose summits are ca. 2000-2500 m in height. The northwesternmost Cascade rocks in Canada

form Vedder Mountain, a northeast trending ridge which crosses the International Boundary and is seen ahead near Exit 87 (21.3 km). It consists of three or four northeast-trending fault slices. From northwest to southeast these are: (1) minor greenstone, marble, and chert with Paleozoic or early Mesozoic radiolarians (F. Cordey, pers. comm., 1992) at the base of the mountain near the concealed northeast-trending Vedder Fault; (2) the Vedder Complex, forming the northwest side of the ridge, which is a polygenetic, variably metamorphosed sequence of amphibolitic gneiss, epidote hornblendite, and local blueschist-bearing greenstone, K-Ar dated at ca. 242 Ma (Bernardi, 1977); (3) a heterogenous assemblage of volcanic sandstone, minor limestone, greenstone, and melange containing chert blocks with radiolarians of Callovian to Tithonian age (ca. 160-150 Ma; F. Cordey, pers. comm., 1991) in a graphitic phyllite matrix; and (4) undated greywacke and conglomerate with granitic clasts at the base of the southeast side, possibly correlative with Jura-Cretaceous Nooksack strata in Washington (McMillan, 1964; Monger, 1989a). Near the International Boundary these rocks are unconformably overlain by the Eocene and younger Huntingdon Formation, which is correlative with the Chuckanut Group and rocks in Vancouver. As reported in Roddick (1965, p. 31), P. Misch suggested that the metamorphic rocks on the northwest side of Vedder Mountain may be a remnant of a (Cascade) thrust plate cut by later northeast-trending block faults; the writers concur with this suggestion.

**26.5 km** Leave Trans-Canada Highway at Exit 92, and drive north (under freeway) on Highway 11. On right, roadcuts in the northeast quadrant of the cloverleaf expose sandstones of the late Paleocene and Eocene Huntingdon Formation (Mustard and Rouse, 1994), which are correlative and along strike with the Tertiary clastics in Vancouver. They dip southwesterly at 10-15°, and on Sumas Mountain to the east overlie granitic rocks U-Pb dated at ca. 164 Ma (Friedman and Armstrong, in press). There are several comparable dates from granitic rocks in the southwestern Coast Mountains. The granitic rocks intrude altered andesites and dacites of the Lower and Middle Jurassic Harrison Lake Formation (Roddick, 1965; Monger, 1993).

**COMMENT** Sumas Mountain is underlain by the southernmost Coast Mountains rocks exposed in Canada. Within it, the Harrison Lake Formation is correlated with the Wells Creek volcanics of the northwest Cascade orogen in Washington, exposed in a window at the base of the stack of thrust sheets north of Mount Baker (Fig. 13; Misch, 1966). Separating Sumas and Vedder mountains is a late Tertiary structural depression bounded by two concealed northeast-trending steep faults, called respectively Sumas and Vedder faults. Tertiary strata at depths of 300 m in the depression,

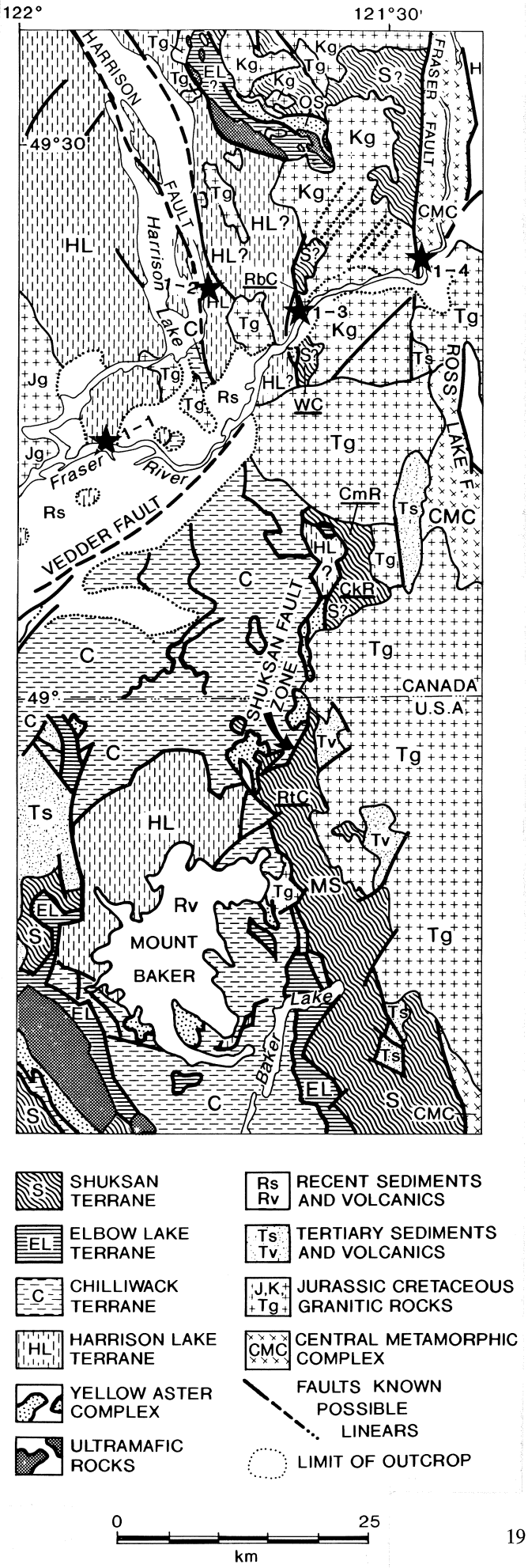



**Figure 13.** Parts of southeastern Coast and northwestern Cascade mountains, showing correlative (?) terranes and rock units that span the physiographic boundary between the two (delineated by the Fraser River) and locations of Stops 1-1, 1-2, 1-3, 1-4. Tentative correlations: HL Harrison Lake terrane = HL? (Slollicum Schist); EL Elbow Lake Formation = EL (Cogburn Group=? Bridge River terrane); S Shuksan and Darrington Phyllite = S? Settler Schist; Geographic localities: CkR, Chilliwack River; CmR, Cheam Range; MS, Mount Shuksan; OS, Old Settler; RbC, Ruby Creek; RtC, Ruth Creek; WC, Wahleach (or Jones) Creek (figure from Monger, 1991a).

below thick Quaternary cover, are correlated with the Huntingdon Formation exposed on flanking Sumas and Vedder mountains. The Sumas and Vedder faults are two of a family of northwest-trending faults in the southern Coast Belt (Monger, 1989a; Journeay, 1990a). These faults appear to control orientation of the upper Fraser Lowland, which is the physiographic boundary between Coast and Cascade orogens west of the town of Hope. Based on ages of dated strata offset by these faults, and ages of intrusions along similarly oriented faults in the region, whose emplacement is apparently controlled by them, the faults are thought to be late Tertiary (ca. 25-14 Ma).

**31.0 km** To east, view of Sumas Mountain. To west and northwest are the south slopes of the hills of the southern Coast Mountains. These reflect the southwesterly dip of the non-conformable surface at the base of the Late Cretaceous-early Tertiary strata, that at least in part reflects Neogene to Recent uplift (Fig. 10).

**38.0-39.0 km** Cross Mission Bridge over Fraser River; on the north side take Highway 7 eastwards to Agassiz, through Mission, along the north side of the Fraser River. To the south is Sumas Mountain. It is not known if the valley containing the Fraser River between Sumas Mountain and the southern slopes of the Coast Mountains is structurally controlled.

**45.0-49.0 km** View ahead of Mount Cheam, on the south side of the Fraser River, underlain by late Paleozoic Chilliwack Group and early Mesozoic Cultus Formation (Chilliwack terrane).

**COMMENT** Rocks on Mount Cheam have the thrust structure and sub-greenschist, local lawsonite-albite metamorphism, typical of the much of the northwest Cascades. In this area (and best seen in the Chilliwack valley to the south; Price et al., 1985, *their* Fig. 85, Monger, 1986, 1989a) there is structural repetition on northwest-vergent folds with northeast-trending, northwest-plunging axes and associated thrust faults. In turn, these structures are refolded/faulted on southwest-vergent, north-northwest trending structures. On Mount Cheam, Permian limestones are repeated three times on the north face of the 2300 m high mountain. The top of Lady Peak, immediately south of Mount Cheam, is a klippe of metabasalt (mainly actinolite-epidote). The discontinuity between the metamorphic grade in the klippe and that of underlying subgreenschist Chilliwack strata, together with the presence of rocks of similar grade correlated with Shuksan Greenschist and Darrington Phyllite at the south end of the Cheam Range and Mount Slesse (Jewett, 1984), suggests that this klippe may mark the trace of the Shuksan fault zone; Fig. 13). The lower ridge east of Mount Cheam is underlain by the Miocene Mount Barr batholith, with K-Ar dates ranging from 16-18 Ma (Monger, 1989a).

**49.0 km** Dewdney Junction; mountains to the north are mainly Middle Jurassic granodiorite, U-Pb dated at ca. 164-167 Ma (Friedman and Armstrong, in press) with minor septa/fault slices of probable Harrison Lake Formation.

**57.0 km** The hill to the south (Chilliwack Mountain) is underlain by Harrison Lake Formation and pelitic rocks that may be equivalent to the Middle and Upper Jurassic Mysterious Creek Formation in the section on the west side of Harrison Lake.

**60.0 km** Nicomen Slough.

**63.0 km** The mountain ahead is Harrison Hill, underlain by probable Middle Jurassic granodiorite (to the north, similar granodiorite is K-Ar dated at 160 Ma; the three U-Pb dates in the area are 164-167 Ma). Roadside outcrops between 71.02-73.0 km of highly fractured granodiorite are part of this body.

**74.0 km** Cross Harrison River. The view ahead is into the northwest Cascades.

**79.0 km** Outcrops of Harrison Lake volcanics, in road cuts on southwestern end of Mount Woodside.

***STOP 1-1:*** **80.1 km** Rest area on right side of road **(TAKE CARE CROSSING ROAD!).** Outcrops on the north side of Highway 7 are typical grey-green, altered, pyritic andesite of the Harrison Lake Formation. Locally, float of volcanic breccia is present; volcaniclastics are probably more typical of the formation than flows, and sandstone and shale are present in places. The Toarcian (ca. 185 Ma) basal conglomerate of the Harrison Lake Formation overlies Middle Triassic rocks of the Camp Cove Formation (Fig. 8), and contains limestone clasts containing fossils that are possibly derived from Permian limestone of the Chilliwack Group (Arthur et al., 1993). Metamorphism of the rocks is variable, and ranges from zeolite facies about 10 km north of here (Beaty, 1974) to greenschist facies; in this area regional penetrative deformation is not present.

The southward view from this stop includes the boundary between Coast and Cascade mountains (Fig. 1). The southernmost outliers of the Coast Mountains (Sumas Mountain, Chilliwack Mountain to the southwest and, to the south, a small hill called Mount Shannon or Cemetery Hill) all contain andesites and dacites of the Harrison Lake Formation (Harrison terrane, southwestern Coast Mountains). South of these, Sumas Prairie is the northeast-trending lowland in a probable graben between Sumas and Vedder mountains, and is bounded by Neogene northeast-trending Sumas and Vedder faults. South of this, the Cascade Ranges, extend from Vedder Mountain in the west to Mount Cheam in the east, and consist of the complexly deformed stack of thrust fault slices of the Northwest Cascades System. The Harrison Lake Formation is not recognized in Canada south of the Fraser River, although probable correlatives, called the Wells Creek volcanics, outcrop still farther south in Washington, in the valley of the North Fork of the Nooksack River, north of Mount Baker (Fig. 13), low in the stack of thrust fault slices forming the Northwest Cascade System (Misch, 1966). This is the first hint on the field trip of the original structural relationship between rocks of southwestern Coast Mountains and the Cascade Ranges.

**80.6-82.4 km** Continuous outcrop of Harrison Lake volcanics, with columnar jointing at 752.9 km.

**82.4 km** Descend to the Fraser River floodplain. At 86.2 km, the knoll on right consists of hornfelsed argillite and siltstone possibly correlative with the lower Callovian Mysterious Creek Formation, which overlies the Harrison Lake Formation on the west side of Harrison Lake.

**88.7 km** Leave Highway 7, and take the road north to Harrison Hot Springs. The ridge on the east is underlain by

Miocene (K-Ar date ca. 19 Ma) granodiorite, which is an older part of the Cascade Magmatic Arc (Fig. 12).

**90.7 km** To the east, small cliffs on Bear Mountain are mostly limestone, and contain crinoids and conodonts; the latter are Mississippian (?late Visean-early Namurian; M.J. Orchard, pers. comm., 1987). As recognized by Crickmay (1930), this limestone is correlative with Carboniferous limestone (which also features large crinoid stems) in the Chilliwack Group, south of Fraser River.

**92.3 km** To the northwest, cliffs of granodiorite, K-Ar dated at ca. 19 Ma, are probably a northwestern apophysis of the Mount Barr Batholith, although separated from the latter by the Fraser River and floodplain sediments (Fig. 1). The granodiorite intrudes the Harrison Lake Formation.

**95.1 km** T-junction at end of road in village of Harrison Hot Springs at south end of Harrison Lake; take road to east. Drive east, and then north, along east side of Harrison Lake.

**96.6-98.2 km** Outcrops by road of grey, locally rusty weathering grits and pelites; these are assumed to be part of the Chilliwack Group, as similar rocks are in apparent stratigraphic contact with the Late Mississippian limestone higher up the mountain. A cleavage is developed towards the north end of outcrop.

**98.2-98.8 km** Covered interval, which is assumed to mark the trace of the Harrison Lake Fault (Fig. 1) and shows up as a linear on the ridge to the south.

**98.8 km** Outcrop of sandstones and pelites, with a strong fabric; these rocks are undated, but are more-or-less continuous with rocks of the Lower Cretaceous Peninsula Formation along strike, to be seen at Stop 1-2.

**COMMENT** About 5 km north of here along the lakeshore and on the tip of Cascade Peninsula within similar rocks are variably deformed clams (*Buchia* sp.; Lowes, 1972; Monger, 1989a). Along strike to the north, on the west side of Harrison Lake, similar, but better preserved forms are Early Cretaceous (Berriasian to early Valanginian; ca. 145-140 Ma; Arthur et al., 1993).

**101.3 km** Drive through Sasquatch Provincial Park; bear right, take east fork and drive north, past rusty outcrops of strongly lineated Peninsula sandstone and pelite.

**102.4 km** Second fork; head left and north up hill; road comes out above Harrison Lake. Scattered outcrops of rusty weathering Peninsula sandstone along road.

***STOP 1-2:*** **104.7 km** The purpose of this stop is two-fold: to examine lineated sandstones of the Peninsula Formation in outcrops across road; and to view and discuss geological features in the vicinity of Harrison Lake, which is close to the boundary between southwestern and southeastern parts of the Coast Mountains (Crickmay, 1930).

The valley containing Harrison Lake and extending north past Pemberton (and crossed in the middle of Day 2) is eroded along a broad fault zone, within which faults are mainly contractional, but also dextral transpressional, and belong to the Coast Belt Thrust System (Fig. 1, 14).

The Harrison Lake Fault is a dextral strike slip fault on the east side of Harrison Lake (Fig. 1, 14), whose probable extension to the north in metamorphic rocks in Big Silver Creek contains muscovite with fabric indicating dextral shear that is Rb-Sr dated at 93.5±11.4 Ma (Parrish and Monger, 1992, p. 104). Near Stop 1-2, the Harrison Lake Fault lies below the inlet, east of Cascade Peninsula on the east side of Harrison Lake (Fig. 1). Kinematic indicators in lineated mylonites on the neck of Cascade Peninsula record a dextral sense of shear (Journeay and Csontos, 1989). On the lower part of the Peninsula, the rocks have little structural fabric and Cretaceous pelecypods (*Buchia*) are undeformed, whereas those on the lakeshore below Stop 1-2 are strongly stretched, as are those on the road about 3 km north of Stop 1-2. The fault zone is characterized by a strong stretching lineation that plunges at angles between 7-45° to the north-northwest (Monger, 1986; Hettinga, 1989).

North-northwest of Stop 1-2, on the east side of Harrison Lake, Long Island is crossed by a fault of that name (Fig. 1, 14); the southwesternmost part of the island contains undeformed pelecypods, whereas most of the island has a strong cleavage. The Long Island Fault is a thrust fault dipping east-northeast at about 70°. Near Harrison Lake, Long Island and Harrison Lake faults mark the boundary between rocks with little penetrative fabric to the west, and rocks to the east with well developed structural fabric to the east.

West of Harrison Lake at this latitude, Middle Triassic to Middle Jurassic rocks of Harrison terrane (Fig. 6, 8; Table 1) are intruded by late Middle Jurassic and younger plutons (Fig. 1), that are part of the suite that extends to the west side of the Coast Mountains and there intrudes Wrangellia. Away from the lake, the plutonic rocks become more voluminous, and together with mid-Cretaceous plutons form much of the southwestern Coast Mountains. In places, the Jurassic plutonic rocks are overlain by Lower Cretaceous strata (Gambier Group) correlative with Peninsula and Brokenback Hill formations (Fig. 8, 14). The metamorphic grade increases from zeolite (laumontite) facies and rarely lawsonite albite facies near the lake, to greenschist facies near the plutons, 15 km west of the lake (Beaty, 1974; M.T. Brandon, pers comm., 1986; Hettinga, 1989).

**COMMENT** (From details in Monger et al., 1990; Monger, 1986, 1989a). To the east of the Harrison Fault, and forming Slollicum Peak northeast of Stop 1-2 are typically greenschist grade, penetratively deformed, locally pillowed flows, volcanic breccias and crystal tuff, and (generally) structurally overlying dark grey graphitic phyllite with (rare) distinctive conglomerate with quartz feldspar porphyry and limestone clasts. These rocks are called the Slollicum series or schist (Fig. 14; Crickmay, 1930; Monger, 1986), and are described in some detail by Lowes (1972) and Hettinga (1989). The range of lithologies and U-Pb dates from this unit of 146 and 102 Ma (reported, respectively, in Hettinga, 1989 and Parrish and Monger, 1992) show that the Slollicum Schist is derived, at least in large part from the Peninsula, Brokenback Hill and(?) Billhook Creek formations, which are integral parts of the footwall of the Coast Belt Thrust System (Journeay and Friedman, 1993). Fabrics in the Slollicum Schist dip steeply to the east; clasts are flattened and elongated downdip, as are metamorphic mineral lineations. We consider these rocks to be the easternmost part of southwestern Coast Mountains.

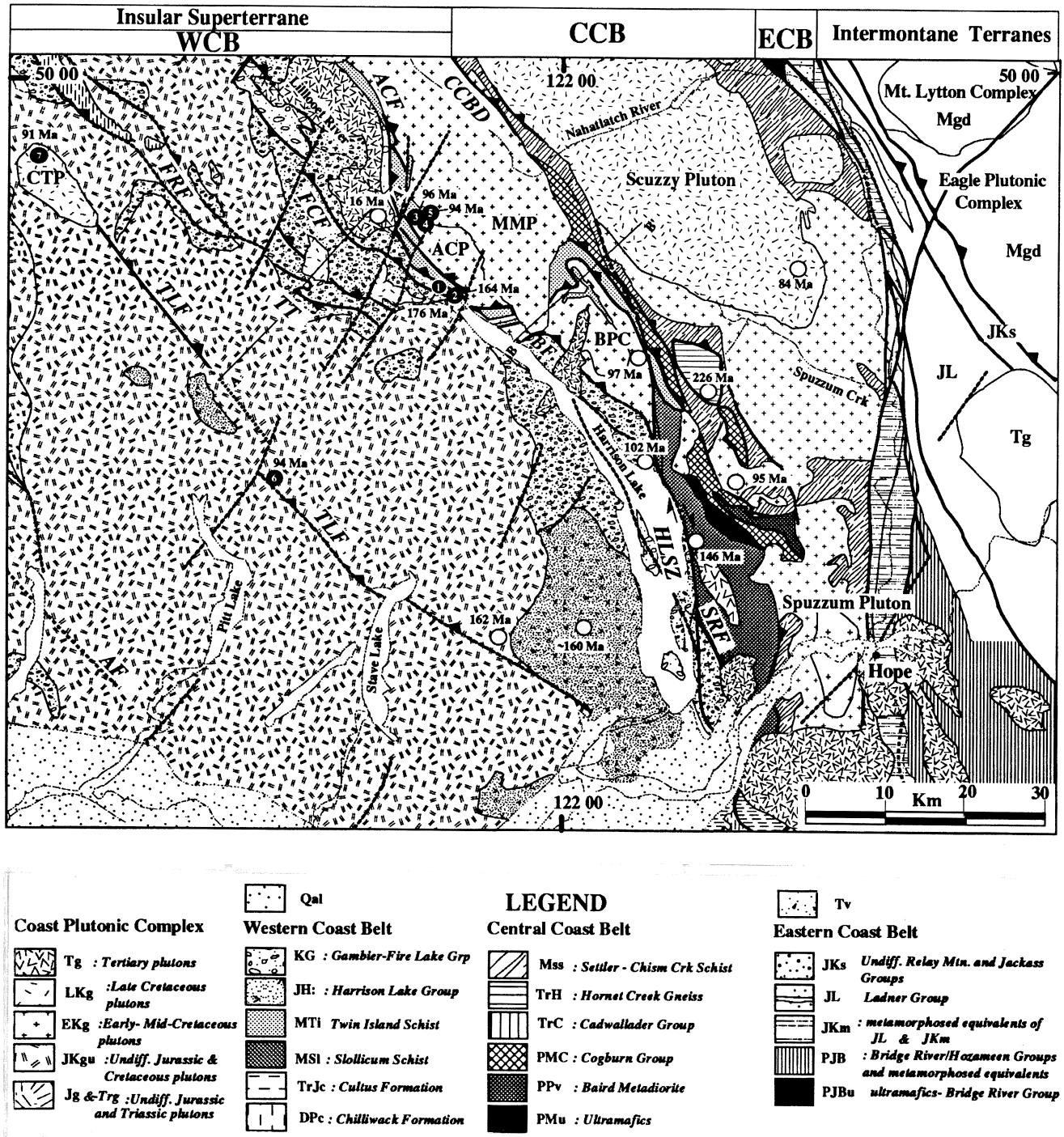



**Figure 14.** Geological map of the Coast Belt thrust system near Harrison Lake, and index to U-Pb sample locations and interpreted dates cited in the text. Solid circles with numbers are keyed to concordia plots in Figure 9. Open circles summarize the results of previous geochronologic studies cited in the text. See Figure 1 for legend to map units. Abbreviations: Ascent Creek Fault (ACF), Ascent Creek Pluton (ACP), Ashlu Creek Fault (AF), Breakenridge Fault (BF), Breakenridge plutonic complex (BPC), Castle Towers Pluton (CTP), Central Coast Belt detachment (CCBD), Fire Creek Fault (FCF), Fitzsimmons Range Fault (FRF), Harrison Lake shear zone (HLSZ), Mt. Manson Pluton (MMP), Slollicum Creek Fault (SCF) Terrarosa Thrust (TT), and Thomas Lake Fault (TLF).

Structurally above them, are rocks more typical of southeastern Coast Mountains. Twelve kilometres northeast of Stop 1-2, on the east side of the Slollicum Schist, the metamorphic grade is of high greenschist-low amphibolite facies (with garnets). East of this, and structurally above (Fig. 14), is a zone featuring undated metachert, amphibolite, marble and alpine-type ultramafic rocks called Cogburn Group by Gabites (1985), as well as the Baird metadiorite, correlated by Lowes (1972) and Gabites (1985) with the Yellow Aster Complex of the Northwest Cascades. In its along-strike structural location and range of lithologies, the Cogburn Group resembles the low metamorphic grade Pennsylvanian to Jurassic Elbow Lake Formation and associated lithologies of the Northwest Cascades. In protoliths it also resembles the Napeequa Schist of the matamorphic core of the Cascades, and the Bridge River Complex to the north.

Structurally above this, 18 km northeast of Stop 1-2, amphibolite grade pelitic and semipelitic schist, quartzo-feldspathic schist, minor quartzite and amphibolite (some with pillows) comprise the Settler Schist. Lowes (1972) and Evans and Berti (1986) correlated the Settler Schist with the Chiwaukum Schist in the metamorphic core of the Cascades. They considered them to be offset from one another by about 180 km across the dextral Fraser-Straight Creek Fault (*see* Stop 1-4). These rocks are poorly exposed along-strike to the southeast on Highway 7 at Stop 1-3, where they are intruded by quartz diorite of the Spuzzum Pluton. On the basis of near-continuity of outcrops, structural position, and range of lithologies, Monger (1991a) considered the Settler Schist to be possibly correlative with the Darrington Phyllite and Shuksan Greenschist of the northwest Cascades (Fig. 13).

In summary, correlation of rock units of the Coast Mountains north of the Fraser River (units underlined) with those of the northwestern Cascade Ranges to the south, suggests that the major rock units, disposed in fault slices, are from base to top: (1) **Harrison Terrane+Gambier Group** = Wells Creek Volcanics+Nooksack Formation; (2) **Slollicum Schist ±Carboniferous carbonate** = unnamed Late Jurassic clastics ±Chilliwack Group+Cultus Formation; (3) **Cogburn Group+Baird metadiorite** = Elbow Lake Formation+Yellow Aster Complex; (4) **Settler Schist** = Darrington Phyllite+Shuksan Greenschist (Monger, 1991a; Journeay and Csontos, 1989). Units (1), (2) are parts of southwestern Coast Mountains; units (3), (4) belong to southeastern Coast Mountains.

To the north of Stop 1-2, the snow-capped ridges of Mount Breakenridge are underlain by mafic orthogneiss dated at 96±1 Ma (Parrish and Monger, 1992), which is similar in age to that of the Spuzzum Pluton (Stop 1-3), and some plutons in the Cascade metamorphic core.

### *Coast Belt Thrust System: (mainly from Journeay and Friedman, 1993)*

The Coast Belt Thrust System (CBTS) is similar in many ways to classical fold and thrust belts developed in zones of large-scale crustal shortening. It is characterized by a foreland belt in the southwestern Coast Mountains, in which rocks are detached from their basement by thrust faults (Thomas Lake, Ashlu Creek faults, Fig. 1, 14), a foreland-hinterland transition marked by an imbricate zone of high-angle, out-of-sequence reverse faults (near Harrison Lake), and an exhumed hinterland belt of high-grade thrust nappes that represent the deep crustal root of the fault system (east of Harrison Lake, Fig. 1, 14, 15).

*Frontal Thrusts*

Thrust faults in the foreland belt of the CBTS are defined by generally northeast-dipping, locally southwest-dipping shear zones, and local stratigraphic discontinuities, that display transitional brittle-ductile behaviour. Fault zones are typically narrow (100-250 m but locally over 2 km wide) and contain a well-developed penetrative foliation and associated downdip stretching lineation (Lynch, 1992; Monger, 1993). The most significant of these structures are, from west to east, the Ashlu Creek, Thomas Lake, Fitzsimmons Range and Fire Creek faults (Fig. 14, 15a).

The Ashlu Creek Fault (Monger, 1993) extends southeast along Ashlu Creek (Fig. 1; *see also* Day 2) and is characterized by downdip lineations. Along most of its strike length, the fault juxtaposes mid-Cretaceous plutons with septa of Gambier Group in the footwall, with mainly Late Jurassic plutons locally capped non-conformably by Gambier strata, and mid-Cretaceous plutons cutting both (Monger, 1993; Monger and McNicoll, 1993). Strain associated with displacement is largely restricted to the fault zone (Monger, 1990; Lynch, 1992).

The Thomas Lake Fault (Monger, 1991b) extends southeast from the Cheakamus River valley, about 4 km south of the village of Whistler (Day 2) to Stave Lake. It is cut by mid-Cretaceous (ca. 91 Ma) granodiorite of the Castle Towers Pluton (Fig. 14). Asymmetric fault zone fabrics in mylonite zones along the trace of the fault record top-to-the southwest displacement (Monger, 1990, 1993). The fault juxtaposes plutonic rocks to the northeast, locally dated at 94 Ma, against Lower Cretaceous clastics of the Helm Formation in the footwall to the southwest.

The Fitzsimmons Range Fault (FRF) extends southeast from Whistler to the headwaters of Snowcap Creek, where it is obscured by Cretaceous plutonic rocks. The fault delineates an important structural discontinuity in the CBTS. Southwest of Whistler, it juxtaposes a hanging wall succession of fossiliferous Upper Triassic shales and siltstones against volcanic and related sedimentary rocks of the Early Cretaceous Gambier Group. In the Long Island region of Harrison Lake a similar fault juxtaposes well-foliated hanging wall phyllite of the Peninsula Formation against undeformed volcanic rocks of the Brokenback Hill Formation (Fig. 14). Folds in the upper plate of these two faults are overturned to the southwest and record a two-stage history of deformation (Journeay and Csontos, 1989).

The Fire Creek Fault (FCF; Lynch, 1990) cuts previously folded and faulted rocks of the Peninsula and Brokenback Hill formations (Fig. 14, 15a). It extends northwest along Fire Creek, through Fire Mountain to Snowcap Creek where it is cut by a younger right-lateral oblique-slip fault and offset to the northeast into the Lillooet River valley. The fault is marked by a zone of chaotic deformation in which asymmetric folds and shear zone fabrics record top-to-the-southwest displacement. It cuts the folded trace of the low-angle Terrarosa thrust and is estimated to have 8-10 km of apparent reverse-slip displacement (Lynch, 1990).

*Imbricate Zone*

The imbricate zone of the CBTS in the Harrison Lake region is defined by low-angle thrust faults that envelop high-grade schist and gneiss of the Breakenridge Complex and Slollicum Schist, and by high-angle reverse faults that cut the boundaries of these allochthonous sheets (Fig. 1, 14, 15a). Low-angle thrust faults are well-exposed along the east side of the Lillooet River valley (Roddick,

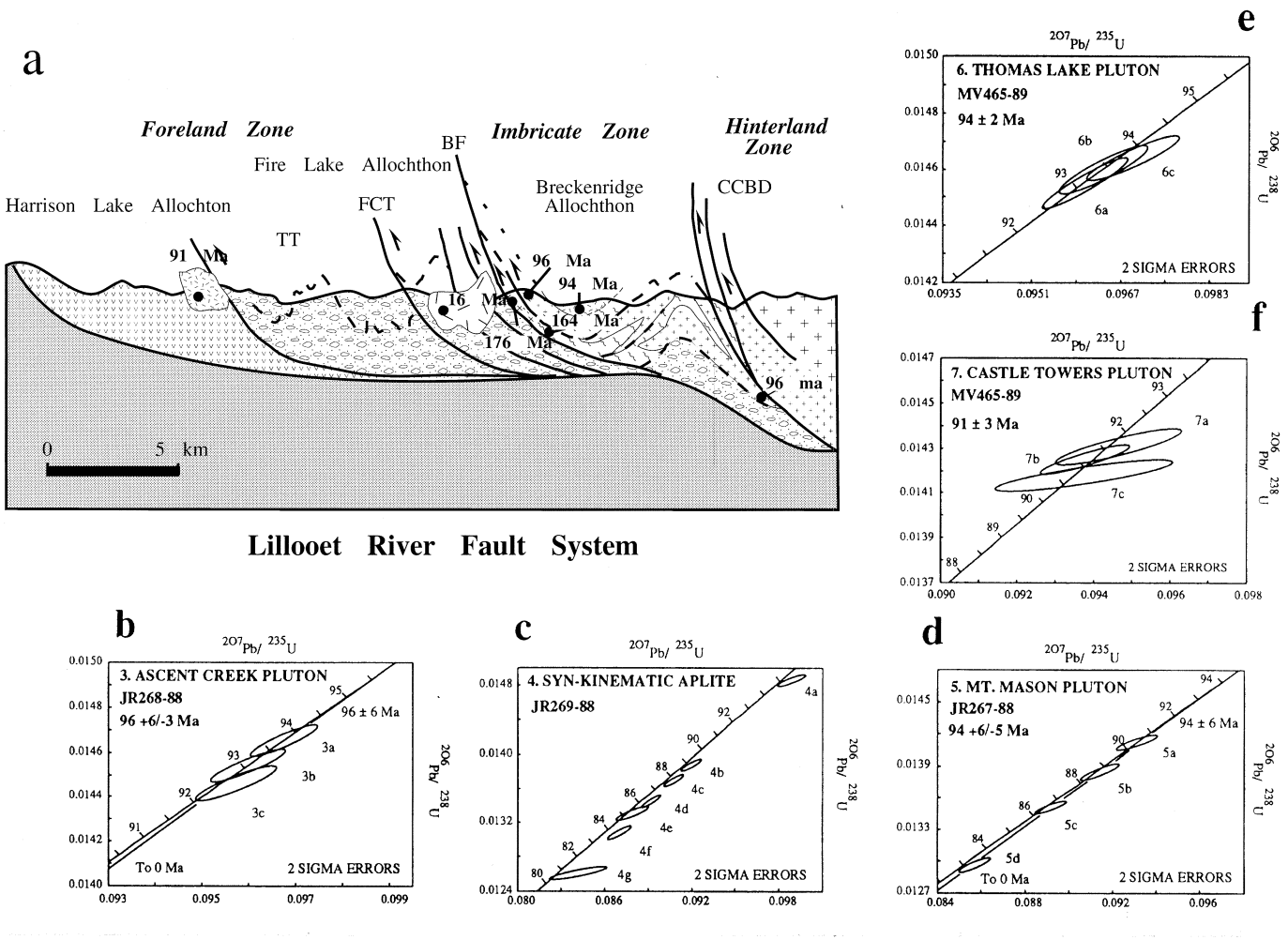



**Figure 15.** Geometry, kinematics and timing of the Coast Belt thrust system. Figure 15a is a crustal profile of the southern Coast Belt showing the geometry and kinematics of the CBTS: location of section on Figure 1. Concordia diagrams (15b, 15c, 15d, 15e and 15f) summarize the results of U-Pb analyses that constrain the timing of movement on the CBTS. See Figure 14 for location of dated samples.

1965; Journeay and Csontos, 1989) and can be traced southwestward into the core of the Breakenridge antiform. The most prominent of these structures are the Tipella and Ascent Creek faults. Both occur as folded thrusts in the upper plate of the Breakenridge Fault and are cut along the east flank of the Breakenridge antiform by high-angle reverse faults of the Central Coast Belt detachment. They duplicate metavolcanic and metasedimentary strata called Twin Island Schist and are marked by localized zones of high strain in which mylonitic fabrics and associated downdip stretching lineations are well-developed. Kinematic indicators in sheared plutonic rocks in the hanging wall of both the Tipella and Ascent Creek faults record top-to-the southwest displacement.

The Breakenridge Fault cuts the trace of folded low-angle thrust faults in both hanging wall and footwall, and marks the eastern boundary of the foreland zone. It dips steeply northeast and separates low-grade metavolcanic and metasedimentary rocks to the west from overlying schist and gneiss to the east (Fig. 14, 15). The fault trace extends southeast along the east side of the Lillooet River valley between Lillooet and Harrison lakes and continues along the east shore of Harrison Lake to Big Silver Creek, where it is offset to the south across right-lateral transcurrent faults of the Harrison Lake Fault (Fig. 14). The counterpart of the Breakenridge Fault on the east side of the HLSZ is the Slollicum Creek Fault (SRF), an east-dipping shear zone (Crickmay, 1930; Lowes, 1972) which marks the base of the Slollicum Creek allochthon (Fig. 14). North of Big Silver Creek, the Breakenridge Fault is mylonitic and contains a well-developed downdip stretching lineation. Asymmetric fault zone fabrics along the trace of the fault record a consistent top-to-the southwest sense of displacement.

*Central Coast Belt Detachment*

The Central Coast Belt detachment (CCBD) marks the eastern boundary of the imbricate zone and is the floor thrust to previously assembled thrust nappes of the overlying Eastern Coast Belt (Fig. 14, 15a). The fault dips steeply to the northeast and is marked by an interlocking network of ductile shear zones. It extends northwestward from Talc Creek to the headwaters of Nahatlatch River, where it is cut by post-kinematic intrusions of the Scuzzy-Mount Rohr plutonic complex. Mylonitic fault zone fabrics and downdip stretching lineations along the trace of the CCBD record an east-side-up displacement history. Locally, the fault zone and associated minor structures are overturned to the southwest, due to back-rotation caused by superimposed shortening and/or displacement above steep northeast-dipping thrust ramps of the CBTS.

Minor structures along the CCBD and in overlying thrust nappes record at least two generations of folding and associated fabric development. Early folds are tight to isoclinal with an axial planar foliation that predates peak conditions of high-pressure metamorphism in the region. Late generation folds are overturned to the southwest and record variable amounts of superimposed flattening and hingeline rotation. Adjacent to the fault zone, these folds are tight to isoclinal with hingelines that are either steeply inclined or reclined. Away from the fault zone, these late generation folds plunge to the northwest and are either upright or overturned to the southwest. They are syn-kinematic with respect to the CCBD and are correlated with southwest-vergent $D_2$ folds in lower plate rocks of the imbricate zone to the west.

Relative timing relationships between fold and fault structures of the CBTS indicate a two-stage history of Late Cretaceous shortening; an early stage of Alpine-style folding and thin-skinned thrusting above a basal decollement, followed by high-angle reverse faulting, out-of-sequence thrusting and southwestward telescoping of the metamorphic hinterland over flanking supracrustal arc sequences of the western Coast Mountains (Journeay and Csontos, 1989; Journeay, 1990a; Journeay and Friedman, 1993). The fold and thrust nappes root to the northeast beneath the high-grade metamorphic and granitic core of the southeastern Coast Mountains, and are interpreted to have formed during impingement and northeastward underplating of the Insular superterrane early in Late Cretaceous time (Journeay, 1990a).

