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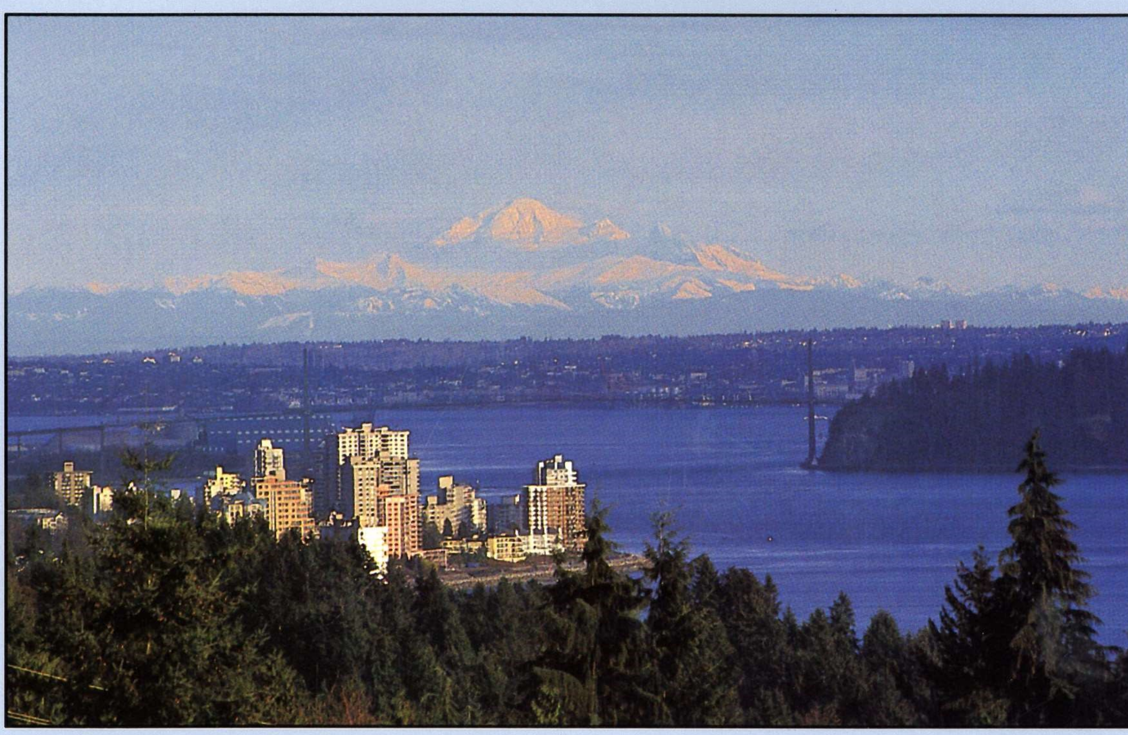
BULLETIN 481

GEOLOGY AND GEOLOGICAL HAZARDS OF THE VANCOUVER REGION, SOUTHWESTERN BRITISH COLUMBIA

Editor
J.W.H. Monger



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Cover description:

View from West Vancouver, from area underlain by Cretaceous plutonic rock at the southern end of the Coast Mountains, across south-dipping Late Cretaceous sandstones containing Oligocene mafic dykes underlying northern Stanley Park and exposed in the cliff at the south abutment of Lions Gate Bridge (near distance), across east Vancouver underlain by early Teritiary to Holocene strata (middle distance), to the North Cascade Range in Washington State capped by the recently active volcanic cone of Mount Baker (far distance). Photo courtesy of T. Turner, taken in March 1991, using 180 mm telephoto lens.

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TABLE OF CONTENTS

Contents

Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia

INTRODUCTION

Greater Vancouver regional district is home to about 1.7 million people in 1994, and projections postulate further rapid growth. This will increase demands on land for building, water for drinking and waste disposal, and sites for waste disposal, and escalate use of transport corridors whose locations are restricted by mountainous terrain. Like much of the Pacific Rim, the Vancouver region is located on an active tectonic plate margin, and because of this is subject periodically to earthquake and volcanic activity. In addition, as it is bordered by mountains and has high winter precipitation, the potential for landslides and debris flows is great in parts of the region. Knowledge of the background geoscience of the region is essential for planners and developers to make prudent use of the landscape and to attempt to avoid potential catastrophes. This volume is intended to provide an introduction to that background.

A barrier to ready acquisition of geoscience information on the region is that information is widely scattered in a variety of public and private documents, most of which deal with specific topics. This volume attempts to lower the barrier by summarizing existing information and synthesizing it with newer research results. Included papers present the latest views on their subjects; most contain extensive bibliographies. The first six papers elucidate the geology of bedrock and surfical deposits and show how the crust of the region formed and how the landscape evolved. The next four papers concern earthquakes, volcanism and landslides, and geological hazards associated with the Fraser River delta. The last two papers discuss groundwater and trace element distribution in the region.

The papers include results from research programs wholly or partly funded by the Geological Survey of Canada, but draw on ongoing university research, studies by provincial government agencies and geotechnical investigations commissioned by other agencies. Several papers (numbers 1, 2, 3, part 7, part 10) were funded initially by the Frontier Geoscience

INTRODUCTION

Le district de la région métropolitaine de Vancouver avait une population d'environ 1,7 million en 1994; selon les projections, la croissance démographique devrait continuer à être rapide. De ce fait, la demande de terres pour les constructions et la demande d'eau pour l'approvisionnement en eau potable et l'élimination des déchets s'intensifieront; il faudra créer davantage de sites de décharge et aussi fortement intensifier l'usage des voies de transport dont la mise en place est limitée par la topographie montagneuse. Comme une grande partie des pays côtiers du Pacifique, la région de Vancouver se situe sur une marge de plaque tectonique active et, de ce fait, elle est périodiquement sujette à une activité séismique et volcanique. En outre, comme elle est bordée par des montagnes et reçoit des précipitations hivernales abondantes, les risques de glissements de terrain et de coulées de débris sont élevés dans des parties de la région. Les planificateurs et les promoteurs doivent absolument connaître les détails géoscientifiques fondamentaux de la région pour prudemment tirer parti de la topographie et tenter de réduire les possibilités de catastrophes. Dans le présent volume, on se propose de donner une introduction concernant les détails fondamentaux.

Un obstacle à l'acquisition rapide d'information géoscientifique sur la région résulte du fait que cette information est très dispersée dans divers documents publics et privés, dont la plupart traitent de sujets spécifiques. Dans le présent volume, on cherche à réduire cet obstacle en résumant l'information existante et en faisant la synthèse de celle-ci avec les résultats de recherches plus récentes. Les opinions les plus récentes sur les thèmes abordés figurent dans les articles inclus, dont la plupart contiennent des bibliographies détaillées. Les six premiers articles expliquent la géologie du substratum rocheux et des dépôts de couverture et montrent comment s'est formée la croûte de la région et de quelle manière le paysage a évolué. Les quatre articles suivants portent sur les séismes, le volcanisme et les glissements de terrain, et sur les risques géologiques associés au delta du Fraser. Dans les deux derniers articles, on étudie la distribution des eaux souterraines et des éléments traces présents dans la région.

Les articles présentent les résultats de programmes de recherche entièrement ou partiellement financés par la Commission géologique du Canada, mais s'inspirent également de la recherche en cours effectuée par les universités, des études entreprises par les organismes des gouvernements provinciaux et des recherches

Program of the Geological Survey to assess hydrocarbon potential of Georgia Basin, the sedimentary basin beneath the Lower Mainland and Strait of Georgia. Information on the nature and structure of the deep crust of the region comes from the southern Canadian Cordillera Lithoprobe transect, funded by the Natural Sciences and Engineering Research Council, and part of Canada's largest geoscience project.

géotechniques faites à la demande d'autres organismes. Plusieurs articles (numéros 1, 2, 3, partie 7, partie 10) ont été financés initialement dans le cadre du Programme géoscientifique des régions pionnières, entrepris par la Commission géologique du Canada pour évaluer le potentiel en hydrocarbures du bassin de Georgia, bassin sédimentaire qui se trouve sous les basses terres du Fraser et le détroit de Georgia. L'information sur la nature et la structure de la croûte profonde de la région provient du transect sud de la Cordillère canadienne effectué dans le cadre du programme Lithoprobe, lequel est financé par le Conseil de recherches en sciences naturelles et en génie et fait partie du plus vaste projet géoscientifique au Canada.

J.W.H. Monger

Basement geology and tectonic evolution of the Vancouver region

J.W.H. Monger[1] and J.M. Journeay[1]

Monger, J.W.H. and Journeay, J.M., 1994: Basement geology and tectonic evolution of the Vancouver region; in Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia, (ed.) J.W.H. Monger; Geological Survey of Canada, Bulletin 481, p. 3-25.

Abstract: The Vancouver region is underlain by three different basements, which form the Coast Mountains to the north, the Vancouver Island Ranges to the west, and the Cascade Ranges to the southeast. These elevated areas are separated by Georgia Depression and Fraser Lowland, low regions in which the Late Cretaceous and younger sedimentary cover is preserved. The basements contain much of the record of the crustal evolution of the region, which can be divided into three stages. (1) During the pre-accretion stage (≥100 Ma), the crustal block forming Vancouver Island and southwestern Coast Mountains was in uncertain paleogeographic relationship to the North American plate margin, and separated from it by basinal terranes in southeastern Coast Mountains. (2) The syn- and postaccretion stage (100-40 Ma) was initiated when the western block was accreted to the plate margin, an event accompanied and followed by crustal thickening, uplift, and erosion centred in southeastern Coast Mountains. (3) During the last 40 million years, the continental Cascade magmatic arc formed on the North American plate margin above a subduction zone, the surface trace of which was, and is west of Vancouver Island. The regional physiography, probably formed in the last 10 Ma, may be related to stress distribution in the plate margin and thermal expansion in the magmatic arc; it was also influenced by the basements, as topographic depressions coincide with basement boundaries.

Résumé : La région de Vancouver comporte trois socles différents qui constituent la chaîne Côtière au nord, les chaînons de l'île de Vancouver à l'ouest, et la chaîne des Cascades au sud-est. Ces zones surélevées sont séparées par la dépression de Georgia et les basses terres du Fraser, régions basses dans lesquelles sont conservées la couverture sédimentaire du Crétacé tardif et une couverture sédimentaire plus récente. Les socles présentent une grande partie de l'histoire de l'évolution crustale de la région; cette évolution se laisse diviser en trois étapes : (1) Pendant l'étape antérieure à l'accrétion (≥100 Ma), le bloc crustal constituant l'île de Vancouver et le sud-ouest de la chaîne Côtière présentait un lien paléogéographique incertain avec la marge de la plaque nord-américaine et était séparé de cette plaque par des terranes formés de bassins dans le sud-est de la chaîne Côtière. (2) L'étape contemporaine de l'accrétion et postérieure à l'accrétion (100-40 Ma) a commencé lorsque le bloc occidental a été accrété à la marge de la plaque, événement accompagné et suivi de l'épaississement, du soulèvement et de l'érosion de la croûte centrés sur le sud-est de la chaîne Côtière. (3) Au cours des 40 derniers millions d'années, l'arc magmatique continental de la chaîne des Cascades s'est formé sur la marge de la plaque nord-américaine au-dessus d'une zone de subduction, dont la trace en surface se trouvait et se trouve encore à l'ouest de l'île de Vancouver. La physiographie régionale, probablement constituée au cours des dix derniers millions d'années, pourrait être associée à la distribution des contraintes dans la marge de la plaque et à l'expansion thermique dans l'arc magmatique; elle a également été influencée par les socles, puisque les dépressions topographiques coïncident avec les limites de ces socles.

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INTRODUCTION

The city of Vancouver is in a region with a complex geological evolution that is part of the tectonically active western North American plate margin. It is built upon Late Cretaceous-early Tertiary south-dipping sedimentary rocks that overlie Mesozoic plutonic rocks exposed in the Coast Mountains north of the city. West of Vancouver, the Strait of Georgia conceals the boundary between the Coast Mountains and the mountains of Vancouver Island. South of Vancouver, the broad valley of the lower Fraser River is underlain by glacial and fluvial deposits, southeast of which lie the Cascade Ranges. The mountainous regions north, west, and southeast of Vancouver each contains geology that differs from that of the other regions.

The rocks exposed in the Coast Mountains, Vancouver Island, and the Cascade Ranges are the three geological **basements** of the region and contain much of the record of the evolution of its crust. The basements are separated, and their boundaries mainly concealed, by a **cover** of mostly younger sedimentary rocks, which underlies part of the Strait of Georgia, the lowlands surrounding it, and the lower Fraser valley. The cover strata are discussed in this volume by Mustard (1994) and Mustard and Rouse (1994). This paper outlines the basement geology and the tectonic evolution of the crust of the region.

Previous work

A geological database extending back for over a century is available for the region surrounding Vancouver. Geological maps and syntheses with references to the early work are provided for Vancouver Island by Muller (1977a, b), for the Coast Mountains by Roddick (1965, 1983) and Roddick and Woodsworth (1975, 1977, 1979), and for the Cascade Ranges by Misch (1966) and Monger (1970, 1989). Earthquake monitoring on Vancouver Island began in 1898, and earlier geophysical studies are reviewed by Berry et al. (1971). A major effort to understand the nature and tectonic evolution of the crust of the region started in 1984, with integrated geophysical and geological studies undertaken during the Lithoprobe Southern Canadian Cordilleran transect (Yorath et al., 1985; Clowes et al., 1987, 1992; Hyndman et al., 1990; Cook et al., 1991; Varsek et al., 1993; Zelt et al., 1993; Friedman and Armstrong, 1990, in press). A large component of the Lithoprobe project is seismic reflection profiling. This results in images showing the disposition of reflectors in the Earth's crust and locally in the mantle, so that by tracing surface geological features such as faults into reflectors, interpretations of the deep structure of the crust can be made (Clowes et al., 1992). In 1989, the Geological Survey of Canada initiated new mapping in the southern Coast Mountains (e.g., Monger, 1993a; Journeay, 1993; Journeay and Friedman, 1993), and carried out sedimentological, stratigraphic, and palynological studies in the sedimentary cover (Mustard and Rouse, 1991, 1992, 1994; Mustard, 1994). This work, combined

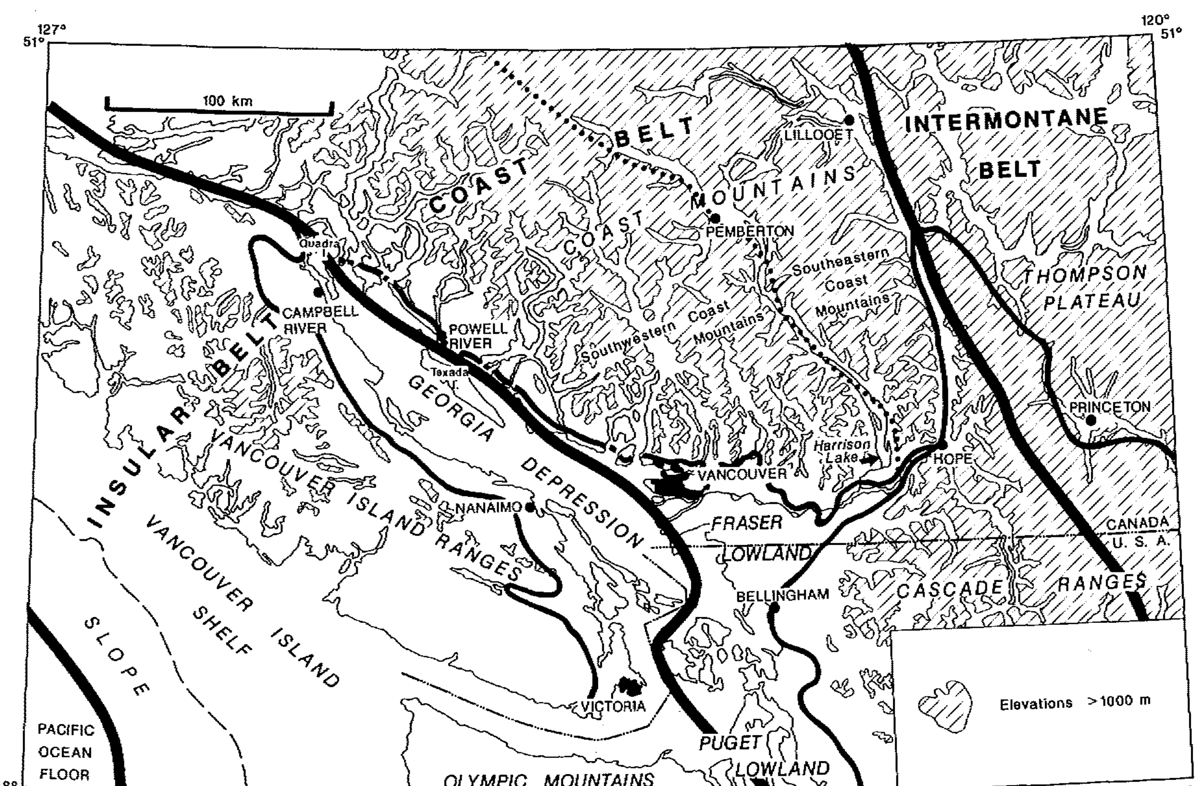

Figure 1. Morphogeological belts (separated by heavy solid lines), and physiographic subdivisions (lighter solid lines; names in italics) of southwestern British Columbia, adjacent parts of Washington State, and offshore regions. Heavy dotted line shows boundary between southeastern and southwestern parts of Coast Mountains.

with Lithoprobe results and ongoing studies in the Cascade Ranges of Washington State (Brown, 1987; Tabor et al., 1989), provides major new insights into the geology and evolution of the region containing Vancouver.

Terminology

Three sets of names are used to define and describe topographic and geological features in the region.

1. **Physiographic units** are areas of homogenous topography distinct from other areas with different physiography (Fig. 1; Mathews, 1986). The topographically high **Vancouver Island Ranges, Coast Mountains**, and **Cascade Ranges** are separated from one another by the low areas of **Georgia Depression** and **Fraser Lowland**. These terms are used herein for geographic reference.

2. **Morphogeological belts** are the traditional major divisions of the Canadian Cordillera. Each belt is defined by lithological, structural, tectonic, and physiographic attributes that together reflect the sum of geological processes that have occurred within it (Gabrielse et al., 1991). The region around the Strait of Georgia comprises the **Insular Belt** to the west and the **Coast Belt** to the east, with the boundary between the two near the eastern margin of the Strait (Fig. 1; Wheeler and McFeely, 1991). The Insular Belt is the continental margin; it includes the Vancouver Island Ranges and much of Georgia Depression, and extends to the base of the continental slope 100 km west of Vancouver Island, in water depths of 2000 m. It comprises sedimentary, volcanic, and plutonic rocks. The Coast Belt is a rugged high-relief region of mainly plutonic and metamorphic rocks which includes both the Coast Mountains and Cascade Ranges. East of the Coast Belt, the **Intermontane Belt** is a lower relief region of volcanic, plutonic, and sedimentary rocks.