The timing of early-stage thrusting is bracketed by the emplacement of syn-orogenic plutonic suites, which yield U-Pb zircon dates of 97 ± 1 Ma and 96 +6/-3 Ma (Fig. 15b-d). Late-stage shortening is bracketed by late- and post-kinematic plutons, which yield U-Pb zircon dates of 94 ± 2 Ma and 91 +4/-3 Ma, respectively (Journeay and Friedman, 1993). In the imbricate zone, this episode of deformation is bracketed by the emplacement of syn- and post-kinematic plutons, which yield U-Pb zircon dates of 96 +6/-3 Ma and 94 +6/-5 Ma, respectively (Fig. 15e-f).

The development of thick-skinned out-of-sequence thrusts in the hinterland of the CBTS signals a northeastward migration of the deformation front in Late Cretaceous time. This shift in the locus of thrusting resulted in southwestward telescoping and structural inversion of the metamorphic hinterland and may account for the complex stacking order and thermal history along the boundary between southwestern and southeastern Coast Mountains (Journeay, 1990a).

## *History of Metamorphism (From Journeay, 1990a)*

Metamorphism along the southeast flank of the Coast Mountains can be linked directly to the imbrication of thin- and thick-skinned thrust nappes in the Coast Belt Thrust System (CBTS) ($M_1$), and to the emplacement of Late Cretaceous plutons ($M_2$) during subsequent ramping and southwestward overthrusting of this thrust stack onto the eastern flank of the Insular Superterrane (Journeay, 1990a,b). The timing of $M_1$ metamorphism is bracketed by syn- and post-kinematic plutons, which give U-Pb dates of 96 +6/-3 and 91.5 ± 2 Ma, respectively (Friedman et al., 1992; Journeay and Friedman, 1993). $M_2$ metamorphism is bracketed by crystallization ages of the Mount Mason and Scuzzy-Mount Rohr plutonic suites, which yield U-Pb dates of 91.5 ± 2 Ma and 86-84 Ma, respectively (Friedman and Armstrong, 1990; Friedman et al., 1992; Parrish and Monger, 1992).

An alternate view to the above, expressed by Miller et al. (1993) and Brown and Walker (1993) is that regional metamorphism in the Harrison Lake area may be in part the product of magma loading. In their model, upward movement and lateral spreading of magma provides the heat, loading and at least some of the metamorphic fabrics seen in high-grade gneisses of the southern Coast Mountains and adjacent Cascades.

*$M_1$ Metamorphism*

Early stage $M_1$ metamorphism in the southern Coast Mountains is characterized by a Barrovian sequence of mineral assemblages that range in grade from lower greenschist facies to middle and upper amphibolite facies (Reamsbottom, 1971, 1974; Lowes, 1972; Pigage, 1976; Bartholomew, 1979; Gabites, 1985). These $M_1$ assemblages are best developed in the imbricate and hinterland zones of

the Coast Belt Thrust System, and in isolated pendants of gneissic rock within late- and post-kinematic plutons to the north.

Matrix phases of chlorite, muscovite, biotite and feldspar are aligned with early $S_1$ flattening foliations and are locally crenulated by younger generation $F_2$ folds. Porphyroblastic phases of garnet, staurolite, hornblende, kyanite and sillimanite are syn- and late-kinematic with respect to the dominant schistosity, and locally record significant post-metamorphic strain. Inclusions of chlorite, biotite and andalusite within these porphyroblasts occur in fault slivers east of Harrison Lake, and imply an earlier history of crystallization that may either be part of the same prograde event, or a relic of an older history of metamorphism.

$M_2$ metamorphism is post-kinematic with respect to structures and associated fabrics of the Coast Belt Thrust System, and is spatially associated with Late Cretaceous plutons of the Scuzzy-Mount Rohr intrusive suite (Hollister, 1966, 1969a,b; Lowes, 1972; Journeay, 1990a). M2 assemblages that occur in the imbricate and hinterland zones of the Coast Belt Thrust System include late stage overgrowths and porphyroblastic phases of andalusite, sillimanite, garnet and hornblende. Equivalent phases in graphitic schists along the north and east flanks of the Scuzzy Pluton include porphyroblastic garnet, staurolite and andalusite. In the vicinity of Kwoiek Needle, $M_2$ assemblages record a history of Buchan-type metamorphism in which metamorphic grade decreases away from the main body of the Scuzzy Pluton (Hollister, 1966, 1969a,b). The sequence and spacing of $M_2$ assemblage zones (staurolite-garnet-andalusite-kyanite-sillimanite) implies a nearly isobaric field gradient, and suggests that emplacement of the Scuzzy Pluton and associated $M_2$ metamorphism most likely occurred at intermediate crustal levels (ca. 10-12 km; Hollister, 1966).

*Conditions of Metamorphism*

Estimates of "peak" metamorphic conditions are based on recalculations of conventional garnet-biotite and garnet-plagioclase thermobarometers (Journeay, 1990a), using mineral compositions reported by Reamsbottom (1974), Pigage (1976), and Bartholomew (1979). Preliminary results of these calculations are summarized in Figure 16.

Assemblages that occur in fault slivers east of Harrison Lake (Reamsbottom, 1974) record pressures of 7.4-9.6 kb and temperatures of between 675-750°C. Equivalent rocks of the Yale Creek region (Pigage, 1976) record pressures of 6.9-8.0 kb at 600-635°C. These data require crustal loads of ca. 20-30 km during peak conditions of metamorphism and are consistent with a model of mid-Cretaceous thrust imbrication and associated tectonic burial, as proposed by Lowes (1972) and Journeay and Csontos (1989).

Assemblages that occur adjacent to post-kinematic plutons in the Mount Breckenridge area (Reamsbottom, 1974), and within sillimanite-bearing rocks of the Yale Creek region (Bartholomew, 1979), record "peak" metamorphic conditions of 4.5-6.1 kb/660-690°C and 3.7-6.0 kb/575-650°C, respectively (Fig. 16). These data are consistent with the results of Hollister (1966), and suggest that the emplacement of Late Cretaceous plutonic suites and associated $M_2$ metamorphism within this part of the Coast Belt occurred at intermediate crustal levels (15-20 km), presumably during southwestward ramping and/or unroofing of thrust sheets in the imbricate and hinterland zones of the Coast Belt Thrust System.

Return down road, through Harrison Hot Springs, to rejoin Highway 7 at 126.3 km.

**128.0 km** Settlement of Agassiz; continue east on Highway 7.

**129.2-130.1 km** Roadcuts of Miocene granodiorite.

**131.2 km** To north, cliffs of Mississippian limestone on western Bear Mountain; a gulley to the east is taken to mark the Harrison Fault. East of this, sandstones and pelites, probably the southward, along-strike extension of the Peninsula Formation, are hornfelsed by small Tertiary intrusions.

**139.6-142.4 km** Roadcuts in Oligocene (24 Ma, K-Ar) granodiorite, coeval with the Chilliwack Batholith near the International Boundary (Fig. 1). At its east end, the pluton cuts pelitic schists with abundant calcsilicates. These may be derived from the along-strike Slollicum Schist, for their composition suggests they originated as calcareous shale and sandstone.

**142.2-142.4 km** Mafic schist and gabbro, part of a body of metagabbro, hornblendite and local talc-schist that extends east from here for about 4 km, along the north side of Johnsons Slough (crossed at 142.3 km). These mafic rocks may delineate the northward extension of the Shuksan Fault zone (Fig. 12), into the southern Coast Mountains, as first recognized by Lowes (1972).

**COMMENT** This is a key location on the field trip, for if there is anything like a suture on the trip, between the Insular Superterrane of southwestern Coast Mountains, and the collage of terranes of southeastern Coast Mountains, it lies in the vicinity of these mafic rocks (Fig. 1, 6, 13)! They are part of a series of aligned bodies of

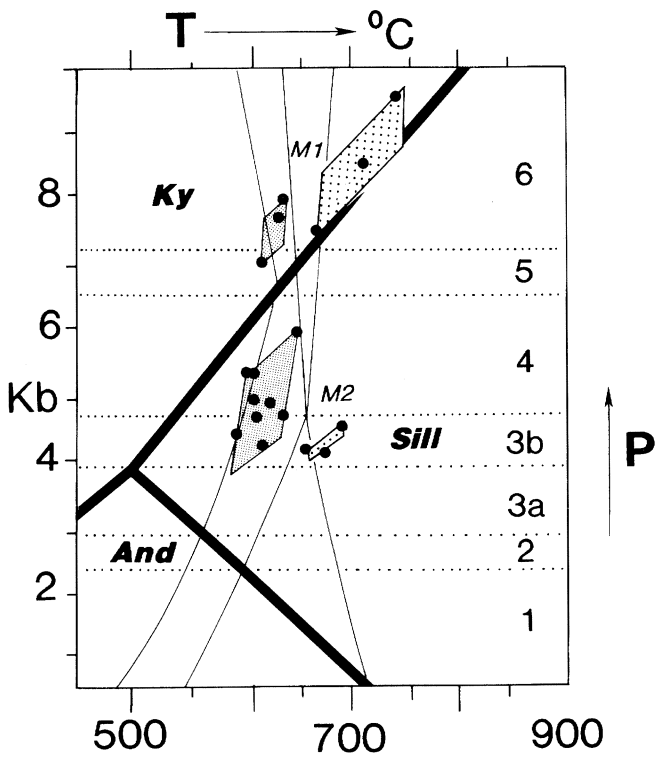



**Figure 16.** P-T diagram summarizing conditions of $M^1$ and $M^2$ metamorphic conditions in the southern Coast Belt (taken from Journeay, 1990a,b). P-T estimates based on mineral chemical data of Reamsbottom (1974) and Pigage (1976).

mafic and ultramafic composition that extend along or close to the trace of the Shuksan Fault from Washington northwards into the southeastern Coast Mountains (Fig. 1, 13). They locally are imbricated with chert and amphibolite (=Cogburn =(?) Elbow Lake =(?) Bridge River). Near the International Border they separate Harrison and Chilliwack terranes to the west from Shuksan Greenschist and Darrington Phyllite to the east, and north of Fraser River, Slollicum Schist, which is derived from strata of southwestern Coast Mountains, from Cogburn Group and Settler Schist typical of southeastern Coast Mountains.

North of this locality, lenses of hornblendite and talc schist occur along Ruby Creek (crossed at 146.2 km) and extend northwards into the area of Old Settler Mountain, ca. 18 km north of Stop 1-3, where they are tectonically mixed with the Cogburn Group (=? Bridge River terrane). East of the Cogburn Group lies pelite, quartz-rich metasandstone and local amphibolite of the Settler Schist (Pigage, 1976; Monger, 1991a). In places, disrupted fabrics in the sandstones are overprinted by high-grade metamorphic minerals (Monger, 1991a). To the south of Fraser River in Wahleach Creek (Fig. 13), greenschist grade, quartz-rich metasandstones, siltstones and pelites with distinctive highly irregular and disrupted structures, flank the mafic rocks on the east. Farther south, in the southern Cheam Range, small bodies of ultramafic rock are associated with chert and clastics. South of this, mafic rocks and disrupted sandstones and pelites are present near Foley and Nesakwatch creeks, in the Chilliwack valley. Just north of the International Boundary, similar clastic rocks mapped as Darrington Phyllite (Jewett, 1984) are intruded by the west side of the Tertiary Chilliwack Batholith. Just south of the International Boundary similar mafic and ultramafic rocks are structurally mixed with metacherts of the Elbow Lake Formation, and lie west (and structurally below) *bona fide* Darrington Phyllite and Shuksan Greenschist.

Monger (1991a) suggested on the bases of composition, near-continuity of outcrop areas, and structural position, that the Settler Schist was correlative with the Shuksan and Darrington rocks. The latter contain blueschists yielding K-Ar and Rb-Sr dates mainly in the 150-120 Ma range (Armstrong and Misch, 1987). This means that, arguably, a subduction complex active into the Early Cretaceous was present in what is now southeastern Coast Belt. The argument is supported by the distinctive disrupted fabrics locally seen in the sandstones, which are similar to those found in modern and ancient accretionary complexes. In places in the Settler Schist these survive overprinting by Late Cretaceous high-grade metamorphic min- erals. Farther north, Settler Schist is correlated with the Cayoosh assemblage and Chism Creek Schist to be seen on Day 2, which locally contains similar disrupted fabrics.

**145.2 km** East end of exposure of mafic rock.

**146.2 km** Ruby Creek. To north about 1 km, in creek bottom, a large block of metachert is associated with talc-schist.

***STOP 1-3:*** **147.0 km** Intrusive contact of Spuzzum quartz diorite with Settler Schist. Park near highway, and walk up Ruby Creek logging road for about 0.5 km, to where biotite garnet schist is intruded by the western margin of the Spuzzum Pluton. In places a northeast plunging mineral lineation is present. The metamorphic grade varies according to the proximity to the Spuzzum quartz diorite. Three kilometres to the northeast, sillimanite is found along the contact, whereas about 4 km north along Ruby Creek, approximately 2 km from the contact, staurolite is the dominant phase.

Return, and continue east along Highway 7.

**147.3 km** Outcrop of Spuzzum hornblende quartz diorite, with northeast plunging mineral lineations.

**COMMENT** A variety of isotopic dates have been obtained from the Spuzzum Pluton. U-Pb dates from this part of the pluton are ca. 100 Ma (Journeay and Friedman, 1993), although farther north ages of 96-97 Ma occur (e.g., Parrish and Monger, 1992). K-Ar dates range from 80-112 Ma. The Spuzzum Pluton continues south of the Fraser River, with no apparent lateral offset, and there is included as part of the Cascade metamorphic core. Plutons similar in age, composition and probable depth of emplacement are present across the International Boundary in the metamorphic core of the Cascades. As noted in the introduction, they also occur in southeastern Alaska and the Idaho Batholith (Zen, 1985).

**155.4-158.8 km** Spuzzum quartz diorite. Near 155.4 km is a xenolith of marble, enclosed by within the quartz diorite.

Also at this locality is one of the paleomagnetic sites sampled by Irving et al. (1985), results from which led to suggestions that these rocks had possibly moved northward by about 2400 km and rotated clockwise since mid-Cretaceous time. As recognized by Irving et al. (1985), and argued strongly by Butler et al. (1989), the absence of paleohorizontal markers in the pluton makes it possible that the anomalous paleomagnetic directions are due to systematic northeast-side-up tilts about north-northwest-trending axes, rather than latitudinal shifts. The argument has been re-opened recently. Ague and Brandon (1992) and Wynne et al. (1993) suggest on the basis of tilt calculations based on geobathymetry and new paleomagnetic data, that early Late Cretaceous rocks in the region (presumably west of the Pasayten Fault; Fig. 1) were about 3000 km farther south of their present position with respect to Cretaceous cratonal North America, than they are today (*see* Cowan, 1994).

There is continuous outcrop of mafic Spuzzum Pluton east of here until the Hope Fault (a segment of the Fraser-Straight Creek fault system; Fig. 1, 17). The country rock hosting the Spuzzum is mainly Settler Schist, although the western part of the pluton cuts both Cogburn Group (=? Bridge River) and Slollicum Schist (=Harrison terrane and overlying Jura-Cretaceous strata).

**159.0 km** Underpass beneath the Trans-Canada Highway. Follow road around to rejoin Highway heading north, then take turn-off to east at 159.7 km on to old highway. Drive north on this.

***STOP 1-4:*** **160.1 km** View of Fraser-Staight Creek fault zone, at outcrop of reddish Eocene conglomerate, dated with palynomorphs by G.E. Rouse (pers. comm., 1986). This is one of a series of outcrops of conglomerate that occurs intermittently along the western, main, strand of the Fraser fault system, here called the Hope Fault (Fig. 17; McTaggart and Thompson, 1967), and along its southward extension in Washington, called the Straight Creek Fault. South of the town of Hope (2 km south-southeast of here) both conglomerate and the fault are intruded by the Oligocene Chilliwack

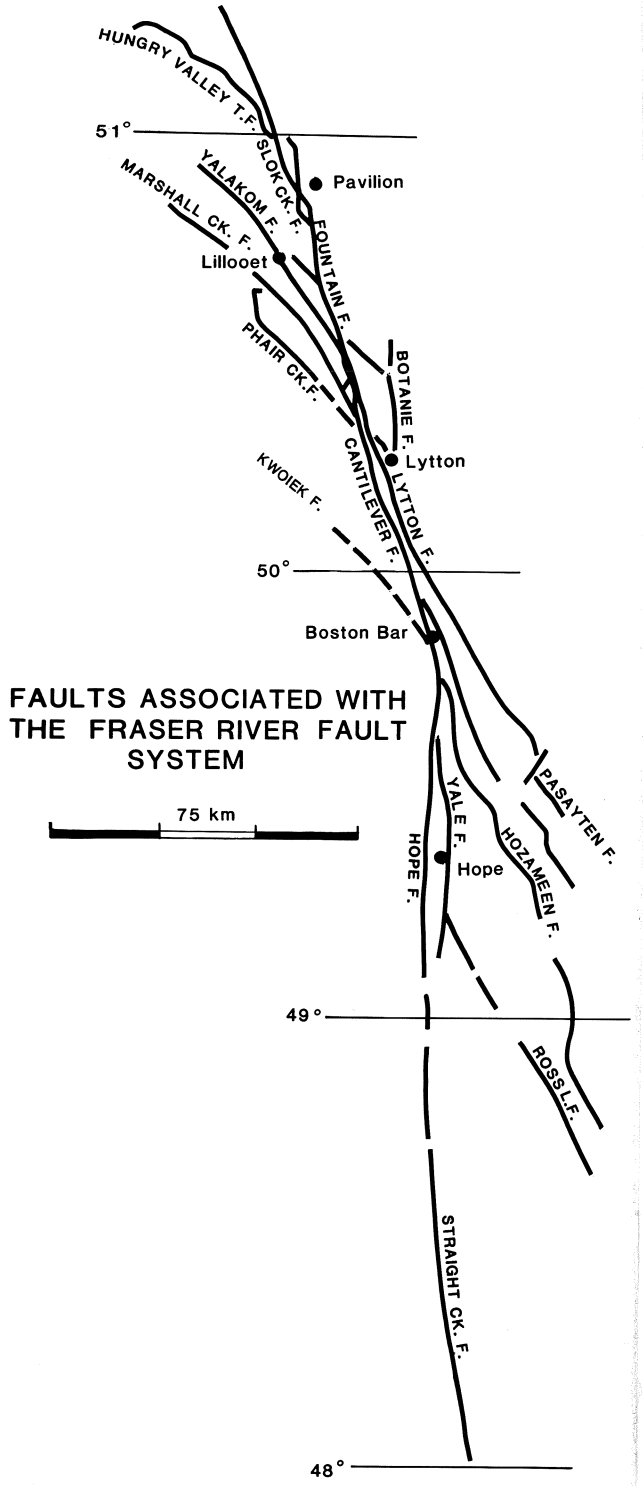



**Figure 17.** Some faults associated with the Fraser-Straight Creek fault system.

Batholith, which is K-Ar dated at ca. 25 Ma in British Columbia, and is as old as 35 Ma in Washington. This date puts a minimum age on dextral displacements on the fault.

Near Stop 1-4, the Hope Fault can be traced to the north by a series of aligned notches part way up the west side of the Fraser River valley, and by small bodies of serpentinite and Tertiary conglomerate along it. The early Late Cretaceous Spuzzum quartz diorite and Settler Schist west of the fault are juxtaposed across it with the latest Cretaceous and(?) early Tertiary Custer Gneiss, here mainly quartzofeldspathic gneiss and pegmatitic intrusions, which underlies the lower part of the valley (Fig. 1). In turn, the Custer Gneiss is faulted to the east across a fault called the Yale Fault near Hope (Fig. 18; McTaggart and Thompson, 1967) with greenschist and lower grade rocks of the Hozameen Complex (=Bridge River terrane). East of Stop 1-4, the Yale Fault is exposed about one third of the way up the eastern wall of the valley (elevation ca. 800 m). It juxtaposes gneissic and higher grade metamorphic rocks to the west with low-grade Hozameen strata to the east; in this the Hope Fault resembles the Petch Creek Fault farther north (Ray, 1986) and the south-southeast-trending Ross Lake Fault farther south.

**COMMENT** The Fraser-Straight Creek Fault is a major, Eocene, dextral wrench fault system that trends north-south, acutely across the north-northeast regional grain of Late Cretaceous-earliest Tertiary age. Movement on the fault is post-47 Ma, the youngest age of rocks and structures demonstrably offset across the fault, and pre-35 Ma, the oldest date from the Chilliwack Batholith which crosscuts the fault south of Hope near the International Boundary.

Estimates of offset across the fault range from about 80 to 190 km (Fig. 18). Although the fault was recognized in the late 1800s in Canada, Misch (1966) and his students probably were the first to identify the Straight Creek segment in Washington as a right-lateral strike-slip fault. Later Misch (1977) suggested that the offset was ca. 190 km, by matching the lithologically similar Settler and Chiwuakum schists across it.

Later workers in Canada (Mathews and Rouse, 1984; Kleinspehn, 1985; Monger in Price et al., 1985) estimated more conservative displacements of between 80-110 km, mainly by matching the following outcrop areas and bounding faults across the fault system: (1) high-grade metamorphic rocks (Settler Schist) southwest of the "Kwoiek Fault" with Custer Gneiss southwest of the Ross Lake Fault; (2) Bridge River and Hozameen complexes (Bridge River terrane), southwest of, respectively, Yalakom and Hozameen faults; and (3) the Jackass Mountain Group (Methow terrane), southwest of the Hungry Valley Fault, west of the river, and Pasayten Fault, east of the river (Fig. 18B).

Recently, Coleman and Parrish (1991) studied extension tectonics associated with Eocene intrusions (ca. 47 Ma) near Lillooet, and from their results matched the Mission Ridge Fault (which lies close to the Yalakom Fault) northwest of Lillooet, with the Ross Lake Fault (whose northward extension appears to merge with the Yale/Petch Creek Fault as it approaches the Fraser River). This gives a maximum displacement of about 140 km (Fig. 18C; restoration juxtaposes the towns of Hope and Lillooet). Matching Eocene rocks and structures across the fault is critical, as these are probably the youngest features to be offset. This displacement also is similar to the amount of offset between Permian rocks of the Mount Lytton Complex and the Farwell Canyon Pluton the Chilcotin River area, about 130 km north of Lillooet (Friedman and van der Heyden, 1992). However, most recently Parrish and Monger (1992) obtained a U-Pb date of 68 Ma from mylonitic granitic rock of the Custer Gneiss near Hope, a date comparable with ages from the latest

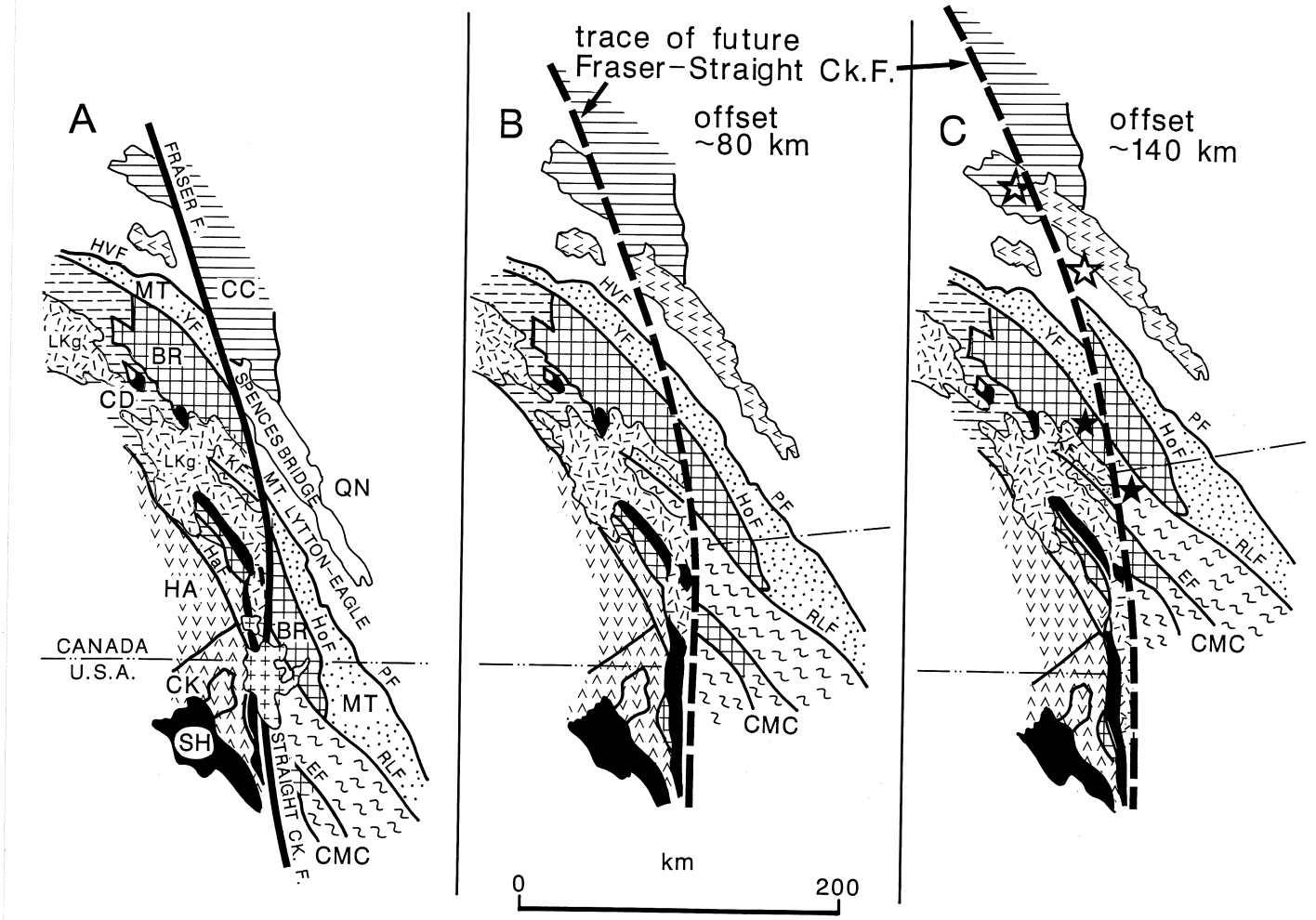



**Figure 18.** Terranes and structures of the southeast Coast Mountains, Intermontane Belt and adjoining parts of the Cascade Ranges, used to demonstrate possible offsets on Fraser-Straight Creek Fault.

A. Present map pattern; terrane abbreviations BR, Bridge River; CC, Cache Creek; CD, Cadwallader; CK, Chilliwack; HA, Harrison; MT, Methow; SK, Shuksan; *Fault abbreviations:* EF, Entiat F., HaF, Harrison Fault; HoF, Hozameen Fault; HvF, Hungry Valley Fault; KF, Kwoiek Fault; PF, Pasayten Fault; RLF, Ross Lake Fault; YF, Yalakom Fault; LKg, Late Cretaceous granite; CMC, Cascade Metamorphic Core.

B. restored dextral offset of 80 km on Fraser-Straight Creek faults.

C. restored dextral offset of ca. 140 km on Fraser-Straight Creek faults; black stars: offset early Tertiary granitic rocks, structures; open stars: offset Permian granitic rocks (Mount Lytton, Farwell Canyon bodies).

Cretaceous Bendor plutons, which are located 15-20 km southwest of the Eocene intrusions near Lillooet; the amount of offset on the Fraser-Straight Creek fault system thus may be about 120-125 km.

On the latter restoration, parts of Bridge River and Methow terranes do not match up, and the Custer Gneiss is juxtaposed with greenschist-grade Bridge River rocks west of the fault near Lillooet. If this restoration is valid, there must be differential **vertical** movement of several kilometres (10 km) to account for the different levels of erosion exposed across the fault system. The Custer Gneiss is far more deeply eroded than matching Bridge River rocks west of the fault (depths of 20+ vs. 10 km; based on the metamorphic grade of rocks exposed at the surface), but because ages of plutons in both Custer and Bridge River rocks southwest of Lillooet are comparable, the Custer may represent a deeper, latest Cretaceous-earliest Tertiary crustal level. In addition, if the terranes are disposed in the thrust sheets as suggested by structural and reflection seismic data, erosion of several kilometres will shift the location of the surface boundaries of terranes, and thus account for the mis-match of different parts of Bridge River and Hozameen terranes.

Note that along the route north to Lillooet, a variety of names are given to individual faults in what is strictly part of the Fraser fault system (Fig. 1, 17). The offset across the Fraser Fault is not a clean break. In general, older, originally north-northwest trending, ductile faults swing into alignment with the more northerly-trending brittle Fraser fault system as they approach it and presumably carry the younger Fraser Fault movements. For example, the north-northwest-trending Ross Lake and Pasayten faults blend into the Fraser fault system.

**161.5 km** Drive north, and rejoin the Trans-Canada Highway, and head north on it.

**161.7-162.8 km** Outcrops of greenschist-grade Hozameen Group occur along the highway at an elevation of about 170 m; to the east, the contact of similar rocks with Custer Gneiss (Yale = Petch Creek = Ross Lake Fault) is at an elevation of about 700 m on the east side of the valley; there appears to be late (post-Yale) vertical movement, as would be expected also from the presence of Eocene conglomerates along the Hope Fault to the west.

**164.3 km** White, highly fractured felsic rock, locally strongly foliated, and correlated with the Custer Gneiss; note juxtaposition of these rocks with the Hozameen rocks just seen. The brittle deformation (and zeolitic alteration) here is characteristic of many rocks in the Fraser Canyon north of here, which is eroded along the trace of the Fraser fault system.

**COMMENT** For the next 43 km the highway runs through similar, generally highly fractured, variably mylonitic granitic rocks, granitic gneiss and local schist, that are part of the Custer Gneiss of Daly (1912). South of Hope, the Custer Gneiss lies between the south-southwest-trending Ross Lake Fault and the south-trending Fraser Fault, and thus broadens to the south (Fig. 1). In crossing the International Boundary, Custer Gneiss changes its name to Skagit Gneiss (Misch, 1966) which is a major component of the Cascade metamorphic core south of the International Boundary. Along the Fraser River, the Custer Gneiss comprises a 4 km wide sliver, sandwiched between the Hope Fault to the west, and the Yale (or Petch Creek Fault of Ray, 1986) to the east. Within this sliver, brittle deformation is superimposed on a mylonitic fabric. Very little isotopic dating has been done on these rocks largely because of their alteration. The only U-Pb date is that of 68 Ma mentioned above (Parrish and Monger, 1992).

Note that west of the Fraser fault system, between Hope and Lillooet, granitic rocks become progressively younger from about 96 Ma (Spuzzum Pluton) near Hope, through 84 Ma (Scuzzy Pluton) to 70-65 Ma (Bendor plutons) to 47 Ma near Lillooet (Fig. 1). The metamorphic grade also decreases northward; south of the Kwoiek Fault rocks are of amphibolite grade and called Settler Schist; north of it they are generally of greenschist grade.

The Yale (or Petch Creek) Fault, which is on the east side of this Custer Gneiss sliver, ends south of Boston Bar. Its offset equivalent near Lillooet appears to be the Lillooet and(?) Mission Ridge faults (Fig. 1). The latter is a northeast-dipping normal fault with subgreenschist grade Bridge River and Cretaceous strata in the hanging wall to the northeast, and high greenschist-lowest amphibolite grade Bridge River and(?) Cayoosh strata in the footwall to the southwest (Coleman and Parrish, 1991).

**173.3 km** Cross Emory Creek; light coloured Custer Gneiss is seen to the east across the Fraser River, with darker coloured Hozameen topographically above it. Scattered outcrops of Custer Gneiss occur along highway.

**180.1 km** Settlement of Yale, which was the upper limit of steam navigation on the Fraser River and a major supply depot during the building of the Canadian Pacific Railway in the 1870-1880s.

The Highway continues through more-or-less continuously outcropping, variably mylonitic and brittley deformed felsic intrusive rock with minor screens of schist, all included in Custer Gneiss.

**180.4 km** Yale Tunnel.

**186.3 km** Saddle Rock Tunnel.

**188.8 km** Sawmill Creek; road cuts of orthogneiss/gneissic quartz diorite of the Custer Gneiss continue.

**190.1 km** On the east side of the Fraser River, gneissic quartz diorite of the Custer Gneiss is truncated by the near-vertical Yale and/or Petch Creek Fault. East of it are greenschist grade phyllite, metachert and minor marble of the Hozameen Complex (Bridge River terrane). The dip of the foliation in rocks on both sides of fault is east-northeast at 75-60°. At this place the fault crosses the bend in Fraser River.

**190.1 km** Dominant quartz diorite gneiss and amphibolite contain stringers of pelitic and siliceous schist (metachert or quartzose sandstone?), and rare marble.

**192.3 km** Sailor Bar Tunnel; granitic rocks predominate. These rocks are included in the Custer Gneiss.

**196.3 km** Spuzzum Creek (higher up the creek is the type locality of Spuzzum quartz diorite). The Hope Fault (the main strand of the Fraser fault system) crosses the road at about this point; if possible look back southwards; the notch marking the trace of fault is visible on the ridge.

**197.3 km** Fractured quartz diorite, west of Fraser fault system; no dating has been done here; a prediction is that these rocks should be appreciably older than granitic rocks at Sailor Bar Tunnel which lie east of the fault (early Late Cretaceous vs. latest Cretaceous-early Tertiary).

**199.6-200.3 km** Biotite quartz diorite, orthogneiss, and local schistose screens. As mapped, these are marginal parts of the Spuzzum Pluton.

**200.3 km** Alexandra Bridge over the Fraser River. To the north and lower down, the suspension bridge carrying the old highway over the river is visible.

**201.9 km** Outcrops of variably fractured quartz diorite and pegmatite.

**203.5-204.5 km** Continuous outcrops of fractured, quartz diorite and pegmatite.

**204.5 km** Alexandra Tunnel; fractured quartz diorite and pegmatite.

**COMMENT** These rocks on the east side of Fraser River appear to be a horse in the Fraser fault system, sandwiched between the Petch Creek Fault to the east and the Hope Fault to the west. However, on lithology, they appear to be more closely related to the early Late Cretaceous Spuzzum/Scuzzy plutons west of the Hope Fault than to Custer Gneiss; as everywhere along Fraser Canyon, more isotopic dating is needed!

**207.8 km** Cross Hope Fault; trace marked by gulley, but no marked change of lithology.

**208.8 km** Cross the southern edge of Hells Gate Pluton,

K-Ar dated at 35 Ma. The pluton is roughly equidimensional. The rocks in it are variably fractured, but generally more massive than the country rocks which they intrude. The pluton is biotite hornblende granodiorite. It is locally porphyritic with feldspar phenocrysts. In contrast with the country rock, the pluton contains few pegmatites. Its eastern boundary appears to be faulted on the Hope Fault; if the existing date is accepted, this means there was at least some (vertical?) movement on the Hope Fault as late as Oligocene time. Again, more work needed!

**209.3 km** Hells Gate Tunnel.

**209.7 km** Ferrabee Tunnel.

**211.7 km** Air tramway to permit viewing of Hells Gate rapids and fish ladder on Fraser River.

**212.7 km** Northern edge of Hells Gate Pluton. North of this is quartz diorite with minor screens of biotite and amphibolite-rich schist, and local pegmatite and quartz feldspar dykes, the latter presumably related to the Hells Gate Pluton.

**214.3 km** China Bar Tunnel; granitic rocks continue north of this for about 2 km.

**216.5 km** Petch Creek; type locality of the Petch Creek Fault of Ray (1986), which here is coincident with the Hope Fault. Between 207.8 km and here, the Hope and Petch Creek (Yale) faults merge into the main strand of the Fraser Fault. For 6 km north of here, the fault follows the Fraser River to a point north of Boston Bar, where it emerges for a short distance on the west side of the river. To the west are mainly granitic and high grade metamorphic rocks; to the east are low grade metamorphic rocks of Bridge River and Methow terranes.

**217.1-217.6 km** Small outcrops of serpentinite mark the boundary between Hozameen and Methow terranes. These are presumably fragmented parts of the Coquihalla serpentinite belt, which is associated with Triassic MORB basalt of the Spider Peak Formation at the base of the Methow terrane (Fig. 8; Ray, 1990).

**220.0 km** Anderson River; this marks the southern end, along the highway, of more-or-less continuous outcrops of phyllitic and slaty argillite and siltstone of the Jurassic Boston Bar Formation of the Ladner Group of Methow terrane (O'Brien, 1987). The Fraser Fault runs below Fraser River to the west; the west side of the valley is underlain by granitic rocks with abundant screens of biotite schist. The granitic rocks are presumably northeasternmost parts of the Spuzzum Pluton. About 10 km west southwest of here the latter is intruded by the granodioritic Scuzzy Pluton, which is U-Pb dated at 84 ± 1 Ma (Parrish and Monger, 1992).

**221.6-224.0 km** Settlement of Boston Bar. Outcrops of thin bedded argillite and siltstone of the Jurassic Boston Bar Formation have a slaty and phyllitic cleavage. At this place, the formation is a narrow sliver about 3 km wide, faulted on the east against tuffaceous shales and volcanic sandstones of the Dewdney Creek Formation of the Ladner Group (Fig. 8). Fossils in the latter, some in flows and tuffs about 5 km east of here, are Middle Jurassic (Aalenian-early Bajocian; 178-172 Ma; O'Brien, 1987).

**225.5 km** Northern end (along the highway) of continuous exposures of Ladner Group.

**COMMENT** To the west of here, the Fraser Fault runs for about 4 km along the west side of Fraser River. Ladner slates (Fig. 8) of Methow terrane are exposed low down in the valley, west of the river. On the western valley wall, (elevation ca. 500 m) they are separated by the Fraser Fault (marked locally by serpentinite) from a 0.6 km wide strip of foliated greenstone, metachert and phyllite of the Bridge River terrane. Above this (elevation ca. 700 m) and separated from it by a shear zone (= Kwoiek Fault) are garnet-biotite schists and quartz-rich biotite schists (=? Settler Schist), that in turn are intruded by granitic rocks exposed near the top of the valley wall.

This is the locality in which the southern Cordillera Lithoprobe deep seismic reflection Line 18 crossed the Fraser River. On the profile, no obvious offset of reflectors is apparent below about 8 s (*see* Fig. 6; Cook et al., 1992; Varsek et al., 1993), although data from this area currently are being re-processed at the University of British Columbia, and appear to show a small offset (R.M. Clowes, pers. comm., 1993). By contrast, magnetotelluric data show a very strong anomaly in the vicinity of the Fraser Fault (Jones et al., 1992).

**230.0 km** Local outcrops of Boston Bar slates, east of which are bluffs of volcanic sandstone of the Dewdney Creek Formation.

**231.6 km** Volcanic sandstone correlated with Dewdney Creek Formation.

**232.1 km** Cross Ainslie Creek.

**236.8-237.6 km** On west side of highway are outcrops of foliated greenstone, graphitic pelite and metachert of the Bridge River terrane. The Boston Bar phyllite is pinched out along the Fraser Fault; to the east are bluffs of Lower Cretaceous sandstone of the Jackass Mountain Group, which is the youngest part of the western Methow terrane.

***STOP 1-5:*** **39.0 km** A narrow fault slice of Ladner Group is sandwiched between the Fraser Fault to the west, here below Fraser River, and the Jackass Mountain Group to the east. It consists of tuffaceous siltstones and shales which contain ammonites (*Dubariceras* sp.; identified by H.W. Tipper, GSC Loc. C-087352) of early Pliensbachian age (ca. 192 Ma).

Directly across the Fraser River are siliceous phyllites that are part of Bridge River terrane.

**COMMENT** Note that the Ladner Group, which overlies the Spider Peak MORB basalts, ranges in age through much of the Jurassic (Fig. 8). Oldest rocks in it are possible Sinemurian strata in Manning Park just north of lat. 49°N, and youngest rocks in it are Late Jurassic (late Oxfordian or Kimmeridgian) *Buchia concentrica*-bearing greywackes in strata that cannot be separated on the basis of lithology from older parts of the Ladner Group by Ray (1990) although farther south O'Brien (1987) reported a major post-Middle Jurassic hiatus. The group consists mainly of fine grained clastic rocks of the Boston Bar Formation with intercalated calc-alkaline/sub-alkaline arc volcanics of the Aalenian-early Bajocian Dewdney Creek Formation (O'Brien, 1987). As such it is mostly coeval with the chert-rich younger part of Bridge River terrane. With the exception of the lowermost Ladner, which locally contains granitic clasts (U-Pb dated at 235 ± 10 Ma; O'Brien et al., 1992),

and the coarse clastics in the Lower Cretaceous Jackass Mountain Group, there is no evidence for margins to the basin in which the Ladner was deposited.

**243.3 km** Outcrop of well-bedded Ladner; probable northward continuation of fault sliver at Stop 1-5.

**245.1 km** Greenish-grey volcanic sandstones of the Lower Cretaceous (Barremian, to Albian; ca. 135 (maybe 145)-100 Ma) Jackass Mountain Group.

**246.4-249.0 km** Continuous outcrop of Jackass Mountain Group.

***STOP 1-6:*** **248.9 km** Pull out on west side of road if possible **(TAKE CARE! VISION OBSTRUCTED!)** to examine an outcrop of greenish interbedded conglomerate, sandstone and siltstone, typical of the Jackass Mountain Group. **CARE CROSSING ROAD!** The conglomerate is lenticular with many lenses terminating abruptly. The clasts are well rounded and include pebbles of granitic rock (some of which are foliated), volcanic rock, argillite and chert, generally supported in a matrix. Some of the finer grained units are pebbly mudstone. Sedimentological studies of the Jackass Mountain Group suggests these rocks were deposited in a submarine fan complex in a fore-arc setting (Kleinspehn, 1985). The related arc is presumably the continental Aptian(?), Albian to Cenomanian Spences Bridge Group and associated granitic rocks (e.g., Thorkelson and Rouse, 1989) which overlaps Cache Creek and Quesnel terranes (Fig. 1, 6, 8), just east of the margin of Methow terrane, although new paleomagnetic interpretations (e.g., Wynne et al., 1993) suggest this may not be the case.

**COMMENT** At this point, the Fraser Fault lies beneath the Fraser River, west of which are siliceous schists of the Bridge River terrane. About 3 km to the north the fault leaves the river as the Cantilever Fault (Fig. 1; 17; named from Cantilever Range north of Kwoiek Creek), and its eroded trace forms a series of aligned notches on the west side of the river. Across the fault, granitic rocks of latest Cretaceous-earliest Tertiary Bendor Intrusions (K-Ar dates, Hb 69 Ma, Bi 63 Ma; U-Pb date 64+11/-2; Friedman and Armstrong, in press) are juxtaposed across the fault against Jackass Mountain rocks. At this place the Methow terrane is a sliver about 2.5 km wide, and is represented only by outcrops of Jackass Mountain Group, sandwiched between latest Cretaceous-earliest Tertiary Bendor Intrusions to the west, and the Quesnellian early Mesozoic Mount Lytton Complex and the overlying Spences Bridge Group east of the Pasayten Fault (Fig. 19).

U-Pb dating of 156 ± 1 Ma of a clast from Albian Jackass Mountain conglomerate in the Manning Park area to the south (O'Brien et al., 1992) supports earlier suggestions that the provenance of the conglomerate in part is the Eagle Plutonic Complex (Greig et al., 1992), which was uplifted together with the Mount Lytton Complex in late Early Cretaceous time (Fig. 18; Monger, 1986). Together, these complexes form the southwesternmost part of Quesnel terrane at these latitudes. Early Cretaceous uplift of the Mount Lytton and Eagle complexes was concomitant with extrusion of the overlying mid-Cretaceous Spences Bridge Group, along the southwesternmost margin of Quesnellia (Fig. 19; Monger, 1986; Thorkelson and Rouse, 1989).

**253.7 km** On the east, the upper part of the valley wall is formed by grey-weathering granitic rock of the Mount Lytton Complex.

**257.0 km** On right, near Fraser River, the two railway bridges are founded on Jackass Mountain strata.

**258.0-262.3 km** Outcrops of Jackass Mountain strata along the highway.

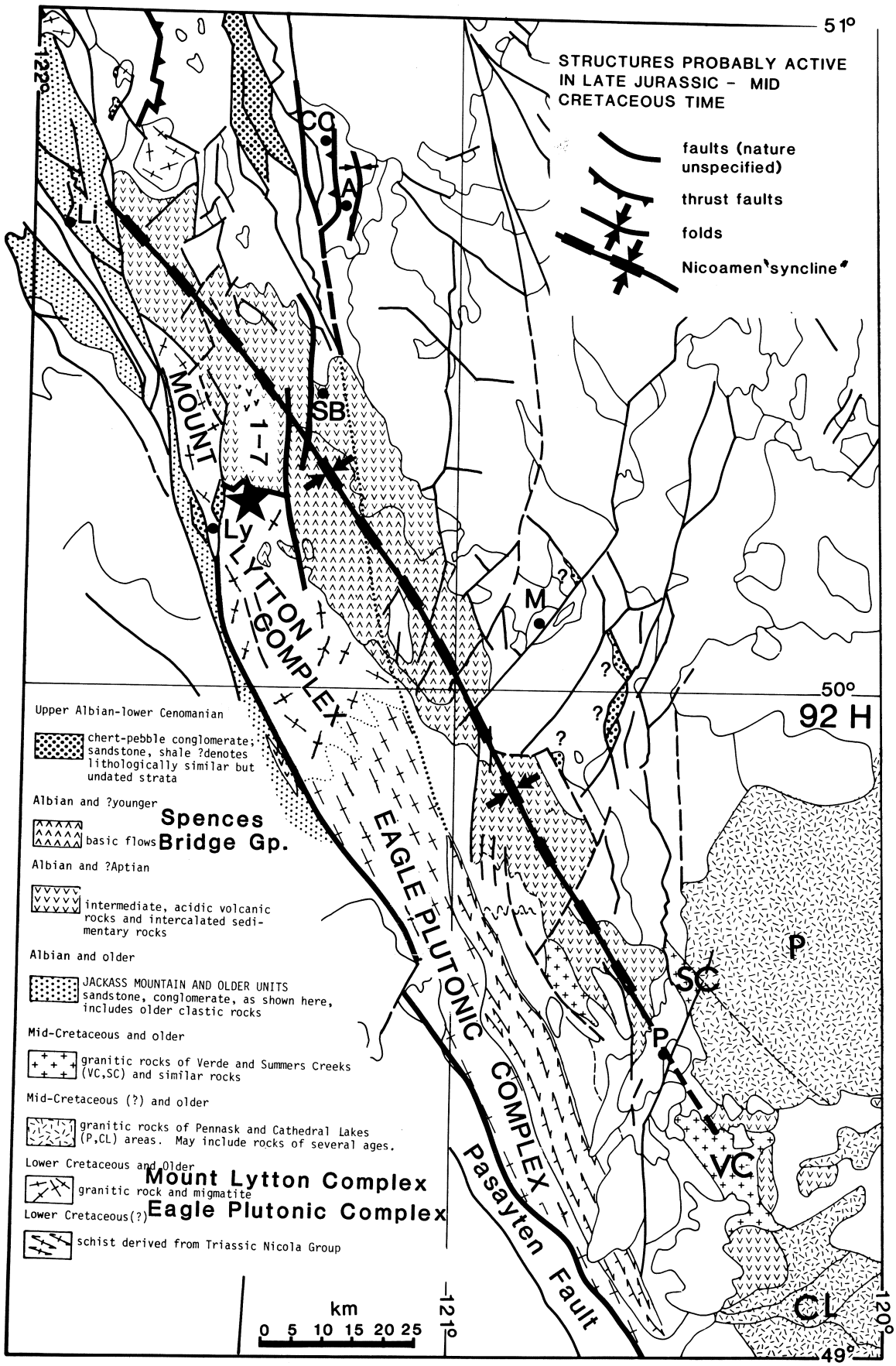
**266.3 km** Continue on Trans-Canada Highway past the southern turn-off to the town of Lytton, which is located at the confluence of Fraser and Thompson rivers. North of the town, there is a brief view of the Fraser River. At the eastern reaction ferry terminal, dark outcrops of Jackass Mountain Group are visible, and about 70 m above these is the lowest outcrop of Mount Lytton Complex. The two are separated by the northernmost extension of the Pasayten Fault, where it merges with the Fraser fault system. Southward from here is continues for over 250 km.

**COMMENT** The Pasayten Fault is the boundary between the Jackass Mountain Group (Methow terrane) and the Mount Lytton and Eagle complexes (Fig. 19). Here, it is more or less aligned with the Fraser Fault, but farther south trends south-southeasterly away from it. The fault has a complex history; to the south on the west side of the Eagle Plutonic Complex, two-mica peraluminous granitic rocks, U-Pb dated at 110 Ma, have fabrics that give a consistent sense of sinistral shear, with an east-side-up component (Greig in Monger et al., 1990; Hurlow, 1993). Rb-Sr and K-Ar dates from intensely deformed rocks along the fault give ages of 97 and 105 Ma. Interpretations of Lithoprobe seismic reflection data show the Pasayten Fault as east-dipping (Cook et al., 1992). There may be Eocene or later vertical movement along it as well, as locally thin slivers of Eocene strata are present (Greig, 1992).

**269.9 km** View to north of the Botanie Creek area which is underlain by Mount Lytton Complex. One phase of this complex, massive diorite cut by pegmatite, is exposed along the highway. Botanie Creek itself marks the trace of a fault within the Mount Lytton Complex, along which are downdropped slivers of middle Albian or younger, red-stained continental chert-feldspar-mica sandstone (elsewhere Cenomanian ages have been given to rocks of similar lithology).

***STOP 1-7:*** **271.8 km** Cross Trans-Canada Highway to former, now closed, part of highway on its north side, to view Mount Lytton Complex (Fig. 19). Visible from the viewpoint are white weathering quartzofeldspathic rocks interlayered with thin dark layers of amphibolite. The latter in places outlines sharp hinged, tightly appressed folds. At this place no fabric is present in the felsic component; the rocks are completely annealed, although elsewhere abundant mylonites

**Facing Page:**
**Figure 19.** Southwestern margin of the Intermontane Belt, showing location of Stop 1-7. The dotted line speculatively connects the eastern boundary of the Eagle Plutonic Complex with the east vergent thrust of Travers (1978) bringing Cache Creek over Quesnellian rocks, near Ashcroft. Settlements: A, Ashcroft; CC, Cache Creek; Li, Lillooet; Ly, Lytton; M, Merritt; P, Princeton; SB, Spences Bridge. Pluton names: CL, Cathedral Lakes; P, "Pennask;" SC, Summers Creek; VC, Verde Creek; (from Monger, 1985).


51°
122°
STRUCTURES PROBABLY ACTIVE IN LATE JURASSIC - MID CRETACEOUS TIME
faults (nature unspecified)
thrust faults
folds
Nicoamen syncline
CC
A
LI
MOUNT
1-7
SB
LYTTON COMPLEX
Ly
M
?
50°
92 H
EAGLE PLUTONIC COMPLEX
Pasayten Fault
SC
P
VC
CL
120°
121°
49°
km
0 5 10 15 20 25
Upper Albian-lower Cenomanian
chert-pebble conglomerate, sandstone, shale ?denotes lithologically similar but undated strata
Albian and ?younger
basic flows
Spences Bridge Gp.
Albian and ?Aptian
intermediate, acidic volcanic rocks and intercalated sedimentary rocks
Albian and older
JACKASS MOUNTAIN AND OLDER UNITS
sandstone, conglomerate, as shown here, includes older clastic rocks
Mid-Cretaceous and older
granitic rocks of Verde and Summers Creeks (VC,SC) and similar rocks
Mid-Cretaceous (?) and older
granitic rocks of Pennask and Cathedral Lakes (P,CL) areas. May include rocks of several ages.
Lower Cretaceous and Older
granitic rock and migmatite
Mount Lytton Complex
Eagle Plutonic Complex
Lower Cretaceous(?)
schist derived from Triassic Nicola Group

are within the complex. Quartzofeldspathic gneiss from the base of the cliff on the north side of the Thompson River below this locality gives a U-Pb date of 225 ± 5 Ma (Parrish and Monger, 1992). The cliff is crosscut by irregular brown dykes that may be feeders to the stratigraphically overlying Albian-Cenomanian Spences Bridge Group. Bedded, reddish-weathering chert pebble conglomerate (correlative(?)) with that in Botanie Creek (there palynologically dated as middle Albian or younger) is visible on the lower shoulder to the north-northwest across river and overlies red-stained Mount Lytton Complex, downdropped on a north-trending fault parallel with the Botanie Creek Fault.

**COMMENT** The Mount Lytton Complex (Fig. 19) probably represents the underpinnings of the early Mesozoic Quesnellian arc which comprises such units as the Nicola Group and Guichon Creek Batholith (Fig. 1, 6, 8, 19). In addition to the rocks seen here, are extensive massive diorites and pegmatites, local marble and, importantly for the correlation made above, crosscutting granodiorite. The latter, 17 km southwest of here, yields a U-Pb age of 212 ± 1 Ma (Parrish and Monger, 1992) which is comparable with an unpublished age of 210 Ma on the Guichon Creek Batholith obtained by N. Mortimer/R.L. Armstrong. The two plutons are separated only by the 25 km wide tract of continental, mid-Cretaceous volcanics of the Spences Bridge Group which overlaps both plutons. The Guichon Creek Batholith was emplaced at a high crustal level within western parts of a Late Triassic-Early Jurassic (ca. 230-190 Ma) volcanic arc composed of Nicola and Rossland groups and co-magmatic granites. In the Guichon Creek body, U-Pb, Rb-Sr and K-Ar isotopic systems give similar dates, whereas in the Mount Lytton granodiorite, a K-Ar date of ca. 186 Ma suggests an early Middle Jurassic cooling age, commensurate with a deeper level of emplacement, and slow cooling.

The Mount Lytton Complex together with the Late Jurassic-Early Cretaceous Eagle Plutonic Complex (Greig et al., 1992) along strike to the south, form the uplifted southwestern margin of Quesnel terrane (Fig. 19). They were uplifted and eroded in late Early Cretaceous time, concomitantly with extrusion of overlying Spences Bridge volcanics and deposition of Jackass Mountain/Pasayten strata (Monger, 1985). There is a red-stained, deeply weathered surface locally below the Spences Bridge volcanics and generally below overlying(?) chert pebble conglomerate and sandstone, some of which can be seen across the river from Stop 1-7. Red staining and deep weathering is seen locally below mid-Cretaceous rocks over this part of the southwestern Intermontane Belt.

A Late Jurassic deformation and plutonic history is recorded in the Eagle Plutonic Complex. Late Jurassic (ca. 160-155 Ma) eastern parts of this complex have steeply west-dipping fabrics in gneissic granodiorite (U-Pb dated at 157 Ma) and meta-Nicola volcanics (Greig et al., 1992). This fabric is more-or-less on strike with the eastern margin of the Cache Creek terrane (exposed 30 km north-northeast of here), although continuity is concealed by the Spences Bridge Group (Fig. 6). The eastern Cache Creek margin was shown by Travers (1978) to be fundamentally an east-vergent fault system on which Cache Creek was overthrust on to Quesnellia possibly in Early and Middle Jurassic time. This age was based on dewatering structures (slaty cleavage, sandstone dykes) in clastic rocks of the Jurassic Ashcroft Formation, although the only unequivocal constraint on time of thrusting is provided by the overlapping late Early Cretaceous Spences Bridge Group. Monger (1985) suggested that, from their alignment, Travers' overthrust and the Late Jurassic Eagle Complex structures record the same event but at different structural levels (Fig. 19). No record of Jurassic, Eagle Complex, deformation has been recognized in the Methow terrane, juxtaposed against it across the Pasayten Fault.

From Stop 7, return back to Highway 5 and head back west towards Lytton.

**275.2 km** Take northern turnoff to Lytton. Drive into and through town, and take Highway 12 north to Lillooet.

**276.8 km** Cross Thompson River; bridge abutments are founded on a fractured, pink-weathering muscovite-bearing granodiorite, lithologically similar to the easternmost part of the Eagle Plutonic Complex to the south-southeast along strike, which there is U-Pb dated at 110 Ma (Greig et al., 1992).

**277.7-282.5 km** Pink muscovite-bearing granite by Highway 12. Below the northern end of this outcrop at river level near the reaction ferry terminal on the east bank, but separated from it by the Pasayten Fault, are Jackass Mountain sandstones and greywacke. Opposite, the west side of the valley is underlain by the Jackass Mountain Group up to an elevation of about 1000 m, above which are latest Cretaceous granitic rocks, separated from the Jackass Mountain by the Cantilever Fault, which here is the main strand of the Fraser-Straight Creek fault system.

**285.7 km** Outcrops of fractured and folded, steeply dipping, red and buff well-bedded sandstone, shale and conglomerate of mid- to Late Cretaceous age occur in a fault slice low in the valley of the Fraser River. The sandstones are distinctive; they feature abundant chert grains together with white mica and feldspar. They probably postdate the Spences Bridge Group, are equivalent to dated (middle Albian or younger) clastics in Botanie Creek to the east, and in the area west of Cache Creek, where they overlie the Cache Creek Complex.

**287.7-288.3 km** Fractured Mount Lytton granodiorite.

**294.7 km** Mount Lytton granodiorite.

**295.8 km** Izman Creek (western end of Lithoprobe seismic Line 12). At this place highly altered green Mount Lytton granodiorites, and chlorite schist containing blue quartz "eyes," feature well developed mylonitic fabrics. The latter trend northwesterly, and are truncated by the more northerly trending Fraser fault system.

**296.7 km** To west, on the west side of the valley, folded, buff weathering sandstone and conglomerate and minor reddish volcanics form a fault slice between Jackass Mountain conglomerate low the valley to the east, and granitic rocks forming the peaks to the west. Plant fossils collected from this unit are probably of Middle Eocene age (Bell in Duffell and McTaggart, 1952).

**300.3 km** Altered, epidotized Mount Lytton granodiorite. Rock from this locality ("northern Mount Lytton") is U-Pb dated at 250 ± 5 Ma and used to determine the amount of displacement on the Fraser fault system, by matching these rocks with similar rocks in Farwell Canyon (U-Pb dated at 258 ± 5 Ma; Friedman and van der Heyden, 1992). Their

estimated dextral displacement is a minimum of 135 km, which is close to the figure derived by Coleman and Parrish (1991) using offsets of Eocene rocks.

**302.9 km** Outcrop of dioritic rock of the Mount Lytton Complex.

**COMMENT** From about this point northwards, faults splay off to the northwest, away from the Fraser fault system (the reverse of the situation south of Boston Bar, where structures fed from the southeast into the Fraser fault system; Fig. 1). Some of these structures are crossed in the beginning of Day 2. From here, westwards, there are four main fault slices (Fig. 1, 17; Monger, 1989a). (1) Below in the valley, across the northernmost equivalent of the Pasayten Fault, the Jackass Mountain Group pinches out, after occupying lower elevations in the Fraser Canyon for the last 40 km. (2) West of it, across a north-south trending fault (the Cantilever Fraser Fault; part of the Fraser fault system) on the west side of the valley, is the northernmost exposure of the Tertiary strata mentioned at 296.7 km. (3) Above this, occupying the lower valley walls, across a fault which may be the continuation of the Yalakom Fault, are subgreenschist grade greenstone, chert and local serpentinite of the Bridge River terrane. (4) Topographically highest and westernmost, across the Lillooet Fault and/or Marshall Creek Fault (which is marked locally by serpentinite) are Eocene granitic rocks of the Texas Creek Pluton (U-Pb date 47.3±1 Ma) with abundant inclusions of rusty schist (Cayoosh assemblage?). The Lillooet Fault appears to be a steep, southward continuation of the Eocene Mission Ridge Fault of Coleman and Parrish (1991). The highest rocks may be the offset equivalents of components of the Custer Gneiss exposed in lower parts of the Fraser Canyon.

**313.9 km** Highly altered (zeolitized and fractured) granitic and dioritic rocks of the Mount Lytton Complex, above which (in fault contact) are Spences Bridge volcanics.

**314.3 km** Cross the (covered) main strand of the Fraser Fault. At this place it leaves the Fraser Canyon, and takes a more northerly trend through Fountain-Cinquefoil valley (Fountain Fault of Fig. 17), and rejoins the Fraser River 9 km northeast of Lillooet.

**314.6 km** Poorly dated, pre-Jackass Mountain Group, Jura-Cretaceous(?), shales and siltstones, with minor greywacke and rare conglomerate of the Lillooet Group of Duffell and McTaggart (1952), are exposed in discontinuous outcrops for the next 15 km, and are cut by small, undated, white-weathering intrusions.

**COMMENT** These rocks are part of Methow Terrane in that they underlie the Jackass Mountain Group. Monger (1989a) correlated them (probably erroneously) with the Jura-Cretaceous Relay Mountain Group to the northwest, for although they may be time equivalent to the Relay Mountain, they are a deeper water, far less fossiliferous unit. Duffell and McTaggart (1952) reported that in the Lillooet Group, belemnites occur rarely and a boulder of argillite with *Buchia* of early Lower Cretaceous (Lower Neocomian age; ca. 140 Ma), is "almost certainly from the lower part of the group." The rocks are possibly correlative with the lithologically similar but more metamorphosed Brew Group, exposed in the mountains to the west (J.B. Mahoney, pers. comm., 1993). This contains earliest Cretaceous fossils, seen on Day 2, and forms the upper part of the Cayoosh assemblage of Journeay and Mahoney (1994). The lower part of the Cayoosh assemblage gradationally overlies the Bridge River terrane. Thus, this unit may be a Jurassic(?)-Early Cretaceous overlap unit linking Methow and Bridge River terranes. Farther south, these rocks may correlate with the youngest (Oxfordian-Kimmeridgian) parts of the Ladner Group of Ray (1990), and the Thunder Lake sequence of O'Brien (1987). Seven kilometres north of here on Fountain Ridge (between Fraser River and Fountain valley) the Lillooet Group is overlain stratigraphically by Jackass Mountain Group. Seven kilometres north of Lillooet, at the confluence of the Bridge and Fraser rivers, the Lillooet Group appears to overlie coarse, turbiditic volcanic sandstones containing probable Middle Jurassic ammonites, that are reasonably correlated with the volcanic-rich Dewdney Creek Formation of the Ladner Group east and southeast of Boston Bar.

**315.4 km** Junction of Highway 12 with road to the north along Cinquefoil and Fountain valley. Continue along Highway 12.

***STOP 1-8:*** **320.7 km** Area of the "Great Slide." Walk along road for 0.5 km to examine Jura-Cretaceous thinly bedded siltstones and argillites of the Lillooet Group of Duffell and McTaggart (1952). Remember these lithologies on Day 2!

Across the Fraser River from Stop 1-8, Lillooet Group strata occur low down on the west bank, and are separated by the southerly extension of the Yalakom Fault from a small knoll of foliated, subgreenschist grade greenstone of Bridge River terrane (Fig. 1). Above this, the steep valley wall is the exhumed surface of the Lillooet Fault (= ? Mission Ridge Fault). West of this, the steep wall and crest consist of Eocene granitic rocks, greenschist grade siliceous phyllites, and local biotite schist, and are physically continuous with the uplifted footwall rocks of the Eocene Mission Ridge normal fault (Stop 2-1; Coleman and Parrish, 1991). West of this again, across the Marshall Creek Fault, the 3000 m high summit of Mount Brew consists of greenschist grade pelite and sandstone with a locally prominent conglomerate horizon, of the Brew Group with local Early Cretaceous (Neocomian) fossils which is the upper part of Cayoosh assemblage.

**321.1-330.2 km** Local outcrops of Lillooet Group occur along the road, within which are small undated granodioritic and dioritic plutons.

**332.9 km** Above road, to east are massive sandstones, local shales and, uppermost, conglomerate-rich units of the late Early Cretaceous Jackass Mountain Group.

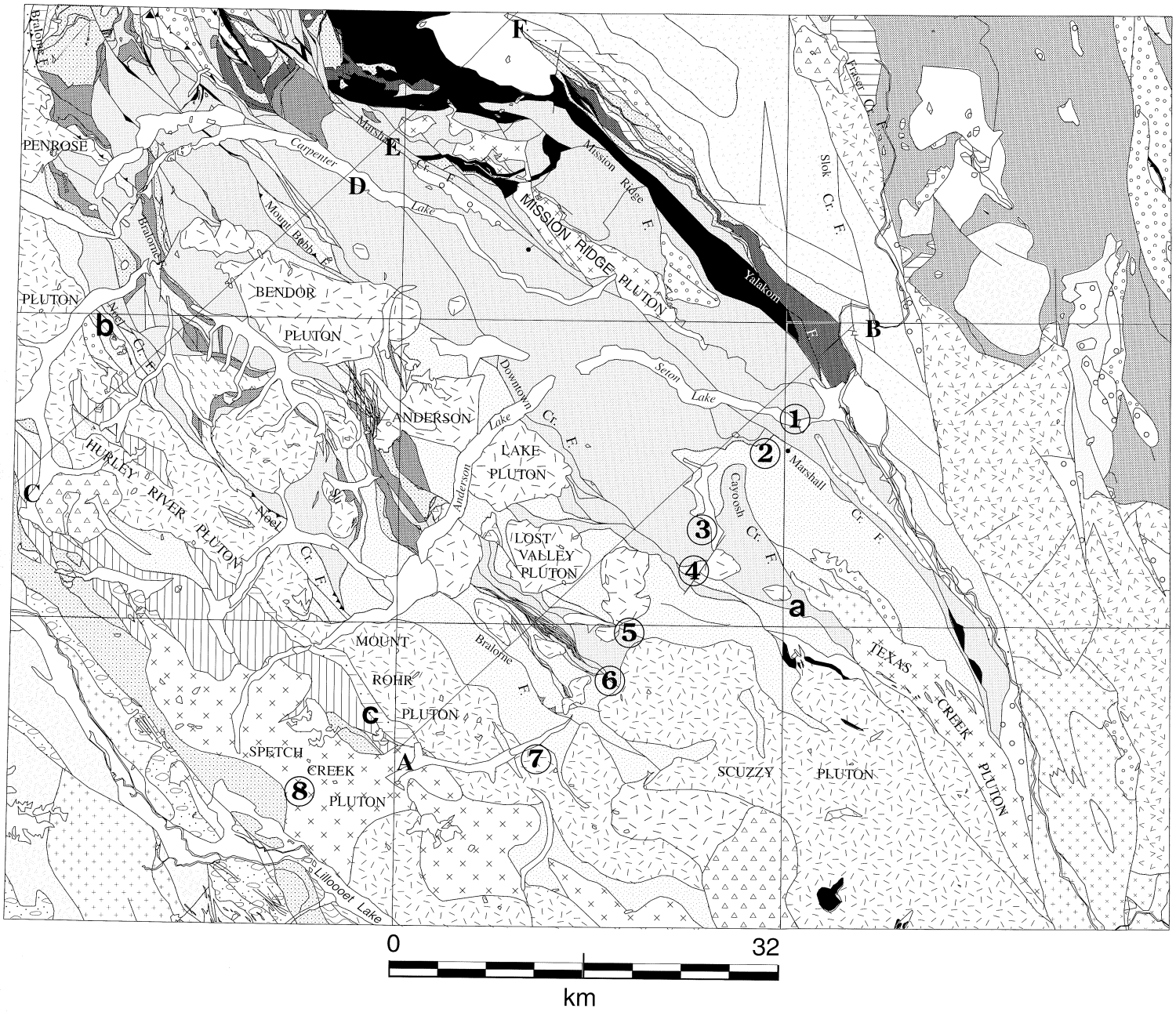
North of here the road continues along a gravel/sand bench; below it, in the banks of the river, Lillooet Group strata are exposed. Above the east wall of the valley (Fountain Ridge) consists of massively bedded, zeolitized strata belonging to the upper part of the Jackass Mountain Group.

**338.1 km** Turnoff to Lillooet.

**338.7-339.3 km** Twenty-three Camels Bridge over Fraser River.

**340.2 km** The Mile 0 Cairn in Lillooet marks the start of the road to the Cariboo gold fields active in the mid-1800s.

End of Day 1.




## Legend

- MPv: Miocene/Pliocene volcanics
- Es: Eocene sedimentary rocks
- lKPC: Powell Creek Formation
- KSq: Silverquick Conglomerate
- KSB: Spences Bridge Group
- KG: Gambier Group
- KT: Taylor Creek Group
- eKJ: Jackass Mountain Group
- JKRM: Relay Mountain Group
- JH: Harrison Lake Formation
- JLC: Last Creek Formation
- JKC: Cayoosh Assemblage
- TT: Tyaughton Group
- TCD: Cadwallader Group
- CJB: Bridge River Complex
- PEL: Bralorne-East Liza Complex
- Sh: Shulaps Complex
- MCC: Chism Creek Schist
- EGd: Eocene plutons
- lKB: Late Cretaceous Bendor Suite
- lKS: Late Cretaceous Scuzzy Suite
- mKgd: mid-Cretaceous plutons
- eKgd: Early Cretaceous plutons
- mlJgd: Mid/Late Jurassic plutons
- MSL: Slollicum Schist
- PT: Permian-Triassic plutons
- PTCC: Cache Creek Complex
- TN: Nicola Group

# FIELD TRIP ROAD LOG AND COMMENTS

# DAY 2: LILLOOET TO VANCOUVER: SOUTHERN COAST MOUNTAINS TRANSECT

J. Murray Journeay and James W.H. Monger

## OVERVIEW

The southeastern Coast Mountains span the boundary between the Intermontane and Insular superterranes of the southern Canadian Cordillera (Fig. 1, 20). They consist of terranes and overlap assemblages of late Paleozoic and Mesozoic age, disposed in complex contractional and strike-slip fault slices. Stitching these rock packages together are Late Cretaceous and early Tertiary syn- and post-tectonic intrusions of the Scuzzy-Mount Rohr and Bendor plutonic suites. Tectonic slivers of the southeastern Coast Mountains are thrust westward over supracrustal arc sequences of the southwestern Coast Mountains along the Coast Belt Thrust System.

The focus for the first half of Day 2 is on the contact relationships between volcanic arc and oceanic assemblages (Cadwallader Group, Bridge River and Bralorne-East Liza complexes) and overlying turbidites and volcaniclastic successions of the Tyaughton and Brew groups and Cayoosh assemblage (Table 1; Fig. 1, 6, 8, 21), and on the crustal structure and history of contractional and strike-slip faulting in the southeastern Coast Mountains. This information bears directly on the history of terrane accretion and on the mechanisms of strain partitioning and crustal deformation in the Coast Belt orogen as a whole (see Introduction). In the last half of Day 2, we cross the boundary between southeastern and southwestern Coast Mountains, and focus primarily on the history of plutonism and deformation in supracrustal rocks that were linked to Insular Superterrane by the Middle Jurassic and that form the foreland to the Coast Belt Thrust System (Fig. 1, 6, 8).

## TECTONIC ASSEMBLAGES

The southeastern Coast Mountains encompass a wide variety of fault-bounded sequences ranging in age from late Paleozoic to early Tertiary (Woodsworth, 1977; Rusmore, 1985; Schiarizza et al., 1989, 1990b). These include: late Paleozoic and younger ophiolitic rocks of the Bralorne-East Liza and Shulaps complexes; Triassic arc volcanics and related sedimentary rocks (Triassic-Lower Jurassic) of the Cadwallader terrane; Mississippian to upper Middle Jurassic (Callovian) oceanic rocks of the Bridge River terrane; Jura-Cretaceous turbidites and clastic successions of the Tyaughton Group; and strata probably correlative in part with those of Methow terrane (Fig. 1, 8, 21). These rocks are structurally interleaved along north and northwest-trending systems of west- and east-vergent thrust faults, and have been shuffled along strike by major dextral and sinistral transcurrent faults.

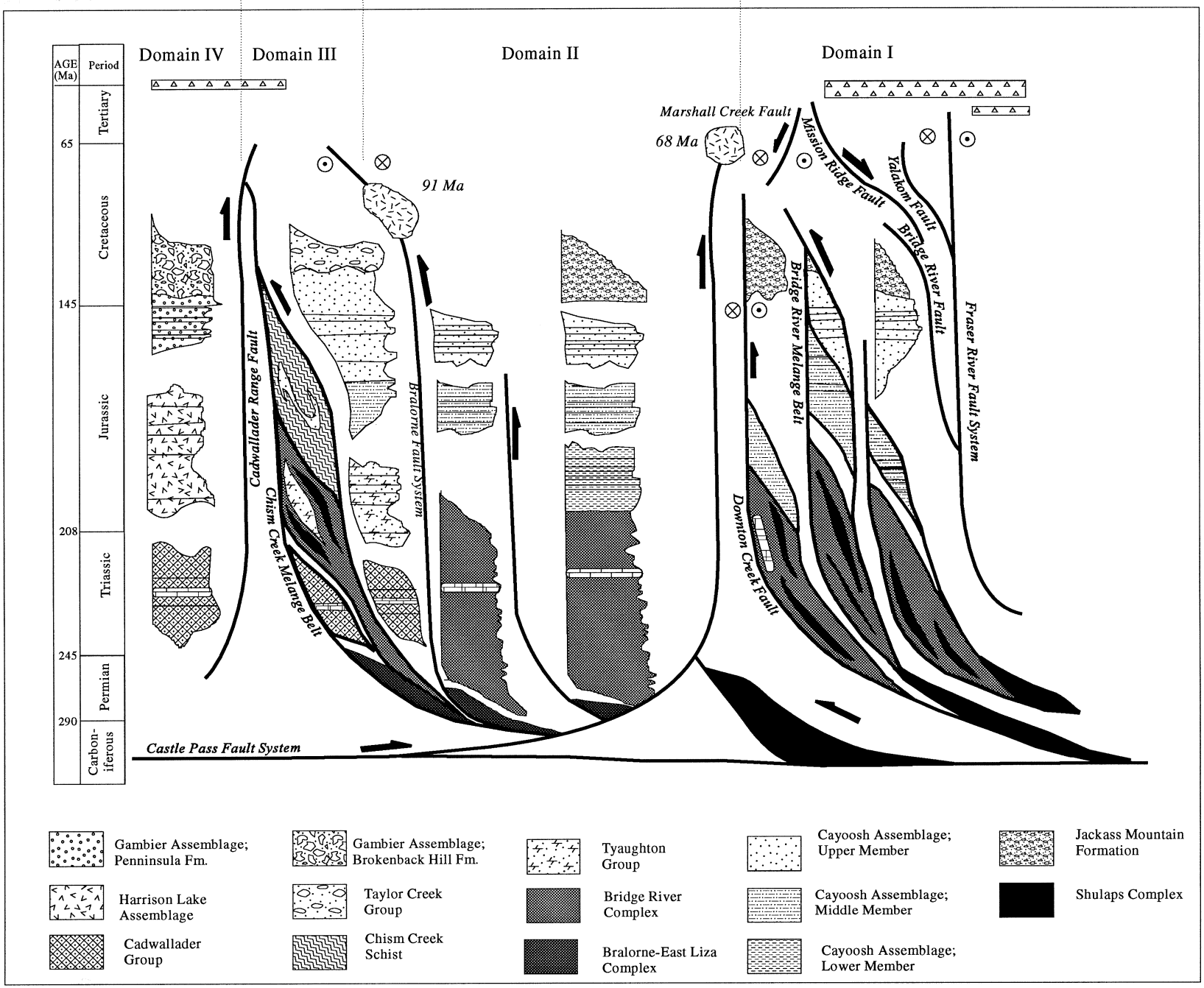
In spite of the structural and stratigraphic complexity, it is possible to subdivide rocks of the southeastern Coast Mountains into the following four structural and stratigraphic domains (schematically summarized in Fig. 21).