3. **Tectonostratigraphic terranes** are areally extensive rock bodies with a geological record distinct from those of adjacent terranes, from which they are separated by faults, extensive plutons, and/or overlapping cover strata (Silberling et al., 1992; Monger, 1993b). The disparate stratigraphic, paleontological, and paleomagnetic records of some Cordilleran terranes make their paleogeographic relationships to one another, and to the North American continent, uncertain or "suspect" (Coney et al., 1980). Their lithologies and chemistry suggest that most western Cordilleran terranes probably originated as intra-oceanic volcanic arcs and ocean floor (e.g., Monger, 1977; Armstrong, 1988; Samson et al., 1989) and were accreted mainly in the Mesozoic to the ancient continental margin of North America.

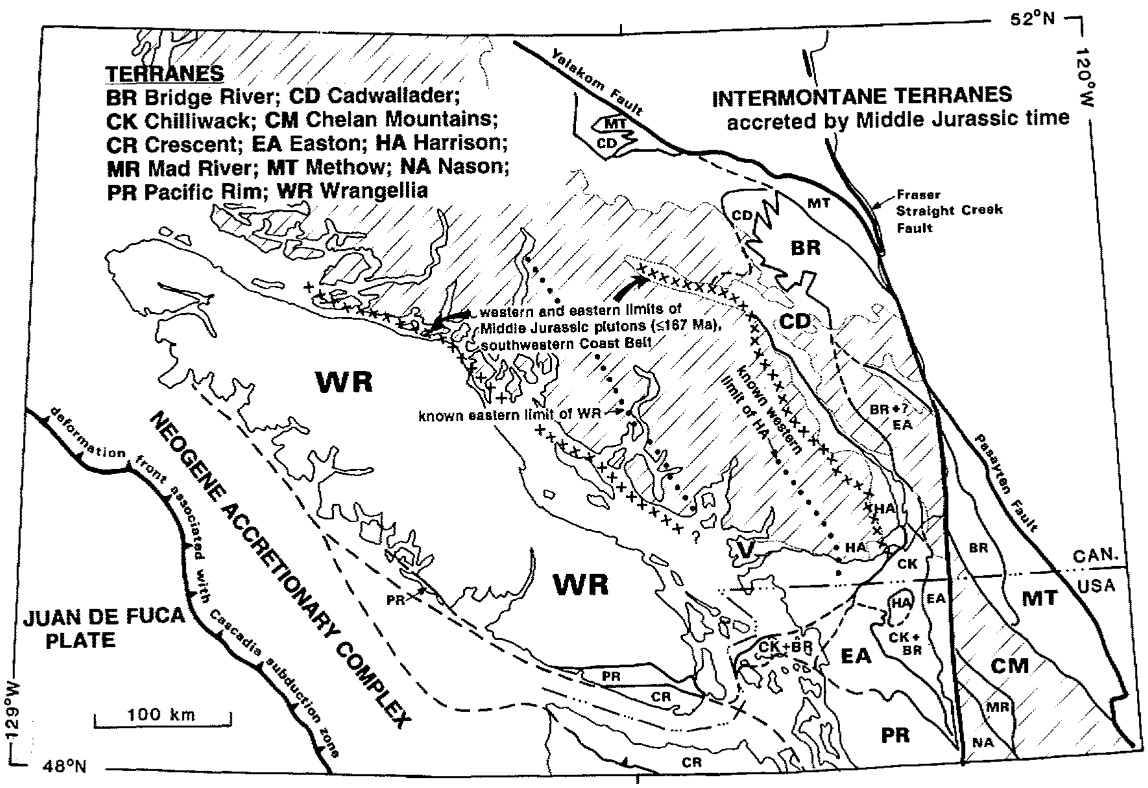

Figure 2. Distribution of terranes in southwestern British Columbia, adjacent parts of Washington State, and offshore regions. Yalakom and Pasayten faults on east, and Neogene accretionary complex on west, give the limits of new crust of mainly continental thickness formed since Early Cretaceous time. Hachured areas are dominated by granitic and high-grade metamorphic rocks, within which terranes occur as metamorphosed septa.

The latter lies in the eastern Cordillera (below the Rocky Mountain Trench near Golden; Cook et al., 1991), and formed in latest Precambrian time (ca. 750 Ma) by rifting and separation of a supercontinent (references in Monger, 1993b). The Intermontane Belt contains terranes accreted by late Early Jurassic (ca. 180 Ma) time to the ancient continental margin. The terranes in the Insular and Coast belts of southwestern British Columbia were accreted by the mid-Cretaceous (ca. 100 Ma), and include **Wrangellia terrane** of Vancouver Island and southwestern Coast Mountains, and **Harrison, Cadwallader, Bridge River, and Methow terranes** within Coast and Cascade mountains (Fig. 2, 3A, B). As terranes in this region are former intra-oceanic lithospheric fragments added to the ancient continental margin, they can be thought of as the building blocks of the crust of the western Cordillera.

The three basements, their relationships to one another, and the crustal evolution of the region are described herein. By the mid-Cretaceous (100 Ma) major terranes of the region were accreted, and in the Late Cretaceous-early Tertiary (100-40 Ma) were thickened, uplifted, and eroded in the main episode of crust building. Between 40 Ma and the present, the **Cascade magmatic arc** formed on the active North American plate margin above the subducting, oceanic **Juan de Fuca Plate** (Fig. 4).

BASEMENT ROCKS OF VANCOUVER ISLAND

Terranes

Vancouver Island and Strait of Georgia are underlain mainly by the extensive Paleozoic and lower Mesozoic terrane called **Wrangellia**, which is present also in the Queen Charlotte Islands and southern Alaska (Fig. 2, 3A, B, 5A). Wrangellian stratigraphy ranges in age from Devonian to Middle Jurassic, and the rocks are typically little metamorphosed. On Vancouver Island, the oldest rocks are Late Devonian (367 Ma) arc-related volcanic and sedimentary strata of the Sicker

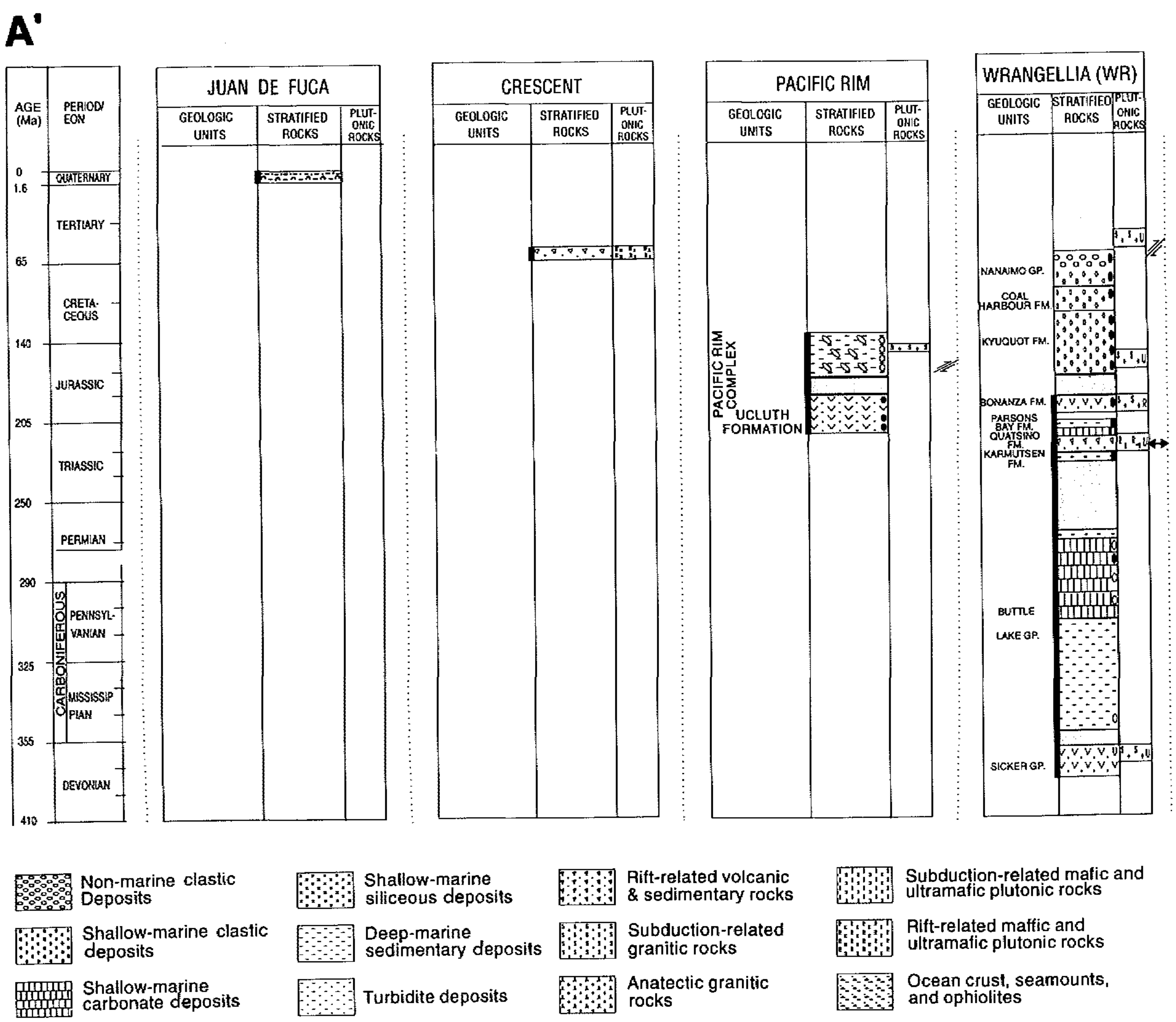

Figure 3A. *Schematic stratigraphic sections of terranes, with names of rock units in terranes and controls on age. Lithological symbols indicate tectonic settings within which the terranes formed.*

Group, and local intrusions associated with the volcanics (Yole, 1969; Muller, 1980; Massey and Friday, 1989; Parrish and McNicoll, 1992). They are overlain by Upper Carboniferous to Lower Permian clastic rocks and carbonates of the Buttle Lake Group, above which are local Middle Triassic black argillites. The most distinctive rock units of Wrangellia terrane are Middle(?) to Upper Triassic tholeiitic basalt of the Karmutsen Formation, which is up to 6 km thick, and overlying shallow-water carbonate of the Quatsino Formation, which grades upwards into Lower Jurassic deeper water clastics of the Parsons Bay and Harbledown formations (Jones et al., 1977). This early Mesozoic sequence may represent a large oceanic plateau, capped by sedimentary rocks deposited initially in shallow, then deeper water. The uppermost units characteristic of Wrangellia terrane are Lower Jurassic arc volcanics of the Bonanza Formation and associated Early Jurassic (176-189 Ma) plutonic rocks (Isachsen, 1984). Cumulative stratigraphic thickness of the Wrangellian succession probably exceeds 10 km.

In addition to Wrangellia terrane, narrow fault slices exposed in westernmost and southern Vancouver Island, constitute Pacific Rim and Crescent terranes (Fig. 2, 3A, B; Rusmore and Cowan, 1985). Pacific Rim terrane consists of Late Jurassic to Early Cretaceous disrupted clastic rocks,

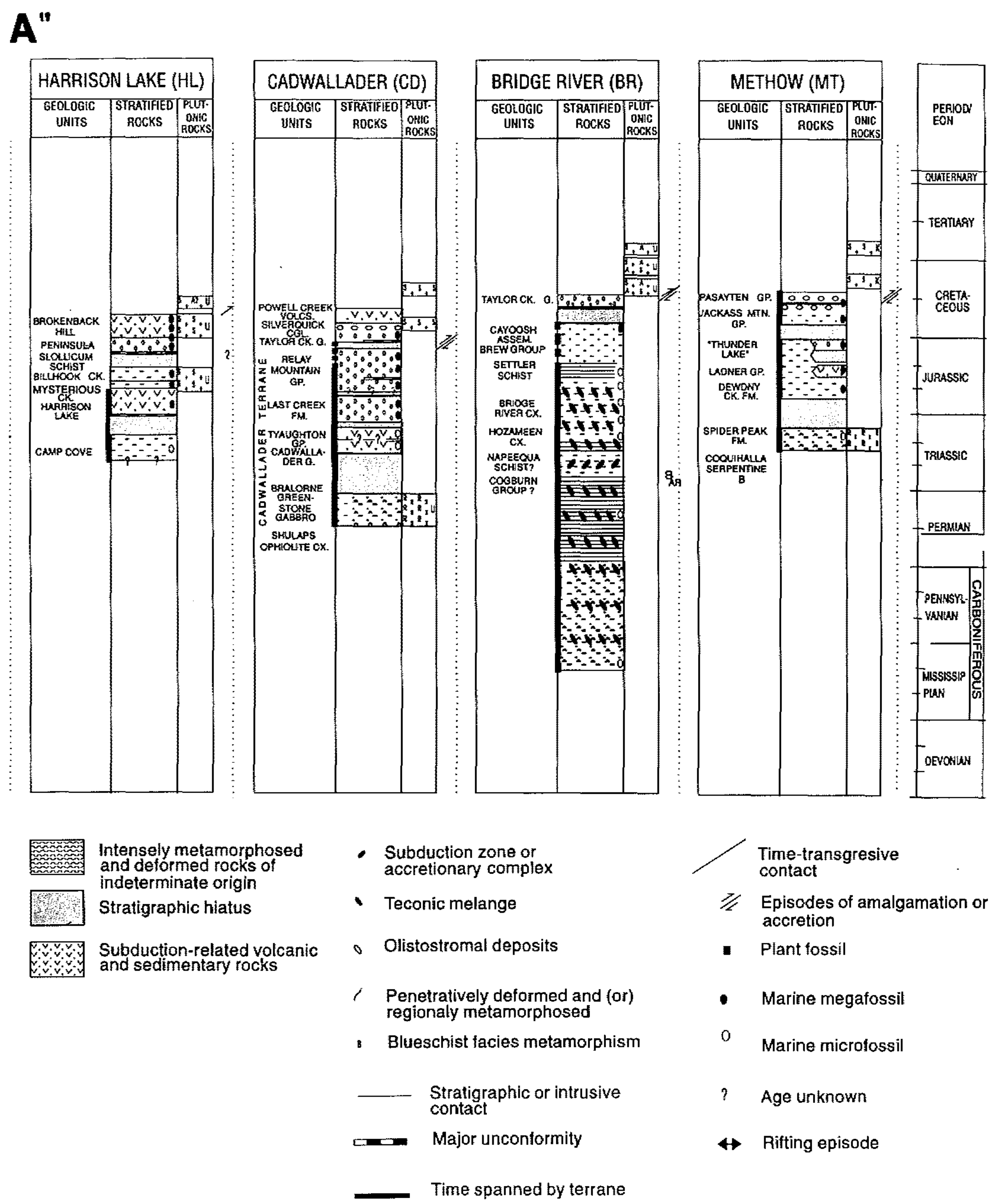

Figure 3A (cont.)

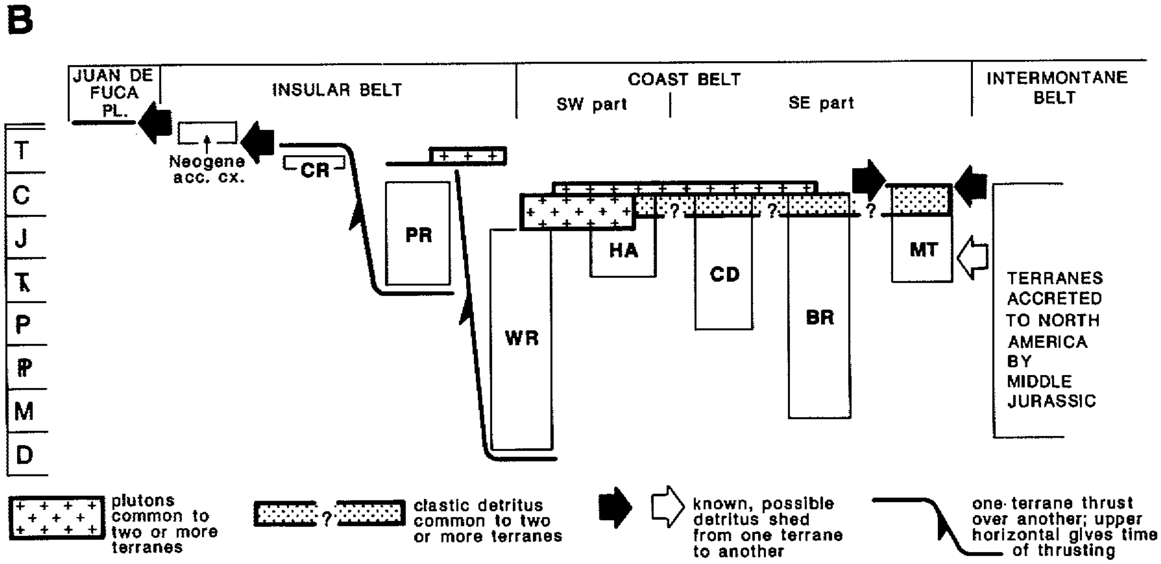

Figure 3B. Times and nature of linkages between terranes shown by overlapping strata, sedimentary detritus shed from one terrane to another, plutonic suites intruding two or more terranes, and dated structures juxtaposing terranes; terrane initials as in Figure 2.

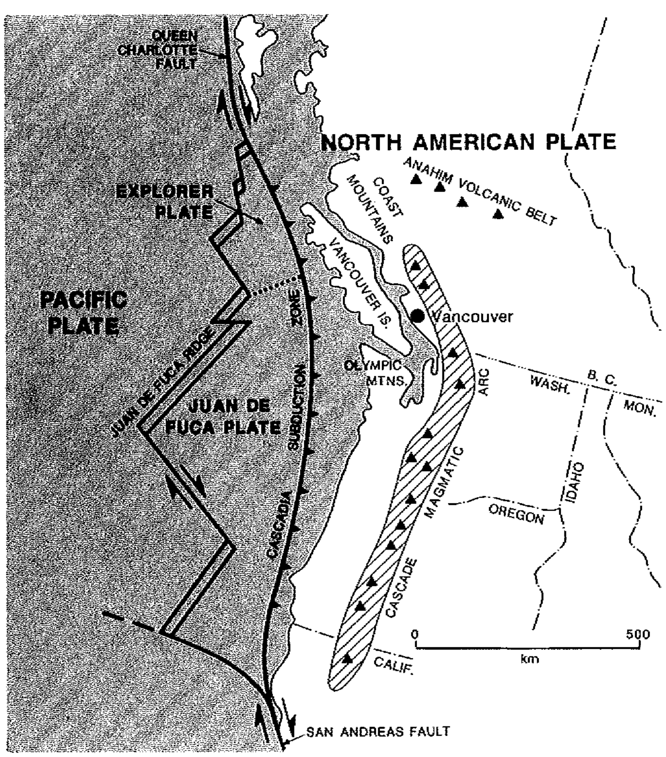

Figure 4. Present-day major tectonic features of parts of western North America, showing the Cascade magmatic arc (hachured) east of the small, subducting Juan de Fuca and Explorer plates (modified from Riddihough and Hyndman, 1991, Fig. 13.1).

chert and tuff, and early Mesozoic volcanic rocks. Muller (1977a) believed these rocks to be trench (or accretionary prism) deposits, but Brandon (1989) considered them most likely to be olistostromal mélanges produced by submarine slumping on to an older volcanic basement and displaced in latest Cretaceous-early Tertiary time from the westernmost Cascade Ranges. Crescent terrane comprises Eocene mafic volcanics and intrusions correlative with the Crescent Formation of the Olympic Peninsula. It possibly formed in a marginal basin (references in Hyndman et al., 1990). The submerged continental shelf and slope is locally underlain by these terranes, but mainly by the Neogene and presently active accretionary complex above the subducting Juan de Fuca oceanic plate (Fig. 2, 4; Hyndman et al., 1990).