**Domain 1.** The eastern melange belt, exposed in the eastern Chilcotin, Bendor and Cayoosh ranges, comprises serpentinite, disrupted ophiolite and oceanic rocks of the Shulaps and Bridge River complexes. It is structurally interleaved with arc-related rocks of the Cadwallader Group and younger Jura-Cretaceous clastic rocks of the Methow terrane (Cayoosh assemblage, Brew Group, Taylor Creek Group) possibly correlated with strata in Methow terrane. It is bounded on the east by dextral transcurrent faults of the Yalakom and Fraser River systems, and on the west by west-directed thrust and dextral strike-slip splays of the Downton Creek Fault (Fig. 20).

**Domain 2.** West of this melange belt, in the western Cayoosh and Bendor ranges, is a relatively coherent domain comprising oceanic rocks of the Bridge River Complex and overlying clastic successions of the Cayoosh assemblage. It is flanked on the east by the Downton Creek Fault and to the west by west-directed thrusts of the Bralorne fault system. New mapping in this domain (Journeay and Northcote, 1992; Journeay et al., 1992) documents a stratigraphic transition between greenstone and chert of the Bridge River terrane and overlying clastic basin facies of the Cayoosh assemblage. The latter contain an important quartz-rich clastic facies that likely records emergence and erosion of volcanic arc and crystalline basement terranes, and deposition of clastic detritus derived from them in the basin underlain by Bridge River terrane in Late Jurassic and Early Cretaceous time.

**Facing Page:**
**Figure 20.** Geologic map of the eastern Coast Belt. A compilation of recent mapping by Schiarizza et al. (1989, 1990a); Coleman (1990); Journeay (1990a); Journeay and Northcote (1992); Journeay et al. (1992), and previous work by Roddick and Hutchison (1973); Woodsworth (1977), and Monger (1989a).

**Following Page:**
**Figure 21.** Structural and stratigraphic framework of the eastern Coast Belt. Regional correlations and terrane descriptions of these rock units are summarized in Figure 6, 8 and Table 1, respectively.


AGE (Ma)
Period
65
145
208
245
290
Tertiary
Cretaceous
Jurassic
Triassic
Permian
Carbon-iferous
Domain IV
Domain III
Domain II
Domain I
91 Ma
68 Ma
Marshall Creek Fault
Mission Ridge Fault
Yalakom Fault
Bridge River Fault
Fraser River Fault System
Bridge River Melange Belt
Dominion Creek Fault
Cadwallader Range Fault
Chism Creek Melange Belt
Bralorne Fault System
Castle Pass Fault System
Gambier Assemblage; Penninsula Fm.
Harrison Lake Assemblage
Cadwallader Group
Gambier Assemblage; Brokenback Hill Fm.
Taylor Creek Group
Chism Creek Schist
Tyaughton Group
Bridge River Complex
Bralorne-East Liza Complex
Cayoosh Assemblage; Upper Member
Cayoosh Assemblage; Middle Member
Cayoosh Assemblage; Lower Member
Jackass Mountain Formation
Shulaps Complex

**Domain 3.** West of, and in the lower plate of the Bralorne fault system, is a second melange belt comprising metamorphosed rocks of the Chism Creek Schist. This belt marks the western edge of the Bridge River terrane. It includes rocks derived in part from the Bralorne-East Liza ophiolite complex, Bridge River and Cadwallader terranes, and overlying Jura-Cretaceous rocks including the Cayoosh assemblage. This belt has been traced southwestward and is correlated with high grade thrust nappes of the Cogburn melange and Settler Schist, east of Harrison Lake.

**Domain 4.** In the footwall of this melange belt is a relatively coherent domain comprising arc volcanics and related sedimentary rocks of the Cadwallader terrane and overlying Jura-Cretaceous volcanic and clastic rocks correlated with Harrison Lake, Relay Mountain/Cayoosh assemblage and Taylor Creek groups. It is bounded to the east by the Bralorne fault system and to the west by imbricate west-directed faults of the Coast Belt Thrust System.

## STRUCTURAL FRAMEWORK

The structural framework of the soiutheastern Coast Belt is defined by a network of northwest-trending contractional and strike-slip fault systems that record a complex history of post-mid-Cretaceous shortening and transpression (Fig. 22). The results of new mapping (Journeay et al., 1992) combined with mapping by Schiarizza et al. (1989, 1990b) allow a partial reconstruction of this deformation history.

**Southwestward-directed thrusting (97-91 Ma)** Folding and associated thin-skinned thrusting of terranes in the southeastern Coast Belt was followed by thick-skinned thrusting and **southwestward** telescoping of these terranes over supracrustal rocks of the southwestern Coast Belt. Structures that were active during this stage of deformation include the Coast Belt Thrust System and out-of-sequence west-directed thrusts of the Bralorne fault system. Faults of the Coast Belt Thrust System root eastward beneath high-grade metamor-

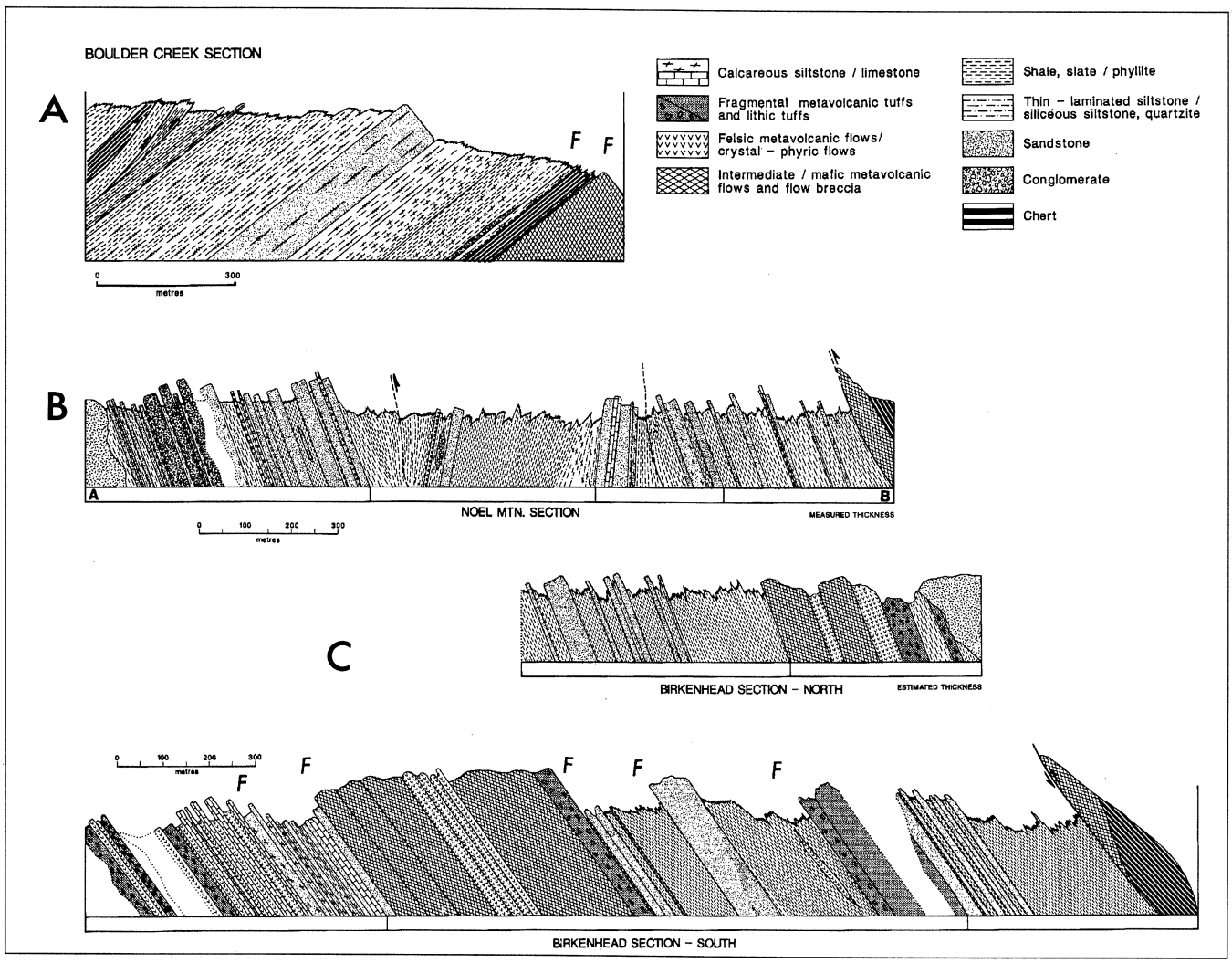



**Figure 22.** Measured structural-stratigraphic sections of clastic successions overlying Bridge River and Cadwallader terranes in the southeastern Coast Belt. Sections are keyed to Figure 20, and include the Bridge River/Cayoosh transition along Boulder Creek (Fig. 22A), middle and upper successions of the Cayoosh assemblage on Noel Mountain (Fig. 22B), and Cadwallader Group/Harrison Lake Formation/Cayoosh assemblage relationships in the Birkenhead Lake region (Fig. 22C).

phic rocks of the Cogburn melange and Settler-Chism Creek Schist. Faults of the Bralorne system root eastward beneath oceanic rocks and overlying basin successions of the Bridge River terrane. Left-lateral oblique slip faults of the Tyaughton Creek System are interpreted to have formed late in this episode of crustal shortening and may have accommodated an important component of orogen-parallel displacement.

**Northeastward-directed thrusting (91-86? Ma)** Northeast-vergent folding and associated thrusting. Structures that are known to have been active during this period of deformation include **northeast-directed** faults and folds of the Castle Pass System (Schiarizza et al., 1990b) and related thrust faults in the western Bendor Range, northern Lillooet and Cadwallader ranges (Journeay et al., 1992). These structures cut older southwest-vergent thrust faults and apparently root westward beneath Wrangellia. They appear to have been active during gold mineralization in the Bralorne district (91-86 Ma; Leitch, 1989). However, we cannot rule out the possibility that northeast-directed faults may have been active at deeper crustal levels over a longer period of time.

**Transpression (86-68 Ma)** Oblique, **southwest-vergent** thrusting and associated(?) dextral strike-slip faulting. The Downton Creek Fault mark the western boundary of the eastern melange belt and are most likely linked to southwest-vergent thrusting in the imbricate zone at the base of the Shulaps Complex. Kinematic linkages between these structures and dextral strike-slip fault systems in the region, including the Marshall Creek and Yalakom faults are plausible but not proven.

**Dextral Transcurrent and Extensional Faulting (68-47 Ma)** Dextral strike-slip (Marshall Creek and Yalakom faults) and extensional normal faulting (Mission Ridge and Marshall Creek faults) along the east flank of the Coast Mountains, resulting in exhumation of an early Tertiary metamorphic complex. This metamorphic complex is interpreted to be the lower grade, offset counterpart of the early Tertiary, eastern, part of the Cascade metamorphic core in northern Washington. The Mission Ridge and Cayoosh Creek faults and steep southwest-dipping brittle shears along the Marshall Creek Fault are outward-dipping extensional faults that flank this metamorphic complex. They cut older dextral strike-slip fabrics of the Marshall Creek and Yalakom faults, and record a two-stage history of faulting; down-to-the-northeast displacement along the Mission Ridge Fault, followed by down-to-the-southwest displacement along the Marshall Creek Fault (Coleman, 1990).

**Dextral Transcurrent Faulting (47-35 Ma)** Dextral transcurrent displacement (90-140 km) along the Fraser River fault system.

The route of the field trip takes us westward across the strike of these fault systems, which in general become older to the west. We begin by examining the relationship between dextral transcurrent and extensional faulting in the early Tertiary, and then focus on the history of Late Cretaceous shortening and related(?) strike-slip faulting.

## ROAD LOG

**0.0 km** Victoria Hotel in downtown Lillooet.

**0.7 km** Junction at the south end of town. Turn left, cross railroad tracks and follow signs for Highway 99. View to the southeast of the Fraser Canyon and the Yalakom Fault.

At this latitude, the Yalakom Fault separates disrupted chert, argillite, basalt, gabbro and serpentinite of the Bridge River and Shulaps complexes along the west side of the canyon from Jurassic and Cretaceous sedimentary rocks of the Methow terrane (see Stop 1-8) along the east side of the canyon. The fault extends southeastward along the Fraser Canyon, where it is cut by and merges with the north-trending Fraser fault system (Fig. 1, 23).

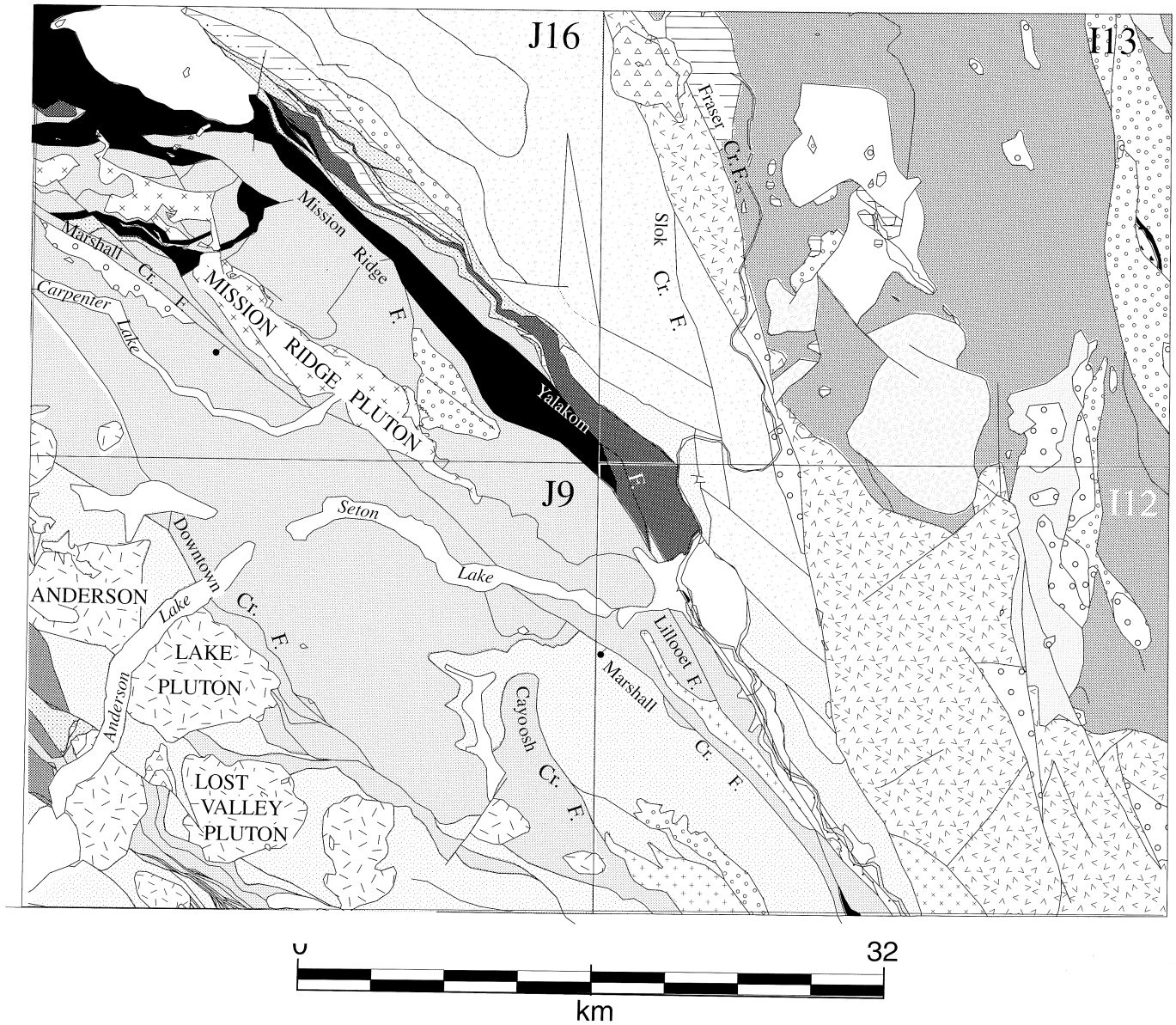
To the northwest, the Yalakom Fault bifurcates into two strands (Schiarizza et al., 1990a,b). The northeast strand, known as the Camelsfoot Fault, separates a steep northeast-facing panel of weakly deformed Middle Jurassic sandstone and shale of the Cretaceous Jackass Mountain Group rocks to the northeast from a structurally complex, northeast-dipping imbricate zone comprising fault slivers of Bridge River Complex and Cadwallader Group to the southwest. Folds in this imbricate zone are overturned to the southwest with axial surfaces that dip moderately (40°-60°) to the northeast (Schiarizza et al., 1990a; Coleman and Parrish, 1991). The southwest strand of the Yalakom Fault, locally called the Bridge River Fault (Schiarizza et al., 1990a), separates this panel of imbricated Bridge River and Cadwallader Group rocks from disrupted oceanic rocks of the Shulaps and Bridge River complexes to the southwest. The fault zone is marked by slivers of serpentinite melange, and locally contains knockers of peridotite, gabbro and diabase believed to be part of the Shulaps ophiolite complex. These two fault strands converge to the northwest, between Blue and Beaverdam creeks.

**COMMENT** The Yalakom Fault (Fig. 23) is one of several major transcurrent faults in the Eastern Coast Belt. It marks the east flank of the Bridge River terrane near Lillooet and extends northwestward more than 250 km into the Taseko Lakes and Mount Waddington map areas (Leech, 1953; Tipper, 1969). A history of post-middle Cretaceous dextral transcurrent motion is documented along the trace of the fault (Glover et al., 1988; Schiarizza et al., 1989) based on a variety of kinematic indicators. Dextral offset of more than 100 km is postulated along the Yalakom Fault based on piercing point correlations. Tipper (1969) proposed 80-90 km of offset based on the correlation of Middle Jurassic volcanic rocks in the Mount Waddington, Anahim Lake and Bella Coola map areas. Kleinspehn (1985) postulated 125-175 km offset by restoring submarine fan facies of the Lower Cretaceous Jackass Mountain Group between Chilko Lake and the Camelsfoot Range. Umhoefer (1989) estimated 100 km offset along the Yalakom Fault by correlating the Shulaps ophiolite complex and the Petch Creek serpentinite belt, a correlation that requires restoration of younger dextral transcurrent

**Facing Page:**
**Figure 23.** Geology of the southern Shulaps Range, based on work of Schiarizza et al. (1990a,b), Coleman (1990), Journeay and Northcote (1992), and previous mapping of Roddick and Hutchison (1973), Woodsworth (1977), and Monger (1989a).

## Legend

- MPv: Miocene/Pliocene volcanics
- Es: Eocene sedimentary rocks
- lKPC: Powell Creek Formation
- KSq: Silverquick Conglomerate
- KSB: Spences Bridge Group
- KG: Gambier Group
- KT: Taylor Creek Group
- eKJ: Jackass Mountain Group
- JKRM: Relay Mountain Group
- JH: Harrison Lake Formation
- JLC: Last Creek Formation
- JKC: Cayoosh Assemblage
- TT: Tyaughton Group
- TCD: Cadwallader Group
- CJB: Bridge River Complex
- PEL: Bralorne-East Liza Complex
- Sh: Shulaps Complex
- MCC: Chism Creek Schist
- EGd: Eocene plutons
- lKB: Late Cretaceous Bendor Suite
- lKS: Late Cretaceous Scuzzy Suite
- mKgd: mid-Cretaceous plutons
- eKgd: Early Cretaceous plutons
- mlJgd: Mid/Late Jurassic plutons
- MSL: Slollicum Schist
- PT: Permian-Triassic plutons
- PTCC: Cache Creek Complex
- TN: Nicola Group



motion (85-100 km) along the Fraser fault system. Conservative estimates of Fraser Fault displacement (Fig. 18B) suggest that the Hozameen Fault of the eastern Cascades is the offset counterpart of the Yalakom Fault east of the Fraser Fault (see Fig. 18; Day 1). The new, larger estimate of displacement (Fig. 18C) suggests that the Yalakom Fault in combination with the Mission Ridge Fault is more likely correlative with the Ross Lake Fault.

**1.5 km** Junction with Highway 12. Continue south along Highway 99, across the trace of the Yalakom Fault.

**1.7 km** Bridge across Seton River. Low knolls to the south are underlain by fault slivers of serpentinized peridotite (Shulaps Ophiolite Complex?) and are part of the Yalakom Fault Zone.

**2.0 km** Junction with Texas Creek road. Continue southwest on Highway 99. Imbricate slivers of serpentinized peridotite outcrop in low hills to the south.

**2.9 km** The Lillooet Fault (Monger, 1989a) is a steep northwest-trending brittle structure that separates imbricate low-grade slivers of Bridge River and Shulaps Ophiolite complexes to the northeast, here, low on the west side of Fraser Valley, from metamorphosed schists of the Bridge River Complex and Cayoosh assemblage to the southwest (Fig. 22, 23, 24). Monger (1989a) interpreted the Lillooet Fault as a strike-slip splay of the Yalakom fault system. Coleman (1990) considered this structure to be the southern equivalent of the northeast-dipping Mission Ridge normal fault. The trace of the fault is marked by prominent topographic notches along the ridge to the north and by steep dip slope surfaces along the west wall of the valley to the south. It crosses the road near this locality. Cliff faces to the northwest are underlain by metamorphosed schists of the Bridge River Complex in the footwall of the Mission Ridge Fault.

**COMMENT** The Mission Ridge Fault (Fig. 22, 23, 24) is a northeastdipping Eocene normal fault that separates low-grade (prehnite-pumpellyite facies) rocks of the Bridge River Complex and unconformably overlying Eocene sedimentary rocks in the upper plate from well-foliated greenschist grade schists of the Bridge River Complex in the lower plate. Barometric discontinuities across the fault indicate normal downdip displacement of 10-18 km (Coleman, 1990). The fault cuts deformed rocks of the Mission Ridge Pluton (47.5 ± 2 Ma; Coleman and Parrish, 1991). It is truncated by the Marshall Creek Fault to west and by dextral strike-slip faults of the Fraser River fault system along strike to the southeast. The Ross Lake Fault, exposed along the east side of the Fraser Fault near the town of Hope (Day 1; Fig. 18), is interpreted to be the offset counterpart of the Mission Ridge Fault (Coleman, 1990). This interpretation implies a minimum right-lateral offset of 90-140 km and places important constraints on the timing of displacement along the Fraser Fault.

**4.7 km** Bridge across Cayoosh Creek.

**4.9 km** B.C. Hydro canal outflow.

***STOP 2-1:*** **6.4 km** View of Seton Lake and surrounding ranges of the Eastern Coast Belt. Polydeformed Late Cretaceous-early Tertiary metamorphic rocks of the Bridge River Complex are visible in cliff faces along the shore of Seton Lake to the northwest and along the Seton River outflow to the northeast. These rocks represent the exhumed lower plate of the Mission Ridge and Marshall Creek faults, and are considered to be the offset, lower grade counterparts to the Custer Gneiss and associated Eocene mid-crustal rocks of the eastern Cascade metamorphic core (Coleman and Parrish, 1991). High peaks visible along the skyline to the northwest (Cayoosh Range) and to the southwest (Mount Brew) are underlain, respectively, by lower grade, rocks of the Mississippian through Middle Jurassic Bridge River Complex, and Lower Cretaceous Brew Group, which occur west of and in the hanging wall of the Marshall Creek Fault.

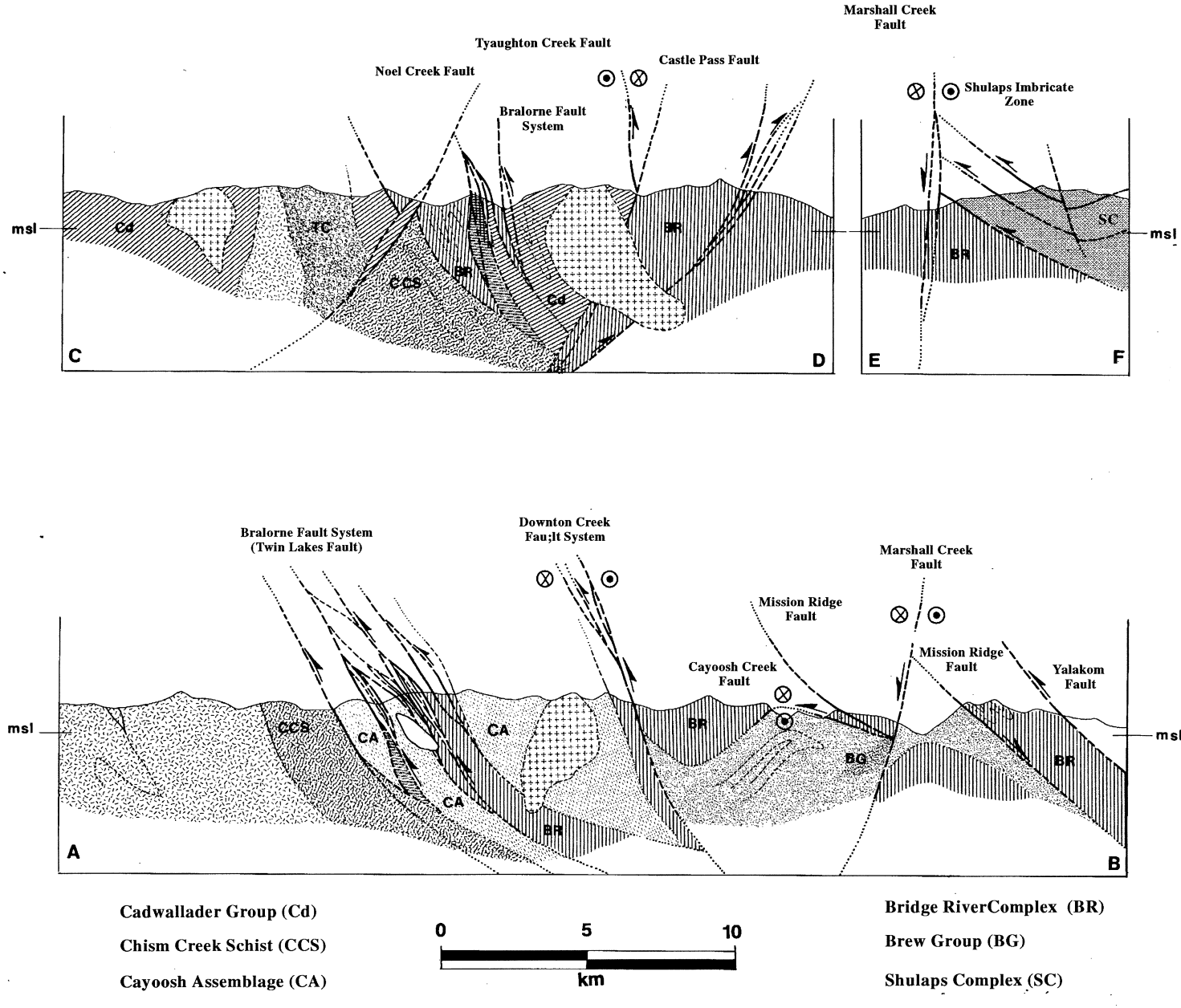
**COMMENT** Metamorphic rocks in the lower plate of the Mission Ridge Fault include strongly foliated and lineated (locally mylonitic) schists of the Bridge River Complex and crosscutting granodiorite of the Mission Ridge plutonic suite. Foliated rocks of the Bridge River Complex are metamorphosed to upper greenschist and lower amphibolite grade, and are characterized by shallow southeast-plunging stretching lineations. Asymmetric fault zone fabrics record a dextral (top-to-the southeast) sense of shear, and are coeval with development of peak metamorphic mineral assemblages (Coleman, 1990). These fabrics are interpreted to have formed prior to movement on the Mission Ridge Fault, and are attributed to earlier (Middle Eocene) dextral transcurrent shearing at mid-crustal levels along an older, shallow-dipping segment of the Yalakom Fault (Coleman, 1990; Coleman and Parrish, 1991).

Crosscutting dykes and sills of the Mission Ridge plutonic suite are folded, boudinaged, and locally contain a well-developed mylonitic foliation. They record a dextral sense of shear and are interpreted to be syn- and late-kinematic with respect to the dominant schistosity. A U-Pb zircon age of 46.5 ± 0.1 Ma (Coleman and Parrish, 1991) from one of these deformed pegmatites indicates that dextral transcurrent faulting at middle crustal levels was ongoing in the Early Eocene.

The Mission Ridge normal fault dips to the northeast and cuts across southeast-directed mylonitic fabrics in footwall schists of the Bridge River Complex. It is interpreted to have formed during late-stage unroofing of Bridge River Schists and associated rocks of the Cascade metamorphic core, prior to offset along the Fraser Fault. A kinematic linkage between this Eocene extensional fault and associated northwest-striking dextral strike-slip faults (Yalakom and Marshall Creek faults) has been suggested (Coleman and Parrish, 1991) but is not proven.

Southeast-directed Eocene structures and related fabrics along the shore of Seton Lake (foreground) overprint an older system of imbricate faults and associated mylonitic foliations (Journeay et al., 1992). These structures occur along the southwest-dipping limb of a regional antiform and are characterized by shallow-plunging lineations and associated fault zone fabrics that record top-to-the northwest displacement. To the west, these structures are cut by dextral strike-slip and southwest-dipping normal faults of the Marshall Creek Fault. They are interpreted to be footwall counterparts to northwest-directed structures associated with the Cayoosh Creek

**Facing Page:**
**Figure 24.** Fault systems of the southeastern Coast Belt. See Figure 20 for location of section lines, and Figures 23 and 28 for details of the geology.


Marshall Creek Fault
Tyaughton Creek Fault
Castle Pass Fault
Noel Creek Fault
Bralorne Fault System
Shulaps Imbricate Zone
msl
Cd
TC
CCS
BR
Cd
BR
SC
BR
msl
C
D
E
F
Bralorne Fault System (Twin Lakes Fault)
Downton Creek Fau;lt System
Marshall Creek Fault
Mission Ridge Fault
Cayoosh Creek Fault
Mission Ridge Fault
Yalakom Fault
msl
CCS
CA
CA
CA
BR
BR
BG
BR
msl
A
B
Cadwallader Group (Cd)
Chism Creek Schist (CCS)
Cayoosh Assemblage (CA)
0
5
10
km
Bridge RiverComplex (BR)
Brew Group (BG)
Shulaps Complex (SC)

Fault, which occurs in the hanging wall of the Marshall Creek normal fault to the west (see Stop 2-2; Fig. 23 and 24).

The age and regional significance of these northwest-directed structures are unclear. They may correlate with a system of westerly-directed contractional faults that occur along the base of the Shulaps Ophiolite Complex north of Carpenter Lake (Fig. 1). These structures are part of an east-dipping duplex in which mantle harzburgite and serpentinite melange of the Shulaps Complex have been internally imbricated and thrust westward over less deformed island arc assemblages of the Cadwallader Group (Schiarizza et al., 1990a). Deformation within this imbricate zone is in part synchronous with the intrusion of hornblende porphyry dykes, which locally yield a $^{40}Ar/^{39}Ar$ plateau age of 77 Ma (Archibald et al., 1991). Alternatively, these northwest-directed structures may be part of the Cayoosh Creek extensional fault system, which roots to the west.

**7.9 km** Discontinuous outcrops of foliated meta-greywacke and phyllitic siltstone in an unmapped fault sliver of the Brew Group. This fault sliver is part of the northwest-directed imbricate zone exposed along the south shore of Seton Lake in the lower plate of the Marshall Creek Fault. Fabrics in this part of the imbricate zone dip steeply to the northeast and are cut by undated granitic dykes that resemble Eocene pegmatites associated with the Mission Ridge plutonic suite.

**9.4-10.2 km** Switchbacks in road climb westward across the unexposed trace of the Marshall Creek Fault. High peaks to the south (Mount Brew) are underlain by Lower Cretaceous (Neocomian) metasedimentary rocks of the Brew Group in the upper (southwest) plate of the Marshall Creek Fault.

**COMMENT** The Marshall Creek Fault extends northwestward from its cut-off with the Fraser Fault in Ashcroft map area (Fig. 23; Monger, 1989a) to its junction with the Relay Mountain and Yalakom faults in the Noaxe Creek map area (Glover et al., 1988; Schiarizza et al., 1990a,b); a strike distance of more than 135 km. North of Carpenter Lake, the Marshall Creek Fault consists of two separate strands. The northeast strand separates greenschist from prehnite-pumpellyite facies rocks of the Bridge River Complex and is estimated to have accommodated more than 10 km of right-lateral strike-slip displacement. The southwest strand locally cuts Eocene sedimentary rocks and is interpreted to be a steep southwest-dipping normal fault. These two fault strands merge south of Carpenter Lake and continue southeastward as a single composite fault zone. Where it crosses Seton Lake, the Marshall Creek Fault comprises an older system of shallow, northeast-dipping shear zones and steep, southwest-dipping brittle faults. Northeast-dipping shear zones record dextral, top-to-the-southeast displacement, similar to that documented in the lower plate of the Mission Ridge Fault to the east (Stop 2-1; Coleman, 1990). These structures may be part of a low-angle fault zone across which displacement was transferred southeastward from the Marshall Creek Fault to the Yalakom Fault.

Normal displacement across the Marshall Creek Fault has down-dropped a segment of the Mission Ridge Fault near Seton Lake (Fig. 23, 24). Restoration of hanging wall and footwall cut-offs suggest a minimum top to-the-southwest displacement of 3.5 km (Coleman, 1990). This displacement is believed to have been tied to uplift and unroofing of Bridge River Schists in Eocene time.

**12.0 km** Outcrops of dark grey phyllitic siltstone and meta-sandstone of the Brew Group in the upper (southwestern) plate of the Marshall Creek Fault. Meta-sandstone units contain shale rip-up clasts; these are also common in clastic rocks of the Lillooet Group.

**COMMENT** The Brew Group comprises a thick succession of thin-bedded shaley siltstone, greywacke, quartzite and boulder conglomerate, locally metamorphosed to lower amphibolite grade (Duffell and McTaggart, 1952). Type localities of the Brew Group, well exposed in the vicinity of Mount Brew and along ridge systems that extend southeastward toward the Fraser River, are flanked to the north and west by the low-angle Cayoosh Creek Fault (Coleman, 1990; Journeay et al., 1992), to the east by the Marshall Creek Fault, and to the south by Tertiary granodiorite (Fig. 23, 24). Although estimated to be more than 2500 m thick (Duffell and McTaggart, 1952), upper and lower contacts of the Brew Group have never been formally defined. Studies in the vicinity of Mount Brew (Mustard, 1983) identified an antiformal syncline with an overturned limb that is several kilometres long. Continuity of map units, minor structures and stratigraphic top indicators reveal that the upper and lower siltstone/quartzite successions (each approximately 1000 m thick), as defined by Duffell and McTaggart (1952), occur on opposite limbs of this syncline and are, therefore, correlative (Mustard, 1983). Boulder conglomerates that core the syncline (Fig. 25) lie stratigraphically above the siltstone/quartzite succession.

**Figure 25.** Matrix-supported conglomerate of the Brew Group, exposed in the core of a southwest-vergent antiformal syncline, south of Mount Brew. This unit overlies fossiliferous early Early Cretaceous calc-arenite of the upper Brew Group and contains pebbles, cobbles and boulders of granite, granodiorite and subordinate siltstone, black chert and minor quartzite. Granite boulders are being processed for U-Pb dating.

The lower succession of the Brew Group is well exposed on the upright lower limb of the syncline and consists primarily of interlayered thin- and thick-bedded quartzite, dark grey siltstone and siltstone-greywacke turbidite. Thin amphibolite lenses near the base of the succession are interpreted to be metamorphosed volcanic flows. Quartzites include a distinctive thin-laminated facies in which centimetre-scale sandstone beds are separated by thin phyllitic partings, and a thick-bedded facies in which sandstone and siltstone layers are 1-5 m thick (Fig. 26). This lower quartzite facies is gradational upsection into a thick sequence of thin-laminated dark grey siltstone, sandy siltstone and quartzite. Siltstone and sandstone laminae are rhythmically interlayered and turbiditic. Higher in the succession, sandstone layers become thicker, compositionally and texturally less mature, and include lenses of calcarenite, fine-grained greywacke, and tuffaceous siltstone. Near the top of this succession are fossiliferous limy sandstones from which Duffell and McTaggart (1952) collected shell fragments identified by Jeletzky as pelecypods (*Aucella* (=*Buchia)* ex. gr. *crassicollis Keyserling*) of early Early Cretaceous age. At the top of this succession are clast and matrix-supported conglomerates containing pebbles, boulders and cobbles of granite, granodiorite and subordinate (20-30%) siltstone, black chert and quartzite (Fig. 25).

Based on similarities in stratigraphy and sandstone composition, we correlate the siltstone/quartz-rich sandstone succession of the Brew Group with quartz-rich clastic rocks of the Cayoosh assemblage (Journeay and Northcote, 1992). Together, these rocks represent a Jura-Cretaceous clastic succession that conformably overlies oceanic rocks of the Bridge River Complex and that is capped by Lower Cretaceous and younger conglomerates.

**13.2 km** Begin descent down into Cayoosh Creek. The road cuts through discontinuous outcrops of deformed, shallow dipping metasedimentary rocks of the Brew Group.

***STOP 2-2:*** **14.5 km** Foliated and lineated metasedimentary rocks of the Brew Group in the lower plate of the Cayoosh Creek Fault. Cliff sections along the road leading down to Cayoosh Creek expose an isoclinally folded sequence of dark grey phyllite and light grey metavolcanic sandstone, locally containing discontinuous lenses of calcareous sandy siltstone and marble. This sequence resembles Lower Cretaceous fossiliferous rocks of the Brew Group exposed along the ridge crest southwest of Mount Brew. Shear bands and asymmetric folds are locally developed in discrete ductile shear zones. These structures record a history of both flattening and northwest-directed rotational strain (Fig. 27) and are interpreted to be footwall splays of the Cayoosh Creek Fault.

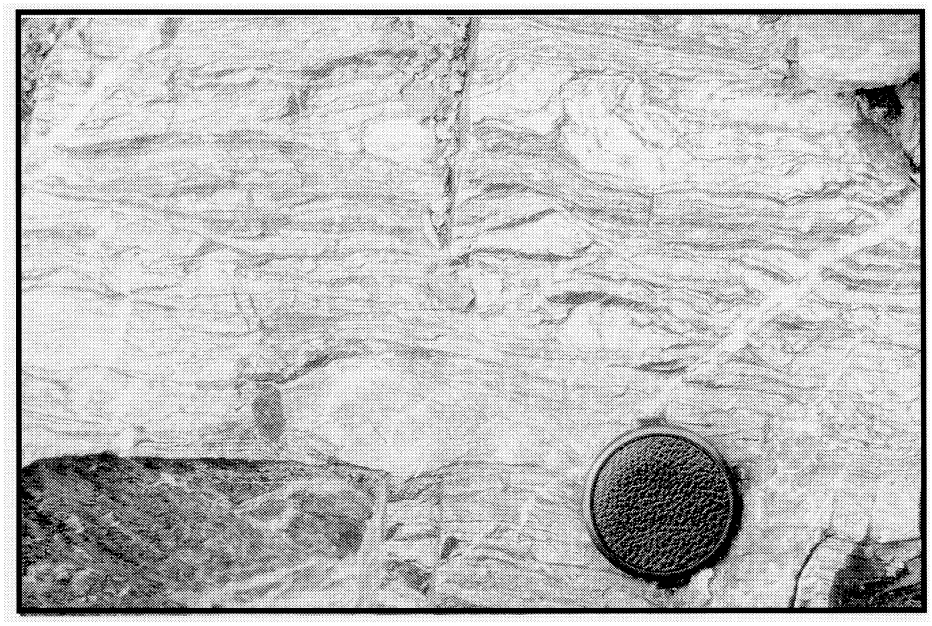


**Figure 26.** Thin-bedded phyllitic quartzite of the lower Brew Group exposed on the ridge southwest of Mount Brew.

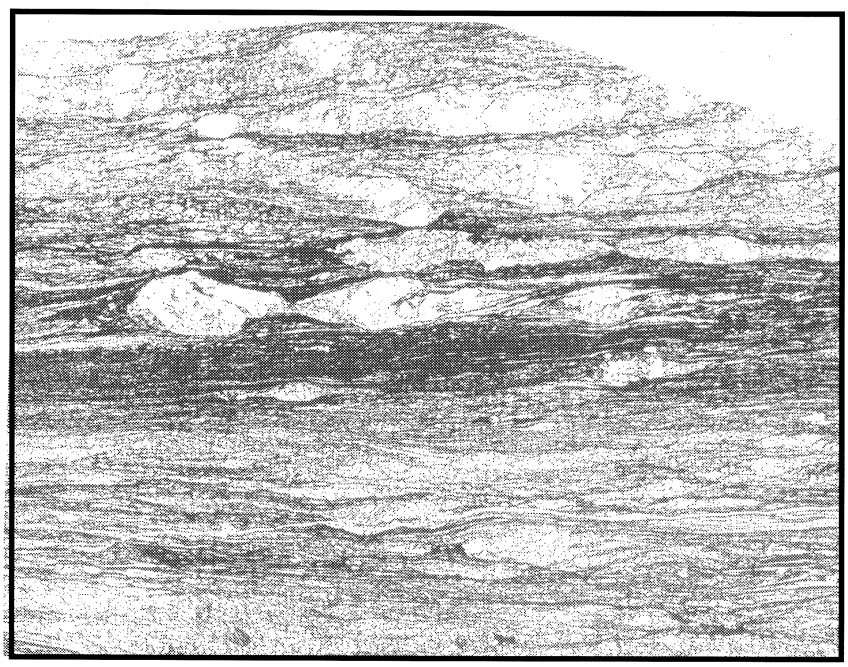
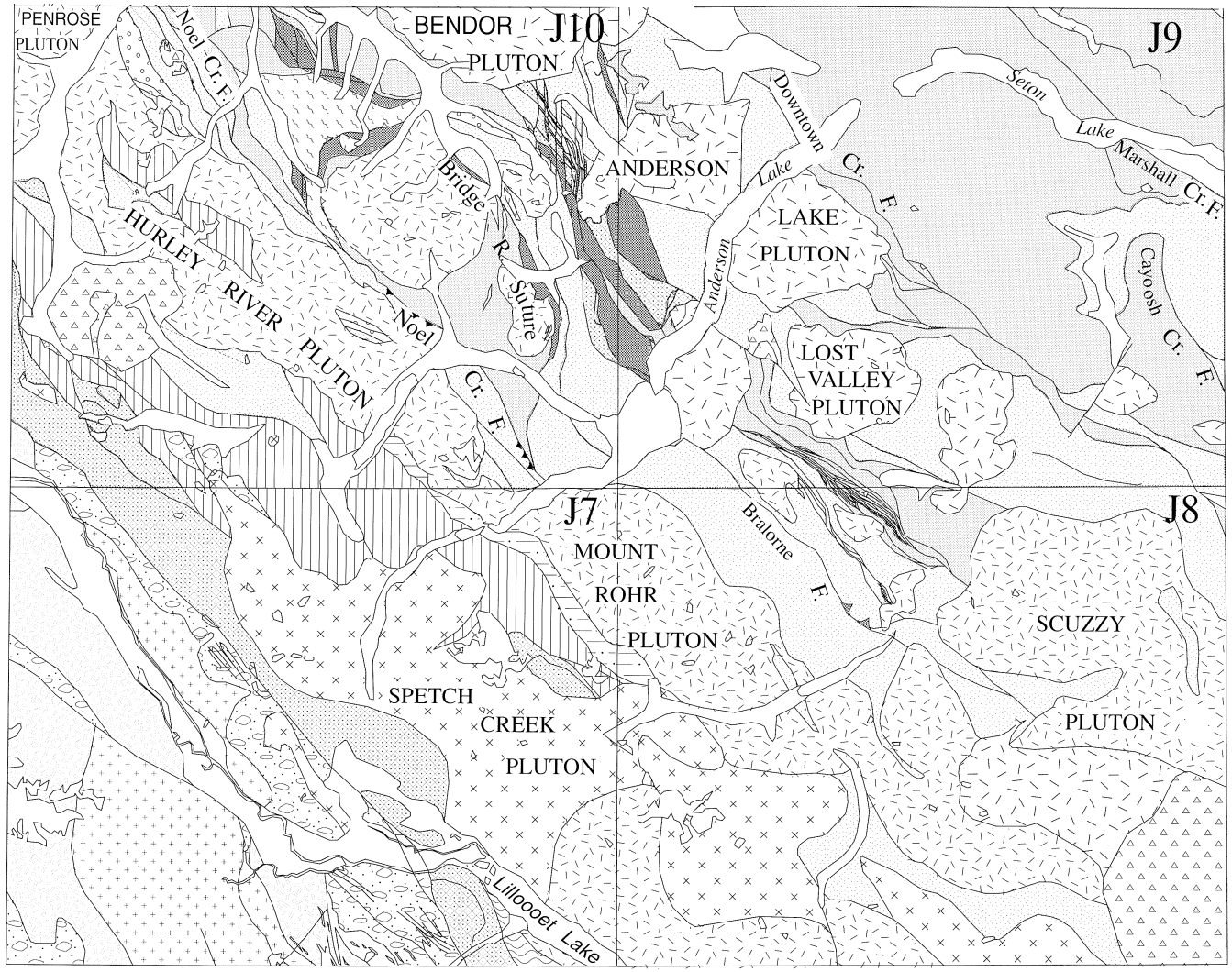


**Figure 27.** Asymmetric fault zone fabrics in the lower plate of the Cayoosh Creek Fault indicate an upper plate to the southwest sense of shear. Common shear sense indicators of this fault zone include shear bands, asymmetric porphyroclasts, pressure shadows and s-c fabrics. This section from mylonitic fault zone rocks near the head of Melvin Creek.

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- PTCC: Cache Creek Complex
- TN: Nicola Group

Undated felsic dykes cut across this shallow-dipping fabric at the west end of the outcrop.

**COMMENT** The Cayoosh Creek Fault (Coleman, 1990) separates deformed and metamorphosed schists of the Bridge River Complex in the upper plate from recumbently folded metasedimentary rocks of the Brew Group in the lower plate (Fig. 23). The fault is curviplanar and includes shallow north- and west-dipping segments mapped by Roddick and Hutchison (1973) and Woodsworth (1977) in cliff sections adjacent to Cayoosh, Copper, Cinnamon and Downton creeks, and a southwest-dipping segment previously mapped as the Phair Creek Fault (Mustard, 1983; Monger, 1989a; Fig. 23).

Structures in the lower plate of the Cayoosh Creek Fault are characterized by mesoscopic and map-scale isoclinal folds, well-exposed in cliff faces along Cayoosh Creek and alpine ridges adjacent to Mount Brew (Mustard, 1983; Coleman, 1990). The largest of these structures is an antiformal syncline exposed south of Mount Brew. The fold duplicates siltstone-quartzite successions of the Brew Group and has an overturned limb length of several kilometres. Minor folds and stretched pebble conglomerates in the hinge region indicate this is an early generation, westerly-verging syncline that plunges gently to the southeast.

Recumbent isoclinal folds in the Cayoosh Creek valley deform an older schistosity and are non-cylindrical along gently dipping axial surfaces. Stretching lineations plunge to the northwest and to the southeast. Asymmetric folds, shear bands and S-C fabrics are well developed in quartz-rich mylonites. In two localities, these fault zone fabrics record top to-the-northwest displacement (Fig. 27). Similar structures, including northwest-directed folds and mylonitic shear zones, are documented in upper plate rocks of the Bridge River Schist both to the east and west of the Marshall Creek Fault (Coleman, 1990; Journeay et al., 1992). These structures are interpreted to be part of the Cayoosh Creek Fault. They are cut by weakly deformed granitic dykes (Mission Ridge Pluton?) and are overprinted by shallow, top-to-the-south-west ductile shear zones related to displacement along the Marshall Creek and Yalakom faults.

The Cayoosh Creek Fault is interpreted to be a low-angle fault across which deformed and metamorphosed rocks of the Bridge River Complex were displaced northwestward over younger basin sequences of the Brew Group (Coleman, 1990; Journeay et al., 1992). It may be part of a Late Cretaceous (about 77 Ma) west-vergent thrust system (Schiarizza et al., 1990a) that roots eastward beneath disrupted oceanic rocks of the Bridge River and Shulaps complexes. However, the available data do not rule out the possibility that the Cayoosh Creek Fault may root northwestward as a low-angle extensional fault. Lithoprobe seismic reflection data support the latter of these two interpretations.

**14.6 km** Bridge over Cayoosh Creek.

**15.4 km-23.4 km** Outcrops of deformed Brew Group in the lower plate of the Cayoosh Creek Fault. Rock types and minor structures are similar to those described at Stop 2-2.

**Facing Page:**
**Figure 28.** Geology of the Cayoosh Range (92J/8, J/9). Based on previous work of Roddick and Hutchison (1973), Woodsworth (1977); and recent mapping of Coleman (1990), Journeay (1990a), Journeay and Northcote (1992), and Journeay et al. (1992).

**25.9 km** Cayoosh Creek Fault dips shallowly to the southwest and crosses the valley near here. Cliff faces to the east and west (above 1300 m elevation) are underlain by upper plate rocks of the Bridge River Complex.

***STOP 2-3:*** **26.7 km** (Pull-off at 26.9 km) Interlayered sequence of metavolcanic greenstone, chert and siliceous siltstone of the Bridge River Complex in the upper plate of the Cayoosh Creek Fault, locally cut by undated dykes of biotite-muscovite granite.

**COMMENT** The Bridge River Complex (Schiarizza et al., 1990b) is characterized by widespread and monotonous exposures of interbedded and structurally imbricated disrupted and broken greenstone and bedded chert with subordinate siltstone, greywacke, metavolcanic tuff, limestone and sheared serpentinite (Roddick and Hutchison, 1973; Potter, 1983; Schiarizza et al., 1989, 1990a). It includes rocks previously mapped in part as the Fergusson Series or Fergusson Group in the western Chilcotin and Bendor ranges (Cairnes, 1937; Church, 1987), and what others have identified as Bridge River Group (Roddick and Hutchison, 1973), Bridge River assemblage (Potter, 1983) and Bridge River Schist (Schiarizza et al., 1990a,b; Coleman, 1990). Radiolarian cherts and siliceous siltstones of the Bridge River Complex range in age from Mississippian to Middle Jurassic (Callovian) and, if deposited as one continuous basin sequence, would represent one of the longest-lived oceans in the geologic record (Cordey, 1988; Cordey and Schiarizza, 1993).

In addition to greenstone-chert melange of the eastern Bridge River Complex, we recognize large tracts of greenstone, chert and related sedimentary rocks in the western Bendor and Cayoosh ranges which, although cut by younger faults, may preserve primary stratigraphic relationships. We informally refer to the greenstone-chert melange and stratigraphically coherent greenstone-chert successions of the Bridge River Complex as the **Eastern** and **Western** assemblages, respectively. The boundary between these assemblages coincides with the Downton Creek Fault (see Stop 2-4; Fig. 28) in the central Cayoosh Range. The Eastern Assemblage of the Bridge River Complex is well exposed in the eastern Cayoosh Range and along the shores of Seton and Carpenter lakes. It consists primarily of discontinuous lenses of mafic greenstone, serpentinite, bedded and massive chert, greenstone-chert melange and minor volcaniclastic siltstone and limestone. The Western Assemblage of the Bridge River Complex is exposed along ridge systems of the western Cayoosh Range and along the east flank of the Bendor Range. Although we have no constraint on the age or stratigraphy of the Western Assemblage, it is possible to map apparently coherent successions of massive and pillowed greenstone (Fig. 29), calcareous greenschist, chert (Fig. 30), and lesser amounts of siliceous siltstone, limestone and fine grained volcaniclastic siltstones at a scale of 1:50 000.

The significance of two distinct assemblages in the Bridge River Complex is uncertain. Middle Triassic blueschists (230 Ma; Archibald et al., 1991) in the central Chilcotin Range (Garver et al., 1989), and greenstone-chert melange throughout much of the Eastern Bridge River Complex suggest a long history of deformation in a subduction zone/accretionary complex setting. Coherent greenstone-chert-siltstone successions of the Western Assemblage are basement to overlying deep water marine sequences of the Cayoosh assemblage, and may represent fragments of a Bridge River ocean that lay west of or within an early Mesozoic accretionary complex.

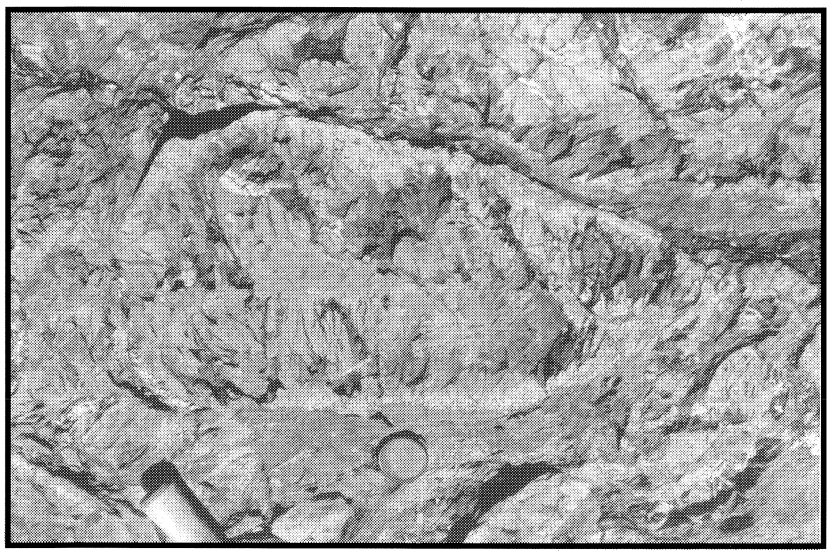


**Figure 29.** Pillowed greenstone of the Bridge River Complex exposed on ridge near headwaters of Downton Creek.

**Figure 30.** Thin-bedded chert of the Bridge River Complex exposed at Twin Lakes, near headwaters of Melvin Creek.

Paleogeographic relationships between the Bridge River terrane and island arc sequences of the Cadwallader terrane are obscured by Late Cretaceous faulting in the Eastern Coast Belt. Based on a comparison of overlying Jura-Cretaceous basin successions (Garver, 1990, 1992), we favour a scenario in which the Cadwallader arc and associated basin successions of the Tyaughton Group lay west of or possibly within the Bridge River ocean prior to the onset of large-scale crustal imbrication east of it, although P. Schiarizza (pers. comm., 1993) would favour a location to the east (Fig. 31).

**27.0 km** Greenstone-chert melange of the eastern Bridge River Assemblage is exposed in cliffs and ridge crests of the Cayoosh Range to the north.

**28.9 km** This northeast-trending section of road follows the trace of a high-angle brittle fault separating granodiorite of the Boulder Creek Pluton (part of the latest Cretaceous-earliest Tertiary Bendor plutonic suite) to the southeast from thinly laminated siltstones and shales of the Cayoosh assemblage to the northwest. Offset markers suggest apparent southeast-side-up normal displacement.

**COMMENT** The Bendor Range Intrusions (Cairnes, 1937) comprise a linear belt of latest Cretaceous-earliest Tertiary plutons that extend southeastward from Carpenter Lake to the Cayoosh River (Fig. 20, 28). From northwest to southeast, these include the Truax, Bendor and Anderson Lake plutons of the Bendor Range, and the Lost Valley, Downton and Boulder Creek plutons of the Cayoosh Range. The Truax, Bendor and Anderson Lake plutons consist primarily of medium and coarse-grained hornblende-biotite granodiorite with subordinate quartz diorite and granite. The Lost Valley, Downton and Boulder Creek plutons show the reverse trend. All three intrusions are made up of medium and coarse grained granite with subordinate quartz diorite and granodiorite.

Both plutonic suites are unfoliated and characterized by contact metamorphic aureoles that range from 300-500 m in width. Index minerals in these contact aureoles include biotite, garnet, andalusite and staurolite. A U-Pb date on igneous zircons collected from the Bendor Pluton indicates a crystallization age of 63 ± 2 Ma (Friedman and Armstrong, 1990). K-Ar cooling dates, ranging from 63-57 Ma (Woodsworth, 1977) provide upper constraints on the age of pluton emplacement and associated contact metamorphism.

**29.8 km** Boulder Creek road turnoff.

**30.4 km** View northwestward into the headwaters of Melvin Creek.

**31.4 km** The Downton Creek Fault is a southwest-directed contractional fault which records a late-stage history of dextral oblique slip displacement (Fig. 32). The fault is well-exposed along strike to the northwest in the headwater region of Downton Creek and along the shore of Anderson Lake (Fig. 24, 28). It crosses the Cayoosh Creek valley near here and extends southwestward into the headwater regions of Boulder, Molybdenite and Texas creeks.

**COMMENT** The Downton Creek Fault represents an important tectonic boundary in the Eastern Coast Belt. It extends northwestward across the Cayoosh and Bendor ranges and is on strike with a system of faults that occur north of Carpenter Lake (Fig. 24). In the Cayoosh and Bendor ranges (Fig. 28), this fault marks the boundary between coherent greenstone-chert-argillite successions of the western Bridge River assemblage and sheared greenstone-chert melange of the eastern Bridge River assemblage. It cuts the overturned limb of a regional southwest-verging syncline and associated southwest-directed thrust faults along strike to the north (Fig. 24, 28), and apparently cuts across northeast-directed thrusts and related folds of

the Castle Pass fault system in the Chilcotin, Bendor and northern Lillooet ranges. East of the Downton Creek Fault is a system of anticlinoria and synclinoria that deform previously folded rocks of the eastern Bridge River Complex. These are upright, gently inclined folds that plunge to the northeast, nearly orthogonal to the trace of the Downton Creek Fault.

Structures that can be linked to motion along the Downton Creek Fault record a complex history of southwest-directed thrusting and dextral strike-slip displacement. Where exposed along the north shore of Anderson Lake, the fault juxtaposes an upper plate assemblage of sheared greenstone and chert against structurally thickened and imbricated lower plate siltstones and graphitic shales of the Cayoosh assemblage. Layer-parallel brittle/ductile shear zones in the lower plate are 1-2 m thick and form a network more than 750 m wide. Stretching lineations are not well-developed. However, long axes of boudinaged siltstone layers and reclined intrafolial folds define a pronounced L-tectonite fabric that is steeply inclined in the dominant schistosity. Shear bands and asymmetric folds observed along several fault strands in the footwall record top-to-the-south-

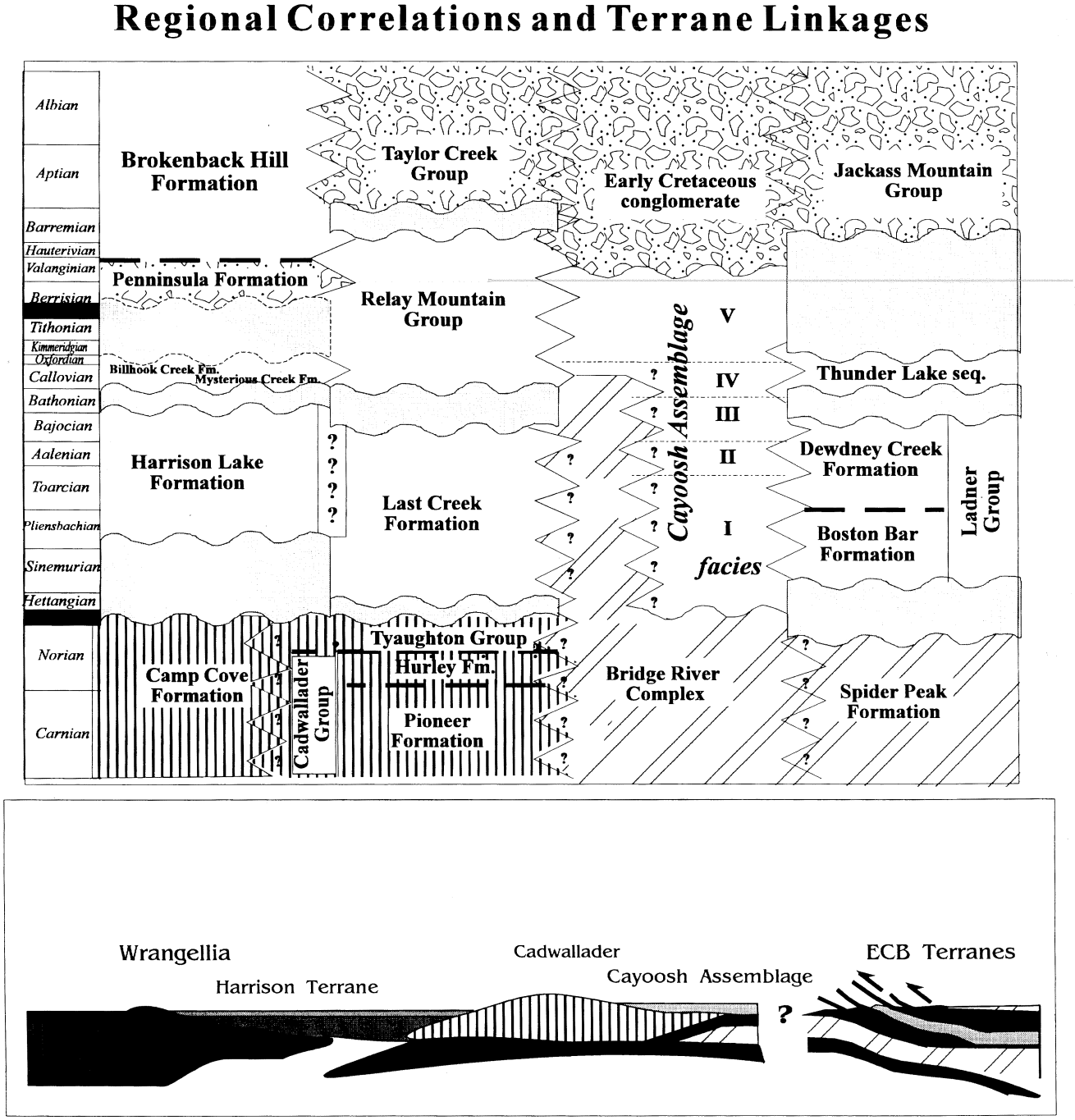
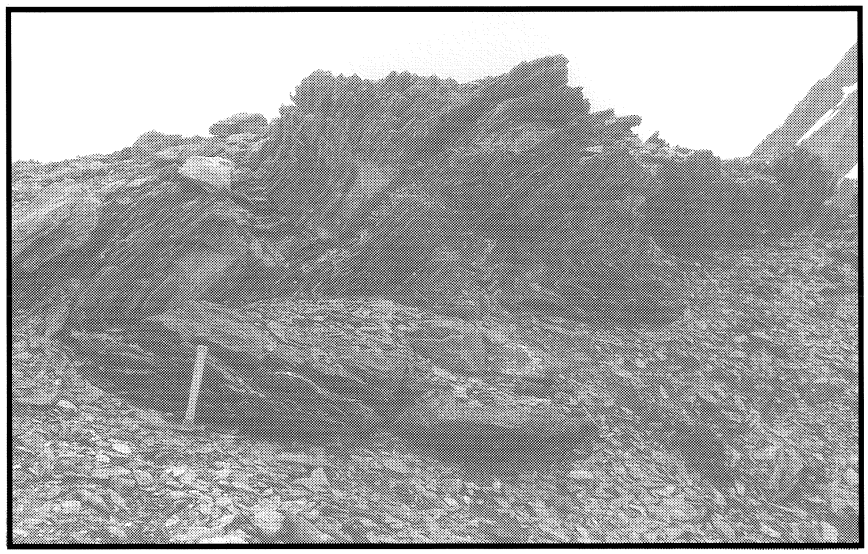



**Figure 31.** Proposed correlations, terrane linkages and paleogeographic relationships for rocks of the southeastern Coast Mountains. Taken from Journeay and Mahoney (1994).

**Figure 32.** Asymmetric folds of foliation in footwall of southwest-vergent thrust faults cut by the Downton Creek Fault. Photo taken along ridge crest at head of Downton Creek.



west thrusting (Fig. 32). However, strike-slip fault zones, containing subhorizontal lineations and asymmetric southeast-verging folds of foliation occur along the full extent of the Downton Creek Fault and record a less pronounced, but regionally significant component of right-lateral displacement.

The timing and magnitude of displacement along the Downton Creek Fault are unknown. It cuts across northeast-directed thrusts of the Castle Pass fault system (91-86 Ma?; Garver et al., 1989), and is intruded by post-kinematic granodiorite of the Boulder Creek Pluton (63-57 Ma; Bendor intrusions). It is considered to be part of the Late Cretaceous thrust system (77 ± 11 Ma; Schiarizza et al., 1990b; Archibald et al., 1991) that roots eastward beneath the Shulaps Complex (Journeay et al., 1992).

**32.1 km** Gott Creek bridge.

***STOP 2-4:*** **32.2 km** On both sides of the bridge across Gott Creek are outcrops of thinly laminated, well-foliated dark grey siliceous siltstone, recrystallized chert, talc-chlorite schist, greenschist, and thin beds of marble. This section represents a transitional facies between greenstones and pelagic sediments of the Western Bridge River Complex and conformably overlying hemi-pelagic siltstones, turbidites and volcaniclastic sandstones of the Cayoosh assemblage. The contact between this transitional facies and overlying basal successions of the Cayoosh assemblage is gradational, and has been mapped along the top of the highest bedded chert horizon in alpine exposures of the western Cayoosh Range (Fig. 33). Thinly-laminated turbidites and graphitic siltstones of the lower Cayoosh assemblage grade upward into a succession of volcaniclastic wackes, shales and quartz-rich clastic rocks. Although undated, this upward-coarsening succession resembles and is tentatively correlated with Jurassic sequences of the Boston Bar Formation (Methow terrane; Journeay et al., 1992), which overly oceanic rocks of the Spider Peak Formation east of the Fraser Fault. Stratigraphic relationships between the Cayoosh assemblage and Jura-Cretaceous rocks of the Tyaughton basin (Relay Mountain Group) are uncertain.

**COMMENT** Contact relationships between oceanic rocks of the Bridge River Complex and overlying Jura-Cretaceous basin successions of the Tyaughton and Methow basins provide an important clue to the history of terrane accretion in this part of the Coast Belt (Fig. 32).

Middle Jurassic clastic rocks of the Relay Mountain Group are widely distributed in the eastern Coast Belt (north of Carpenter Lake) and record a relatively simple history of deformation. They are structurally interleaved with island arc successions of the Cadwallader Group, and with oceanic rocks of the Bridge River and Shulaps complexes, some of which record a polyphase history of folding and associated fabric development. Based on these and other observations, Rusmore et al. (1988) proposed that terranes of the Eastern Coast Belt were amalgamated and deformed along the western edge of the Intermontane Superterrane prior to deposition of the Relay Mountain Group, which they interpreted to be a Middle Jurassic overlap assemblage. This interpretation is consistent with the megaterrane hypothesis of van der Heyden (1992), which states that terranes of the western and eastern Coast belts were amalgamated and accreted to the continental margin in Middle Jurassic time, prior to the development of a Jura-Cretaceous magmatic arc and associated intra-arc rift basins.

Recent mapping in the Chilcotin Range reveals that contacts between the Relay Mountain Group and older terranes of the eastern Coast Belt (Cadwallader and Bridge River terranes) are everywhere faulted; there is no known stratigraphic tie between these Jurassic clastic rocks and older terranes (Schiarizza et al., 1990b). In addition, there is no evidence of widespread Middle Jurassic deformation in the Eastern Coast Belt (Schiarizza et al., 1990b). Lower Jurassic shales of the Last Creek Formation and overlying Jura-Cretaceous siltstones and clastic rocks of the Relay Mountain Group (both considered to be part of the Tyaughton basin) share the same history of folding and fabric development (P. Schiarizza, pers. comm., 1993). Local variations in structural style that exist between Jurassic rocks of the Relay Mountain Group and older terranes are not unique; similar variations occur in rocks of the Cadwallader Group and Bridge River Complex, and most likely reflect a history of inhomogeneous deformation.

It remains unproven whether Bridge River and Cadwallader terranes were amalgamated by Late Jurassic time. We favour a model in which the Cadwallader terrane lay west of, or perhaps within the Bridge River ocean (Fig. 31). In the western Cayoosh Range, pelagic sedimentary rocks of the Bridge River Complex grade upwards into hemi-pelagic turbidites and graphitic siltstones of the lower Cayoosh assemblage (Journeay and Northcote, 1992). Although structurally interleaved, the contact between these two units is well-exposed and is apparently conformable. Furthermore, there is no evidence of polyphase deformation: minor structures and associated fabrics record a single-stage history of southwest-vergent folding and associated thrusting, and are well-developed in both the

Western Bridge River Complex and in conformably overlying sedimentary rocks of the Cayoosh assemblage. Based on these observations, we conclude that the Cayoosh assemblage most likely represents the youngest stage in the evolution of a long-lived Bridge River ocean basin. Upward-coarsening clastic rocks of the upper Cayoosh assemblage most likely record the final closure of this ocean basin and the amalgamation of Bridge River and Cadwallader terranes.

**32.8 km** Foliated greenstone, greenschist, and phyllite of the Bridge River Complex.

**33.4 km** Bridge over Cayoosh Creek. Cherts and greenstones of the Bridge River Complex are interlayered with phyllites and siliceous siltstones.

**34.2 km** Slow down and/or stop. The contact between greenstone/chert successions of the Bridge River Complex (to the southeast) and thin-laminated siltstones and turbidites of the Cayoosh assemblage (to the northwest) is well-exposed along the ridge to the northeast (see Fig. 33). The contact dips to the southeast and is structurally inverted along the overturned limb of a southwest-verging syncline (Fig. 24, 28). The syncline is cored by volcaniclastic sandstones and pyritiferous shales of the Cayoosh assemblage and is cut by a system of southwest-directed thrust faults.

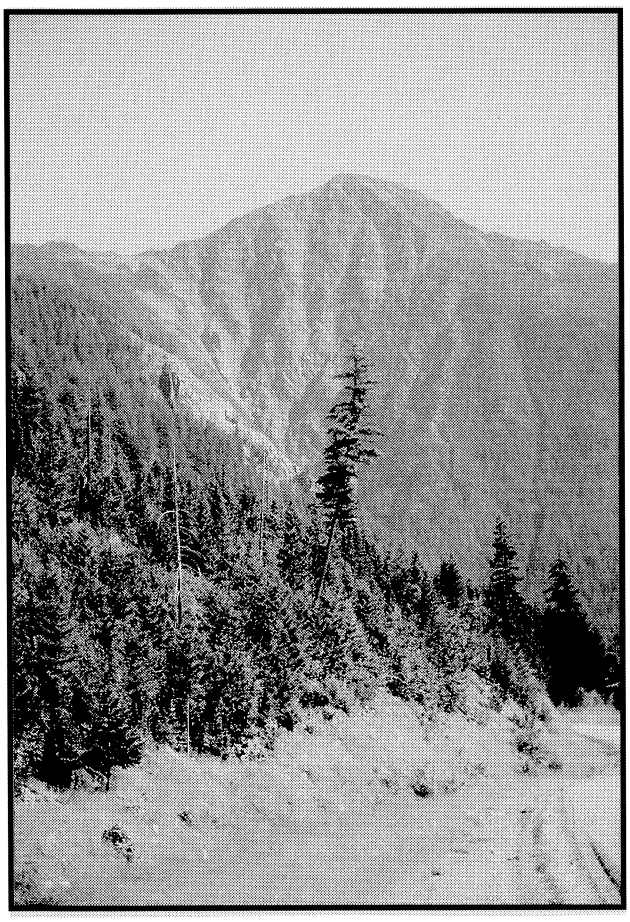


**Figure 33.** View is to northeast from Duffey Lake Road. Transitional facies of interlayered greenschist, siltstone, and chert along the contact (dotted line) between the Bridge River Complex (massive cliffs to right) and conformably overlying Cayoosh assemblage (recessive low saddle to left). The section here is structurally inverted on the overturned limb of a regional southwest-verging syncline.

**34.5 km** Float blocks of coarse-grained biotite granite derived from the Downton Creek Pluton; one of several in a northwest-trending belt of Late Cretaceous (63-57 Ma) intrusions known as the Bendor plutonic suite.

***STOP 2-5:*** **38.1 km** Bridge across Cayoosh Creek. Outcrops to the east of the bridge include dark grey siltstones and phyllites of the lower Cayoosh assemblage. These rocks are well foliated and occur on the upright lower limb of a regional southwest-verging syncline. To the west, these rocks are cut by imbricate thrusts of the Twin Lakes Fault Zone and are structurally interleaved with greenstone-chert successions of the Bridge River Complex and ultramafic rocks of the Bralorne-East Liza Complex.

**COMMENT** The Cayoosh assemblage, which is part of a widespread Jura-Cretaceous succession that may include rocks of the Ladner, Lillooet, and Brew groups, comprises a thick succession of turbidite, thin-bedded siltstone, sandstone and shale (Fig. 35). These rocks stratigraphically overly and are structurally interleaved with greenstone and chert of the Bridge River Complex.

The unit is well exposed in ridge systems of the west-central Cayoosh Range (Fig. 34), where it forms a northwest-trending belt more than 10 km wide and 50 km long. It extends northwestward into the Bendor Range, where it overlaps and is structurally imbricated with rocks of the Bridge River and Bralorne-East Liza complexes, and southeastward into the Lillooet ranges where it is intruded by Late Cretaceous granodiorite of the Scuzzy Pluton (Fig. 28). Imbricate thrusting, isoclinal folding and transposition of bedding have structurally thickened the assemblage and have obscured primary sedimentary structures. However, facies relationships are preserved and permit a partial reconstruction of the stratigraphic succession (Fig. 34).

The Cayoosh assemblage can be subdivided into five distinct lithofacies (Fig. 35). In ascending stratigraphic order, these include; graphitic phyllite, siltstone and sandstone (Unit 1), tuffaceous phyllite, graphitic phyllite, minor lapilli tuff and tuff breccia (Unit 2), graphitic phyllite, siltstone, limestone and volcaniclastic sandstone (Unit 3), graphitic siltstone, shale, phyllite, arkosic sandstone and quartzite (Unit 4), and thin-bedded graphitic phyllite, siltstone and volcaniclastic sandstone (Unit 5).

This stratigraphy represents a more detailed subdivision of the Cayoosh assemblage than that initially proposed by Mahoney and Journeay (1993). Neocomian sandstones and siltstones of Unit 5, initially included in the upper siltstone/ quartz-rich sandstone succession of the Cayoosh assemblage, are here recognized as a distinct lithofacies. Fluvial conglomerate, initially included as the uppermost member of the Cayoosh assemblage, is interpreted to unconformably overlie fossiliferous sandstones of Unit 5 in the vicinity of Mount Brew. This conglomerate represents a fundamental change in depositional environment, and has been split out as a separate stratigraphic unit. Clast compositions and the occurrence of Late Jurassic granitoid boulders (ca 151 Ma; V. McNicoll, pers. comm., 1992) suggest correlation with Early Cretaceous conglomerate of the Jackass Mountain Group. Detailed descriptions of the Cayoosh assem-

**Figure 34.** Ridge section of volcanic sandstone, siltstone and shale of the Cayoosh assemblage exposed in headwater region of Melvin Creek.