Structure of Vancouver Island

Earlier mapping on Vancouver Island showed a seemingly random pattern of deformation, typically featuring steep faults with little vertical offset, and no penetrative deformation on a regional scale (Muller, 1977b). New mapping and Lithoprobe seismic reflection profiling across Vancouver Island and offshore regions shows that some northeast-dipping thrust faults mapped at surface align with reflectors at depth (Fig. 5A, B; Yorath et al., 1985; Clowes et al., 1987; Cook et al., 1991; Varsek et al., 1993). Pervasive northeast-dipping reflection fabrics are present in, from top to bottom, Wrangellia, Pacific Rim, and Crescent terranes, the Neogene to Recent accretionary prism, and the presently underthrusting/subducting oceanic Juan de Fuca Plate. Thus, Wrangellia terrane appears to be the uppermost sheet of a stack of southwest-vergent thrust sheets which forms the entire crust of Vancouver Island.

The stack is internally imbricated on northeast-dipping listric thrust faults that involve the Upper Cretaceous Nanaimo Group cover. In part, these structures form the Cowichan fold and thrust system of probable Eocene age, and may have originated during the time of accretion of Pacific Rim and Crescent terranes to Wrangellia terrane at ca. 50-40 Ma (Fig. 5; England and Calon, 1991; Mustard, 1994). In part they may be younger. Although Eocene deformation clearly is responsible for some of the northeast-dipping seismic fabric in Wrangellia terrane, structures in lower and younger (Neogene) rocks are presumably due to Neogene to Recent underthrusting of the Juan de Fuca Plate. In addition, Vancouver Island may be tilted gently eastward by underthrusting of the Juan de Fuca Plate, because originally deeply buried rocks called the West Coast Complex, which were the root of the Bonanza arc (Isachsen, 1984), are exposed on the west side of the island, and the shallowest rocks, the Nanaimo Group, which overlie the Wrangellian succession (Fig. 3A), occur widely on the east side of the island.

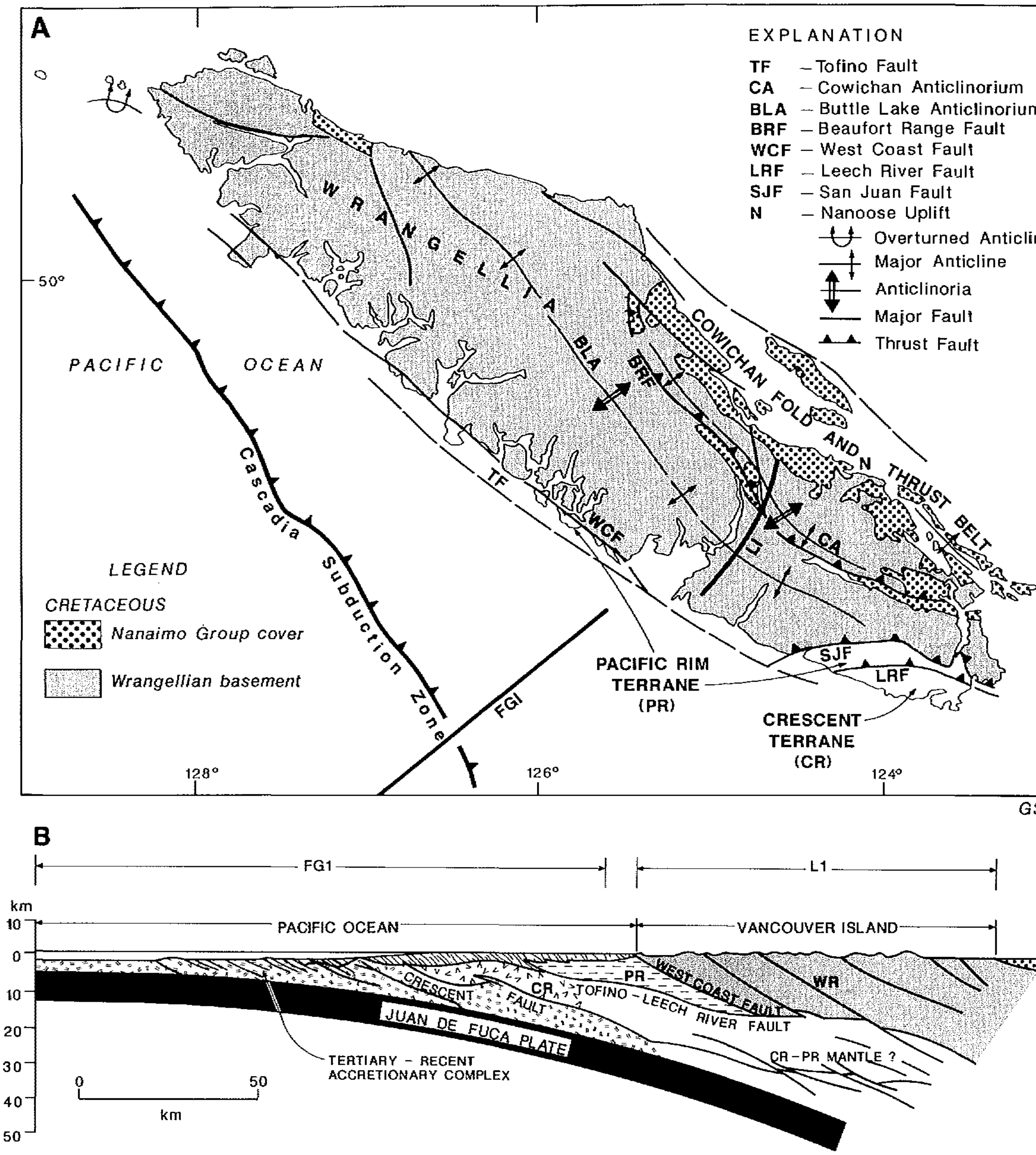

*Figure 5. A) Map of Vancouver Island showing distribution of terranes, cover of Late Cretaceous Nanaimo Group, major structural features, and location of seismic reflection lines used in constructing the crustal cross-section. **B)** Crustal cross-section, showing probable vertical distribution of terranes, faults, and the presently subducting oceanic plate (modified from Gabrielse, 1991, Fig. 17.6, 17.7).*

Relationships of Vancouver Island, Coast and Cascade basements

The eastern part of Wrangellia terrane is intruded by the suite of Middle and Late Jurassic (ca. 170-145 Ma) plutons which form a major component of southwestern Coast Mountains (Fig. 6; Friedman and Armstrong, 1990, in press). The boundary between Insular and Coast belts lies mainly beneath the Strait of Georgia, but is exposed on Quadra Island at the north end of the Strait (Fig. 1), and on West Thurlow and Hardwicke islands in Johnstone Strait to the northwest of it. There, Middle and Late Jurassic plutonic rocks of the westernmost Coast Mountains intrude and metamorphose Wrangellian rocks (Kunioshi and Liou, 1976; Nelson, 1979; Monger, 1991a; Monger and McNicoll, 1993). The evidence supports Nelson's (1979) contention that the western boundary of the Coast Belt at these latitudes is the western limit of Middle to Late Jurassic intrusions, that possibly were localized along pre- or synplutonic faults.

Farther south, surface evidence, in the form of exposed Wrangellian strata versus more-or-less continuous plutonic rock, favours a boundary towards the east side of the Strait of Georgia, between Mitlenatch and Hernando islands, and between Texada and Thormanby islands and the mainland (Fig. 1; Monger, 1991a). South of Vancouver, location of the boundary is unknown, but projects to lie below the delta of Fraser River.

Shallow seismic profiling in the Strait of Georgia provides evidence for local young faults, such as that east of Galiano Island (White and Clowes, 1984). These do not seem to have very large offsets on them, as northeast-dipping reflectors below eastern Vancouver Island (Fig. 5B) can be projected downdip to link up with similarly-dipping reflectors recorded to depths of about 50 km beneath westernmost Coast Mountains south of Powell River (Fig. 7; Cook et al., 1991; Varsek et al., 1993). Seismic refraction studies show the disposition of rock velocities within the Earth. The results of refraction profiling across Strait of Georgia suggested to White and Clowes (1984) that Coast Belt plutonic rocks extended westwards beneath the Strait of Georgia. However, as their profiles end at more than 5 km from the eastern side of the strait, close to the boundary of Wrangellia terrane and Coast Mountains located by surface exposures, it is possible that their profiles were entirely within Wrangellia terrane.

At the south end of the Strait of Georgia, a very different relationship from that described above exists between Wrangellia terrane and rocks in the San Juan Islands of Washington State (Fig. 8). There, the northwesternmost structures of the Cascade Ranges, called the San Juan-Cascades nappes, appear to have been thrust northwest in the early Late Cretaceous over Wrangellia terrane (Brandon et al., 1988). Erosion of the emergent nappes provided detritus to the

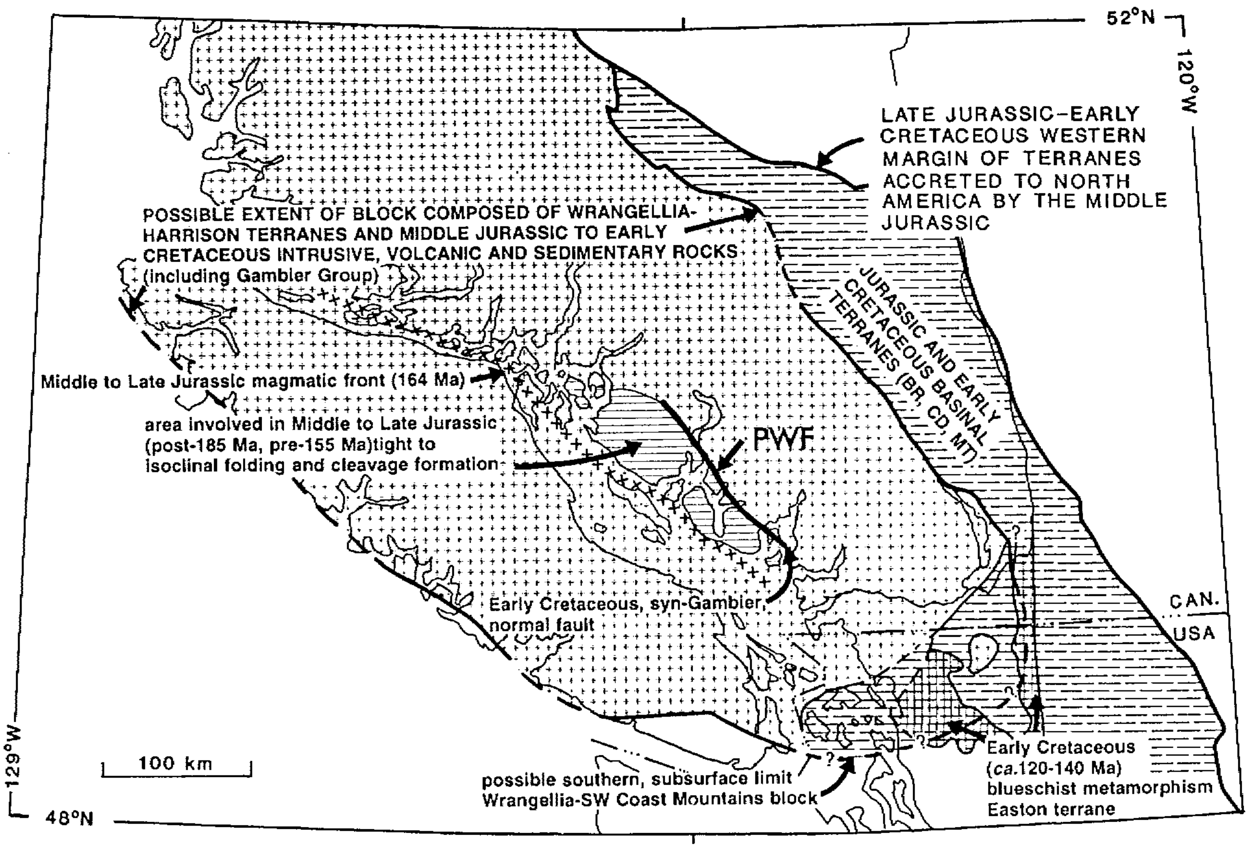

Figure 6. Distribution of pre-mid-Cretaceous (≥100 Ma) geological features in the region. These features existed prior to mid-Cretaceous (≥100 Ma) accretion of Wrangellia and southwestern Coast Belt to North America. PWF, Prince of Wales Fault.

Nanaimo Group (c.f., Mustard, 1994). The southward, along-strike extension of the Eocene Cowichan fold and thrust belt is not recognized in the San Juan Islands.

BASEMENT ROCKS OF THE COAST MOUNTAINS

The dominance of plutonic rocks in the Coast Mountains (ca. 80%) is reflected by the use of the names "Coast Range batholith" (LeRoy, 1908), "Coast Crystalline Belt" (Roddick, 1966), and "Coast Plutonic Complex" (Wheeler and Gabrielse, 1972), although the latter term is now restricted to plutonic and metamorphic rocks (Woodsworth et al., 1991). Except for Neogene-Quaternary volcanic rocks that belong to the Cascade magmatic arc, and Late Cretaceous-early Tertiary sedimentary cover rocks lapping onto its western and southern margins, stratified rocks in the Coast Mountains occur mainly as septa and fault slices surrounded by plutonic rocks.

The southern Coast Mountains consist of two parts, with the boundary between them marked approximately by the north-northwest-trending Harrison Lake-Pemberton valley (Fig. 1). This division, recognized by early workers (Crickmay, 1930; Misch, 1966, his Fig. 7-17), is confirmed by new mapping, U-Pb dating, and seismic reflection and refraction profiles (Friedman and Armstrong, 1990, in press; Journeay and Friedman, 1993; Monger, 1993a; Varsek et al., 1993; Zelt et al., 1993). At surface, southwestern Coast Mountains consist of Middle Jurassic to mid-Cretaceous (U-Pb dates of ca. 167-91 Ma) quartz diorite, granodiorite, and minor diorite, with minor septa and fault slices of Triassic and Jurassic strata belonging to Wrangellia and Harrison terranes and the overlapping Lower Cretaceous Gambier Group (Fig. 2, 3A, B). The septa are mainly of greenschist metamorphic facies. Southeastern Coast Mountains consist of mid-Cretaceous to early Tertiary (U-Pb dates 103-47 Ma) quartz diorite and granodiorite and upper Paleozoic and/or Triassic through Lower Cretaceous strata of Cadwallader, Bridge River, and Methow terranes, which are highly deformed and metamorphosed in places up to high grades (Fig. 2, 3A, B). Mid-Cretaceous (ca. 103-91 Ma) plutonic rocks are present in both parts.

The two-fold division recognized at the surface appears to extend to the base of the crust. The seismic refraction record shows generally higher velocity material in southwestern Coast Mountains than in southeastern Coast Mountains. Its interpretation is that Wrangellia terrane forms much of the lower and middle crust (between depths of 10-35 km) of southwestern Coast Belt, and ends within southeastern Coast

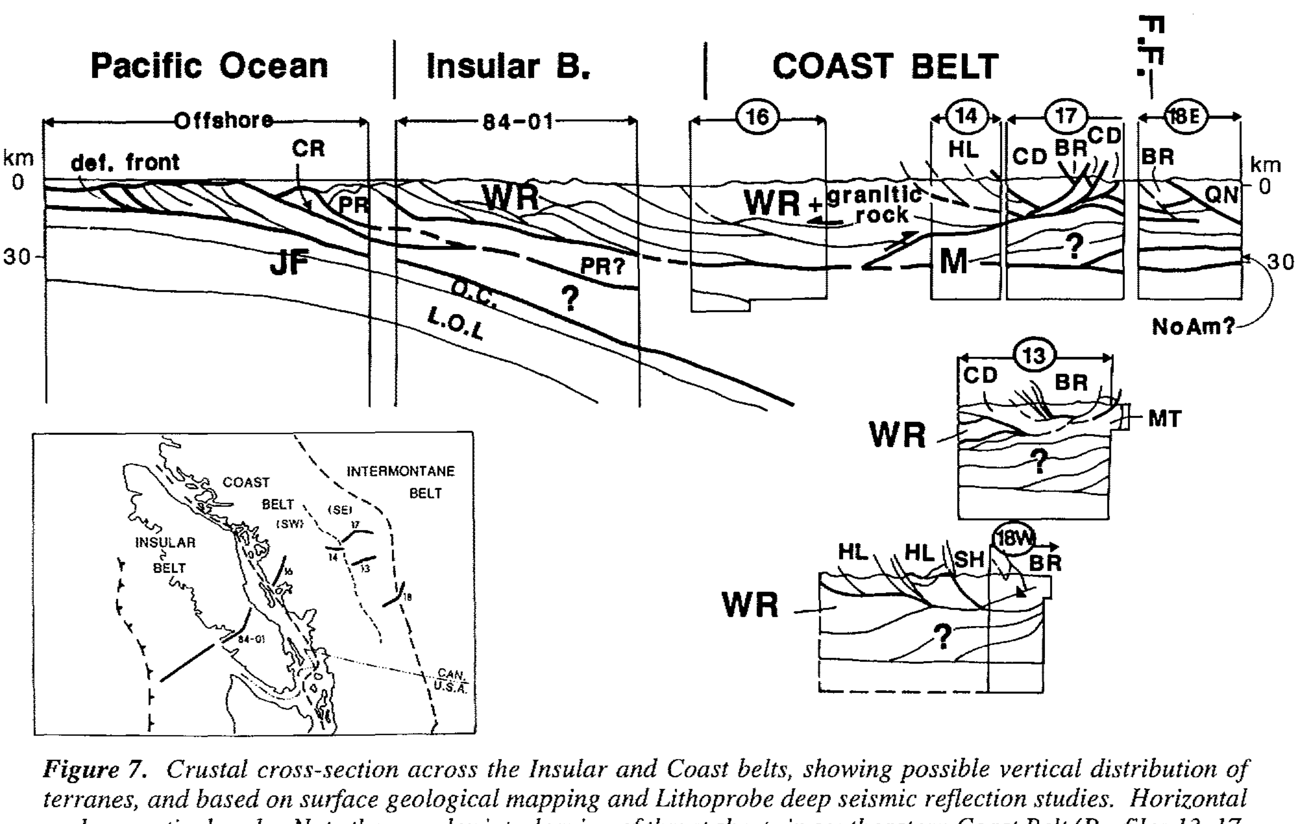

Figure 7. Crustal cross-section across the Insular and Coast belts, showing possible vertical distribution of terranes, and based on surface geological mapping and Lithoprobe deep seismic reflection studies. Horizontal scale = vertical scale. Note the complex interleaving of thrust sheets in southeastern Coast Belt (Profiles 13, 17, 18E). This section, drawn by Journeay, differs in detail from those of Cook et al. (1991) and Varsek et al. (1993). No granitic rocks are shown; terranes are lettered as in Figure 2, except WR + granitic rock denotes the dominant granitic rocks emplaced in Wrangellia terrane in western Coast Belt. M = reflection Moho, or base of crust; O.C. = oceanic crust; LOL = lower oceanic lithosphere, J.F. Juan de Fuca Plate. Inset map shows location of Lithoprobe reflection profiles used to construct the cross-section.