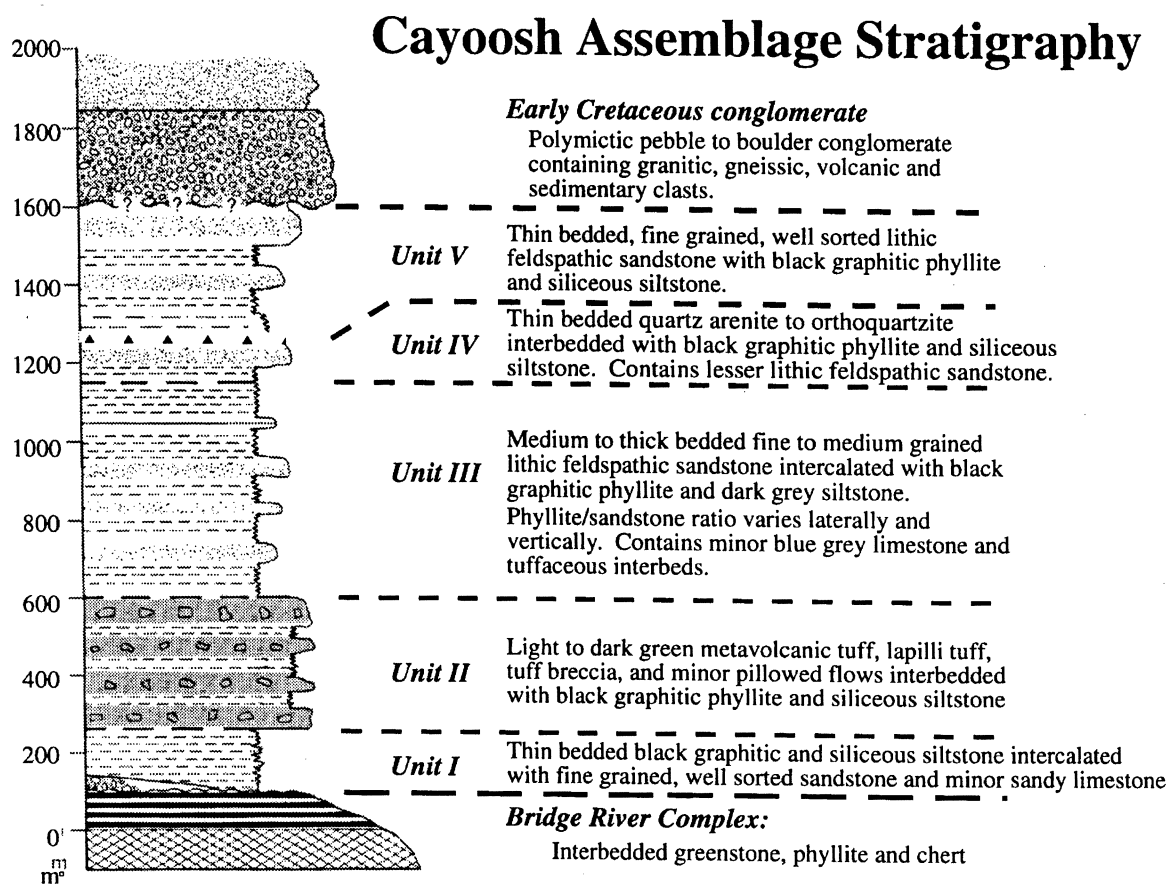



**Figure 35.** Schematic stratigraphic section of the Cayoosh assemblage, as compiled from measured structural sections in the Melvin and Mount Brew regions. See Figure 22 for details of the stratigraphy.

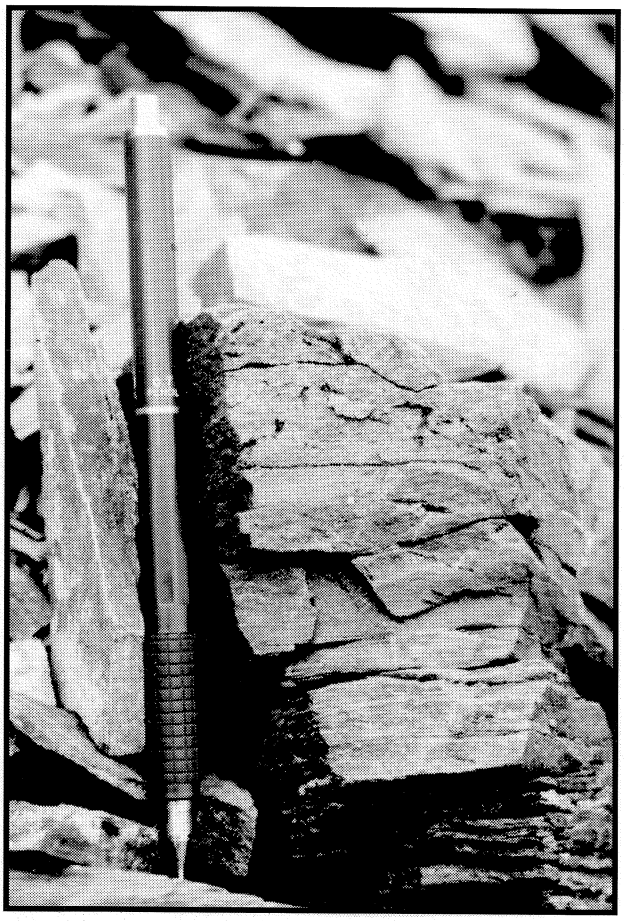


**Figure 36.** Thinly laminated turbiditic siltstones of the lower Cayoosh assemblage.

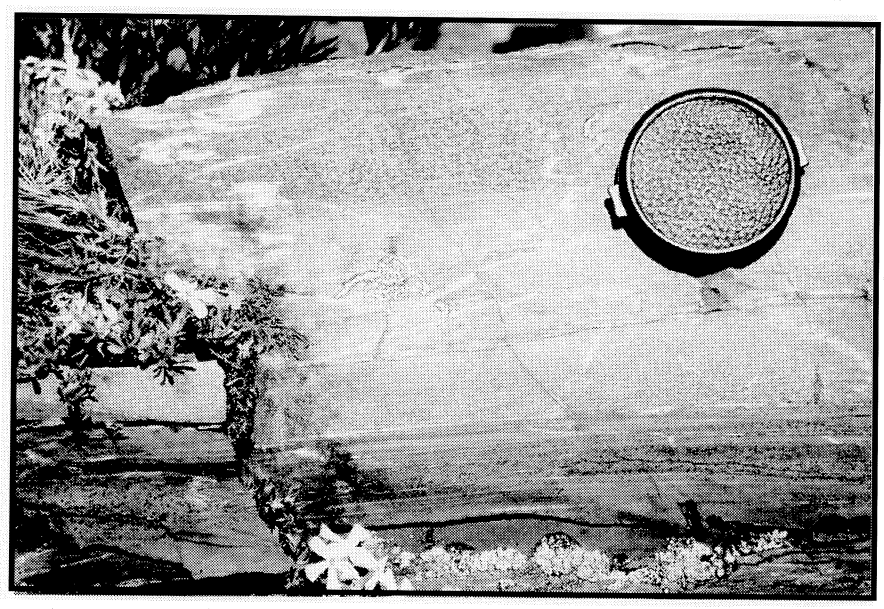


**Figure 37.** Graded volcaniclastic sandstone and siltstone of the Cayoosh assemblage exposed along ridge crest north of Duffey Lake.

blage in its type locality are reported by Mahoney and Journeay (1993). The following is a brief synopsis.

Interlayered greenstone, ribbon chert, limestone, calcareous greenschist and graphitic phyllite of the Bridge River Complex (Fig. 29, 30) grade upward with apparent conformity into a succession of graphitic siltstone, phyllite, greywacke and thin-bedded turbidites (Fig. 36) of the Cayoosh assemblage (Unit 1). The basal contact of the Cayoosh assemblage is defined at the top of the stratigraphically highest chert horizon, and is locally marked by a thin intra-formational pebble conglomerate containing clasts of limestone, siliceous argillite and chert. Thinly laminated tuffaceous siltstone occurs near the top of this succession and signals an important change in provenance.

Unit 2 is characterized by thick (2-60 m) horizons of metamorphosed light to dark green, medium and thick-bedded tuff, lapilli tuff, and tuff breccia, which occur as distinct layers in a succession of graphitic phyllite and siltstone. These pyroclastic rocks represent a minor (10-25%) but stratigraphically important component of the section. They record distinct pulses of volcanic activity in a basin characterized by thick accumulations of fine grained sediments. Both lower and upper boundaries of this succession are gradational.

Unit 3 comprises a thick (500-600 m) succession of fine to coarse grained volcanic sandstone and siltstone (30-35%; Fig. 37) interlayered with graphitic phyllite (55-65%) and thin bands of limestone (5%). Sandstone horizons consist primarily of interlayered feldspathic arenite and greywacke with intercalations of thinly laminated siltstone. Horizons of quartz-rich feldspathic arenite occur near the top of the unit (Fig. 38), suggesting that the Cayoosh assemblage may have received detritus from multiple source regions. Sandstone intervals vary in thickness (50-100 m) and lateral extent throughout the unit, and are interpreted to be lenticular in cross-section and lobate in plan view. They define a cyclic coarsening and thickening upward/fining and thinning upward succession, with the highest proportion of sandstone occurring near the middle of the unit. Pyritiferous shale and siltstone near the top of Unit 3 grade upward into feldspathic sandstone, similar in appearance to those of Unit 1, but containing a higher proportion of detrital quartz.

Unit 4 is characterized by the presence of a quartz-rich clastic facies (Fig. 39) in what is, otherwise, a thick and monotonous succession of graphitic siltstone, shale and phyllite. Thin, light grey and white laminae, interpreted as tuffaceous ash layers, are locally abundant and attest to continued volcanic activity. Sandstone horizons consist primarily of thin and thick-bedded, fine to coarse grained quartz-rich feldspathic arenite, orthoquartzite and thinly laminated phyllitic quartzite. Feldspathic arenite beds are 5-8 m thick, locally graded and contain basal shale rip-up clasts. Quartzite horizons, which represent an unusual but very distinctive component of the upper Cayoosh assemblage, consist primarily of quartz with lesser amounts of detrital feldspar and hornblende. The appearance of compositionally

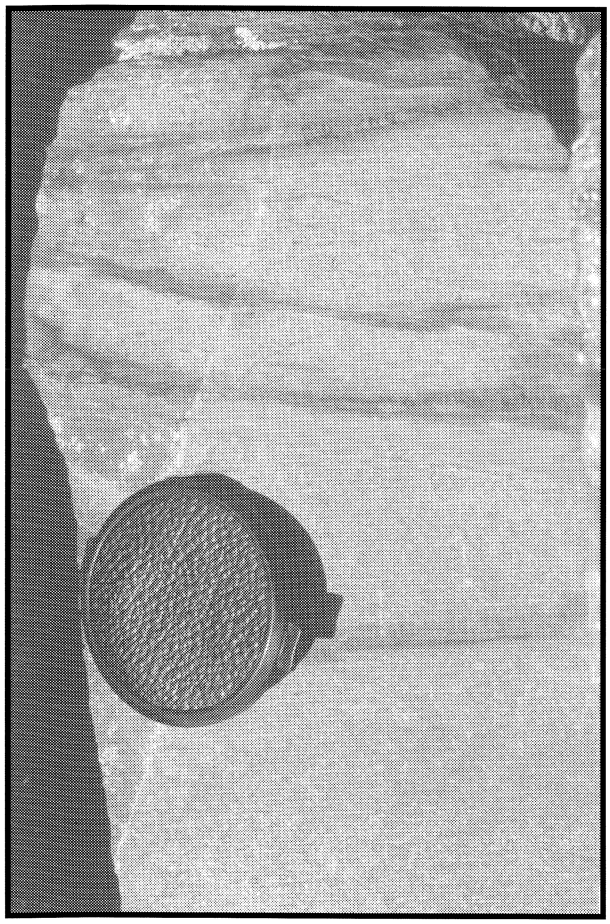


**Figure 38.** Graded feldspathic wacke of the Cayoosh assemblage (Unit III) exposed near headwaters of Downton Creek.

mature quartz-rich detritus in this part of the section suggests a continental influence. Provenance studies, including systematic petrography and U-Pb analysis of detrital zircon, are underway.

Quartz-rich clastic rocks of Unit 4 grade upward into a thick (200-300 m) sequence of thin-bedded graphitic phyllite, siltstone and fine and medium grained volcaniclastic sandstone (Unit 5). Near Mount Brew, this section coarsens upward into a sequence of interlayered calcareous sandstone, phyllite and green lithic feldspathic arenite. This upper clastic section is interpreted to be unconformably overlain by fluvial conglomerate, tentatively correlated with Early Cretaceous conglomerates of the Jackass Mountain Group (Fig. 25). The upward coarsening trend and the occurrence of fossiliferous shallow marine sandstone near the top of Unit 5 signal an important change in depositional environment, and may reflect the end of marine sedimentation in the Cayoosh assemblage.

Available age constraints indicate that the Cayoosh assemblage probably ranges from Early Jurassic(?) to Early Cretaceous (Journeay and Northcote, 1992; Mahoney and Journeay, 1993). Basal turbidites in the type locality of Melvin Creek overlie Upper Triassic (Norian) limestone-bearing oceanic rocks of the Bridge River Complex with apparent conformity. The uppermost clastic facies of the Cayoosh assemblage (Unit 5) contains shallow marine volcanic sandstones that yield bivalves (*Buchia*) of Early Neocomian age (Duffell and McTaggart, 1952).

For reasons given earlier, the Cayoosh assemblage is considered to be correlative with strata in Methow terrane (Fig. 31); it represents a deep water marine succession that overlaps rocks of the Bridge River and Bralorne-East Liza complexes, and that elsewhere includes Jura-Cretaceous turbidites and clastic successions of the Ladner, Brew, and Lillooet groups. The occurrence of thin-laminated quartz-rich clastic facies, locally characterized by chaotic intrastratal deformation, suggests that these rocks may have been deposited as distal turbidites along the Late Jurassic-Early Cretaceous continental margin. Upward-coarsening volcanic and quartz-

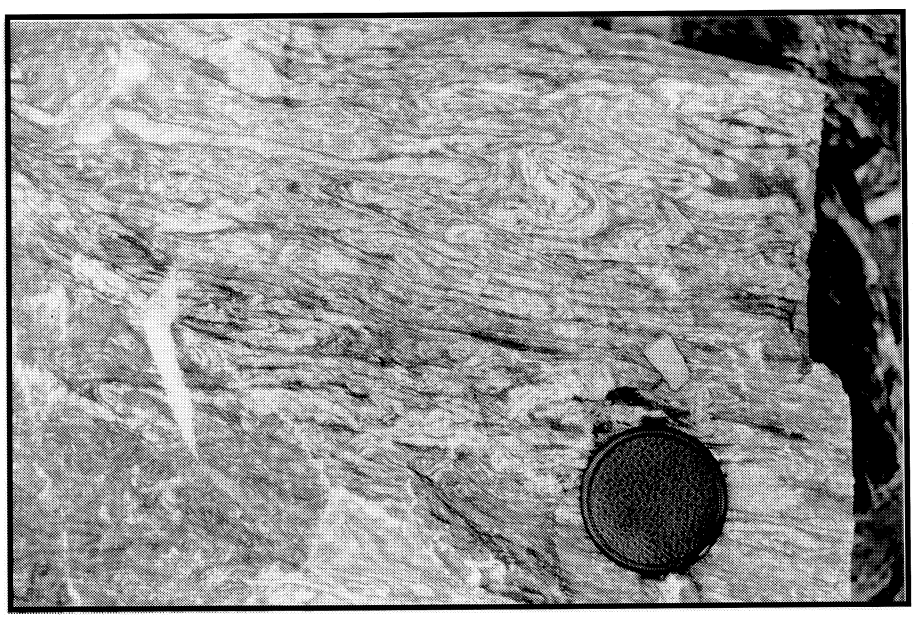


**Figure 39.** Thinly bedded phyllitic quartzite of the upper Cayoosh assemblage featuring complex intraformational(?) deformation structures, similar to those described by Monger (1991a) in the Settler Schist and Darrington Phyllite. This distinctive lithofacies is well-exposed along north shore of Anderson Lake and in the Lillooet Range, south of Cayoosh Creek Road. It is also recognized in higher-grade fault slivers of the Chism Creek melange (Fig. 21), exposed in the footwall of the Bralorne fault system to the east (see Fig. 42), which can be traced southward into type exposures of the Settler Schist (Journeay, 1990a).

rich clastic rocks in the upper part of the Cayoosh assemblage record either the final stages of terrane amalgamation or the emergence of volcanic arc and crystalline basement terranes in the Late Jurassic and Early Cretaceous. Stratigraphic relationships between this succession and age-equivalent rocks of the Tyaughton basin (Relay Mountain Group) and Gambier Assemblage are uncertain.

**39.4 km** Interlayered greenstone, greenschist, chert and dark grey siliceous siltstone of the Bridge River Complex in cliff exposures to the north, and in scattered outcrops along the road for the next 2.5 km. High ridges to the south are underlain by coarse-grained granodiorite of the Scuzzy Pluton.

**COMMENT** The Scuzzy Pluton (Roddick and Hutchison, 1973) is one of several batholithic scale igneous complexes of the southern Coast Mountains. The main body of the pluton is exposed in the Hope map area (92H; Monger, 1989a). It extends northeastward and cuts across the Twin Lakes Fault zone (Fig. 20, 28). South of Cayoosh Creek, the Scuzzy Pluton consists of medium and coarse-grained, unfoliated biotite granodiorite. Contacts with adjacent rocks of the Bridge River Complex and overlying Cayoosh assemblage are sharp and marked by a narrow Buchan-style metamorphic aureole defined by post-kinematic porphyroblasts of biotite, garnet, hornblende, andalusite and staurolite. Igneous zircons collected from the south end of the pluton yield a crystallization age of 84 Ma (Parrish and Monger, 1992).

**42.6 km** Channel Creek bridge.

**42.9-44.4 km** Discontinuous outcrops of interlayered greenstone, greenschist, chert and dark grey siliceous siltstone of the Bridge River Complex.

**44.7 km** Uppermost strand of the Twin Lakes Fault Zone crosses the road near here.

**COMMENT** Ultramafic rocks and associated greenstone-gabbro complexes occur as fault slivers throughout the Twin Lakes Fault zone and are interpreted to be part of the Bralorne-East Liza Complex. Metavolcanic greenstones of this complex include pillowed flows and flow breccias, lapilli tuffs and fine-grained amygdaloidal basalts and basaltic andesites with geochemical signatures that have affinities with modern island arc tholeiites (IAT) and mid-ocean ridge basalts (MORB). Mafic and ultramafic rocks include variably altered dunite, harzburgite, lherzolite and peridotite in which primary igneous cumulate textures are locally preserved. Adjacent to fault zones, these rocks are altered to serpentinite and talc-carbonate schist. In the Bralorne district, these ultramafic rocks and associated metavolcanic greenstones are intruded by hornblende quartz diorite and plagiogranite (soda granite) of the Bralorne suite, which yield igneous zircons of Early Permian age (270 Ma; Leitch, 1989).

***STOP 2-6:*** **45.1 km** Roof thrust to the Twin Lake Fault Zone. Disembark and walk 0.6 km eastward across imbricate footwall slivers of disrupted greenstone-chert-siltstone (Bridge River Complex), phyllite and volcaniclastic sandstone (Cayoosh assemblage), serpentinite and talc-carbonate schist (Bralorne-East Liza Complex).

**45.7 km** Pull-off for bus.

**COMMENT** Detailed (1:25 000 scale) mapping in the western Cayoosh Range (Journeay et al., 1992) has identified a system of northwest-trending imbricate faults that we refer to collectively as the Twin Lakes Fault Zone (Fig. 24, 29). The fault zone extends southeastward from Darcy, across the Cayoosh Range and into the northern Lillooet ranges, where it is cut by post-kinematic granodiorite of the Scuzzy Pluton. It represents a link in the system of southwest-directed thrust faults that define the western edge of the Bridge River terrane in the southern Coast Mountains. These include the Bralorne-McGillivray Pass and Kwoiek Creek faults (Fig. 1).

Individual faults of the Twin Lake segment (Fig. 24, 28) are marked by narrow melange zones (1-5 m thick) that contain exotic slivers of sheared ultramafic rock (serpentinite and talc-chlorite-carbonate schist) and lenticular pods of greenstone, chert and metasedimentary rocks derived from both upper and lower plate domains. The lower plate comprises polydeformed metamorphic rocks of the Chism Creek Schist, which include distinctive micaceous and phyllitic quartzites, phyllitic siltstones, metavolcanic schists and marble derived in part from the Cayoosh assemblage.

The axial part of the fault system is defined by two distinct zones of imbricate thrusting that separate stratigraphically coherent panels of low-grade sedimentary rocks (Fig. 24). The lower fault panel comprises dark grey and black graphitic siltstone, shale, thin-bedded turbidite, greenish-grey tuffaceous siltstone and greywacke, and is most likely derived from the lower Cayoosh assemblage. The upper fault panel consists primarily of thin and thick-bedded volcanic sandstone, dark grey siltstone and shale that resembles the middle sequence of the Cayoosh assemblage. Graded bedding and asymmetric cleavage relationships indicate that both fault panels are structurally overturned.

Arrays of imbricate thrust faults that flank these two panels have the overall geometry of a hinterland-dipping duplex, and define the main locus of displacement for the fault system (Fig. 24). Floor and roof thrusts contain a well-developed downdip stretching lineation and are locally mylonitic. Shear bands and asymmetric flattening foliations record top to-the-southwest displacement (Fig. 40).

Structural studies of the Bralorne-McGillivray Pass-Twin Lakes Fault Zone (Fig. 28) reveal a complex history of west-directed thrusting and left-lateral oblique-slip faulting (Rusmore, 1985; Leitch, 1989; Schiarizza et al., 1990a; Journeay et al., 1992). The earliest history of displacement is represented by north and northwest-trending thrust faults and related folds. These include the Eldorado and Fergusson faults of the Bralorne-Gold Bridge district (Rusmore, 1985; Leitch, 1989), and several unnamed faults along strike to the southeast in McGillivray Pass (Fig. 28; Journeay et al., 1992). Asymmetric folds that occur in both upper and lower plates of these faults are inclined along steep northeast-dipping axial surfaces, and are overturned to the southwest. In multi-layered sequences, these folds are tight to isoclinal and are cut by an axial planar cleavage. This early cleavage represents the dominant schistosity throughout upper structural levels of the fault zone. Stretching lineations are developed only along faults that flank the eastern and western margins of the fault zone and are steeply inclined. The timing of early thrusting (100-85 Ma) is bracketed by imbrication of the Taylor Creek Group and by post-kinematic intrusions of the Coast Plutonic Complex (Rusmore, 1985).

Vein systems and shear fractures that cut the dominant schistosity are the principal structures controlling gold mineralization in the Bralorne-Gold Bridge district. These structures are interpreted to be part of a conjugate riedel shear system that formed in a left-lateral

strike-slip regime (Leitch, 1989). They predate the main stage of gold mineralization, dated at 91-86 Ma (Leitch, 1989), and are interpreted to be linked to major left-lateral oblique slip along the Tyaughton Creek Fault and related structures (Schiarizza et al., 1990a).

We consider the Twin Lakes, McGillivray Pass and Bralorne Fault Zones to be part of a regional southwest directed thrust system that formed in response to large-scale crustal shortening and imbrication along the eastern margin of Wrangellia in Late Cretaceous time. In the Harrison Lake region, these structures are known to have been active between 97 and 91 Ma (Journeay and Friedman, 1993). Constraints on the timing of deformation in the Bralorne Fault Zone (Leitch, 1989) are consistent with this hypothesis, and suggest a possible link between southwest-directed thrusting and oblique left-lateral displacement along the boundary between the Insular Superterrane and terranes of the Eastern Coast Belt.

**46.3 km** Volcaniclastic sandstones and siltstones of the Cayoosh assemblage are well exposed in logging cuts and alpine ridges to the northwest (Hurley River Silver mine road), and in scattered outcrops along the road for the next 2.3 km. The sequence includes thin- and thick-bedded, immature volcaniclastic sandstone, dark grey slate and phyllite. Sandstone horizons are fine to medium grained and locally well-graded. Coarse sandstone beds range from 10 cm to several metres thick. They are locally channelled and contain both shale rip-up clast and pebbly mudstone horizons. This section is correlated with the upper clastic facies of the Cayoosh assemblage. Bedding-cleavage relationships and sedimentary top indicators in graded sandstones reveal that this panel is upside down. Disrupted and broken shell fragments collected from calcareous siltstones near the top of this succession are not diagnostic.

**47.0 km** Discontinuous outcrops for the next 1.5 km of volcaniclastic sandstone, siltstone and thin-bedded turbidite of the lower Cayoosh assemblage. These rocks occur in an overturned panel near the base of the Twin Lakes Fault Zone, and are separated from structurally overlying rocks of the upper Cayoosh assemblage by a tectonic melange containing imbricate slivers of serpentinite, talc-carbonate schist, chert, marble and greenstone of the Bralorne-East Liza and Bridge River complexes.

**48.6 km** Footwall thrust of the Twin Lakes Fault Zone. The fault separates hanging wall rocks of the lower Cayoosh assemblage from polydeformed footwall rocks of the Chism Creek Schist. Along strike to the northwest, this fault cuts the lower of two imbricate zones that make up the Twin Lake fault system. This imbricate zone is well-exposed in alpine ridges to the north, and is marked by a tectonic melange containing exotic slivers of disrupted diorite, gabbro, talc-carbonate schist, chert, marble and thinly laminated phyllitic quartzite. Stretching lineations and asymmetric fault zone fabrics record a top to the southwest sense of shear.

**48.9 km-50.8 km** Freshly blasted, nearly continuous outcrops of Chism Creek Schist in the lower plate of the Twin Lakes Fault Zone. The section here is dominated by metavolcanic wackes, phyllites and immature quartz-rich clastic rocks, believed to be derived in part from the upper Cayoosh assemblage. Phyllites and shaley siltstones near the east end of Duffey Lake contain late and post-kinematic porphyroblasts of garnet, staurolite ± andalusite (Fig. 41).

**COMMENT** The name Chism Creek Schist was coined by Rusmore (1985) to facilitate mapping of structurally interleaved amphibolite grade metamorphic rocks of the eastern Cadwallader Range. As defined, the Chism Creek Schist is not a stratigraphic unit, but a composite tectonic assemblage comprising imbricated slivers of bedded metachert, amphibolite, mafic and ultramafic rock, pelitic schist, micaceous and phyllitic quartzite and minor marble (Rusmore, 1985, Fig. 21). It represents the high-grade upper plate of the Coast Belt Thrust System and is well-exposed in a narrow belt that extends more than 50 km southeastward across the Cadwallader and Lillooet ranges. This belt is cut by post-kinematic granodiorite and diorite of the Coast Plutonic Complex. Pendants of Chism Creek Schist extend southward to Nahatlatch Creek and are correlated with rocks of the Cogburn Group and Settler Schist exposed in high-grade thrust nappes east of Harrison Lake (Fig. 1; Journeay, 1990a; Monger, 1991a).

As noted by Rusmore (1985), protoliths of the Chism Creek Schist are varied. Thin-bedded metachert and associated amphibolite are most likely derived from upper plate rocks of the Bridge River Complex. Mafic and ultramafic metaplutonic rocks, including meta-peridotite, serpentinite and talc-carbonate schist are presumably derived from upper plate

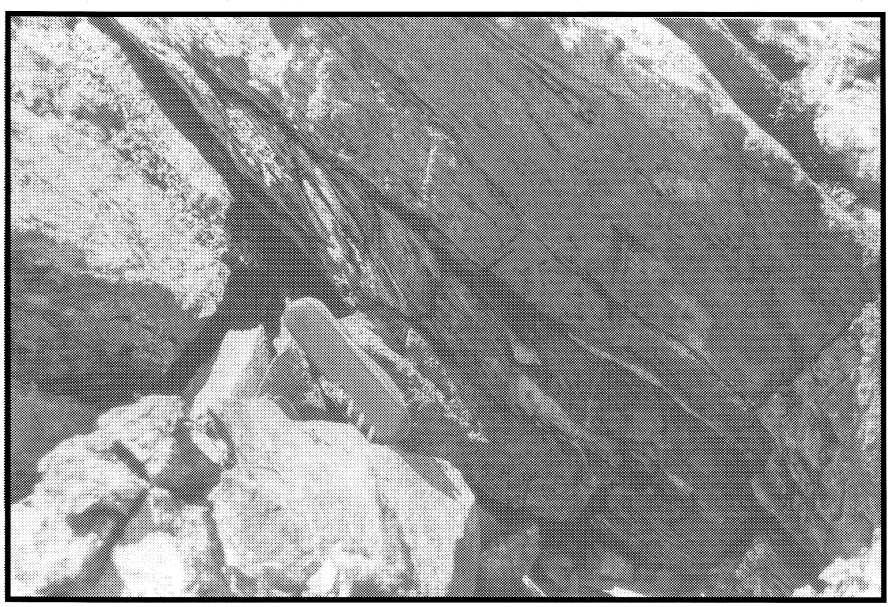


**Figure 40.** Asymmetric fault zone fabrics along the roof thrust of the Twin Lakes fault zone. Hanging wall rocks are sheared talc-carbonate "listwanite" schists of the Bralorne-East Liza Complex. Near Twin Lakes at headwaters of Melvin Creek.

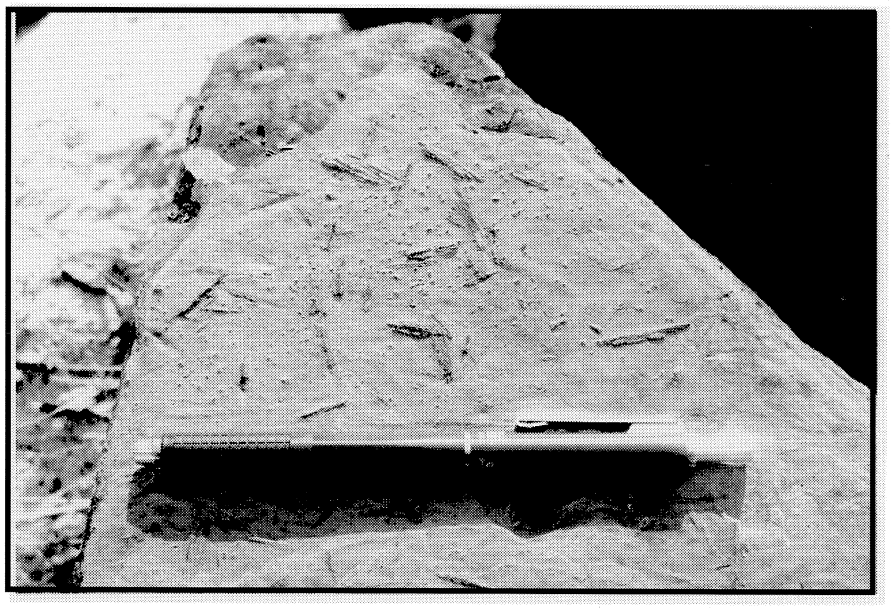


**Figure 41.** Staurolite-garnet-andalusite schists of the Chism Creek Schist exposed along ridge north of Duffey Lake.

**Figure 42.** Thinly laminated phyllitic quartzite of the Chism Creek Schist. Minor structures record a polyphase history of folding and fabric development exposed along ridge crest north of Duffey Lake. Similar structures occur in Settler Schist and possibly Darrington Phyllite.

ophiolites of the Bralorne-East Liza and Shulaps complexes. Calcareous metasedimentary rocks and associated marble horizons that occur near Chism Creek in the eastern Cadwallader Range are likely derived in part from the Hurley Formation of the Cadwallader Group. Massive and thin-bedded, locally disrupted, quartzite, pelitic schist and graphitic phyllite are by far the most abundant and distinctive rocks of the Chism Creek Schist (Fig. 42). For reasons outlined above, we consider these rocks to be part of the Cayoosh assemblage. This interpretation implies that quartz-rich clastic rocks of the Cayoosh assemblage were widespread in the southern Coast Belt prior to Late Cretaceous shortening and associated thrust imbrication.

**51.0 km** Outflow of Duffey Lake.

**51.4-52.9 km** New roadcuts of phyllitic quartzite, graphitic phyllite, metavolcanic sandstone, greenschist and (minor) greenstone of the Chism Creek Schist. Spectacular views to the west into high peaks of the Joffre Group along the southern margin of the Spetch Creek Pluton. The massif on the north is Joffre Peak. The jagged peak to the south is Mount Matier.

**53.5 km** Thin-laminated phyllitic quartzites of the Chism Creek Schist.

***STOP 2-7:*** **54.6 km** New outcrops excavated during recent construction of the Duffey Lake road expose the contact between the Mount Rohr Pluton and metasedimentary rocks of the Chism Creek Schist. Outcrops of Chism Creek Schist are characterized by a sequence of thin-laminated phyllitic quartzite, locally disrupted thin-bedded quartzite and graphitic schist (Fig. 42). Graphitic schist layers locally contain porphyroblasts of euhedral garnet and staurolite. Minor structures include two generations of tight asymmetric folds, and late-stage brittle gouge zones. Early generation folds are tight to isoclinal, steeply inclined and contain a well-developed flattening foliation. These structures are refolded by second generation asymmetric folds and overprinted by an $S_2$ axial planar crenulation cleavage. The contact between these polydeformed schists and granodiorite of the Mount Rohr Pluton is exposed at the west end of this outcrop.

**COMMENT** Integration of near-surface geological data (Fig. 24) and Lithoprobe seismic reflection data (Fig. 43, 44) provides valuable new insights on crustal structure and the history of deformation in the southern Coast Belt (see Fig. 7 and Varsek et al., 1993). Interpretations of these data (Fig. 6, 43, 44) portray the Eastern Coast Belt as a crustal-scale accretionary complex comprising an imbricate stack of tectonic flakes. Structures within this accretionary complex record a long history of west-directed thrusting (Bralorne-Duffey Lake, Downton Creek and Cayoosh Creek faults; 97-68 Ma), east-directed thrusting (Castle Pass System; 91-86 Ma?), and coeval orogen-parallel displacement along both sinistral (Tyaughton Creek Fault) and dextral (Downton Creek, Marshall Creek and Yalakom

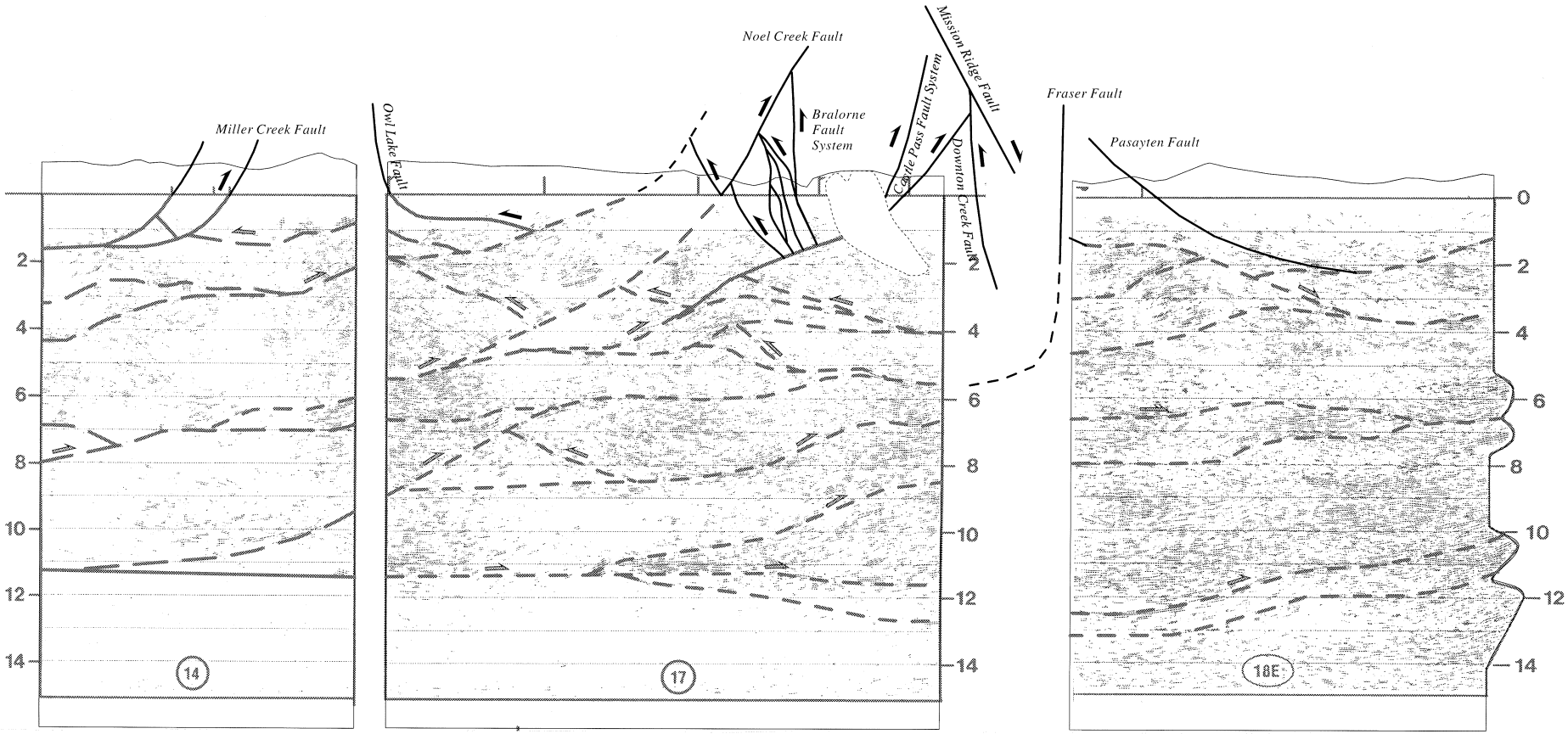



**Figure 43.** Lithoprobe seismic reflection data and interpretations of Lines 14 and 17 along the Carpenter Lake-Hurley River Road, showing integration of geological and geophysical data. Interpretation by Journeay in Monger et al. (1990). See Figure 24 for geological cross-sections and Varsek et al. (1993) for alternate interpretation of reflection data.