Belt (Zelt et al., 1993). Seismic reflection data indicate that the crust of Vancouver Island, derived from Wrangellia, Pacific Rim, and Crescent terranes, dips eastward and extends at depth into southwestern Coast Belt (Fig. 7; Cook et al., 1991; Varsek et al., 1993). In southeastern Coast Belt the seismic reflection fabric is complex. East-dipping reflectors, some of which can be traced up into west-vergent thrust faults at surface, appear to be cut and offset at depth by west-dipping reflectors (Fig. 7; Cook et al., 1991; Varsek et al., 1993). The fabric suggests that the crust of southeastern Coast Belt, which is ≥30 km thick, is a stack of interleaved thrust sheets derived from Wrangellia, Harrison, Cadwallader, Bridge River, Methow, and Intermontane terranes, as well as plutonic rocks.

Southwestern Coast Mountains: terranes, overlap units, plutons

Wrangellia terrane is represented by septa of metavolcanic and metasedimentary rock within Middle Jurassic to Early Cretaceous plutonic rocks, in the westernmost part of the Coast Mountains and on islands in the northern Strait of Georgia ((Fig. 2, 3A) Roddick and Woodsworth, 1977, 1979; Monger, 1991a). The septa comprise pillow basalt, amphibolite, and marble, which are correlated on the basis of lithology with Karmutsen and Quatsino formations, and schistose volcaniclastic and fine grained clastic rocks of the Bowen Island Group, which is dated as Early Jurassic from fossils and from a U-Pb date of 185 Ma on felsic volcanics (Friedman et al., 1990). The Bowen Island Group is probably a facies of Parsons Bay(?), Bonanza, and Harbledown formations on Vancouver Island (Fig. 3A).

Harrison terrane is exposed on the west side of Harrison Lake as the lower part of a Middle Triassic through Lower Cretaceous sequence mainly of subgreenschist metamorphic grade (Fig. 2, 3A; Crickmay, 1930; Arthur et al., 1993). The oldest rocks, Middle Triassic cherty argillites and mafic volcanics of the Camp Cove Formation, are overlain by thick Lower and Middle Jurassic intermediate to felsic flows and volcaniclastics of the Harrison Lake Formation. Toarcian basal conglomerate of the Harrison Lake Formation contains carbonate clasts with a fauna at least superficially similar to that in Permian limestone of the Chilliwack terrane in the western Cascade Ranges, although more work is needed to prove this correlation (Arthur et al., 1993). Harrison terrane is probably not part of Wrangellia terrane as the distinctive Triassic stratigraphy of Wrangellia terrane is absent, and the Harrison Lake Formation within it is generally younger than the Bonanza Formation of Vancouver Island.

Middle Jurassic (≤167 Ma) and younger plutons occur across the entire southwestern Coast Mountains and intrude Wrangellia and Harrison terranes, thus linking them by the Middle Jurassic (Fig. 3B; Friedman and Armstrong, 1990, in press; Monger, 1991a). The plutons are associated spatially and temporally with Lower Cretaceous and locally Middle to Upper Jurassic strata, the volcanic components of which may represent effusive equivalents of the plutons. Lower Cretaceous volcanic and sedimentary strata collectively are included in the Gambier Group (Roddick and Woodsworth, 1979; Monger, 1991a; Lynch, 1992). West of Harrison Lake, Middle and Upper Jurassic sedimentary Mysterious Creek and volcanic Billhook Creek formations, are overlain apparently disconformably by the Lower Cretaceous sedimentary Peninsula and volcanic Brokenback Hill formations (Fig. 3A; Arthur et al., 1993; Lynch, 1991). Elsewhere, shallow marine strata as old as Hauterivian (ca. 135 Ma) nonconformably overlie Jurassic plutons as young as ca. 145 Ma (Mathews, 1958; Monger, 1993a) indicating rapid uplift and erosion of the underlying plutons. Middle Jurassic and Early Cretaceous plutonic, volcanic, and sedimentary rocks represent a magmatic arc (or arcs) overlying and intruding Wrangellia and Harrison terranes.

The youngest plutons in southwestern Coast Belt range in age from 110-91 Ma and overlap in age with the oldest plutons (ca. 103 Ma) in southeastern Coast Belt (Friedman and Armstrong, 1990, in press; Monger and McNicoll, 1993). Rocks in southwestern Coast Belt as young as late Early Cretaceous appear to have formed a discrete crustal block that by the end of Early Cretaceous time interacted with terranes to the east (Fig. 3B).

Southeastern Coast Mountains: terranes, overlap units, plutons

Southeastern Coast Mountains contains, generally from southwest to northeast, Cadwallader, Bridge River, and Methow terranes (Fig. 2, 3A). All appear to have been founded on oceanic basements and many rocks in them were probably deposited in deep-water basins. They were strongly deformed and metamorphosed locally up to amphibolite grade in Late Cretaceous-early Tertiary time, and are least metamorphosed, generally subgreenschist and lowest greenschist facies, between the settlements of Pemberton and Lillooet.

Cadwallader terrane comprises a basement of oceanic arc tholeiite associated with gabbro, diorite, trondhjemite, and alpine-type ultramafic rocks of Permian age. This is overlain by Upper Triassic arc volcanics, carbonate, and distinctive carbonate-volcanic-plutonic clast conglomerates of the Hurley Formation, and (probably) by Lower and Middle Jurassic fine grained clastics of the Last Creek Formation (Rusmore, 1987; Leitch et al., 1991; Cordey and Schiarizza, 1993).

Bridge River terrane comprises the Bridge River Complex of highly disrupted Mississippian to late Middle Jurassic (ca. 340 to 160 Ma) radiolarian chert, argillite, and basalt, subordinate siltstone and greywacke, minor Late Triassic carbonate blocks, and associated alpine-type ultramafic and mafic intrusive rocks. It is mainly of subgreenschist metamorphic grade, although in places of greenschist and amphibolite grades, and contains rare Triassic blueschist (Potter, 1983; Monger, 1989; Schiarizza et al., 1990; Archibald et al., 1991; Journeay and Northcote, 1992; Cordey and Schiarizza, 1993). Its range of lithologies is characteristic of strata deposited in ocean basins.

The ocean was probably large because of the time (ca. 180 million years) spanned by pelagic sedimentary facies (radiolarian chert) within it. Its distinctive disrupted structural style, with mélange, broken formation, and local blueschist, suggests that initially it was deformed as an accretionary or subduction complex (analogous to the complex forming today west of Vancouver Island above the subducting Juan de Fuca oceanic plate), prior to overprinting by intense, Late Cretaceous-early Tertiary deformation and local high-grade metamorphism.

In places, chert of the Bridge River Complex is overlain gradationally by the Cayoosh assemblage, composed of fine grained clastic rocks with beds of distinctive and locally disrupted quartz-grain sandstone, conglomerate with some granitic clasts, and volcaniclastics. Cayoosh assemblage herein is included in Bridge River terrane, but may in part correlate with strata in Methow terrane (Fig. 3B; Journeay and Mahoney, 1994)). As the upper part of the Cayoosh assemblage is Early Cretaceous, and oldest parts may be early Mesozoic, and as there are Middle Jurassic cherts in the Bridge River Complex, the basal Cayoosh assemblage appears to be time-transgressive. It appears to represent younger, clastic, deposits of the Bridge River ocean (Mahoney and Journeay, 1993). The disrupted fabrics locally within it suggest that parts may have been involved in accretionary processes, like the Bridge River Complex. Lithologically similar but undated, high-grade metamorphic equivalents of the Bridge River Complex and Cayoosh assemblage occur east of Harrison Lake, and there are called, respectively, Cogburn Group and Settler Schist (Monger, 1989, 1991b; Journeay and Northcote, 1992).

Methow terrane in easternmost Coast Mountains, west of the Fraser-Straight Creek Fault, comprises predominantly fine clastic rocks of Jurassic-Cretaceous age, possibly correlative with the Cayoosh assemblage, and upper Lower Cretaceous (Aptian-Albian) greywacke, conglomerate, and shale. The most complete stratigraphic record of this terrane is in the eastern Cascade Ranges and is described in the section on the Cascades.

Quartz dioritic, granodioritic, and minor dioritic plutons in southeastern Coast Mountains form four north-northwest-trending, northeast-younging belts dated at (1) 103-92 Ma; (2) 86-84 Ma; (3) 75-64 Ma; and (4) 48-46 Ma (Friedman and Armstrong, 1990, in press; Coleman and Parrish, 1991; Journeay and Friedman, 1993). As noted earlier, the oldest, southwesternmost belt overlaps in age with the youngest plutonic rocks in southwestern Coast Mountains. Involvement of these plutons in structures of southeastern Coast Mountains closely documents the times of deformation in the southern Coast Belt (Journeay and Friedman, 1993).

Structure, metamorphism, and sedimentary linkages

Some structures in the Coast Mountains are older than the dominant Late Cretaceous-early Tertiary contractional deformation discussed below. In the southwestern Coast Mountains, the Lower Jurassic Bowen Island Group was penetratively deformed and locally isoclinally folded prior to intrusion of a pluton U-Pb dated at 154 Ma (Fig. 6; Monger, 1991a; Friedman and Armstrong, in press). Along Jervis Inlet, a 2 km wide dyke swarm may mark the syn-Gambier Prince of Wales northeast-side-down normal fault, which separates Middle Jurassic-Early Cretaceous plutons and Wrangellian strata to the west from mid-Cretaceous plutons and Lower Cretaceous Gambier strata to the east (Fig. 6; Monger, 1991a; Lynch, 1991). In southeastern Coast Belt, distinctive disrupted fabrics and local blueschist metamorphism of the Bridge River terrane, and disrupted fabrics in parts of the Cayoosh assemblage and correlatives to the south (Monger, 1991b), are typical of fabrics produced in accretionary complexes and predate Late Cretaceous-early Tertiary deformation.

The dominant structures of the southern Coast Mountains are of mid-Cretaceous to early Tertiary age (Fig. 8). They formed during and following mid-Cretaceous closure of the basin floored by Cadwallader, Bridge River, and Methow terranes in southeastern Coast Belt. Compression and transpression were concentrated in the southeastern Coast Mountains between the southwestern Coast Mountains-Wrangellia block and the Intermontane Belt.

In the southwestern Coast Mountains, deformation typically is concentrated along discrete, contractional shear zones that are north-northwest-trending, and dominantly west-southwest-vergent, although locally they have reversed sense (Monger, 1991a; Lynch, 1992). Rocks between the shear zones in tracts up to ≥10 km wide may be little deformed (Monger, 1991a, 1993a). Rocks as young as mid-Cretaceous (96 Ma; Monger and McNicholl, 1993) are crosscut by the shear zones. The Late Cretaceous Castle Towers pluton, U-Pb dated at 91 Ma, crosscuts the Thomas Lake shear zone, which features rocks as young as 94 Ma in its hanging wall (Fig. 8; Friedman and Armstrong, 1990, in press; Monger, 1991a; Journeay and Friedman, 1993). In places, there is a gradual increase of intensity of deformation to the northeast, but elsewhere, as near southern Harrison Lake, little deformed strata to the west are juxtaposed with schist to the east, across the dextral strike-slip Harrison Fault (Fig. 8; Monger, 1989; Journeay and Friedman, 1993).

By contrast, penetrative deformation on all scales is characteristic of the southeastern Coast Mountains. In the boundary zone between southeastern and southwestern Coast Mountains, initial post-Albian, southeast-vergent thrusting and folding is overprinted by southwest-vergent thrusting (Lynch, 1990, 1992). The latter is constrained by U-Pb dating to have occurred between 97-91 Ma (Journeay and Friedman, 1993; Journeay, 1993). Thrust faulting may have overlapped in time with dextral strike-slip deformation, as a Rb-Sr date of 93.5 ± 11.4 Ma is interpreted by Parrish and Monger (1992) to date movement on a splay of the Harrison Lake Fault. Farther north along the east side of the Coast Mountains, northeast-vergent thrusting appears to postdate southwest-vergent thrusting, but took place prior to 84 Ma (Rusmore and Woodsworth, 1991). Subsequent deformation involves both contraction and dextral displacement on such north-northwest-trending faults as the Downton and Yalakom faults, which are located near the northeastern margin of the

Coast Mountains near Lillooet (Fig. 8). This took place between 84 and 68 Ma, and may have continued into the early Tertiary. In addition, the easternmost Coast Mountains underwent extension associated with dextral strike-slip faulting on the Yalakom Fault in Eocene time (ca. 46 Ma, which locally exhumed mid-crustal rocks (Fig. 8, 9; Coleman and Parrish, 1991).

The relative importance and amounts of contraction versus transcurrent displacement in this part of the Coast Belt in mid-Cretaceous to early Tertiary time are strongly debated but undetermined (cf. Brown and Talbot, 1989; McGroder, 1991). In southeastern Coast Mountains, both contractional and dextral strike-slip faults of mid-Cretaceous to early Tertiary age are recognized (Fig. 8). Restorations of fragmented Early Cretaceous sedimentary basins in the eastern Coast Mountains and Cascades were used by Kleinspehn (1985) to estimate a total of ca. 260 km of post-Albian dextral offset. Estimates of latitudinal, south-to-north, displacements based on paleomagnetic data are an order of magnitude greater (ca. 2000-3000 km; Irving and Wynne, 1990; Ague and Brandon, 1992; Wynne et al., 1993). Some of these estimates are questioned by Butler et al. (1989) who suggest that the paleomagnetic data may record tilting, rather than latitudinal translation and rotation. Reconstructions of plate motions along the western edge of the North American Plate, deduced from magnetic anomaly patterns on the ocean floor and hot spot locations, agree reasonably with the structural evidence. The reconstructions suggest that strong orthogonal relative plate movements in mid-Cretaceous time were followed by dextral relative motions that started at about 85 Ma and continued into the early Tertiary (Engebretson et al., 1985; D.C. Engebretson and K. Kelley, pers. comm., 1993).

Metamorphism in the southern Coast Belt shows a wide variation. In the southwestern Coast Belt typically it is of greenschist facies, and probably largely due to emplacement of large masses of plutonic rock. The grade drops to subgreenschist facies near Harrison Lake, then rises rapidly through greenschist to kyanite-sillimanite grade east of the lake. Farther north, lower grade rocks are present along strike in the region between Lillooet and Pemberton. Pressure-temperature-time estimates on metamorphic rocks in the area between Harrison Lake and the Fraser River suggest that they were uplifted from depths of 20-30 km to 15-20 km between ca. 94 Ma and 84 Ma (Journeay, 1990). Uplift can be explained as the result of crustal thickening on opposing interleaved systems of thrust faults (Fig. 7; Varsek et al., 1993; Journeay and Friedman, 1993), although an alternative or additional mechanism proposed by Brown and Walker (1993) is that the crust was thickened by magma emplacement. Uplift of 5-15 km of the region of former basinal terranes in southeastern Coast

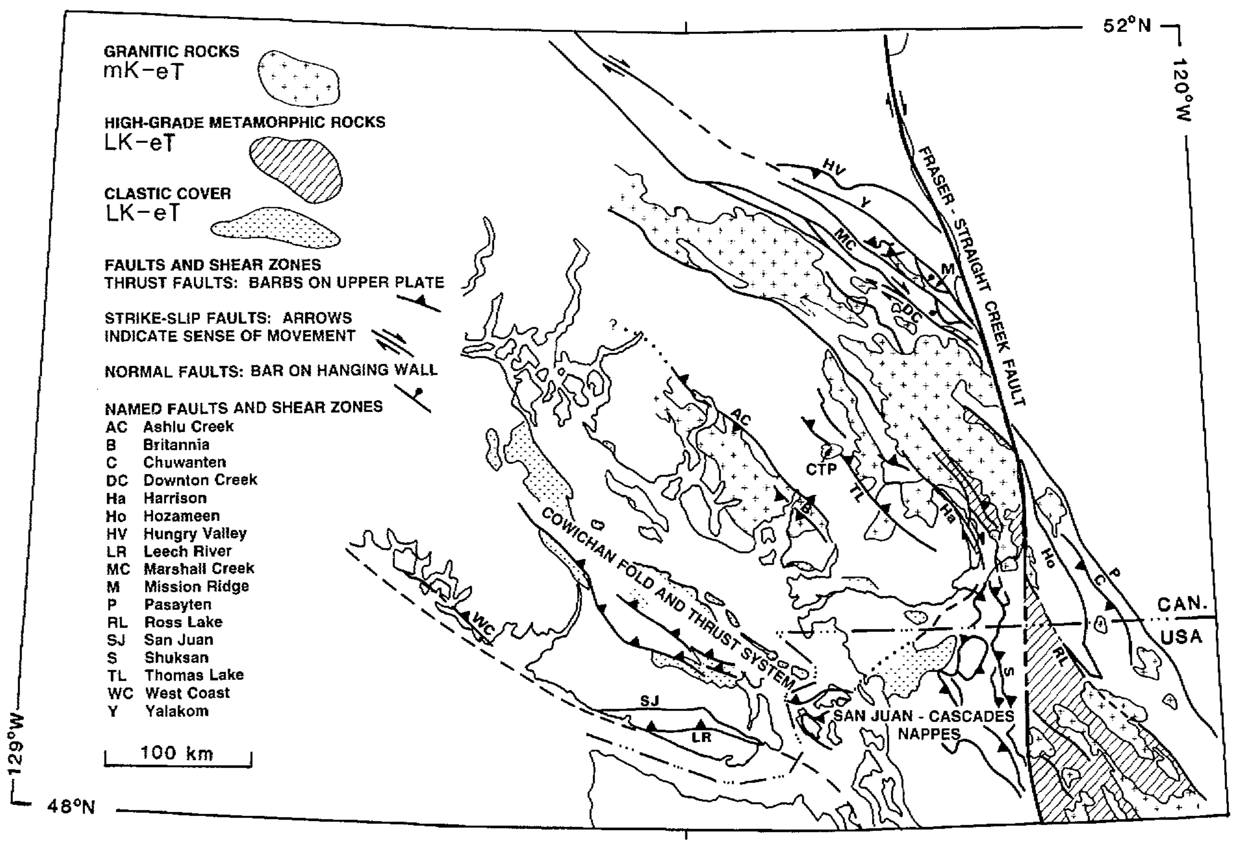

Figure 8. Distribution of mid-Cretaceous to early Tertiary (ca. 100-40 Ma) geological features in the region. These formed during the major period of deformation, metamorphism, and crustal thickening in the region; CTP Castle Towers Pluton.

Mountains in early Late Cretaceous time resulted in erosion. The region became a major source of clastic detritus for the Nanaimo Group (Mustard, 1994), whose oldest known basal deposits are early Turonian (ca. 90 Ma; Haggart, 1992).

By the mid-Cretaceous (Albian-Cenomanian; 112-95 Ma), sedimentary linkages can be made across the entire region from Insular to Intermontane belts (Fig. 3B; Garver, 1992; O'Brien et al., 1992). Earlier, the Cayoosh assemblage may have linked Cadwallader, Bridge River, and Methow terranes by Early Cretaceous time and possibly even before that (Fig. 3B; Journeay and Mahoney, 1994). However, the presence of pelagic facies as young as late Middle Jurassic (Callovian; ca. 160 Ma) in the Bridge River terrane suggests that up until that time, at least, there was either a very wide basin between the Wrangellia-southwestern Coast Mountains block to the west and the Intermontane Belt to the east, or else the western block was not present and open ocean existed to the west. Monger et al. (1994) argued that in the Middle Jurassic, the western block was part of the active North American plate margin, but was located farther north. The plate margin was cut by orogen-parallel sinistral faults and part was displaced southward by ≥800 km in the Early Cretaceous, trapping Bridge River, Methow, and Cadwallader terranes east of it. The intervening basinal areas were closed and inverted in the mid- and Late Cretaceous, a process accompanied by the deformation, metamorphism, intrusion, and regional uplift that was concentrated in southeastern Coast Belt.

Relationship of Coast and Cascade basements

The physiographic boundary between the Coast and Cascade mountains is the Fraser River (Mathews, 1986). North of the town of Hope this follows the trace of the north-trending Fraser-Straight Creek Fault, and west of Hope it follows the traces of northeast-trending Vedder and Sumas faults (Fig. 9, 10; Monger, 1989, 1991b). The faults are Tertiary, and superimposed on terranes, plutons, and structures common to both southeastern Coast and Cascade mountains.

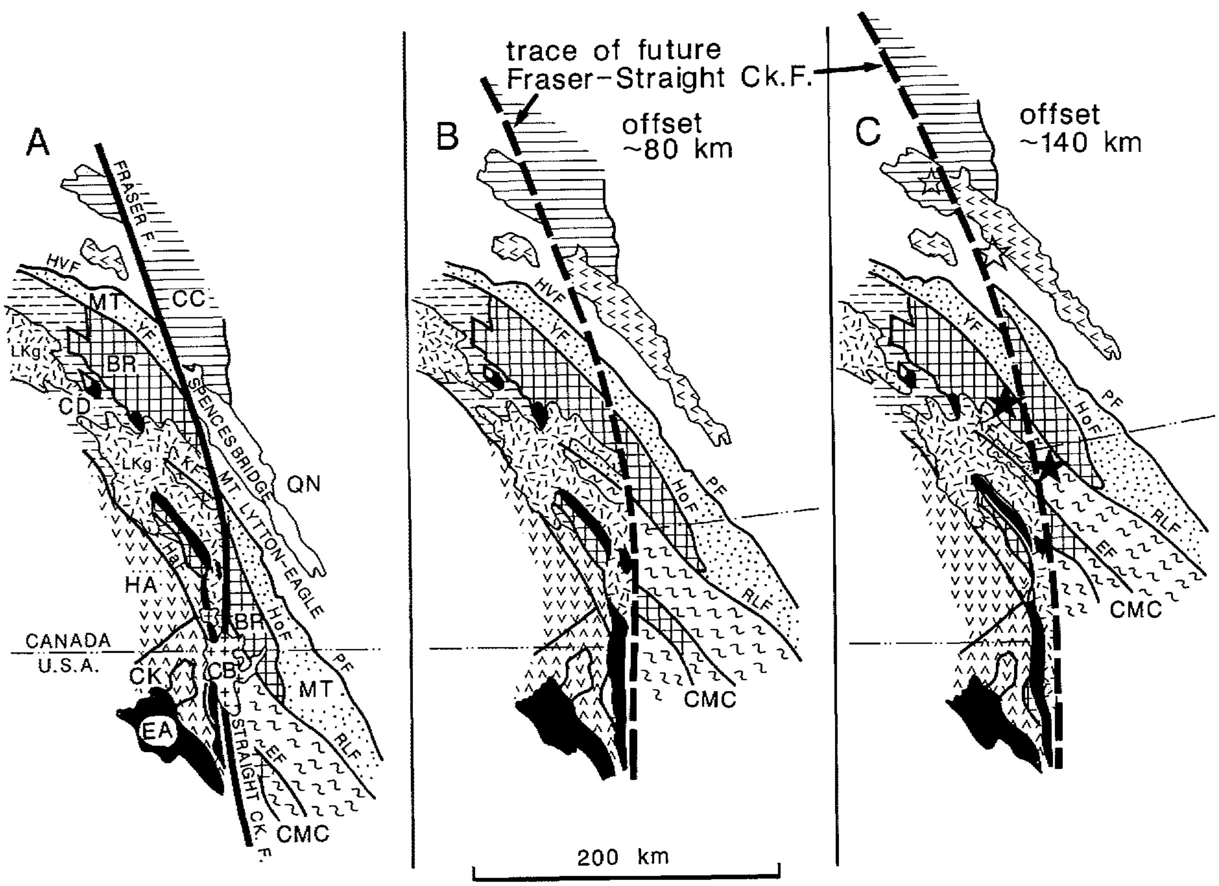

Figure 9. Fraser-Straight Creek Fault; A) Present map pattern; terrane abbreviations as in Figure 2, except that EA is within BR in Canada in Figure 2; CB – Chilliwack Batholith. Fault abbreviations: EF – Entiat Fault; HaF – Harrison Fault; HoF – Hozameen Fault; HvF – Hungry Valley Fault; KF – Kwoiek Fault; PF – Pasayten Fault; RLF – Ross Lake Fault; YF – Yalakom Fault. LKg – Late Cretaceous plutons; CMC – Cascade metamorphic core. ***B)*** *Restored dextral offset of 80 km on Fraser-Straight Creek Fault.* ***C)*** *Restored dextral offset of ca. 140 km on Fraser-Straight Creek Fault; black stars: offset early Tertiary granitic rocks, structures; open stars: offset Permian granitic rocks (Farwell Canyon on west; Mount Lytton on east).*

The **Fraser-Straight Creek Fault** (Fig. 7, 9) is a dextral strike-slip fault active between 46 Ma and 35 Ma. The youngest features cut and offset by the fault are Eocene intrusions and structures near Lillooet (Coleman and Parrish, 1991). The Chilliwack batholith, whose oldest date is 35 Ma, is emplaced across the fault (Fig. 9A). Estimates of displacement on the fault range between 80 (Fig. 9B) and 190 km (Misch, 1977; Mathews and Rouse, 1984; Kleinspehn, 1985; Monger in Price et al., 1985). Most recently, a restoration of about 140 km of displacement was proposed (Fig. 9C; Coleman and Parrish; 1991; Friedman and van der Heyden, 1992), which aligns Ross Lake and Yalakom faults, dated plutons, and Eocene extensional structures, and is favoured by the writers. However, in this restoration other features such as terranes do not align, and high-grade metamorphic rocks southwest of the Ross Lake Fault are juxtaposed with greenschist and subgreenschist rocks southwest of the Yalakom Fault. The differences can be explained by vertical offsets of several kilometres on the Fraser-Straight Creek Fault. Subsequent erosion laterally moved boundaries of terranes originally disposed in relatively thin thrust sheets.

The **Vedder and Sumas faults** (Fig. 10) are members of a family of northeast-trending faults that occur in the Coast and Cascade mountains and western parts of the Intermontane Belt (Fig. 10). They appear to be among the youngest structures in the region. Faults with this orientation offset Paleocene and Eocene strata on Sumas and Vedder mountains (Fig. 10; Vedder and Sumas faults; Mustard and Rouse, 1994). They form magma conduits for the 16 Ma Crevasse Crag volcanic complex (Fig. 10; Coish and Journeay, 1992), and may control locations of some hot springs (G.J. Woodsworth, pers. comm., 1994). The Coquihalla Fault (Fig. 10) appears to offset the 22 Ma Coquihalla volcanics.

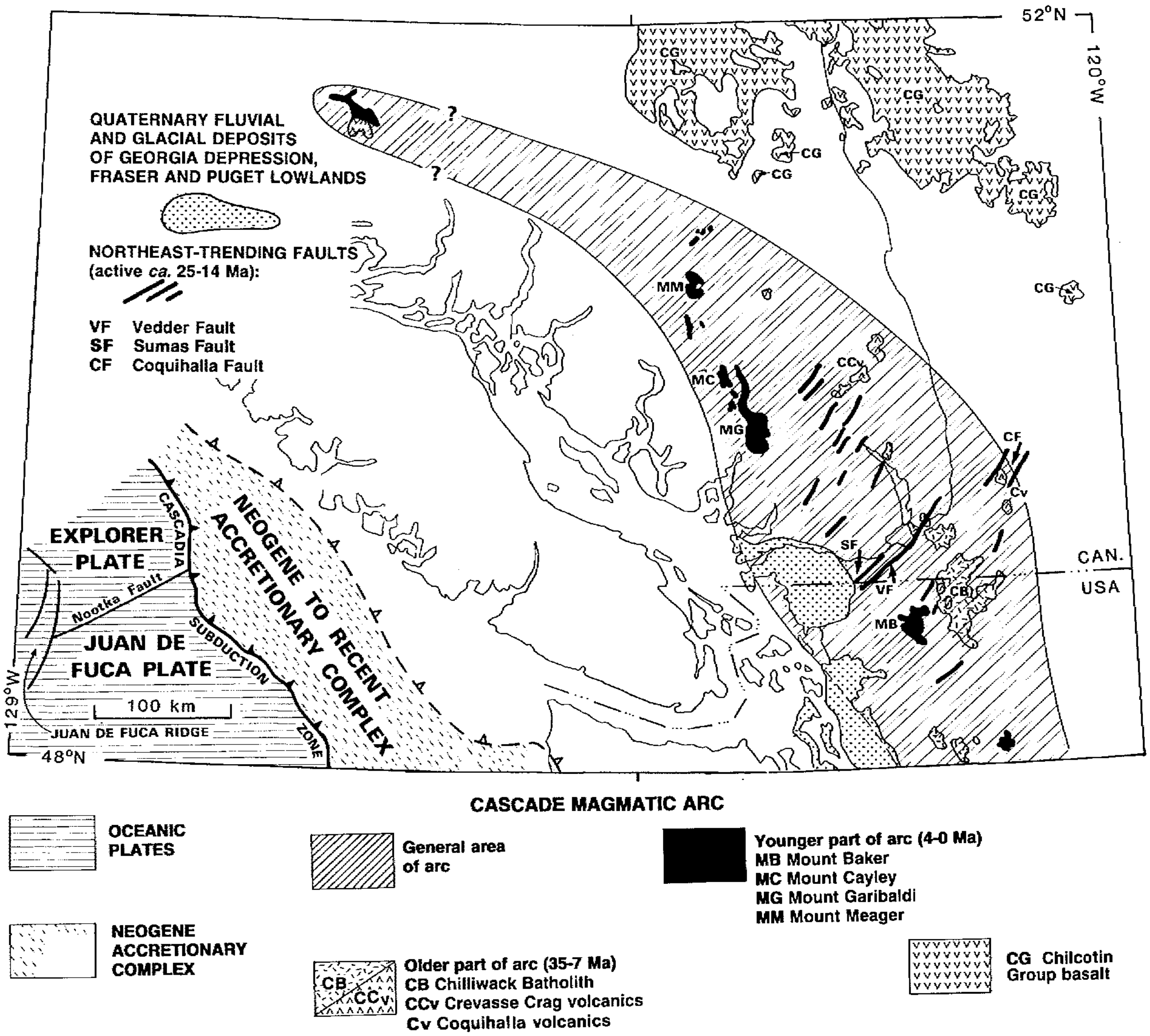

__Figure 10.__ Distribution of Tertiary to Recent features formed during the regime of the Cascade magmatic arc (40-0 Ma).

(Monger, 1989). These northeast-trending faults formed during earlier stages of the Cascade magmatic arc, discussed below.

West of Hope for a distance of 25 km, elements of southwestern Coast Mountains cross the Fraser River into the Cascades with no lateral but possible minor vertical offset (Monger, 1989). Farther west, the boundary between Coast and Cascade mountains is concealed below surficial deposits of Fraser Lowland, but must lie between Vedder Mountain, with geology typical of the Cascades, and Sumas Mountain, whose geology is typical of southwestern Coast Belt. As suggested by Misch (in Roddick, 1965, p. 31), the boundary appears to be fundamentally a Cretaceous thrust fault that is cut by the steeper and younger northeast-trending Vedder Fault.

BASEMENT ROCKS OF THE CASCADE RANGE

The three traditional divisions of the Cascade Range are: (1) the Northwest Cascades system; (2) the Cascade metamorphic core; and (3) the eastern Cascade Ranges (Misch, 1966; Tabor et al., 1989). All divisions are bounded by Tertiary faults (Fig. 9A). (1) On the west, mainly subgreenschist grade stratified rocks of Devonian to Cretaceous age form the **Northwest Cascades System**. According to Misch (1966), the rocks were deformed by a mid-Cretaceous arcuate system of west-vergent thrusts that is widest at about latitude 48°30′N. Misch's (1966) structural picture was questioned by Brown (1987), who recognized within it both low- and high-angle faults formed by coeval strike-slip and thrust movements. The northwest Cascade System is separated from rocks to the east by the Fraser-Straight Creek Fault. (2) In the centre of the Cascade Ranges, the **Cascade metamorphic core** contains greenschist to amphibolite-grade metamorphic rocks of Precambrian(?) through Mesozoic protolith ages, that were metamorphosed and intruded by plutonic rocks in Late Cretaceous and early Tertiary time (Tabor et al., 1987, 1989). These are separated from rocks to the east by the early Eocene north-northwest-trending Ross Lake Fault. (3) On the east, the **eastern Cascade Ranges** consist of mainly subgreenschist late Paleozoic to Cretaceous strata and minor early Tertiary plutons, deformed by Late Cretaceous east-vergent thrust faults and folds (McGroder, 1991). The eastern boundary of the Cascade Range is the Pasayten Fault, east of which is the Intermontane Belt.

Terranes and plutons of the Cascade Ranges

Four terranes can be identified in the **Northwest Cascades System** on the south side of Fraser Lowland and near the International Boundary (Misch, 1966; Tabor et al., 1989; Monger, 1991b). Structurally lowest and exposed only in a structural window north of Mount Baker, the Wells Creek volcanics and Nooksack Formation are correlated with **Harrison terrane** and overlying Late Jurassic and Early Cretaceous strata (Misch, 1966; Monger, 1991b; R.W. Tabor, written comm., 1993). With the exception of mid-Cretaceous plutons, these rocks are the only known equivalent in the Cascade Range of rocks in southwestern Coast Mountains. Structurally above these is a succession of rocks well exposed south of the Fraser River, near the Chilliwack valley in Canada, called **Chilliwack terrane** by Monger and Berg (1987) and Monger (1991b). Chilliwack terrane comprises Devonian to Permian arc volcanics, clastic rocks and limestones of the Chilliwack Group, stratigraphically overlain by Upper Triassic and Lower Jurassic fine grained clastics of the Cultus Formation, and(?) Upper Jurassic pelites (Monger, 1989). In Washington State, **Grandy Ridge terrane** (which includes Harrison terrane) has been used for correlative rocks (Tabor et al., 1989). Chilliwack terrane extends north of Fraser River and east of the south end of Harrison Lake as a narrow structural slice imbricated with the Jura-Cretaceous strata that overlie Harrison terrane (Crickmay, 1930; Monger, 1989). A possible Early Jurassic linkage between Harrison and Chilliwack terranes was mentioned earlier in the description of Harrison terrane, and correlatives of the Chilliwack Group probably are present in the San Juan Islands to the west. Structurally above and/or imbricated with Chilliwack terrane is a complex zone of Pennsylvanian to Jurassic chert, volcanic sandstone, and basalt called Elbow Lake Formation by Brown (1987). The Elbow Lake Formation is widely imbricated with alpine-type ultramafics, and locally with the metaplutonic Yellow Aster Complex (Misch, 1966; Tabor et al., 1989). Similar rocks can be traced north of the International Boundary through the Cheam Range, across Fraser River, into southeastern Coast Belt, where their possible metamorphosed equivalent is the Cogburn Group with its associated ultramafic and metaplutonic rocks (Monger, 1991b). In lithology and age range, the Elbow Lake Formation is similar to rocks of Bridge River terrane in Canada.

The structurally highest unit of the Northwest Cascades System is **Easton terrane** (Tabor et al., 1989) which comprises Darrington Phyllite and Shuksan Greenschist. Isotopic dating suggests the probable protolith age of these rocks is Jurassic, and they contain 140-120 Ma blueschists (Armstrong and Misch, 1987). Monger (1991b) argued on the bases of similar lithologies, structural position, and near continuity of outcrop areas, that Shuksan-Darrington rocks correlate with the high-grade metamorphic Settler Schist east of Harrison Lake. In turn, that is correlative with Cayoosh assemblage farther north. Thus, the Easton terrane and underlying Elbow Lake Formation may be equivalent of Bridge River terrane, including Cayoosh assemblage, in southeastern Coast Mountains.

Farther west, in the San Juan Islands, rocks of the Northwest Cascades System are complexly imbricated on mid-Cretaceous thrust faults and form the San Juan-Cascade nappes. Following thrust faulting and uplift, they were a major source of clastic sediments to the Nanaimo Group to the northwest (Brandon et al., 1988; Mustard, 1994).

In the westernmost foothills of the Cascades and some of the San Juan Islands, mélange terranes include Jura-Cretaceous clastics, serpentinite and metagabbro, and local greenstone (Fig. 8; Frizzell et al., 1987; Tabor et al., 1989). In part at least, these may be equivalent to rocks of **Pacific Rim terrane**, that structurally is below Wrangellia terrane. Although in the western Cascades such rocks are juxtaposed on faults with possible Bridge River equivalents, in Canada the two are separated by the Wrangellia-southwest Coast Mountains block.

Terranes in the Cascade metamorphic core close to the International Boundary contain probable equivalents of rocks in the Northwest Cascades system and in southeastern Coast Mountains. The **Chelan Mountains terrane** of Tabor et al. (1989) includes quartzite (= metachert?), quartz-rich mica schist, amphibolite and ultramafic pods (Napeequa Schist), and granitic clast conglomerate and other clastic and volcanic rocks (part of Cascade River schist) associated with Triassic metaplutonic rocks (Marblemount quartz diorite). Tabor et al. (1989) suggest that the clastic rocks overlie Napeequa schist. This relationship, and lithologies, exclusive of the Triassic granite, are not so different from the Bridge River and overlying Cayoosh assemblage in southeastern Coast Mountains, or, as suggested by Tabor et al. (1989) if the Triassic granite is included, a possible Bridge River-Cadwallader package. The **Nason terrane** includes Chiwaukum Schist, that long has been correlated with Settler Schist of southeastern Coast Belt (Evans and Berti, 1986), which in turn is correlated with Cayoosh assemblage.