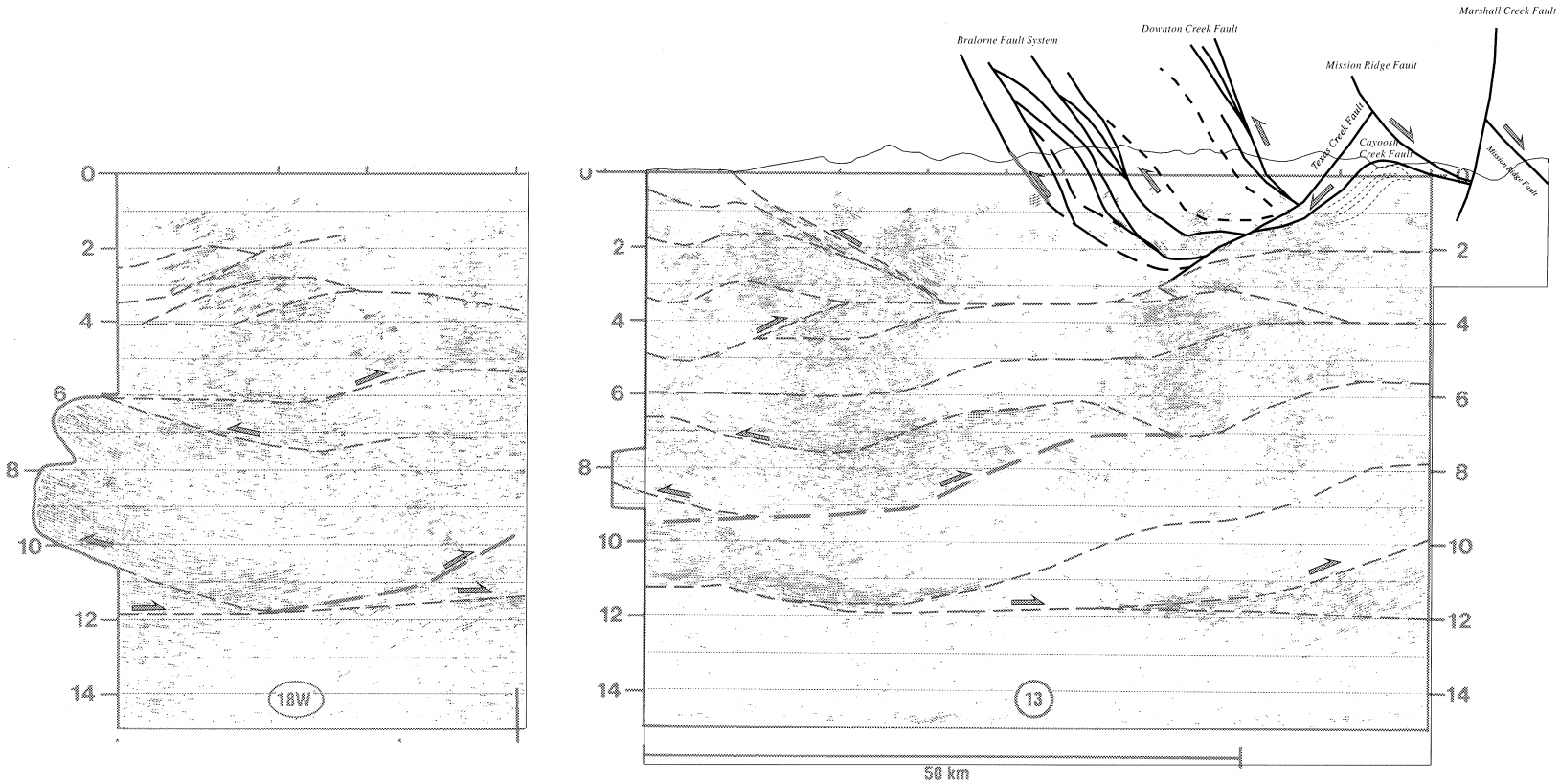



**Figure 44**. Lithoprobe seismic reflection data and interpretations of Line 13 along the Duffey Lake Road, showing integration of geological and geophysical data. Interpretation by Journeay in Monger et al. (1990). See Figure 24 for geological cross-sections and Varsek et al. (1993) for alternate interpretation of reflection data.

faults) transcurrent fault systems (Schiarizza et al., 1989, 1990a,b; Journeay, 1990a; Journeay and Northcote, 1992).

The western margin of the southeastern Coast Mountains accretionary complex is the Coast Belt Thrust System, which is interpreted to be part of a crustal-scale ramp across which assembled terranes of the southeastern Coast Mountains were thrust southwestward over supracrustal arc sequences of the Insular Superterrane in Late Cretaceous time (Journeay, 1990a; Journeay and Friedman, 1993). High-pressure (7-9 kb) metamorphic rocks in folded thrust sheets east of Harrison Lake indicate that imbricate faults of the Coast Belt Thrust System must have rooted eastward at deep crustal levels (25-30 km). This thrust system most likely represents the basal decollement for age-equivalent structures of the southeastern Coast Mountains.

East-dipping thrust faults and corresponding crustal reflections that define the fabric of this accretionary complex in the Cayoosh and Bendor ranges (Bralorne-Twin Lakes-Kwoiek Creek faults) can be traced to 3-4 s (TWTT; about 10-12 km), where they are apparently truncated by a system of subhorizontal and west-dipping reflections. On Line 14, these west-dipping reflections extend into shallow crustal levels (1 s TWTT) and can be correlated with east-directed thrusts of the Cadwallader River-Castle Pass fault system, which cut older west-directed thrusts of the Bralorne fault system (Journeay et al., 1992). Based on these observations, we interpret similar west-dipping reflections at mid-crustal levels to be part of this east-directed thrust system. Geometrical evidence in support of this interpretation includes: 1) the ramp-flat trajectory of mid-crustal reflections in Lines 13, 14 and 17 and, 2) folded reflections (Coast Belt Thrust System?) above these ramp-like structures (as seen at west end of Line 13 at 3.5-4.5 s TWTT). These structures cut across the basal decollement of the Coast Belt Thrust system, and apparently root westward beneath Wrangellia.

Shallow-dipping mylonites in the transfer zone between the Marshall Creek and Yalakom faults (Stop 2-1) suggest that these and other transcurrent faults of the Eastern Coast Belt may be listric at depth. Relative timing relationships indicate that some of these transcurrent faults were active during both west- and east-directed thrusting, although direct kinematic linkages have not been proven. As a working hypothesis, we propose that some of these contractional and strike-slip structures may share common detachment horizons at depth, thereby allowing both orthogonal shortening and orogen parallel displacement in an overall transpressional regime. Subhorizontal reflections in the upper and middle crust may correspond with these detachment horizons.

The southeastern Coast Mountains accretionary complex is similar in both scale and overall geometry to that formed in the upper plate between the present Cascadia subduction zone and Wrangellia terrane on Vancouver Island, in which slivers of oceanic plateau and ocean floor and(?) arc complexes (Crescent and Pacific Rim terranes) have been imbricated and underplated beneath the Cretaceous continental margin represented by Wrangellia (Fig. 6). We interpret the southwestern Coast Mountains (part of Insular Superterrane) as a relatively rigid crustal block that was incorporated into the footwall of this accretionary complex some time in the late Early Cretaceous. Its western margin forms the coherent upper plate of a Late Cretaceous to Holocene subduction zone. Incorporation of this rigid crustal block into the upper plate of the Cascadia subduction zone appears to have had a buttressing effect on subsequent deformation in the southeastern Coast Mountains, leading to renewed shortening, detachment and eastward backthrusting of the Insular Superterrane along systems of both west- and east-vergent contractional faults and to orogen-parallel displacement along systems of both sinistral and dextral strike-slip faults.

**55.0 km** End of outcrop. Pull-out for bus.

**56.4 km** Discontinuous outcrops of granodiorite along the northeast margin of the Mount Rohr Pluton.

**COMMENT** The Mount Rohr Pluton (Roddick, 1987) is contiguous with and believed to be part of the Late Cretaceous Scuzzy plutonic suite (Fig. 28). It is an elongate body of medium and coarse grained biotite-hornblende quartz diorite that extends 20 km northwestward from Scudamore Creek, across Duffey Lake into the western Cayoosh Range. The pluton is post-kinematic and intrudes the boundary between polydeformed metamorphic rocks of the Chism Creek Schist to the east and low-grade volcanic and sedimentary rocks of the Cadwallader Group to the west. Biotite and hornblende-rich inclusions throughout the pluton (Roddick, 1987) are elongate parallel to the regional northwest-trending structural grain of the southern Coast Belt. Laths of subhedral epidote (Roddick, 1987; Michael and Russell, 1989), suggest magmatic crystallization at relatively high pressure. However, the contact aureole along the southeast flank of the pluton is characterized by the assemblage biotite-garnet-staurolite ± andalusite, indicating final emplacement at mid-crustal levels. These data, together with a recently reported U-Pb zircon date of 86 Ma (Friedman, 1990) suggest that the Mount Rohr Pluton most likely represents an older and deeper phase of the Scuzzy Pluton. This belt of Late Cretaceous plutons extends northwestward and includes the Hurley River quartz diorite (Roddick and Hutchison, 1973), which yields a U-Pb zircon age of 87 ± 1 Ma (Friedman and Armstrong, 1990).

**58.3 km** Bridge over Van Horlick Creek.

**58.5 km** Fine-grained banded gneiss and metasedimentary rocks of the Chism Creek Schist are exposed in an elongate pendant along the east flank of the Mount Rohr Pluton. Pendants along strike to the south in the Lillooet Range contain a distinctive suite of thin-laminated phyllitic quartzite, massive orthoquartzite, graphitic schist, minor amphibolite and thin lenses of marble. Pods of altered mafic and ultramafic rock occur in several of these pendants. The association of rock types, together with similarities in the style of deformation and history of metamorphism, suggest that these pendants are part of a continuous metamorphic belt, which includes high-grade thrust nappes of the Cogburn Group and overlying Settler Schist. These thrust nappes represent the upper plate of the Central Coast Belt detachment, and track northwestward toward Bralorne (Rusmore, 1985). Variations in metamorphic grade in the upper plate of the Coast Belt detachment reflect the regional northwestward plunge in this part of the Coast Belt.

**59.5 km** Bridge over Cayoosh Creek.

**61.4-65.3 km** Discontinuous outcrops of medium and coarse grained biotite-hornblende quartz diorite of the Mount Rohr Pluton.

**67.6 km** Bridge across upper Cayoosh Creek.

**68.4 km** Approximate location of the contact between the

Mount Rohr and Spetch Creek plutons. High peaks to the northwest are underlain by granodiorite of the Spetch Creek Pluton. The pluton intrudes Triassic rocks of the Cadwallader Group and overlying Middle Jurassic volcanic and related clastic rocks of the Harrison lake Formation and Cayoosh assemblage (?) along the southwest flank of Cayoosh Mountain (Fig. 28). To the east, this pendant of fossiliferous clastic rocks is intruded by Late Cretaceous quartz diorite of the Mount Rohr Pluton.

**70.9 km** Medium and coarse-grained biotite tonalite of the Spetch Creek Pluton near the entrance to the Joffre Lakes recreational area.

**71.9 km** Bridge over Joffre Creek.

**76.9-79.7 km** Discontinuous outcrops of Spetch Creek Pluton. Good views southwestward into rugged peaks at the head of northern Joffre Creek. High peaks flanking the Joffre Creek valley are underlain by massive granodiorite of the Spetch Creek Pluton. High peaks to the south (Joffre Peak and Mount Matier) are underlain by undated diorite and quartz diorite of the Lillooet Range plutonic suite.

***STOP 2-8:*** **80.6 km** View northwest across the Pemberton Valley. The valley is underlain by thrust faults that separate Lower Cretaceous Peninsula and Brokenback Hill formations of the Gambier Group to the southwest from metavolcanic rocks correlated with the Jurassic Harrison Lake Formation and Triassic Cadwallader Group to the northeast (Riddell, 1991). These faults are part of the Coast Belt Thrust System (imbricate zone) and delineate the boundary between southwestern and southeastern parts of the Coast Mountains. Shallow northeast dipping reflections, imaged along the west end of Lithoprobe Line 13 (Fig. 44) can be traced southwestward into this zone of imbricate thrusts, and are interpreted to be part of a crustal ramp that roots eastward beneath high-grade metamorphic rocks of the Chism Creek Schist.

**COMMENT** The Cadwallader Group, as defined by Rusmore (1987), comprises a lower metavolcanic sequence of mafic greenstone (Pioneer Formation) and a conformably overlying succession of limestone, conglomerate, turbidite and siltstone (Hurley Formation). Metavolcanic greenstones of the Pioneer Formation are characterized by green to purplish-weathering basaltic flows and flow breccias with major and trace element geochemical signatures that are transitional between modern island arc tholeiites and mid-ocean ridge basalts. Conformably overlying sedimentary rocks of the Hurley Formation define an upward fining sequence of thin-bedded volcaniclastic wacke, greywacke-siltstone turbidite, and calcareous siltstone (Rusmore, 1987). Intercalated with these fine-grained sediments are limestone, distinctive pebble and cobble conglomerates containing fragments of limestone, felsic volcanic and granitoid rocks presumably derived from an adjacent arc complex. Fossils collected from limestones near the top of the Pioneer Formation and from overlying rocks of the Hurley Formation range in age from late Carnian to middle Norian (Rusmore, 1985).

**81.0 km** Scattered outcrops of the Spetch Creek Pluton are exposed for the next 4 km on the descent into the Lillooet River valley.

**COMMENT** The Spetch Creek Pluton is 33 km long and up to 16 km in width. It consists primarily of homogeneous, medium and coarse-grained, biotite-hornblende granodiorite. A nearly concordant U-Pb zircon date of 102 ± 1 Ma from the south end of the pluton is interpreted to be the age of crystallization (Friedman and Armstrong, 1990). Conventional K-Ar dates of 86.2 ± 3.1 Ma and 89.8 ± 3.4 Ma on biotite, and 76.7 ± 3.9 Ma on hornblende reflect a complex history of cooling.

**84.0 km** Junction with Lillooet River road along the shore of Lillooet Lake. Continue northward along Highway 99 (dirt road). High peaks to the west are underlain by plutonic rocks of the Pemberton Complex.

**84.3 km** Foliated quartz diorite border phases of the Spetch Creek Pluton along the northeast shore of Lillooet Lake. Views to the southeast are along strike of the Coast Belt Thrust System (Fig. 45). Low hills to the west are underlain by thrust faults that separate weakly deformed arc sequences of the Gambier Assemblage in the lower plate from more complexly deformed metavolcanic rocks of the Cadwallader Group. High peaks to the east are underlain by late- and post-kinematic granodiorite of the Lillooet Range plutonic suite (96-91 Ma) and by pendants of amphibolite grade metasedimentary rocks of the Slollicum Schist which are metamorphosed equivalents of Jura-Cretaceous rocks west of Harrison Lake.

**85.0 km** Bridge over Lillooet River at north end of Lillooet Lake. Deposition of suspended sediment from the Lillooet

**Figure 45.** Aerial view to the southeast, showing Lillooet Lake along strike of the Coast Belt thrust belt imbricate zone, and the boundary between eastern and western Coast belts.

River, the principal stream feeding the lake, is estimated to be 1.1 million cubic metres per year since 1913. This corresponds to a denudation rate in the basin upstream of 280 mm/millennium (Gilbert, 1973, 1975). An acoustic reflection horizon detected in sediments of the lake bottom is correlated with the Bridge River Ash (2400 B.P.), erupted from a centre near Meager Mountain, 40 km upstream of the Lillooet River.

**86.9 km** Mount Currie rodeo grounds.

**92.4 km** The Owl Creek Fault (Fig. 28) is a steep north-northwest-striking reverse fault that separates Triassic volcanic and related sedimentary rocks to the east from Cretaceous volcanic and sedimentary rocks of the Gambier Assemblage to the west. The fault is characterized by a zone of brittle deformation that is nearly 2 km wide. Slivers of altered Gambier Assemblage and Cadwallader terrane volcanics occur throughout the fault zone, along with brecciated granodiorite and altered metasedimentary rocks of uncertain correlation. Kinematic indicators within the fault zone include a well developed system of conjugate shears, and quartz-filled extension fractures. The orientation of these brittle structures with respect to the shear zone boundary record east-side up reverse displacement.

**COMMENT** Regional uplift and unroofing of the southwestern Coast Mountains is recorded by a regional erosional unconformity of latest Jurassic-earliest Cretaceous age. The youngest known rocks below the unconformity belong to the latest Jurassic (U-Pb dates of 147, 145 Ma) Cloudburst quartz diorite near Black Tusk mountain (km 149), where they are overlain by the Cheakamus Formation with Hauterivian (ca. 135 Ma) fossils (Monger, 1993). Near Harrison Lake, lower Oxfordian (ca. 155 Ma) Billhook Creek Formation is overlain by the Lower Berriasian (ca. 145 Ma) Peninsula Formation (Arthur, 1986; Arthur et al., 1993).

Conformably overlying the earliest Cretaceous clastic succession near Harrison Lake is a late Early Cretaceous volcanic arc sequence comprising upper Valanginian to middle Albian volcanic flows, breccias and associated pyroclastic rocks of the Brokenback Hill Formation (=Fire Lake Group; Arthur, 1986). These and related volcanic rocks, included in the the Gambier Group and coeval plutonic rocks, define a widespread zone of Early Cretaceous arc magmatism in the Western Coast Belt. Middle Albian volcanics near the top of this succession indicate that the arc remained active until mid-Cretaceous time, when it was overridden from the east by high-grade thrust nappes of the Coast Belt Thrust System.

**94.3 km** Village of Mount Currie. Turn left and continue northward along Highway 99. Prominent north-northeast-striking linear scarps, highlighted by snow-filled gullies, are visible on high alpine ridges of Mount Currie to the west. These structures offset tundra along the crest of the ridge, and are interpreted to be gravitational collapse structures. The orientation of these structures appears to have been controlled in part by a system of late Tertiary right-lateral strike-slip faults.

**96.1 km** Low hills to the northeast are underlain by folded and faulted arc sequences of the Gambier Group in the lower plate of the Coast Belt Thrust System. The style of deformation is similar to that documented in the Fire Lake region (foreland belt) along strike to the southeast (see Fig. 13; descriptions of Day 1).

**99.1 km** Bridge over the Lillooet River.

**99.4 km** Airport turnoff road.

**101.3 km** Y-junction near the village of Pemberton. Turn left and continue southwestward along Highway 99.

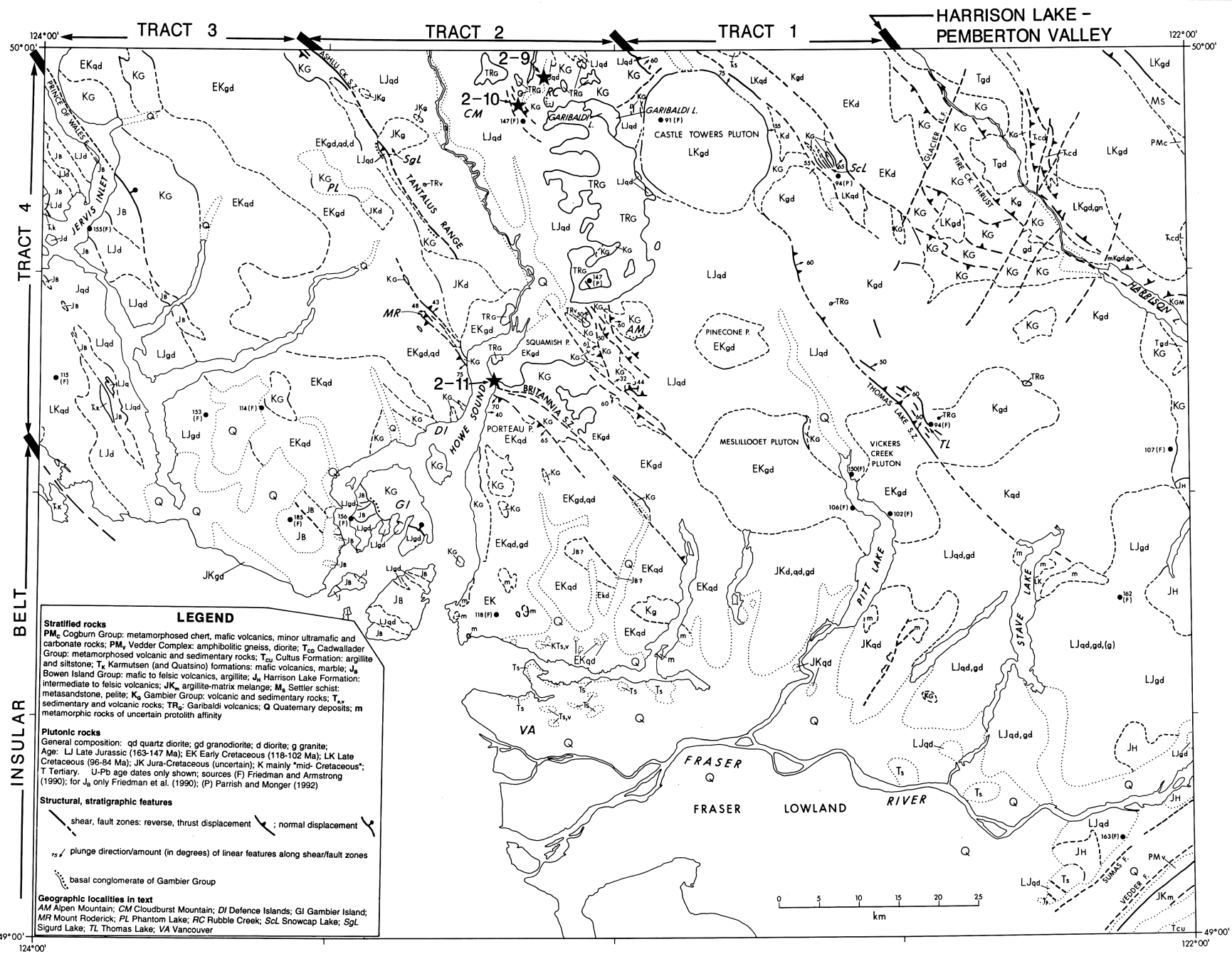
**COMMENT** A question to be pondered on this last leg of the field trip is whether Middle Jurassic-Early Cretaceous (ca. 167-110 Ma) magmatic rocks of the southwestern Coast Mountains are an integral part of Insular superterrane (e.g., Monger et al., 1994), or whether they belong to a long-lived, generally eastward-younging, Jurassic to Tertiary magmatic arc (Armstrong, 1988; van der Heyden, 1992). It is clear that mid-Cretaceous (ca. 110-90 Ma) intrusions are common to both eastern and western parts of the southern Coast Mountains. (Fig. 1, 6; Friedman and Armstrong, 1990, in press). After that time, (and before Cascade magmatism; 40 Ma) magmatism is largely restricted to southeastern Coast Mountains.

In contrast to the regionally penetrative deformation typical of southeastern Coast Mountains, the deformation pattern in most of southwestern Coast Mountains generally is concentrated in discrete major shear zones, that separate tracts in which the rocks are far less deformed (Roddick and Woodsworth, 1975; Monger, 1991b, 1993). Thus, although the southwestern Coast Mountains act as a foreland to the early Late Cretaceous Coast Belt Thrust System (CBTS), it was not a completely rigid foreland. Eastern parts, located northwest and southeast of Harrison Lake, were penetratively deformed and form the lower, external part of the CBTS, as discussed at Stop 1-2. In addition, the time of intrusion of the widespread mid-Cretaceous (110-90 Ma) plutons overlaps the time of compression on the CBTS, and some plutons contain fabrics which show their emplacement during compression. The vergence of most structures in southwestern Coast Mountains is towards the southwest, but some have the opposite sense.

Southwest of the Harrison-Pemberton valley, the Coast Mountains can be divided into four northwest-trending "tracts" separated by shear zones, within each of which plutonic rocks of a particular age predominate (Fig. 46; Monger, 1991a, 1993). From northeast to southwest, in the order traversed on the field trip, these are:

**Tract 1** contains widespread mid-Cretaceous granitic rocks, which intrude Harrison terrane and overlying Jura-Cretaceous strata, and with Middle Jurassic plutons towards its southeastern end (Fig. 46). The southwestern boundary of this tract is the mid-Cretaceous (post-94 Ma, pre-91 Ma) Thomas Lake shear zone (Fig. 1). The rocks west of the southern part of Harrison Lake are relatively little deformed, but those northwest of the northern part of the lake are involved in tight folds and thrust faults of mid-Cretaceous age, that are the frontal thrusts of the Coast Belt Thrust System (see Fig. 15, "Fire Lake allochthon;" Lynch, 1992; Journeay and Friedman, 1993). On the field trip, Tract 1 is crossed between the Pemberton

**Facing Page:**
**Figure 46.** The four "tracts" of the southwestern Coast Mountains, with location of stops 2-9 to 2-11 indicated by dots. The tracts generally have structural boundaries.




## LEGEND

**Stratified rocks**

**$PM_c$** Cogburn Group: metamorphosed chert, mafic volcanics, minor ultramafic and carbonate rocks; **$PM_v$** Vedder Complex: amphibolitic gneiss, diorite; **$T_{cd}$** Cadwallader Group: metamorphosed volcanic and sedimentary rocks; **$T_{cu}$** Cultus Formation: argillite and siltstone; **$T_k$** Karmutsen (and Quatsino) formations: mafic volcanics, marble; **$J_B$** Bowen Island Group: mafic to felsic volcanics, argillite; **$J_H$** Harrison Lake Formation: intermediate to felsic volcanics; **$JK_m$** argillite-matrix melange; **$M_s$** Settler schist: metasandstone, pelite; **$K_g$** Gambier Group: volcanic and sedimentary rocks; **$T_{s,v}$** sedimentary and volcanic rocks; **$TR_g$**: Garibaldi volcanics; **Q** Quaternary deposits; **m** metamorphic rocks of uncertain protolith affinity

**Plutonic rocks**

General composition: qd quartz diorite; gd granodiorite; d diorite; g granite; Age: LJ Late Jurassic (163-147 Ma); EK Early Cretaceous (118-102 Ma); LK Late Cretaceous (96-84 Ma); JK Jura-Cretaceous (uncertain); K mainly "mid- Cretaceous"; T Tertiary. U-Pb age dates only shown; sources (F) Friedman and Armstrong (1990); for $J_B$ only Friedman et al. (1990); (P) Parrish and Monger (1992)

**Structural, stratigraphic features**

shear, fault zones: reverse, thrust displacement ; normal displacement

plunge direction/amount (in degrees) of linear features along shear/fault zones

basal conglomerate of Gambier Group

**Geographic localities in text**

*AM* Alpen Mountain; *CM* Cloudburst Mountain; *DI* Defence Islands; GI Gambier Island; *MR* Mount Roderick; *PL* Phantom Lake; *RC* Rubble Creek; *ScL* Snowcap Lake; *SgL* Sigurd Lake; *TL* Thomas Lake; *VA* Vancouver

valley (ca. 101 km) and a location just south of the village of Whistler (ca. 145 km).

**Tract 2** features widespread Middle and Late Jurassic (ca. 162-147 Ma) Cloudburst intrusions, which are overlain non-conformably in places by Lower Cretaceous clastic and volcanic strata of the Gambier Group, and locally intruded by mid-Cretaceous (106-91 Ma) plutons. The southwestern boundary of this tract is the mid-Cretaceous Ashlu Creek shear zone, crossed near the town of Squamish (ca. 195 km), which in places has northeast vergence (Monger, 1993).

**Tract 3** consists mainly of mid-Cretaceous (U-Pb dates of 103-95 Ma) plutonic rocks that belong to the Howe Sound Batholith, and subordinate septa of the Early Cretaceous Gambier Group. It contains the Britannia shear zone, which involves Gambier strata and plutons as young as ca. 96 Ma. The southwestern boundary of Tract 3, in part obscured by granitic intrusions as old as 114 ± 2 Ma (Friedman and Armstrong, 1990), extends along Prince of Wales Reach of Jervis Inlet, where little deformed Gambier Group rocks to the east are juxtaposed against tightly folded Lower Jurassic Bowen Island strata of Wrangellia terrane to the west. At the north end of Prince of Wales Reach, the exposed fault is a 2-km wide zone that locally contains up to 30% of mafic (Gambier-related?) dykes (Monger, 1991b). On Gambier Island, fault zone fabrics indicate northeast-side-down normal displacement (Lynch, 1991). Arguably, the southwestern boundary of Howe Sound tract is a normal fault coeval with Gambier Group deposition. On the field trip, the boundary of Tract 3 is obscured by granitic rock, but is probably crossed on the highway along the east side of Howe Sound, about 7 km north of Horseshoe Bay (ca. 235 km in road log).

**Tract 4** comprises mainly granitic rocks of Middle to Late Jurassic (164-153 Ma) and Early Cretaceous (118-102 Ma) ages. No Gambier rocks have been recognized in this tract in Vancouver map area, but it contains pendants of undated metabasalt (locally pillowed amphibolite) and marble, which are probably metamorphosed equivalents of the Triassic Karmutsen and Quatsino formations on Vancouver Island, and Early Jurassic tuffaceous pelite, tuff, minor flows, and conglomerate of the Bowen Island Group, which is probably correlative with Bonanza and Harbledown formations of Vancouver Island (Fig. 8; Roddick and Woodsworth, 1979; Friedman et al., 1990; G.J. Woodsworth, pers. comm., 1990). Isoclinal folds in the Bowen Island Group (a felsic tuff from which provides a U-Pb date of 185 Ma) are crosscut by 155 Ma plutons, and demonstrate Early to Middle Jurassic contractional deformation in southwesternmost Coast Belt (Monger, 1991b), an event not recognized farther east. The southwestern limit of this tract, the western boundary of the southern Coast Belt, is a Middle Jurassic (ca. 164 Ma) magmatic front, which on islands in the north end of Georgia Strait intrudes Wrangellia terrane, and whose location is possibly controlled in part by syn-plutonic faults (Nelson, 1979; Monger and McNicoll, 1993).

Tertiary-Quaternary arc volcanics of the Garibaldi volcanic belt, seen on this part of the field trip, are the northernmost and youngest part of the **Cascade magmatic arc** (Fig. 12), which extends south from this region to northern California. Older, and more deeply eroded parts of this arc are represented by the Miocene and Oligocene (ca. 19, 24 Ma) Mount Barr and Chilliwack batholiths (Fig. 1), parts of which were crossed on Day 1, west of Hope (92.3; 140 km).

**102.2 km** The Miller Creek Fault marks the eastern boundary of the Pemberton plutonic complex (Fig. 1, 28). It dips moderately to the southwest, and extends along strike for more than 50 km. At this locality, the fault carries foliated rocks of the central Pemberton Diorite Complex northeastward over a pendant of metavolcanic rocks possibly belonging to the Gambier Group. This pendant is one of many that occur within what appears to be an eastern marginal facies of the Pemberton Complex. Crosscutting plutonic rocks in this part of the complex range in composition from hornblende diorite to biotite granite, and yield a nearly concordant U-Pb zircon age of 113 ± 2 Ma (Friedman and Armstrong, 1990, in press).

Rocks included in the Gambier Group at this locality comprise well-foliated, feldspar-phyric, meta-volcanic flows and talc-sericite schists, similar to those exposed in the Alta Lake pendant near Whistler Village. Cutting through the section is a network of ductile shear zones which locally contain a well-developed downdip stretching lineation. Asymmetric fault zone fabrics, including shear bands and recrystallized porphyroclast tails (both sigma and delta-type) indicate top-to-the-northeast sense of shear.

**COMMENT** The Miller Creek Fault is part of a family of steep, high-angle, northeast-directed thrust faults that occur in the southwestern and southeastern Coast Mountains. These faults may be imaged as prominent west-dipping reflections in Lithoprobe Lines 13, 14 and 17 (Fig. 43, 44), and are interpreted to root westward beneath Wrangellia (Fig. 6).

**103.8 km** Weakly foliated hornblende diorite of the Pemberton Complex. Along its eastern margin, the Complex is extremely heterogeneous, but is composed primarily of diorite and quartz diorite. Throughout the complex abundant inclusions and schlieren of amphibolite and recrystallized meta-sedimentary and meta-volcanic rocks, are believed to have been derived from the adjacent Gambier Group(?). Crosscutting dykes and sills of hornblende granodiorite and granite occur throughout the complex, and appear to be syn- and late-kinematic with respect to the earliest recognized structures.

Deformation within the complex is inhomogeneous. Steeply-dipping zones of well-foliated (locally mylonitic) gneiss are 5-10 m wide, and record both east-side-up and east-side-down displacement.

Plutonic rocks of the Pemberton Complex consist primarily of plagioclase and hornblende (locally altered to chlorite) with minor biotite, magnetite and sporadic quartz. In thin-section, plagioclase ($An_{40-48}$) is irregular in shape and poorly zoned. Many grains contain relic calcic plagioclase ($An_{60}$), which in places forms skeletal outlines of former zoned cores (Roddick, 1987).

The heterogeneity of the Pemberton Complex is reflected in part by its U-Pb systematics. Two zircon splits from a sample of diorite from this locality yield the following dates:

| | | |
|---|---|---|
| $^{206}Pb/^{238}U$ | 115 M | 116 Ma |
| $^{207}Pb$-$^{235}U$ | 121 Ma | 127 Ma |
| $^{207}Pb$-$^{206}Pb$ | 246 Ma | 338 Ma |

More recent analyses of multiple zircon splits yield a

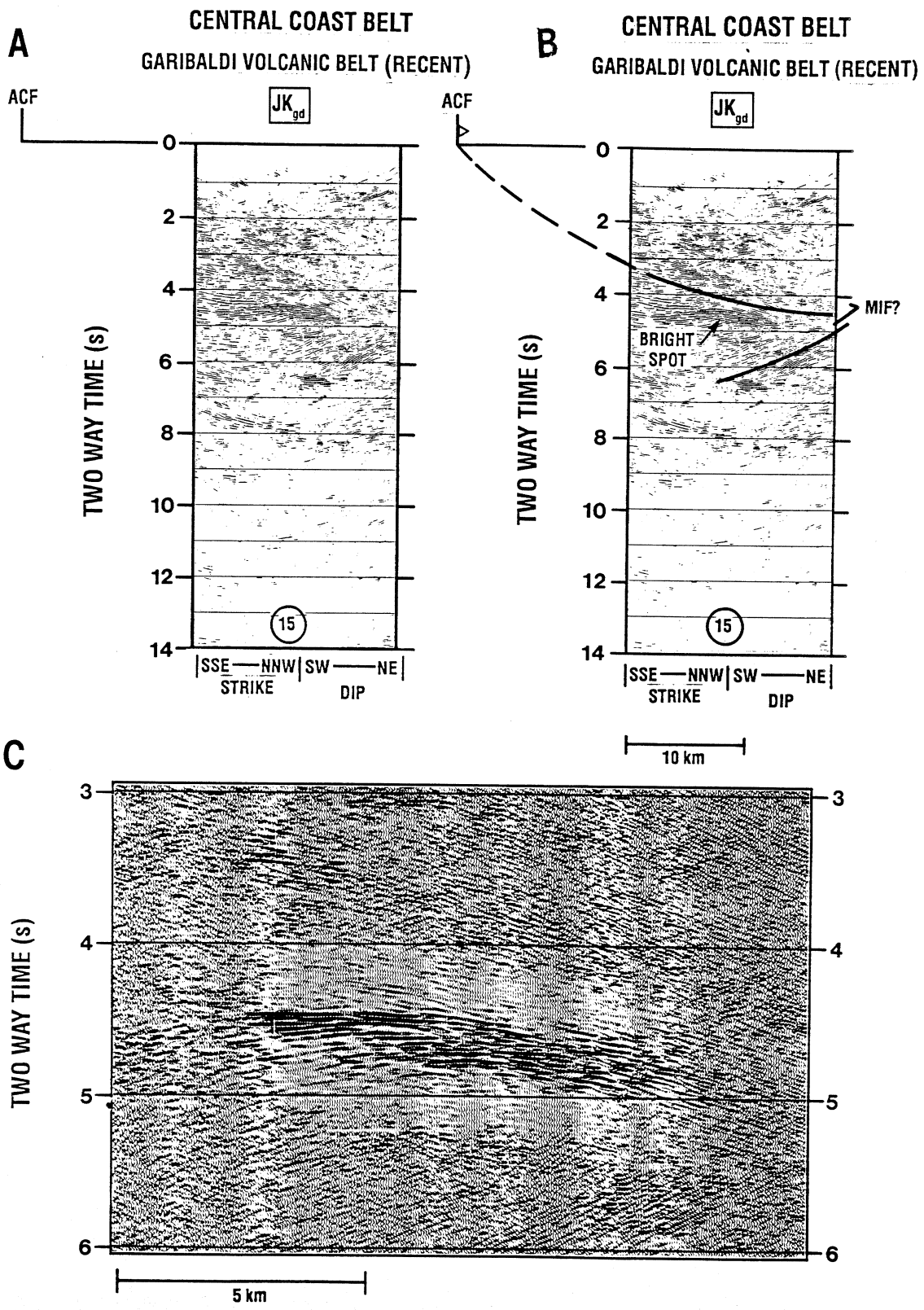



**Figure 47.** Lithoprobe seismic reflection data and interpretations of Line 15 illustrating high amplitude reflections beneath the Mount Cayley volcanic complex. Taken from Varsek et al. (1993).

nearly concordant date of 113 ± 2 Ma (Friedman and Armstrong, 1990), which is interpreted to be the age of crystallization for this part of the complex.