Both Nason terrane and Chelan Mountains terrane were metamorphosed in greenschist to upper amphibolite facies in Late Cretaceous to Eocene time, and contain synkinematic plutons (Tabor et al., 1989). Towards the east side of the core, the migmatitic Skagit gneiss contains rocks U-Pb dated at 75-65 Ma. Small plutons within the Skagit Gneiss along the east margin of the core, yield Eocene (ca. 50 Ma) dates. The plutonic rocks have the same age range, and the same tendency to become younger eastwards, as the plutons in southeastern Coast Mountains.

Terranes in the eastern Cascades include Bridge River and Methow terranes. **Bridge River terrane** comprises the Hozameen Complex east of Fraser River (Monger, 1989), is lithologically identical to the Bridge River west of Fraser River, and contains fossils ranging in age from Pennsylvanian to (probably) Early Jurassic. **Methow terrane** contains Triassic basalt of the Spider Peak Formation, which is of mid-oceanic ridge basalt chemistry, and is associated with serpentinized ultramafic and gabbroic rocks of the Coquihalla serpentinite belt (Fig. 3A, B; Ray, 1986). This is overlain by Early to (?)Late Jurassic pelite, siltstone and tuff of the Boston Bar Formation of the Ladner Group. It laterally interfingers with Middle Jurassic (Aalenian-Bajocian; ca. 175 Ma) arc-related volcanics of the Dewdney Creek Formation of the Ladner Group, and Upper Jurassic (Oxfordian to Tithonian; ca. 153 Ma) volcanic sandstone and argillite of the Thunder Lake sequence of O'Brien (1987; O'Brien et al., 1992). Uppermost is thick (4500 m) Lower to lowermost Upper Cretaceous (Hauterivian to Turonian; ca. 135-90 Ma) marine and nonmarine clastic rocks of Jackass Mountain and Pasayten groups (Kleinspehn, 1985; McGroder, 1991). In late Early Cretaceous time, this terrane may have been the fore-arc of the continental arc represented by the Spences Bridge Group and coeval plutons built on Intermontane Belt terranes (Thorkelson and Smith, 1989). However, interpretation of new paleomagnetic data (Wynne et al., 1993) from strata probably correlative with those of youngest Methow terrane, suggested that these rocks were far to the south of the Spences Bridge Group and the two are not related.

Cascade structure

The classical picture of Cascade geology is of a western, west-vergent thrust system, a metamorphic core, and an eastern east-vergent thrust system (Misch, 1966). This is still true in part (McGroder, 1991), although there is abundant evidence for orogen-parallel displacement, particularly within the metamorphic core (Brown, 1987; Brown and Talbot, 1989). All of these structures were produced between mid-Cretaceous and Eocene time. In late Eocene time, the major Fraser-Straight Creek fault cut acutely across the north-northwest-trending grain of the orogen, truncating the core and displacing rocks on the west side of the fault northward to their present location within southeastern Coast Belt (Fig. 9).

SUMMARY: VANCOUVER ISLAND, COAST AND CASCADE BASEMENTS

Both Crickmay (1930) and Misch (1966, his Fig. 7-17) recognized that structures in the Cascade Ranges continue north into the southeastern Coast Mountains and Vancouver Island, bifurcating around southwestern Coast Mountains. More recent work supports their insights. Mid-Cretaceous through early Tertiary plutons and metamorphic rocks occur both in southeastern Coast Mountains and the Cascade Ranges along strike to the south (e.g., Tabor et al., 1989). Bridge River and Methow terranes and correlatives are present widely in the Cascades (e.g., Monger, 1991b). However, neither the voluminous pre-mid-Cretaceous plutonic rocks of southwestern Coast Mountains nor Wrangellia terrane are recognized in the Cascades. The sole representatives of southwestern Coast Mountains rocks in the Cascades are Jurassic-Cretaceous rocks correlated with Harrison terrane and overlying Gambier strata, exposed in the window near Mount Baker, and mid-Cretaceous plutons.

The nature of Bridge River terrane including at least part of the overlying Cayoosh assemblage, and the Cretaceous part of Methow terrane suggests that they represent, respectively, a Mesozoic accretionary complex and a Cretaceous forearc basin. If so, they delineate the Early Cretaceous North American Plate boundary. The accretionary-forearc complex was profoundly modified by mid-Cretaceous accretion of Wrangellia terrane (Davis et al., 1978, their Fig. 8) together with the Middle Jurassic and Early Cretaceous rocks of southwestern Coast Mountains (Monger et al., 1994). Following accretion, Late Cretaceous through early Tertiary compression, transpression, extension, metamorphism, and intrusion was centred in the former accretionary complex-forearc region. This became uplifted, eroded, and was a major source of detritus to the cover of Nanaimo Group (Mustard, 1994). Following accretion, the plate boundary moved to a position west of Vancouver Island, and Pacific Rim and Crescent terranes were accreted, possibly causing Eocene contraction on Vancouver Island.

THE TECTONIC SETTING OF THE LAST 40 MILLION YEARS

There was a major change of tectonic activity along this part of the plate margin ca. 40 million years ago, from the Late Cretaceous-early Tertiary dextral transpressional/transtensional tectonics, whose youngest expression may be the Fraser-Straight Creek Fault, to a convergent regime dominated on land by the **Cascade magmatic arc** (Fig. 4, 10). The change may reflect response of this part of the North American plate margin to replacement of the northward-moving, oceanic Kula Plate by the more orthogonally convergent Juan de Fuca Plate (Engebretson et al., 1985). Although products of this phase of the crustal evolution are volumetrically minor in comparison with those of earlier phases, associated ongoing tectonism is important as it includes earthquakes and volcanoes, which are potential major hazards in the region.

Today, the city of Vancouver lies well within the western margin of the **North American Plate**, which converges with, and overrides at rates of about 45 $mm \cdot a^{-1}$ (millimetres per year), the northern part of the oceanic **Juan de Fuca Plate** and the small **Explorer Plate** (Fig. 4, 10; Riddihough, 1984; Riddihough and Hyndman, 1991). The boundary along which the oceanic plate dives eastwards beneath the North American Plate is the **Cascadia subduction zone**, whose surface trace is at the base of the continental slope ≥100 km west of Vancouver Island in water depths of 2000 m. The distribution of earthquake hypocentres suggests that the subduction zone is located about at a depth of about 70 km below the city of Vancouver (Rogers and Horner, 1991, their Fig. 5).

Several features in the region demonstrate the active nature of this part of the plate margin. (1) Faults and folds are forming in the submerged **Neogene to recent accretionary complex**, which consists of sediments scraped off the descending oceanic plate (Fig. 4, 10; Davies and Hyndman, 1989; Hyndman et al., 1990). (2) Earthquakes, caused by strain release on faults, are located within both the upper, North American Plate at depths mostly above 25 km, and also within the upper part of the subducting Juan de Fuca Plate (Rogers, 1988; Duncan and Kulm, 1989; Rogers and Horner, 1991; Rogers, 1994). (3) Vancouver Island, Georgia Strait, and westernmost Coast Mountains are the forearc region of low crustal heat flow. There is a sharp transition, located near the heads of the Coast Mountain fiords, to a region of high heat flow, coincident with the dormant volcanoes of Mount Baker, Mount Cayley, Mount Garibaldi, and Mount Meager (Fig. 10; Lewis et al., 1992). These are the northernmost centres of the chain of active and dormant volcanoes called the **Cascade magmatic arc**, which extends southward into northern California and lies 200-300 km east of the Cascadia subduction zone (Fig. 4). Neogene (15-2 Ma) basalt of the Chilcotin Group (Bevier, 1983; Mathews, 1988) is in a backarc setting in the Intermontane Belt (Fig. 10). (4) Geodetic, long-term tidal monitoring, and gravity measurements show that the North American plate margin is deforming. The maximum horizontal strain, whose rate averages 0.25 $mm \cdot km^{-1} \cdot a^{-1}$ on Vancouver Island, is oriented east-west to northeast-southwest, in part parallel with the convergence vector between Juan de Fuca and North American plates. For vertical motion measurements, it is necessary to separate movement caused by tectonics from that caused by crustal rebound following melting of the continental ice sheet about 13 000 years ago (Riddihough, 1982). If this is done, measurements suggest that the floor of southern Strait of Georgia, which is a forearc depression, is subsiding at a rate of 1 $mm \cdot a^{-1}$. Parts of the Coast Mountains, including its coastline east of northern Vancouver Island, are rising at rates of between 3-4 $mm \cdot a^{-1}$ (Fig. 11; Dragert, 1987; Holdahl et al., 1989). Because of the high rates, the measured strain is probably short-term and dominated by movements related to stress build-up across a locked subduction zone above a young, buoyant oceanic plate (H. Dragert, pers. comm., 1993). Stress may be episodically released during major earthquakes.

Features similar to those of the present tectonic setting apparently existed in the region between latitudes 40°-51°N for about the last 40 million years. This is deduced from the longevity of the latest Eocene to Quaternary (ca. 37-0 Ma) Cascade magmatic arc, and from interpretations of Pacific basin plate configurations in this time interval, using magnetic anomaly isochrons and bottom topography (Engebretson et al., 1985; Atwater, 1989; Christiansen and Yeats, 1992).

However, within this interval there are differences whose origins are obscure. These are most clearly seen in the Vancouver region by the different natures and trends of older and younger parts of the Cascade arc (Fig. 10). The older (35-7 Ma) part of the arc has been called the Pemberton Volcanic Belt by Souther (1991), and comprises plutons such as the Chilliwack Batholith, and scattered remnants of volcanic centres. These rocks were uplifted and deeply eroded in places before effusion of pre-, syn-, and postglacial Pleistocene and Holocene (3.8 Ma to 1340 BP) volcanic rocks (Fig. 10) of the Garibaldi Volcanic Belt of Souther (1991). The northeast-trending fault system (Fig. 10), noted earlier, was apparently active during the 25-14 Ma interval, and probably controlled emplacement of the older arc volcanic and plutonic rocks (Coish and Journeay, 1992).

The configuration of this part of the plate margin probably changed in the Neogene (≤20 Ma). Brandon and Calderwood (1990) and Walcott (1993) discuss changes of about this age in the region of the Olympic Mountains, where the Neogene accretionary complex is exposed on land. Walcott (1993) suggests that Neogene extension in Nevada, and consequent northwestward movement of northern California caused shortening, crustal thickening, and uplift of rocks in the Olympic Peninsula as they were compressed against the salient of Vancouver Island.

Finally, there are data and speculations to account for the present distribution of elevated and depressed areas in the region. Phase changes in the descending oceanic slab (Rogers, 1983) are invoked as a cause of the forearc subsidence forming Georgia Depression. In addition, or alternatively, the lithospheric flexure model formulated by Yorath and Hyndman (1983) to explain the origin of the Queen Charlotte Basin may apply to Georgia Depression. Vancouver Island appears to be tilted eastward by underthrusting of the young, buoyant Juan de

Fuca Plate, because the deepest rocks, the West Coast Complex, are exposed on the west side and the youngest rocks, the Nanaimo Group, are on the east side. Several features show the Coast Mountains were elevated in the Neogene. There is a gradual increase in elevation of Late Miocene Chilcotin Group lavas (Fig. 10) from about 1000 m in the Intermontane Belt to about 2500 m at the eastern front of Coast Belt (Mathews, 1991). Different pre- and post-late Miocene floras on the east side of the Coast Mountains indicate a transition from a wetter to a dryer climate as a rain shadow was formed by the uplifting Coast Mountains (Mathews, 1991; and references therein). Parrish (1983) combined the geological reasoning used above with fission track dates on zircon and apatite collected at high and low altitudes in the mountains (Fig. 11) in an attempt to quantify the amount of uplift. He suggested that the southern Coast Mountains underwent very slow apparent uplift at rates <0.1 km•Ma^{-1} between 30-10 Ma, and far greater rates of >0.6 km•Ma^{-1} between 10-0 Ma. He concluded that total uplift in the last 5 million years of the southern Coast Mountains was more than 3 km over a wide area, and suggested that this might be due to thermal expansion in the high heat flow region of the Coast Mountains. Finally, basement features apparently influence physiography, as depressed areas coincide with the boundaries of the three basements. Georgia Depression follows the Middle Jurassic magmatic front of the Coast Belt. Fraser Lowland is close to the southern limit of the Wrangellian-southwest Coast Mountains block; its orientation is at least partly controlled by northeast-trending Neogene faults.

CONCLUSION: CRUSTAL EVOLUTION OF THE VANCOUVER REGION

The crust of the Vancouver region evolved through three tectonic stages (Fig. 12A-C).

1. **Pre-accretion stage (≥100 Ma)** Before the mid-Cretaceous, the North American plate margin probably lay within the eastern Coast Mountains and Cascade Ranges, which are located on the site of a long-lived accretionary complex represented by the Bridge River terrane (Fig. 12A). The paleogeographic relationship between the crustal block composed of Wrangellia terrane and rocks of southwestern Coast Mountains, upon which Vancouver is built, and the Cretaceous plate margin is uncertain for this stage.

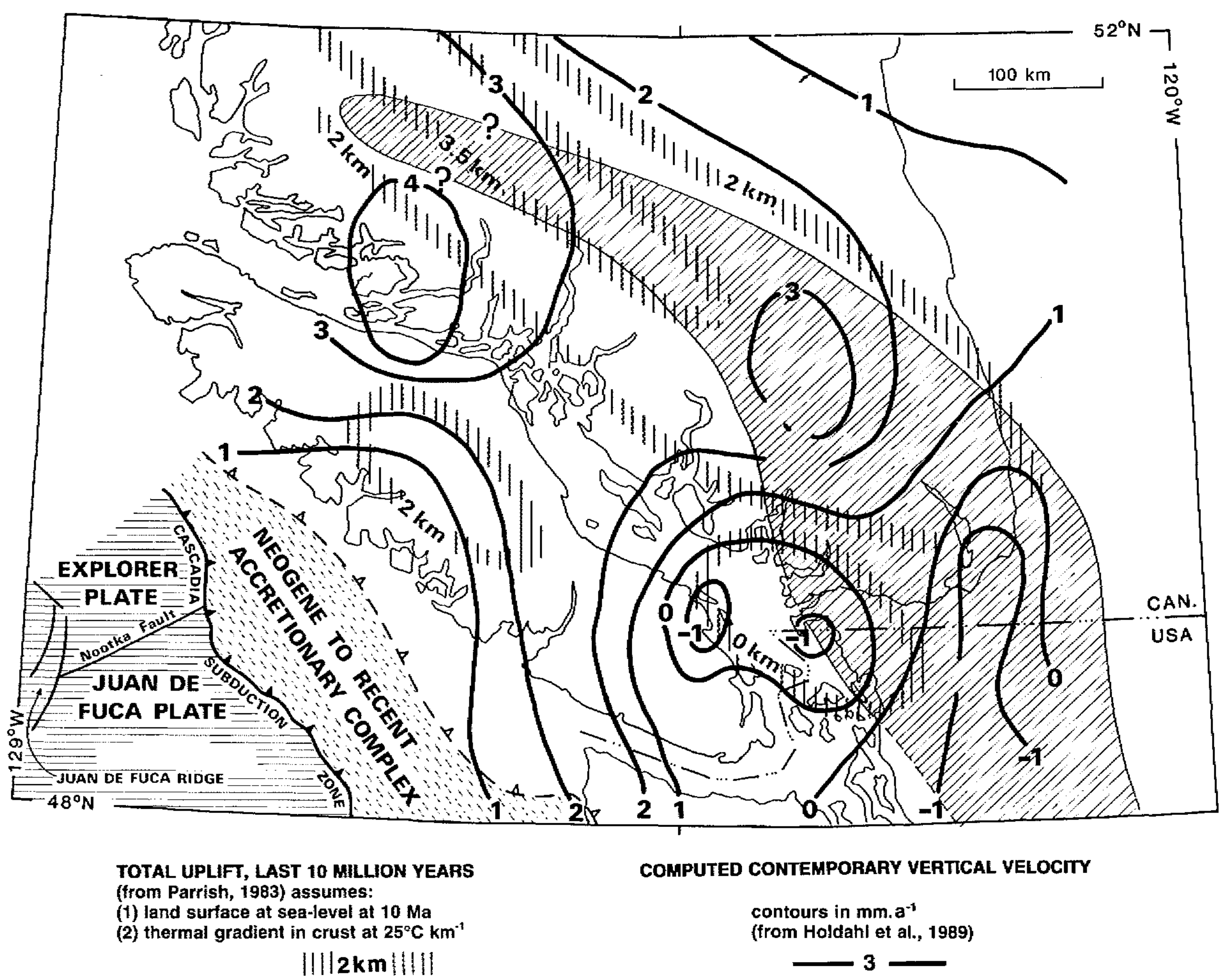

Figure 11. Amount of amount of uplift in last 10 Ma in kilometres and rates of present uplift in millimetres per year (mm•a^{-1}); patterns as in Figure 10.

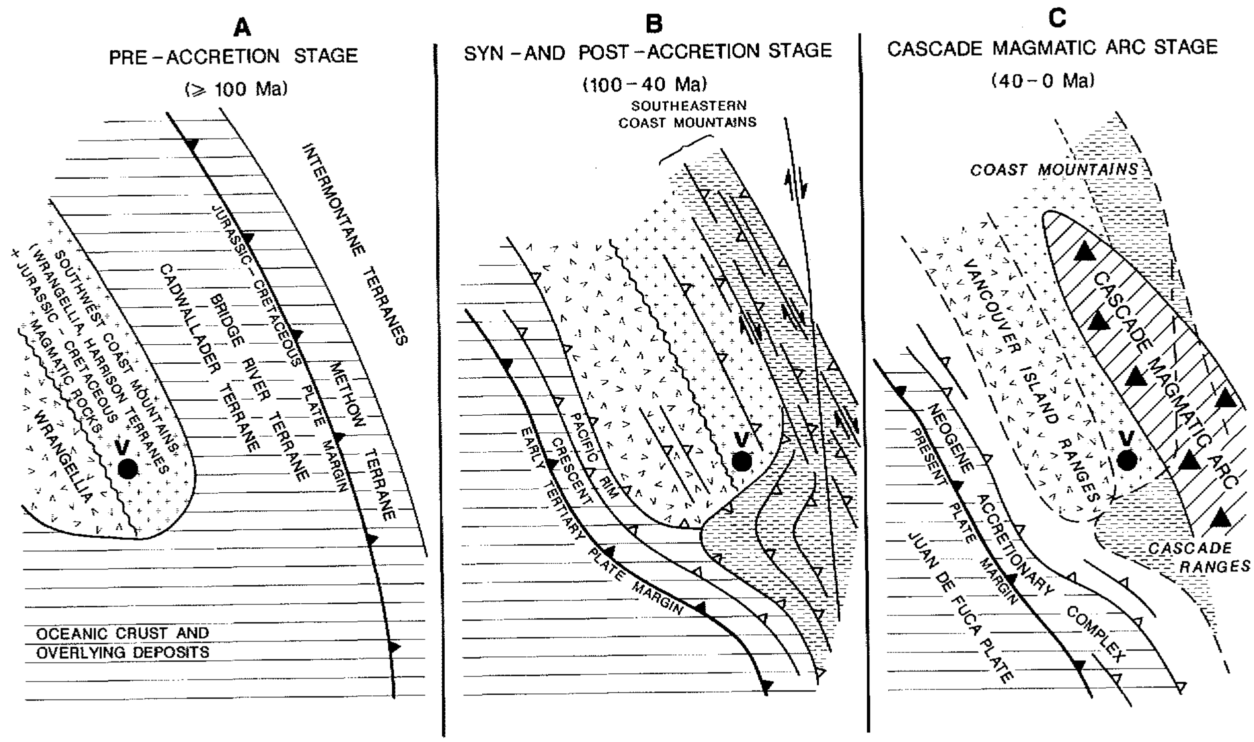

Figure 12. *Tectonic evolution of region containing Vancouver, whose location is shown by V. A) shows basinal terranes (Bridge River, Cadwallader, Methow) presently in southeastern Coast Mountains in the west to east paleogeographic indicated by Garver (1992). The western, Wrangellia-southwestern Coast Mountains block possibly was emplaced by sinistral, orogen-parallel faults in Early Cretaceous time (Monger et al., 1994).*

2. **Syn- and post-accretion stage (ca. 100-40 Ma)** This is the major crust building stage. It appears to have been initiated by mid-Cretaceous (≥100 Ma) accretion of Wrangellia terrane and rocks in southwestern Coast Mountains 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) intraplate contraction and dextral transpression, accompanied by metamorphism and intrusion, were located near the former plate boundary in the southeastern Coast Mountains and Cascade Ranges (Fig. 12B). The crust was thickened to ≥30 km, uplifted, and eroded. The former region of basinal terranes in southeastern Coast Mountains became a major source of detritus for the Late Cretaceous clastic cover.

3. **Cascade magmatic arc stage (40-0 Ma)** The tectonics of the region today, and probably for the last 40 million years, are dominated by plate convergence, with subduction of the oceanic Juan de Fuca Plate beneath the North American Plate, a process accompanied by arc magmatism, whose northernmost volcanoes occur in the region (Fig. 12C).

The physiography mainly formed in Neogene to Quaternary time (≤10 Ma), with uplift of the basement areas. Cover strata are preserved as erosional remnants in depressed areas that coincide with boundaries between the different basements.

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The Upper Cretaceous Nanaimo Group, Georgia Basin

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Mustard, P.S., 1994: The Upper Cretaceous Nanaimo Group, Georgia Basin; in Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia, (ed.) J.W.H. Monger; Geological Survey of Canada, Bulletin 481, p. 27-95.

Abstract: The Nanaimo Group comprises up to 4 km of sedimentary rock of Turonian to Maastrichtian age, forming the lower part of the Late Cretaceous to Neogene Georgia Basin of southwest British Columbia. This Upper Cretaceous succession was deposited in a single elongate basin deformed by Eocene compression into a fold and thrust belt. Eleven formations are recognized, comprising conformable and laterally intertonguing successions with sandstone-conglomerate units separated by mudstone and fine grained sandstone formations. Initial alluvial and coastal marine deposits formed on a rugged unconformity. Coal-bearing facies formed in coastal and marginal marine back-barrier environments, associated with fluvial and shallow marine facies. Most of the Nanaimo Group was deposited in marine, generally outer neritic to bathyal depths, by gravity flows and generally as submarine fan deposystems. Initial detritus was from local basement, but most sediment came from the Coast Belt to the east and northwest Cascades, although by latest Cretaceous time the eastern Cordillera was also a source. A forearc basin setting for the Nanaimo Group is only correct in that deposition occurred oceanward of a partly coeval magmatic arc. A foreland basin model is preferred because basin initiation and sedimentation was a direct result of contemporaneous thrusting in the Coast Belt and north Cascades. Nanaimo Group coal resources were historically important, but are now exhausted in most areas. Kaolin-rich deposits on the unconformity may be economically viable. Oil and gas potential is poor, although coalbed methane could be locally present.

Résumé : Le Groupe de Nanaimo comprend jusqu'à 4 km de roches sédimentaires d'âge turonien à maastrichtien, qui constituent la portion inférieure du bassin de Georgia (Crétacé tardif à Néogène) dans le sud-ouest de la Colombie-Britannique. Cette succession du Crétacé supérieur s'est déposée dans un bassin allongé unique, déformé par la compression subie à l'Éocène en une zone de plissements et de chevauchements. Onze formations ont été identifiées, notamment des successions concordantes et latéralement interdigitées, avec des unités de grès et conglomérat séparées par des formations à mudstones et à grès fins. Les premiers dépôts alluviaux et littoraux marins se sont formés sur une discordance accidentée. Des faciès carbonifères se sont constitués dans des milieux côtiers et des milieux marins marginaux d'arrière-cordon littoral, associés à des faciès fluviatiles et épicontinentaux. En majeure partie, le Groupe de Nanaimo s'est déposé en milieu marin, généralement en milieu infranéritique ou bathyal, sous l'effet de glissements par gravité et en constituant généralement des réseaux de cônes sous-marins. Les premiers débris provenaient du socle local, mais la plupart des sédiments provenaient de la ceinture Côtière à l'est et de la partie nord-ouest de la chaîne des Cascades, bien que dès le Crétacé terminal, la Cordillère orientale ait aussi été une source de sédiments. Un contexte de bassin d'avant-arc pour le Groupe de Nanaimo ne convient que dans la mesure où la sédimentation s'est produite du côté océanique d'un arc magmatique partiellement contemporain. On préfère un modèle de bassin d'avant-pays parce que le commencement de formation du bassin et la sédimentation étaient le produit direct de charriages contemporains survenus dans la chaîne Côtière et le nord de la chaîne des Cascades. Les ressources en charbon du Groupe de Nanaimo ont autrefois été importantes, mais sont maintenant épuisées dans la plupart des secteurs. Des dépôts riches en kaolin présents au niveau de la discordance pourraient être économiquement exploitables. Le potentiel en pétrole et en gaz est faible, bien que localement il puisse y avoir du méthane provenant de couches de charbon.

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INTRODUCTION

The Upper Cretaceous Nanaimo Group has been the subject of extensive study since coal was first mined in the Nanaimo area in 1852, from strata later described by Newberry (1857). Early research is reviewed in detail in Muller and Jeletzky (1970) and Usher (1952). The first extensive geological investigations were by Richardson (1872, 1873, 1878) in the Comox region and Dawson (1887, 1890) who proposed the name Nanaimo Group for the Upper Cretaceous succession of eastern Vancouver Island. Formations were first established by Clapp (1912a, 1914a) and Clapp and Cooke (1917). Early paleontological descriptions include Whiteaves (1879, 1895, 1901, 1903) and Dawson (1889), with early biostratigraphic zonations based on ammonites by Usher (1949, 1952), plant remains by Bell (1957), and foraminifera by McGugan (1962, 1964). Regional mapping and detailed biostratigraphic studies by the Geological Survey of Canada during the 1960s resulted in a unified stratigraphic nomenclature and the first formation-scale regional synthesis of Nanaimo Group geology (Muller and Jeletzky, 1970; Muller, 1983). Thirty-six undergraduate and graduate student theses (compiled separately in Appendix C) have resulted in surprisingly few published articles (major articles are Usher, 1952; Scott, 1974b; Ward, 1976b, 1978a, b; Pacht, 1984; England and Calon, 1991; and England and Hiscott, 1992). Other major papers in the last 20 years are by Sliter (1973), Ward and Stanley (1982), McGugan (1979, 1981, 1982, 1990), Haggart and Ward (1989), and numerous provincial and federal government publications (referenced in the text).

Several aspects of the Nanaimo Group remain subject to differing interpretations. Formation nomenclature for the Nanaimo Group is controversial, with several proposed lithostratigraphic schemes. Early interpretations of Nanaimo Group sedimentation proposed fluvial-deltaic and shallow marine depositional models almost exclusively. Post-1980 interpretations abandoned deltaic models and emphasize submarine fan facies models for the majority of Nanaimo Group deposition, a practice continued herein. The tectonic setting of the Nanaimo Group also has been debated with forearc, foreland, and even strike-slip basin settings proposed by different researchers. This paper, and Appendices A-C, incorporate the wealth of available data (much never published), with observations and data from the author's fieldwork in 1990 and 1991, into a synthesis of Nanaimo Group stratigraphy, structure, depositional history, and tectonic setting.

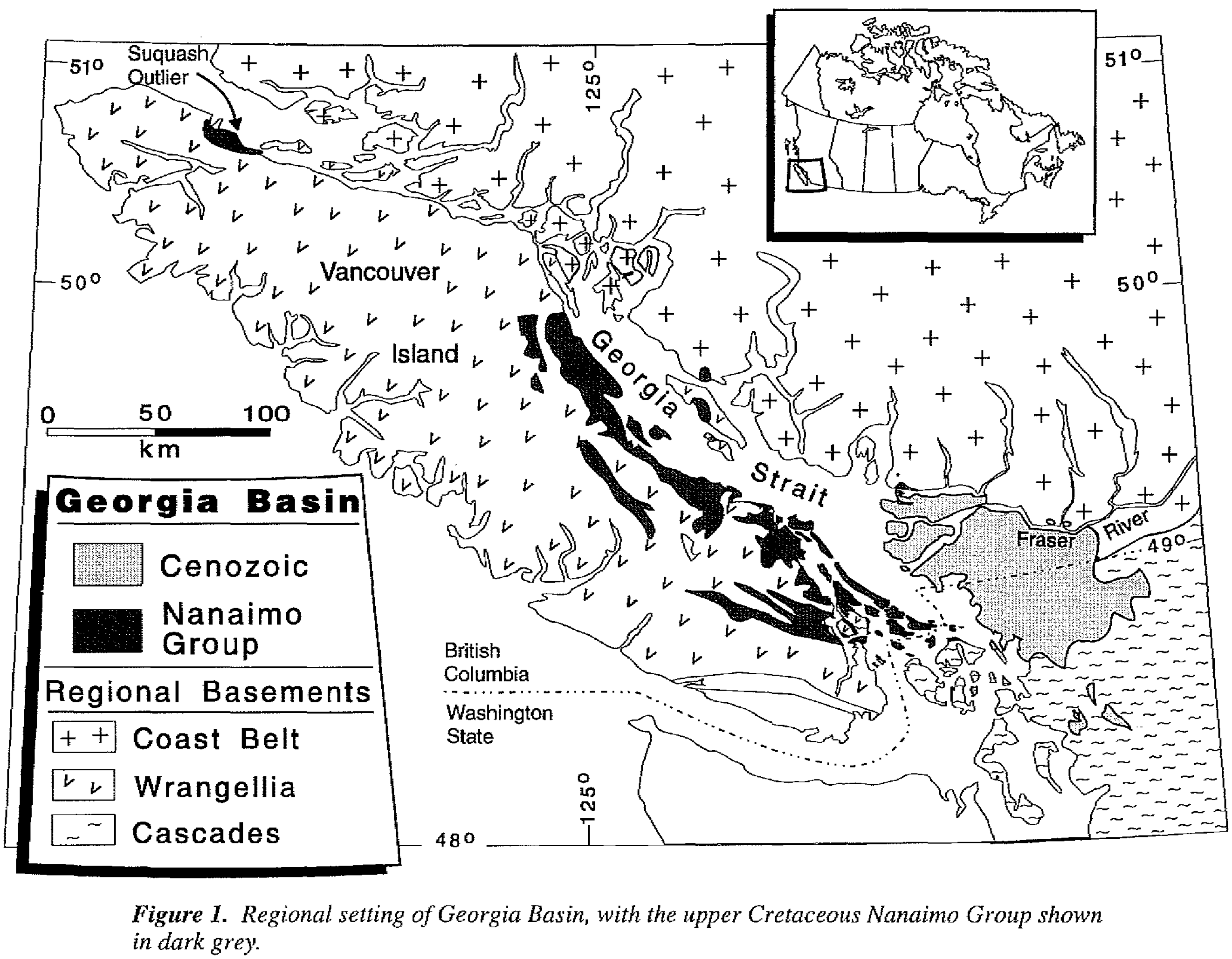


Figure 1. Regional setting of Georgia Basin, with the upper Cretaceous Nanaimo Group shown in dark grey.

REGIONAL SETTING

The Nanaimo Group is part of the Georgia Basin, a northwest-trending neotectonic structural and topographic depression which includes Georgia Strait, eastern Vancouver Island, the Fraser River lowlands of British Columbia, and northwest Washington State (Fig. 1). The present Georgia Basin is an erosional remnant and its configuration is largely the result of postdepositional deformation. For much of its depositional history the basin extended considerably farther to the west and perhaps to the east as well. Sedimentary rocks of the Georgia Basin comprise two main packages: the Upper Cretaceous Nanaimo Group exposed mainly on the east side of Vancouver Island and on the islands of Georgia Strait; and Paleocene to Miocene age rocks exposed mainly in the Vancouver area and northwest Washington State.

The Nanaimo Group unconformably overlies Wrangellia terrane on its west side, the Coast Belt on its east side, and is in inferred fault contact with the San Juan thrust system (part of the northwest Cascades) to the southeast (Fig. 1). Monger and Journeay (1994) reviewed the geology of these basements to the Nanaimo Group. Pre-Tertiary Wrangellia terrane comprises stratified rocks of Devonian to Jurassic age, and intrusions of mainly Early to Middle Jurassic and locally Devonian age (ca. 200-175 Ma and 360 Ma or older, Armstrong, 1988; Parrish and McNicoll, 1992). In contrast, the southern Coast Belt comprises predominantly Middle Jurassic to Eocene granitic rocks, with minor septa and fault slices of mainly Triassic, Jurassic, and Cretaceous volcanic and sedimentary rocks. It is divisible into two parts. The western part features Jura-Cretaceous granitic rocks, septa of greenschist grade which include Wrangellian rocks, and deformation concentrated in discrete, widely spaced zones. The eastern part contains Late Cretaceous and early Tertiary granitic rocks, with metamorphism ranging from zeolite to high-amphibolite facies, and penetrative deformation (Crickmay, 1930; Armstrong, 1988; Friedman and Armstrong, 1990, in press; Journeay and Friedman, 1993; Monger and Journeay, 1994). The San Juan thrust system (shown on Fig. 5), which is the northwest part of the Northwest Cascades System of Brown (1987), affects Devonian to Upper Cretaceous strata and metamorphic rocks dissimilar from those of Wrangellia terrane (Brandon et al., 1988). Structurally and possibly stratigraphically, these rocks are comparable to some in the eastern Coast Belt (Monger, 1991a, b).

The main structural control on the sub-Georgia Basin rocks, and to some extent Georgia Basin itself, has been the complex history of mainly transpressional deformation that took place from Late Jurassic to Holocene time as a response to the interaction of the North American plate with the Kula and subsequently the Farallon oceanic plates (the Farallon now the Juan de Fuca plate) as reviewed in Yorath et al. (1985) and Monger (1993). Critical to the initial formation of the Late Cretaceous Georgia Basin is a mid- to Late Cretaceous west-vergent thrust system preserved at the southeast margin of the Georgia Basin (Brandon et al., 1988; McGroder, 1991) and in the eastern Coast Belt (Journeay et al., 1992; Journeay and Friedman, 1993). Structures appear to wrap around from the San Juan Thrust System into the eastern Coast Belt as first recognized by Crickmay (1930). This deformation event includes thrusts active during both basin formation and major periods of Nanaimo Group sedimentation. Dextral strike-slip faults may have influenced depositional patterns during the Tertiary stage of Georgia Basin fill (Johnson, 1984). The basin has also been affected by early Tertiary compression, which resulted in southwest directed thrusting that included the Nanaimo Group (England and Calon, 1991) and northwest-plunging and -trending folds in the Tertiary Chuckanut Formation (Johnson, 1984).

The original extent of the Late Cretaceous Georgia Basin is not well constrained. The stratigraphic architecture, facies relationships, and provenance evidence summarized herein indicate that the original basin axis was oriented northwest-southeast and probably roughly centred over the present Georgia Strait or eastern Vancouver Island. The present preserved extent of the Nanaimo Group in the northwest-southeast direction is about 230 km. A preserved northeast-southwest basin width of about 90 km is restored to about 120-130 km when effects of early Tertiary shortening are subtracted. However, these are minimal basin dimensions. To the north, Upper Cretaceous strata present on northern Vancouver Island (commonly termed the Suquash outlier and shown in Fig. 1) have been correlated with the Nanaimo Group by some (e.g., Muller and Jeletzky, 1970; Kenyon, 1991), although such correlations are based on broad similarities of rock type but poor age constraints. For this review the Suquash outlier is not considered as part of the Nanaimo Group. Most of the upper Nanaimo Group comprises relatively deep marine strata preserved on the islands of Georgia Strait. Facies relationships and paleocurrent patterns for these rocks suggest the basin continued for an unknown distance to the west. More precise estimates of original basin extent are obscured by two postdepositional effects. First, several kilometres of uplift occurred on both the eastern and western sides of the basin (reviewed in Monger, 1990). On the east, rapid late Neogene uplift of the Coast Belt (Parrish, 1983) may have resulted in erosion of most Nanaimo Group strata in this area. In the west, eastward tilting and uplift of Vancouver Island, probably due to Juan de Fuca plate underthrusting, has resulted in erosion of any Nanaimo Group sediments deposited west of the outcrop areas. Second, the mid- to Late Eocene compression which deformed the Nanaimo Group into a fold and thrust belt includes a thrust contact between the southern Nanaimo Group and the San Juan terranes (England and Calon, 1991, shown on Fig. 5). This obscures the original southern boundary of the basin.