**104.5 km** Turn-off to Nairn Falls campground.

**104.7 km** Railroad crossing.

**106.8 km** Railroad bridge over Highway 99. Sharp S-turns.

**108.7 km** Bridge over Rutherford Creek.

**109.8 km** Quarried deposits of alluvial gravel at the southwest end of Tisdale valley.

**110.2 km** Scattered outcrops of Pemberton Diorite Complex for the next 3 km.

**111.0 km** Bridge over the Green River.

**113.1 km** The contact between rocks of the Pemberton Complex and quartz diorite of the Wedge Mountain suite crosses the road at this point, but is not exposed in the valley bottom.

**COMMENT** The Wedge Mountain quartz diorite is part of a northwest-trending belt of undated plutonic rocks which flank the western margin of the Pemberton Diorite Complex and intrude meta-volcanic rocks of the Alta Lake pendant. To the southeast, these rocks are juxtaposed with Late Triassic argillites and siltstones along the Fitzsimmons Creek Fault.

**114.2 km** Landslide blocks of the Pemberton Diorite Complex. At this point, the road crosses the nose of a landslide that has come down from the east side of the Green River valley. The blocks are from a heterogeneous "migmatitic" phase of the Pemberton Complex, in which most of inclusions are elongate, and include both amphibolite and fine-grained diorite. The plutonic rock consists mainly of plagioclase ($An_{36}$) and hornblende, along with secondary biotite and accessory magnetite. Epidote occurs both in vein fillings and as scattered grains in the matrix.

**116.3 km** Railroad crossing. Cliffs flanking the valley for the next 3 km (116-119 km) are underlain by quartz diorite of the Wedge Mountain suite.

**122.6 km** Sheared contact between Wedge Mountain quartz diorite and metavolcanic rocks of uncertain age. The fault zone strikes to the northwest, and is cut to the south by brittle structures of the Green Lake Fault Zone. It is interpreted to be the northwest extension of the Fitzsimmons Creek Fault that has been offset to the northeast across the Green Lake Fault. The Fitzsimmons Creek Fault is the floor thrust to polydeformed rocks of the Fire Lake Allochthon (see Day 1 and Fig. 15 for details).

**122.9 km** View to northeast up to high peaks of Wedge Mountain on the left and Parkhurst Mountain to the right. Both peaks are underlain by massive, weakly foliated quartz diorite of the Wedge Mountain plutonic suite.

**123.4-124.1 km** Scattered outcrops of Wedge Mountain quartz diorite, locally cut by northwest-trending splays of the Fitzsimmons Creek thrust and brittle fault splays of the Green Lake Fault Zone.

**126.8-128.1 km** Outcrops of Wedge Mountain quartz diorite cut by brittle structures of the Green Lake Fault Zone.

**COMMENT** The Green Lake Fault is believed to be one of a regional system of northeast-striking, at least partly right-lateral oblique-slip, faults of Neogene age that occur within the southern Coast and northern Cascade mountains. Vedder and Sumas faults east of Vancouver, discussed on Day 1, belong to this system. Locally, these faults are associated with, and appear to structurally control emplacement of arc-related plutonic suites and related volcanic complexes of early and possibly late parts of the Cascade magmatic arc (Coish and Journeay, 1992; Monger, 1993). They may have formed in response to upper plate shortening (pure shear component) associated with eastward subduction of the Juan de Fuca Plate (Journeay and Csontos, 1989), and may be related to coeval low-angle thrust faults developed on western Vancouver Island. The pattern of faulting is similar to that documented in the heel region of modern subduction-zone complexes, where it apparently reflects across-strike variations in orientation of intermediate principal stresses, that are caused by the internal dynamics of the accretionary wedge.

**129.7 km** Bridge over 19 Mile Creek.

**131.0 km** Railroad tracks.

**133.7 km** Stop lights in village of Whistler.

**134.4 km** Scattered outcrops for the next 3.7 km of well-foliated chlorite-sericite schist, talc schist, phyllite, metavolcanic breccia and minor limestone in the high strain zone of the Fitzsimmons Range Fault. Tentatively these rocks are correlated with the Gambier Group.

**137.8 km** Stop lights in Whistler village. Unexposed trace of the Fitzsimmons Creek Thrust crosses road near here.

**COMMENT** West of Whistler is the young Mount Cayley volcanic complex, one of the eruptive centres of the northermost Cascade magmatic arc. The high-amplitude reflections recorded on Lithoprobe Line 15 from just west of Mount Cayley (Fig. 47) can be interpreted to indicate either the presence of a magma chamber at depth, a mafic sill of very different composition from the country rock, or a concentration of fluids trapped at this depth (Lewis et al., 1992).

**138.1 km** Scattered outcrops of foliated granodiorite of(?) Late Jurassic Cloudburst quartz diorite exposed for the next 7.1 km.

**141.4 km** Pass Cheakamus Lake turnoff.

**143.3-145.2 km** Foliated granodiorite and volcaniclastics of the Gambier Group(?). The granitic rocks here are undated but are believed to be part of the Late Jurassic Cloudburst Pluton (see Stop 2-10).

**COMMENT** The Thomas Lake Fault (Fig. 1, 46) probably crosses the highway near here, and is manifested by highly broken, green altered quartz diorite exposed in road cuts. On the ridge near Corrie Peak, 12 km to the southeast, undated granitic rocks are thrust southwestwards over metamorphosed sandstone and argillite of the Cretaceous Helm Formation. The dip of the fault there is 60° to the northeast. Fault zone fabrics and downdip stretching lineations record northeast side up displacement (Fig. 48). Time of movement on the fault appears to be well constrained. About 16 km southeast of here, the Castle Towers Pluton, U-Pb dated at 91 ± 3 Ma, intrudes

the Thomas Lake Fault (Fig. 1). Still farther southeast, near Thomas Lake, granitic rocks U-Pb dated at ca. 94 Ma occur in its hanging wall (Fig. 46).

**147.1 km** Columnar jointed basalt exposed by small lake is part of an olivine basalt flow (Cheakamus valley basalt) that descends Callaghan Creek valley to the northwest for at least 20 km, and continues down the main Cheakamus valley for at least another 10 km (Green et al., 1988). Intercalated glacial tills and fluvioglacial sediments yield radiocarbon dates of 34 000 B.P. This flow is well-exposed in roadcuts along Cheakamus valley for the next 10 km and lies on ice-scoured surfaces above chloritic schists (?meta-Gambier Group), Jurassic granitic rocks and fluvioglacial sediments.

**147.3 km** Cross Callaghan Creek.

**147.9 km** Columnar jointed flow. The sharp peak known as The Black Tusk is visible on the ridge to the southeast (left), at a distance of 11 km and an elevation of 2500 m. It comprises volcanic products of two stages of volcanism; older andesitic lavas, K-Ar dated at ca. 1.2 Ma, and younger hypersthene andesite lavas and a dome that probably rose above the Pleistocene ice sheets. The younger lavas are K-Ar dated at ca. 0.09-0.21 Ma (Green et al., 1988).

**148.5-150.2 km** More-or-less continuous outcrops of columnar jointed Cheakamus valley basalt.

**151.1 km** Brandywine Falls Park.

**151.8 km** Highway crosses the British Columbia Railway track. South of here, along the railroad and along the highway to about 193.4 km, a sequence of chlorite schist, schistose metasandstone with local conglomerate, and thin discontinuous marble horizons is part of the Gambier Group.

**COMMENT** These rocks can be traced across Daisy Lake to the east and up on to the ridges on the east side of Cheakamus valley, where they are underlie to the young volcanics of the Black Tusk. Between the valley floor and an elevation of about 2000 m they lose their penetrative foliation and are continuous with rocks of the Cheakamus, Empetrum, and Helm formations of the Gambier Group (Mathews, 1958). The Cheakamus Formation consists of coarse, green, locally crossbedded arkosic sandstone, lesser argillite and local conglomerate that in places contains well-preserved large clams (*Inoceramus* cf. *paraketzovi* Efimova, of probable Hauterivian age (re-collected with W.H. Mathews in 1991, and identified by J.W. Haggart, pers. comm., 1992; GSC Catalogue # C-187629)). The Cheakamus appears to of shallow marine origin. The overlying Empetrum Formation is a distinctive olistostromal(?) unit comprising coarse conglomerate and breccia, containing granitic, dioritic and volcanic clasts (a granitic clast was U-Pb dated by R.M. Friedman (pers. comm., 1992) at 161 ± 7 Ma)), and large blocks of limestone up to about 100 m long. The Helm comprises graded, pyritic, argillaceous quartzitic sandstones, and slaty argillite.

Similar strata are observed to non-conformably overlie Cloudburst quartz diorite of latest Jurassic (147 Ma) age in three localities (Fig. 46): near Rubble Creek (Stop 2-9; Mathews, 1958); on Alpen Mountain, 15 km east of Squamish (Lynch, 1991); and near Sigurd Lake in the Tantalus Range, 30 km northwest of Squamish (Monger, 1993). This relationship demands that uplift, erosion and deposition of shallow marine strata took place in an interval of about 12 million years (Hauterivian = ca. 135 Ma). The lithologies of the Early Cretaceous strata suggest that subsidence continued after this time; the Empetrum Formation possibly was produced by mass flow from bounding fault scarps, and the Helm Formation appears to be a deep water deposit. Although no normal faults of Early Cretaceous age have been recognized in this area, they may be present farther west and south, along Prince of Wales Reach of Jervis Inlet, and on Gambier Island (Fig. 46). These relationships suggest that emplacement of the Late Jurassic(?)-Early Cretaceous plutons and deposition of rocks of the Gambier Group was in an extensional (or transtensional setting). This contrasts with the contractional and transpressional regime during which the mid-Cretaceous through-early Tertiary plutons were emplaced.

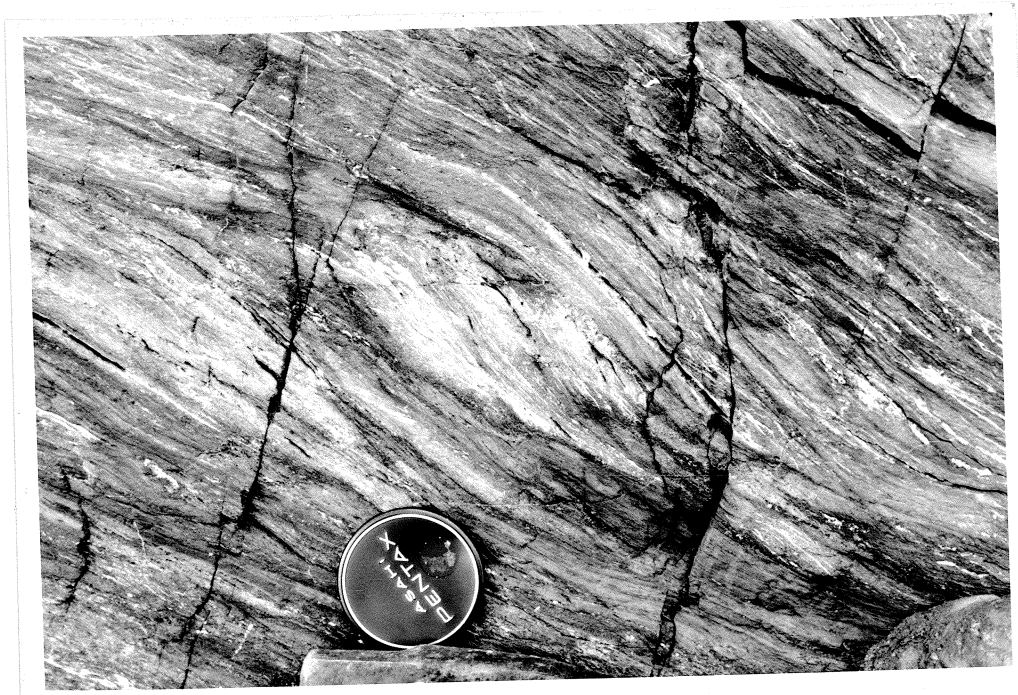



**Figure 48.** Sigmoidal structure in foliated amphibolite indicating movement to southwest (top to left). Corrie Ridge, 6.5 km east-northeast of The Black Tusk, Garibaldi Park.

**154.7-158.7 km** Cheakamus valley basalt occurs along highway; locally its can be seen to overlie chlorite schist and granitic rock.

**158.7 km** Cross Cheakamus River just below Daisy Lake dam. Water from the lake runs through a penstock below Cloudburst Mountain to the southwest to generate electricity at its outflow on the Squamish River 12 km to the southwest; the hydraulic head is about 330 m.

**159.3 km** Sharp turnoff to the northeast, then turn at 159.7 km to the southeast on a sideroad leading to the trailhead parking lot for the trail which goes to Garibaldi Provincial Park (Black Tusk Meadows). The road is over boulders of the Rubble Creek slide that formed in the winter of 1855-56 by collapse of 30 x $10^6$ $m^3$ of a lava flow of the Mount Price Complex exposed in The Barrier above Stop 2-9 (Mathews, 1958).

***STOP 2-9:*** **162.1 km** From the parking lot, walk upstream by Rubble Creek for about 100 m to outcrops of Cheakamus conglomerate. The basal contact of these rocks, nonconformable on Cloudburst quartz diorite, is exposed on the west side of the knoll 1 km south of here, across Rubble Creek. Most of the detritus in the conglomerate is derived from underlying Cloudburst quartz diorite.

To the east, up Rubble Creek, the cliff called The Barrier is composed of young (≥ 6000 B.P.) lavas, that were ponded against ice occupying the Cheakamus valley; subsequent melting of the ice created an unstable, oversteep mass. One of these flows collapsed to fill the valley with the bouldery rubble crossed on the way to Stop 2-9 (Mathews, 1958). Comparable features occur in the Cheakamus valley south of here for the next 30 km.

From Stop 2-9, return to rejoin Highway 99.

**165.2 km** Cross Rubble Creek. South of here the road crosses the boulder fields from the Rubble Creek landslide; logs caught in the debris flow are visible in the Cheakamus River bed to the right.

**167.4 km** Mylonitic Cloudburst quartz diorite.

**168.6-171.9 km** Continous roadcuts of mylonitic Cloudburst quartz diorite in roadcuts in Cheakamus canyon.

***STOP 2-10:*** **169.2 km** (Second pullout). Outcrop of Cloudburst quartz diorite, with mylonitic fabrics, first recognized by Mathews (1958). The mylonites have a northeast dipping foliation and only a weakly developed, at best, linear fabric, but generally record a sense of northeast-side up movement. The foliation trends north-northwesterly and dips steeply to the northeast. The age of mylonitization is unknown, but at this place it has a sense consistent with that of the early Late Cretaceous contraction faults.

**COMMENT** From near this locality a U-Pb date of 145 ± 2 Ma was obtained by Friedman and Armstrong (1990). Twenty kilometres south of here a precise U-Pb date of 147.0 ± 0.5 Ma was determined by Parrish and Monger (1992) from the southerly extension of the same body. These latest Jurassic plutonic rocks underlie most of the region. Gneissic fabric within them, 5 km northwest of Stop 2-10 indicate that some, at least, probably formed at depth.

Continue up hill; all roadcuts have well developed mylonitic fabric.

**171.9 km** Southern limit of zone of mylonites within Cloudburst quartz diorite. From this point on the highway southwards, mylonitic zones not recognized in the Jurassic plutonic rocks.

**173.2-176.2 km** Continuous outcrop of Cloudburst granitic rocks.

At 214.3 km, view to the south Sky Pilot Mountain, with the crest formed of Gambier volcanics that are intruded by Cretaceous granitic rocks of Tract 3.

**176.6 km** Cross Culliton Creek. Upper parts of this contain lavas similar to those at The Barrier that were ponded against ice.

**177.1-187.2 km** Outcrops of Late Jurassic granitic rock, featuring dark mainly Gambier(?) dykes, and rare brown weathering dykes probably related to the young, Garibaldi, volcanics.

**187.7 km** Cheekye River crossing. South of here the route is across the gently south-sloping surface of the Cheekye Fan, built of debris from the collapse of Mount Garibaldi, a Quaternary (ca. 0.22-0.26 Ma) volcano built through the Wisconsin ice (Mathews, 1958; Green et al., 1988). Melting of the ice filling the valley of Cheekye River led to collapse of the unsupported western flank of Mount Garibaldi. Smaller debris flows down the the Cheekye River valley occurred as recently as 1958. The 2900 m high volcanic edifice of Mount Garibaldi, located 12 km northeast of the crossing, is founded on an irregular surface lying at elevations of 1500-2000 m composed mainly of latest Jurassic granitic rock.

**188.3 km** Pass Alice Lake turnoff.

**190.8 km** Pass Brackendale turnoff. South of here, view to southeast, with notch formed by the valley of the Stawamus River. This follows the trace of a strand of a fault system aligned with the Ashlu Creek shear zone, which is the boundary between tracts 2 and 3. East (left) of the valley are mainly Late Jurassic plutonic rocks, locally nonconformably overlain by Gambier volcanics (and traversed since ca. 145 km); west (right) of it, are mainly mid-Cretaceous plutonic rocks with minor septa of Gambier volcanics (Fig. 1).

To the southwest is the high, rugged Tantalus Range is composed of mainly of Late Jurassic granitic rock (U-Pb date of 150 Ma; V. McNicoll, pers. comm., 1994), with a narrow, discontinuous band of Gambier Group strata crossing from its south side, west of Squamish, to its northeast side in the upper part of Ashlu Creek. Near Sigurd Lake, in the northwestern Tantalus Range, Gambier strata locally nonconformably lie on Late Jurassic quartz diorite and both are intruded by mid-Cretaceous granitic rock (K-Ar date of 94.7 Ma; P. Hunt, pers. comm., 1994) (Fig. 46). Although the Ashlu Creek shear zone is a major boundary, the sense of movement along it is equivocal. Most strain indicators, mainly broken and rotated feldspars in both Jurassic plutons and Gambier Group strata, indicate southwest-side up movement. In the Stawamus valley, local subhorizontal slickensides on fault surfaces indicate local dextral movement.

**193.7 km** Garibaldi Way crossing.

**195.2 km** Mamquam River crossing. The bottom of the valley of the Mamquam River to the east of here contains a young postglacial flow (post-6670 B.P.; Green et al., 1988).

From this point, is a good view to the southeast of the "half dome" of Stawamus Chief Mountain, composed of massive granodiorite, U-Pb dated at ca. 101 Ma (Friedman and Armstrong, in press). About 7 km east of here on the north side of the Mamquam River valley is latest Jurassic Cloudburst quartz diorite (U-Pb dated at 147.0 ± 0.5 Ma; Parrish and Monger, 1992). It is cut in places by Gambier(?) dykes; on the south side of the Mamquam valley, the dykes are more abundant, and on Alpen Mountain, about 20 km east of here, the pluton is nonconformably overlain by Gambier volcanics (Lynch, 1991).

**198.2 km** Junction with turnoff to town of Squamish to south.

**198.6 km** Cross Mamquam Blind Channel.

**199.9 km** Cross Stawamus River. Above (to southeast) the local rock climber's mecca is the west face of Stawamus Chief Mountain. Outcrops of this mid-Cretaceous granodiorite occur by highway at 240.0 km.

Older maps (Roddick and Woodsworth, 1979) show this as a discrete intrusion called the Squamish Pluton. However, Monger and McNicoll (1993) argued, on the basis of U-Pb dating, that it was merely an apophysis of a much larger, mid-Cretaceous (ca. 103-96 Ma) "Howe Sound Batholith" and which also contains the partly fault-bounded "Porteau" (= "Furry") plutons crossed further south along Howe Sound. In dimension, the batholith is at least 120 km long, and 40 km wide. It was emplaced in part during the period of contraction which formed the Coast Range Thrust Belt.

**202.2 km** Shannon Falls on east side.

**202.6-208.7 km** More-or-less continuous outcrops in mid-Cretaceous quartz diorite, which locally is very mafic and contains magnetite layers. This intrudes Early Cretaceous Gambier Group at 247.7 km (1.6 km south of Murrin Provincial Park); the contact is crossed on the hill descending to Britannia Beach.

**209.6 km** The settlement of Britannia Beach is the former townsite for the now inactive Britannia mine, which was discovered in 1888, worked from 1905 to 1974, and presently is the home of the B.C. Museum of Mining. The mine yielded primarily copper, with lesser amounts of zinc and lead, and minor gold and silver, derived from stringer lodes and replacements of pyrite, chalcopyrite and sphalerite, that probably formed as volcanogenic massive sulphides (Payne et al., 1980). The deposits are located in chlorite and sericite schist derived from Gambier Group tuff and sediments and formed within complex northwest-trending shear zones which mostly dip to the southwest (Lynch, 1991; Monger, 1993).

***STOP 2-11:*** **210.3 km** Brief stop at pullout on left to examine deformed volcanogenic strata of the Gambier Group.

**COMMENT** A narrow 1 km wide sliver of deformed Gambier strata along strike to the northwest across Howe Sound is probably the continuation of the Gambier rocks at Britannia Beach. The northern contact of this sliver is intrusive, whereas the southern contact is a southwest-dipping (at about 75°) reverse fault, marked by mylonite, across which are granitic rocks of the "Porteau Pluton" discussed below. Fabrics in the granitic rock feature not only planar

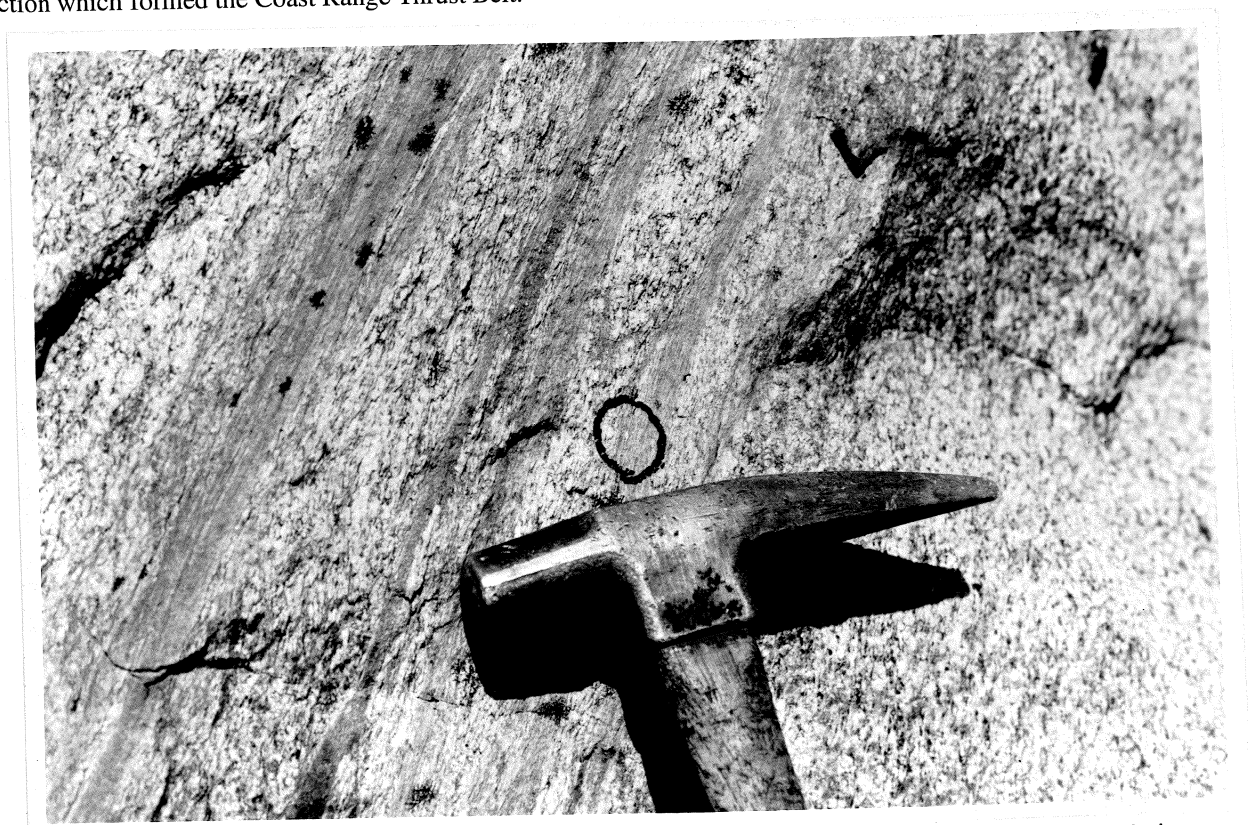


**Figure 49.** Northeast-vergent, southwest-dipping mylonite/sheared granitic rock; barely visible but circled asymmetric feldspar indicates left-side-up movement. From mid-Cretaceous granite exposed near summit of Mount Roderick (Fig. 46). This is the probable northwest extension of the Britannia shear zone, west of Howe Sound.

fabrics produced by aligned mafic minerals and elongate xenoliths flattened in the plane of foliation, which were presumably produced when the rock was being intruded, but also cataclastic fabrics with broken feldspars and mylonites with the same planar orientation. Similar southwest-dipping fabrics occur 8 km along strike to the northwest near the summit of Mount Roderick, entirely within plutonic rock, as the sliver of Gambier strata dies out (Fig. 49). The rocks there are U-Pb dated at 95.8±1 Ma (Monger and McNicoll, 1993). In addition, crosscutting northeast-dipping bands of ultramylonite are present within the granitic rock (Fig. 50; Monger, 1993).

Speculatively, these fabrics, and perhaps some fabrics along the Ashlu Creek shear zone, formed during emplacement and cooling of the Howe Sound Batholith, U-Pb ages of which range from ca. 103-96 Ma. These ages overlap with the time of movement on the Coast Belt Thrust System (97-91 Ma), showing that the batholith was emplaced in part, perhaps by injection, during a period of strong regional compression. The early intrusive fabrics in the batholith were overprinted by cataclastic, mylonitic fabrics and ultimately by ultramylonites as cooling continued.

South of a covered interval here, near the shoreline the contact between Gambier strata to the north and the Porteau Pluton to the south is exposed (Fig. 46). It appears to be a reverse-faulted intrusive contact, that dips to the southwest at 70° and is comparable to the same contact discussed above, along strike, across Howe Sound (Lynch, 1991; Monger, 1993).

**212.3 km** Northern exposures of massive well-jointed quartz diorite which are the part of the Howe Sound batholith called Porteau or Furry plutons. It yields a preliminary U-Pb date of 100 ± 2 Ma (R.L. Armstrong, in Irving et al., 1985). The pluton, exposed more-or-less continuously along the highway south of here, contains many dark inclusions, and locally has an east-dipping (at about 40°), locally cataclastic foliation, within which lie flattened dark xenoliths and amphibolitic layers.

**218.0 km** Porteau Cove. From this part of the Porteau quartz diorite, Symons (1973) and Irving et al. (1985) obtained paleomagnetic data that were interpreted as showing paleolatitudes that are anomalous with respect to mid-Cretaceous paleolatitudes for cratonic North America. Together with other data from coeval plutons (e.g., Spuzzum, visited on Day 1; Mount Stuart in the North Cascades), these data were used to suggest that the rocks moved northward about 2400 km and rotated clockwise since mid-Cretaceous time. As noted by Irving et al. (1985) and argued strongly by Butler et al. (1989), the absence of paleohorizontal markers in these rocks makes it possible that the anomalous paleomagnetic directions result from systematic southwestward tilting, rather than in northward displacements. The tilting of these granitic rocks in the southern Coast Belt may be extremely complex. The Spuzzum was at great depths (ca. 28 km) and then rapidly uplifted in early Late Cretaceous time (ca. 92-84 Ma), whereas the Porteau Pluton was probably emplaced at relatively high levels (±5 km?), yet both indicate a similar paleomagnetic direction. In addition, a complex transpressional (rotational?) history is recorded in the eastern Coast Belt. Note also that this part of the Coast Belt probably was tilted southwestwards

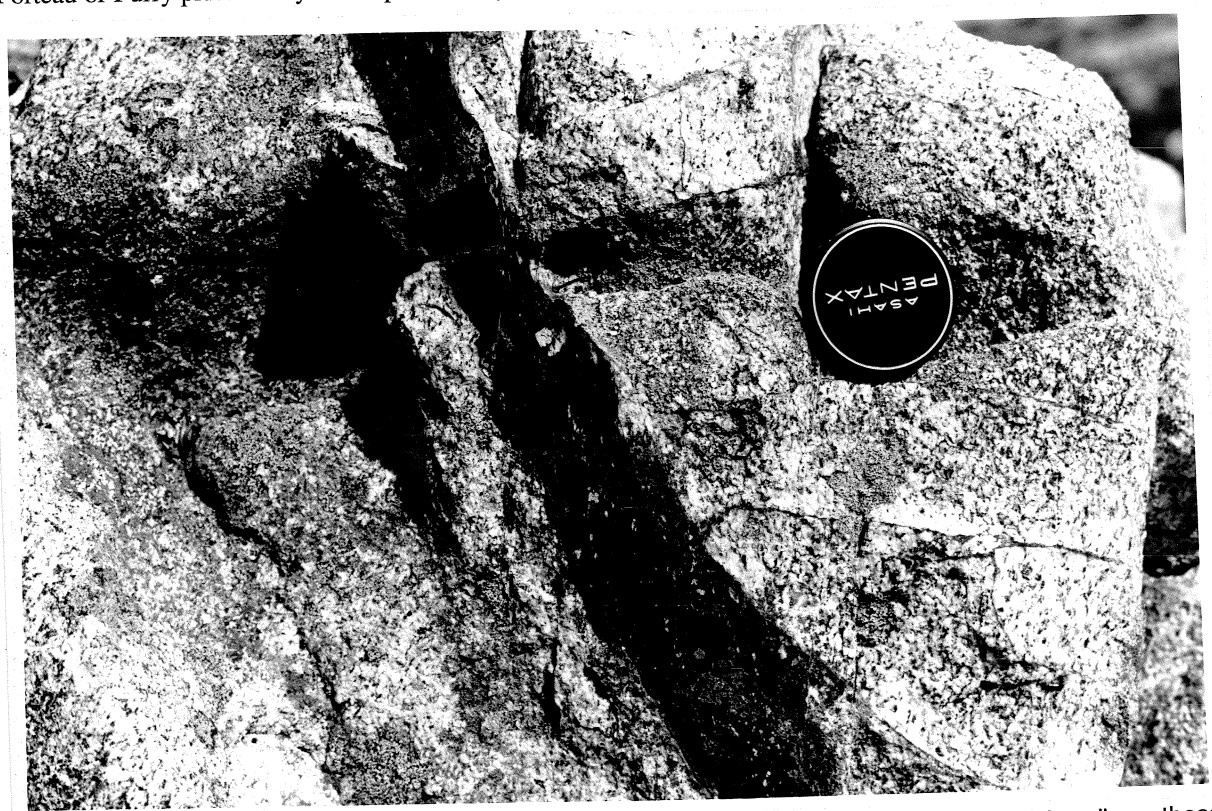



**Figure 50.** Strike-normal view of pseudotachylite (ultramylonite) dyke in quartz diorite; these dykes dip northeast and have southwest-vergence, in apparent conjugate relationship with mylonite zones in same area, shown in Figure 49. From mid-Cretaceous granite exposed near summit of Mount Roderick (Fig. 46), in northwest extension of the Britannia shear zone west of Howe Sound.

during Neogene uplift of 3-4 km of the axial part of the Coast Mountains (Fig. 10; Parrish, 1983).

**219.0 km** South of Porteau Cove, the pluton is in local fault contact (elsewhere intrusive contact) with well-bedded silicified argillites and sandstones of the Lower Cretaceous Gambier Group. Near Brunswick Point (4 km south of contact) the sedimentary rocks contain the ammonite *Cleoniceras (Grycia?) perezanium* of probable middle Albian (ca. 105 Ma) age (H.W. Tipper, pers. comm., 1989).

**220.0-230.7 Ma** More-or-less continuous road cuts in Gambier Group. The northern 5 km of road cuts are in thin-bedded, locally ripple marked, tuff and argillite gently folded into an open, west northwest-trending anticline. The southern 5 km comprises massive altered volcaniclastics including breccias and crytal lithic tuffs. The rocks are altered, probably because the contact with the pluton to the east runs roughly parallel with the slope of the eastern side of Howe Sound; the pluton is exposed higher up, about 2 km to the east.

**230.7 km** Enter the village of Lions Bay. Note the extensive (and expensive) engineering that has been undertaken to control debris torrents in creeks in this area.

To the west is Gambier Island, the northern two-thirds of which comprise Gambier Group, and the southern part is penetratively deformed Lower Jurassic Bowen Island Group, intruded by the granitic Thornborough Intrusions U-Pb dated at 156 ± 2 Ma (Friedman and Armstrong, 1990). The contact, long thought to be stratigraphic, is at least in part a northeast-side-down normal fault, which is probably the continuation of the normal fault to the northwest in Prince of Wales inlet (Fig. 1). It marks the southwestern boundary of Howe Sound tract. Exactly where the boundary crosses to the east side of southern Howe Sound is not clear. Gambier volcanics persist south of Lions Bay for about 3 km; the contact is probably near Newman Creek, about 3 km to the south of here.

**232.5-233.5 km** Altered Gambier volcanics.

**234.3 km** Cross Newman Creek; southernmost outcrop of Gambier Group. About 200 m south of this (on railway below) are outcrops of altered, green granodiorite of unknown age.

**235.6-237.3 km** Scattered outcrops of altered granitic rock.

**237.3 km** Cross Montizambert Creek.

**239.5-240.5 km** Outcrops of fine-grained amphibolite, gneiss and small granitic, dioritic and pegmatitic bodies, and local breccia with clasts of these lithologies in a plagioclase-mafic matrix. These rocks are described by Roddick (1965) and included in his Twin Islands Group. According to Roddick (1965) the protolith apparently was andesitic tuff; its age and stratigraphic affinity is unknown but speculatively could be Bowen Island Group; they clearly are of higher metamorphic grade than the Gambier Group to the north.

**241.1 km** Outcrop of granodiorite. This appears to be the western marginal phase of a granodioritic body that about 4 km east southeast of here (on the Cyprus Provincial Park road, elevation ca. 250 m) was U-Pb dated at 118 ± 2 Ma by Friedman and Armstrong (1990). Plutons of about this age are widespread in the westernmost part of the Coast Mountains northwest of here.

**242.5 km** Junction with turnoff to Horseshoe Bay and B.C. Ferries terminals. Bowen Island, west of here is largely composed of fine gained altered andesites(?), thinly laminated to massive white, grey, green and rarely pink, fine-grained siliceous tuff, and light and dark grey laminated chert and argillite (Friedman et al., 1990). The unit is U-Pb dated at about 185 Ma (zircon from felsic tuff from Mount Elphinstone, on the west side of southern Howe Sound), and by Early Jurassic fossils (probably Sinemurian; ca. 200 Ma) in septa of similar lithology west of Jervis Inlet.

The Penrose Conference field trip ended here, with the discussion part of the conference held on Bowen Island.

The return to Vancouver from Horseshoe Bay along the Trans-Canada Highway (continuation of Highway 99) is through granitic rocks that are probably Early Cretaceous, based on the Cypress Park U-Pb date quoted above.

To the southwest, across Georgia Strait, Vancouver Island and the Gulf Islands are visible. Lapping up on to older rocks of Vancouver Island (Wrangellia terrane) and locally on to the Coast Mountains on the east side of the Strait, and forming most of the Gulf Islands, are Late Cretaceous marine and nonmarine clastics of the Nanaimo Group. These appear to be foreland basin deposits or "molasse" derived mainly from uplift and erosion of the eastern Coast Mountains and the North Cascade orogen in Late Cretaceous time (Brandon et al., 1988; Mustard and Rouse, 1991), although (as noted in the introduction) England and Calon (1991) regard these rocks as deposits of a forearc basin.

Georgia Strait occupies a Neogene to Holocene structural depression, that is one of a series of depressions that extend from southernmost Alaska to southern Oregon (and beyond) and lie about 150 km east of the margin of the North American Plate.

To the east, the southern slopes of the Coast Mountains dip south at about 10°, a dip roughly conformable with the dip of the base of the overlying Late Cretaceous-early Tertiary molassic strata. The surface is presumably a Late Cretaceous erosion surface, upwarped in Late Miocene to Recent time (Roddick, 1965; Parrish, 1983). To the southeast are the North Cascade ranges, topped by Mount Baker.

Sandstones and shales of Late Cretaceous age form the southern abutment of Lions Gate Bridge (Mustard and Rouse, 1991, 1992), where they are intruded by mafic intrusions of Oligocene age (K-Ar whole rock dates of 31-34 Ma; R.L. Armstrong, pers. comm., 1980). About 1 km south of the abutment (projected along strike from exposures along the shore to the west) they are overlain by lithologically similar sandstones of Eocene age, which are slightly older than the Huntingdon strata seen on Day 1. To the south, these rocks underlie the city of Vancouver, south of Burrard Inlet. As noted at the beginning of the trip, they thicken southwards (beneath the Quaternary cover) towards the International Boundary, and are correlative (in part) with the Chuckanut Formation near Bellingham, Washington, which lies on a North Cascade basement.

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