GENERAL STRATIGRAPHY

The Nanaimo Group comprises a stratigraphic thickness of more than 4 km of sedimentary rocks of Turonian to Maastrichtian age. The major outcrop areas are shown in Figure 2. The strata form separate erosional remnants, with the main outcrop belt along the east side of Vancouver Island extending from Campbell River in the north to the Saanich Peninsula in the south, and including both northern and southern Gulf Islands and the northern San Juan Islands in Washington

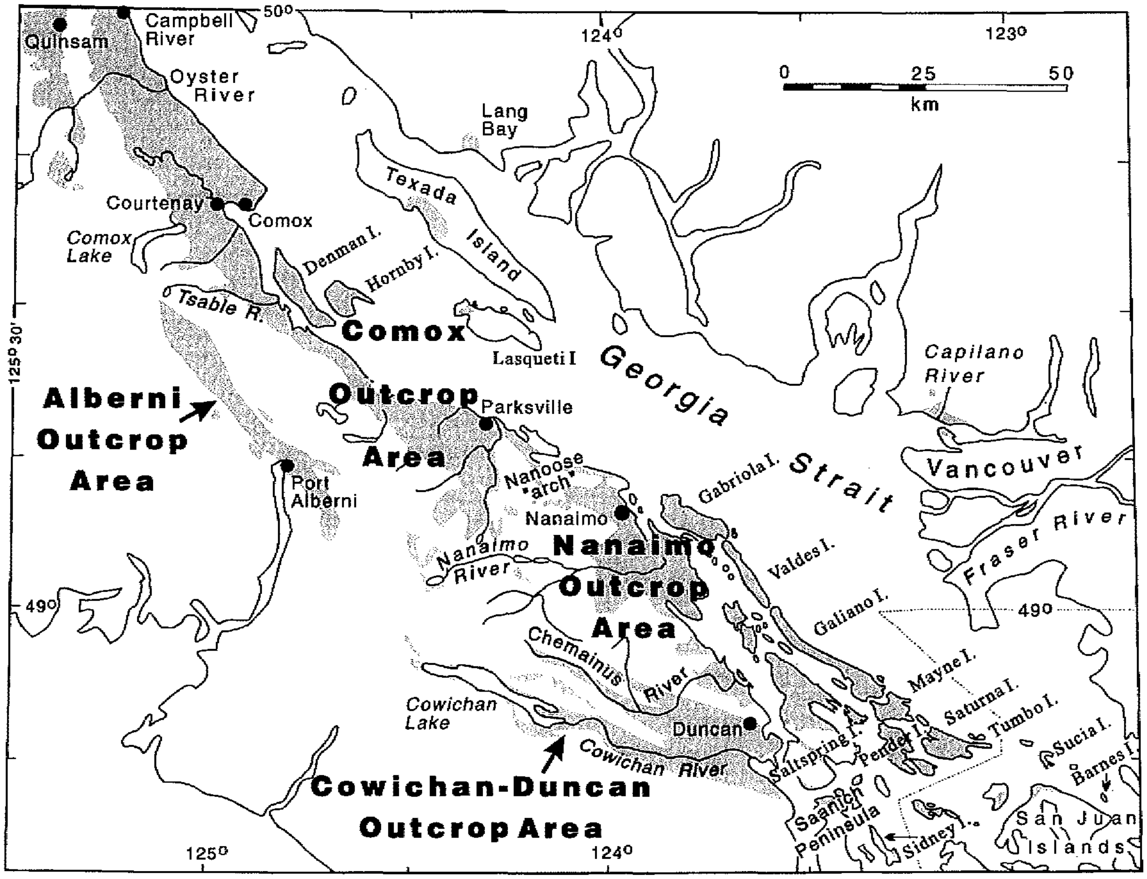


Figure 2. Major outcrop areas of the Nanaimo Group (shown in dark grey). Place names of towns (black dots), and other geographic features mentioned in text are also shown.

State. Significant outliers also occur in the Alberni Valley, with smaller outliers on Texada and Lasqueti islands and on the mainland at Vancouver and Lang Bay.

In the last fifty years the stratigraphic nomenclature of the Nanaimo Group has undergone a complex evolution. Figure 3 illustrates the formation nomenclature proposed in major published studies. Several others have been presented in government reports and graduate theses. Most early workers interpreted the separate erosional remnants as the products of deposition in separate basins, but most recent regional studies, including this one, consider the Nanaimo Group to have been deposited in a single basin, and use a single formation nomenclature, as is done here and shown in Figure 4 (based on Muller and Jeletzky, 1970 as modified by Ward, 1978a). The main objections to this nomenclature come from a continued belief that the Comox and Nanaimo outcrop areas (shown on Fig. 2) do in fact represent originally discrete depositional basins and thus two lithostratigraphic schemes are appropriate (England, 1989; McGugan, 1990; England and Hiscott, 1992). Appendix B reviews the evolution of Nanaimo Group stratigraphic terms and the main reasons suggested for the several proposed schemes, including the unified formation nomenclature used here.

Figure 5 provides a simplified geological map of the Nanaimo Group, showing major formation boundaries and structures. Formation-scale revisions suggested by England (1989) for the entire basin and the recent formation names of England and Hiscott (1992) for the upper Nanaimo Group strata of the outer southern Gulf Islands are rejected as unnecessary for the reasons outlined in Appendix B. In addition, the proposed redefinition of the Gabriola Formation contact on Galiano Island to include strata previously considered to be upper Geoffrey and Spray formations is not supported by the author's observations, who agrees with the detailed mapping of Carter (1977). This demonstrates that the perceived fault, which England and Hiscott (1992) show offsetting the Gabriola Formation on Galiano Island, does not offset well-exposed strata on the north and south coasts of the island. This suggests that the original interpretation of the mudstone interval at Sturdies Bay as the top of the Spray Formation is more likely (as shown on Fig. 5).

Haggart (1994) has informally proposed a new formation (his Sidney Island Formation) for the Turonian age strata preserved in the southernmost Gulf Islands, suggesting these strata should be considered separate and probably older than the Comox Formation in which they had previously been included. This new formation name is shown on Figure 4,

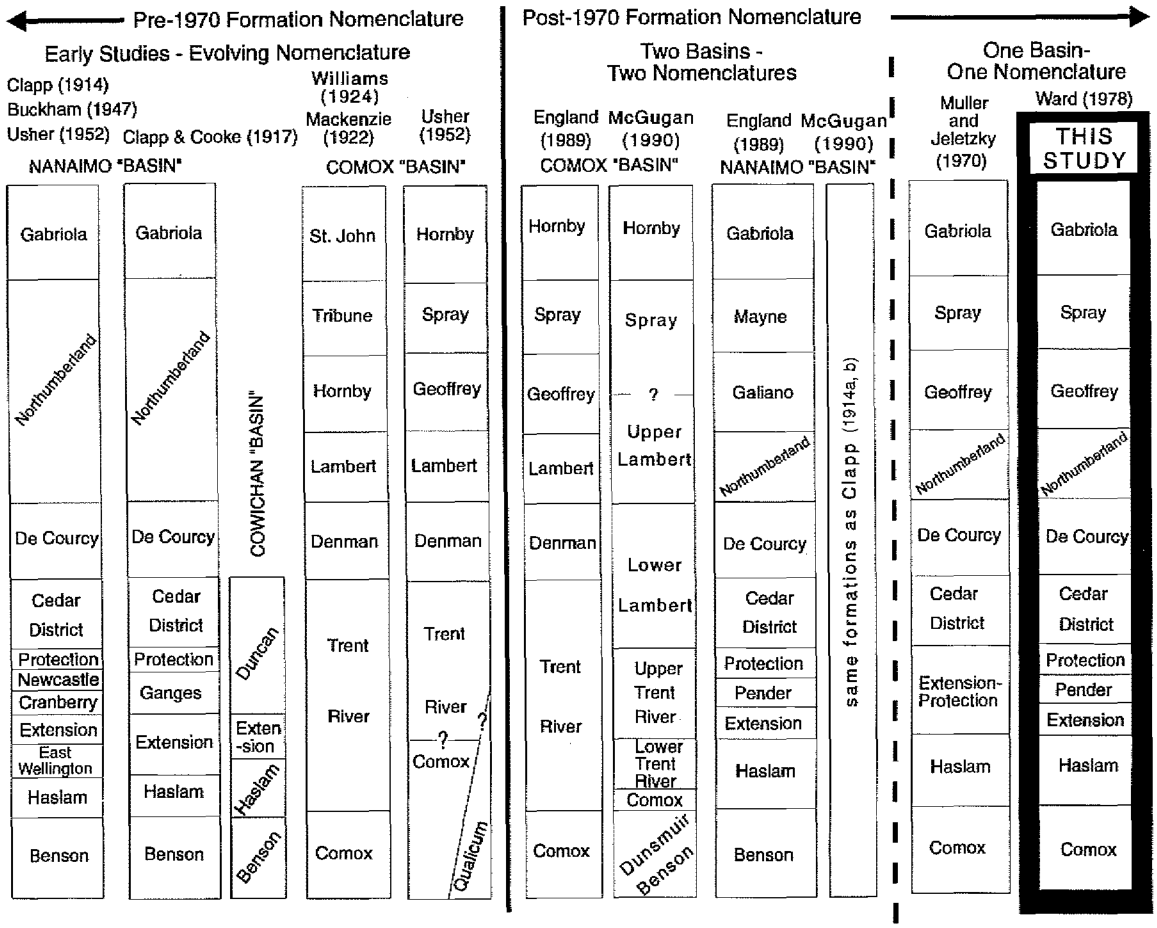


Figure 3. Major formation nomenclatures proposed for the Nanaimo Group. Members proposed for several of the formations by various workers are discussed in the text.

but is not used in this review as the proposed name is presently informal, has not been widely published, and thus not had the opportunity to be accepted or rejected by other researchers.

Several member subdivisions have been proposed for different formations and different areas, most recently by Bickford and Kenyon (1988), England (1989), Bickford et al. (1990), Cathyl-Bickford (1992a), and England and Hiscott (1992). This is especially relevant for the basal Comox Formation which shows significant lateral variation, and for the areas where coal occurs in economically significant quantities and detailed lithostratigraphic subdivisions are useful. Specific member names are shown where considered relevant in the tabulated formation summaries (Appendix A, Tables 1 to 11).

BIOSTRATIGRAPHY AND AGE OF THE NANAIMO GROUP

Molluscan macrofossil, microfossil, and radiometric dating studies of the Nanaimo Group provide an age range of early Turonian to Maastrichtian (about 91 ± 3 Ma to 66 ± 2 Ma on the Harland et al., 1990 time scale used in this review). Formations are slightly diachronous laterally, but no significant time gaps are apparent with the exception of a possible slight (about 2 Ma) diachronous contact between Turonian and Santonian strata in some places (described below and in Haggart, 1994). In general, formation boundaries are conformable and gradational, with intertonguing relationships common. In some places however, dramatic facies changes have resulted in locally sharp and erosive contacts between formations, which some workers have interpreted as unconformities (e.g., Muller and Jeletzky, 1970), although no biostratigraphic gaps are apparent.

The lower age limit of the Nanaimo Group is provided by the presence of early Turonian ammonites and bivalves in the most southerly exposures of Nanaimo Group at Sidney Island (Haggart, 1991, 1994; located on Fig. 2) and possible late Turonian ammonites in mudstones on Barnes Islands in the northern San Juan Islands (P.D. Ward, pers. comm., 1985; cited in Garver, 1988) which may be part of the Nanaimo Group. Elsewhere in the basin, the lowest fossils found in the Nanaimo Group are generally Santonian age macro- and microfossils, although the basal Comox Formation throughout the basin is unfossiliferous and could be older (Jeletzky in Muller and Jeletzky, 1970; McGugan, 1962, 1979). Thus, although more study is needed, there appears to be a slight

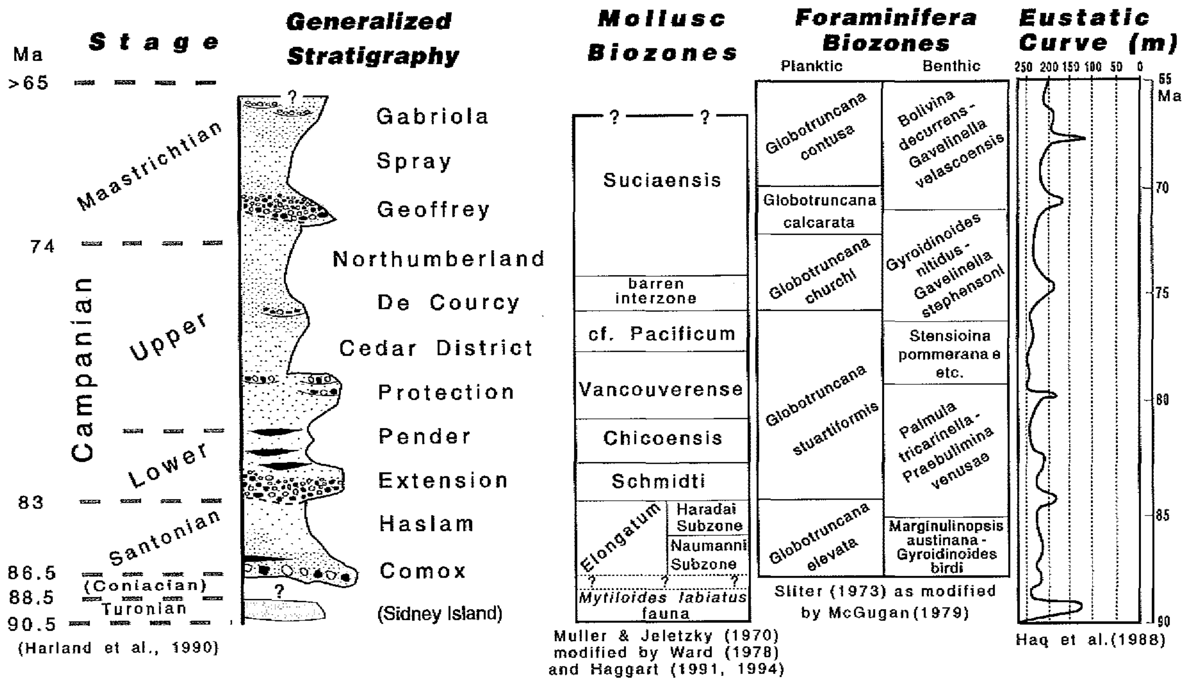


Figure 4. General stratigraphy of the Nanaimo Group showing the most recent biostratigraphic zonations based on molluscan and foraminifera fossils. The time scale used here and in text is from Harland et al. (1990); note that stage boundary age values have margins of error ranging from 2 to 3.6 million years, formations are diachronous laterally, and the upper age of the Gabriola Formation is not well constrained. Eustatic sea level variations as suggested by Haq et al. (1988) are shown at right.

gap representing Coniacian time (about 2 Ma) between these strata and the Santonian fossils present in the Comox Formation elsewhere in the basin. The only direct radiometric date from the Nanaimo Group is a U-Pb zircon age of 82.5 ± 1 Ma from a tuffaceous layer in the upper Comox Formation (Dunsmuir Member) in the Quinsam area near the northern edge of preserved outcrop (Kenyon et al., 1992, located on Fig. 2). This age is within the range of the age of the Santonian-Campanian boundary of 83 ± 3.6 Ma (Harland et al., 1990). It is compatible with the mid- to upper Santonian age typically cited for the Comox Formation and the Santonian age for the basal Nanaimo Group strata in all but the southern (Turonian) part of the basin (Muller and Jeletzky, 1970; Ward, 1978a).

The upper age limit of the Nanaimo Group is not well constrained. The top of the Gabriola Formation has yielded a few *Baculites* and other molluscs (Haggart, 1989), but only from Galiano Island. A gradational contact with the underlying Spray Formation, from which Maastrichtian foraminifera have been documented (McGugan, 1979, 1982), suggests the Gabriola Formation is Maastrichtian. Paleocene sandstones unconformably overlie the Nanaimo Group at several places, including Lasqueti Island (Mustard and Rouse, 1991) and Tumbo and Sucia islands (Mustard and Rouse, 1992). Detrital zircons collected a few metres above the base of the Gabriola Formation at its type area are dominated by large euhedral forms which give concordant U-Pb ages of 72-73 Ma (Mustard et al., in press), suggesting the Gabriola Formation in this area is less than 72 Ma old and probably upper Maastrichtian (about 70 to 66 Ma). Thus, there may not be a major time gap between the Cretaceous and Tertiary parts of Georgia Basin. Preliminary palynological data from mid- to upper Gabriola Formation strata on Hornby and Gabriola Islands suggest an early Paleocene age for these strata as well (G. Rouse, pers. comm., 1993), which implies a conformable Cretaceous-Tertiary boundary in at least these areas.

Biostratigraphic studies of the Nanaimo Group have concentrated principally on ammonites and inoceramid bivalves. Early studies are summarized in Muller and Jeletzky (1970). An ammonite biozonation of Usher (1952) was superseded by a zonation developed by Jeletzky (in Muller and Jeletzky, 1970), based on ammonites and inoceramids. This recognized four major biozones and several subzones ranging from mid-Santonian to early Maastrichtian age. Ward (1978a) refined this molluscan biochronology, adding additional zones. His molluscan biozone scheme is shown in Figure 4, with modifications from more recent work, most importantly by Haggart (1991, 1994) who documented Early Turonian strata at the base of the Nanaimo Group in the southern Gulf Islands.

Nanaimo Group microfossil studies are restricted to foraminifera and palynomorphs. McGugan (1962) proposed the first foraminifera biozonation, which was modified by Sliter (1973) and refined by McGugan (1979) to give the scheme shown in Figure 4. Other important published foraminifera studies are Scott (1974b) and McGugan (1981, 1982, 1990). Palynological studies demonstrate the Late Cretaceous age of eastern Nanaimo Group outliers (Crickmay and Pocock, 1963; Bradley, 1973; Rouse et al., 1975; Mustard and Rouse, 1991), and the early Tertiary age of strata previously correlated with the upper Nanaimo Group in the southeast outcrop area (Mustard and Rouse, 1992).

STRUCTURE

Much of the Nanaimo Group is characterized by moderate to shallow northeast-to east-dipping beds. Locally, these reverse their dip over northwest-trending folds. Beds of eastern outliers tend to dip shallowly to the southwest. A major northwest-trending fault set and a younger set of minor northeast-trending faults cuts both the entire Nanaimo Group and underlying strata (Fig. 5).

Muller and Jeletzky (1970) and Muller (1977a, b, 1983) provided the first structural synthesis of the entire Nanaimo Group. They interpreted major fault sets as high-angle normal and reverse faults, suggesting a block faulting model with both vertical and strike-slip components of movement. Deep seismic profiles across Vancouver Island resulting from the LITHOPROBE program of the mid-1980s clearly demonstrated that the major northwest-trending faults are northeast-dipping, listric thrust faults and that the dominant structural style for both Wrangellia terrane and the overlying Nanaimo Group reflects west-to southwest-directed compression, a response to underthrusting of the Farallon (now Juan de Fuca) oceanic plate beneath the North American plate (Yorath et al., 1985; Sutherland-Brown and Yorath, 1985).

Post-LITHOPROBE geological mapping and structural interpretations of eastern Vancouver Island reflect this current understanding of the regional structural evolution. Most northwest trending faults are now recognized to be part of a major southwest vergent thrust system called the Cowichan Fold and Thrust Belt in the comprehensive regional structural interpretation by England and Calon (1991, also see authors listed in Fig. 5 caption). England and Calon (1991) interpret the folds and thrusts cutting the Nanaimo Group and underlying Wrangellia as a thick-skinned, linked fold and thrust system, developed in about late Eocene time. They calculate about 20-30% shortening due to southwest-vergent thrusting and folding, interpreting the system as a large-scale imbricate fan with northeast dipping listric thrusts which root in a moderately northeast dipping sole fault within Wrangellian basement. Folds in most of the thrust system are commonly open, upright and slightly asymmetric to the southwest, subhorizontal to gently plunging, and generally cylindrical, although in western areas non-cylindrical folds are more common. The pattern of fold and fault geometries, orientation of cover rock sequences, and fault kinematic indicators all support essentially orthogonal compression with insignificant strike-slip movement.

The fold and thrust system is most evident in the Nanaimo, Cowichan-Duncan, and Alberni outcrop areas and less well-developed in the Comox outcrop area, although thrusting in the Nanoose-Parksville area is well documented (Sutherland-Brown et al., 1986) and southwest-directed thrusts are present in the Quinsam area (Kenyon et al., 1992). Most of the Comox outcrop area is roughly homoclinal, dipping generally <15° to the northeast, with minor northwest-and northeast-trending faults. England and Calon (1991) suggested the Comox outcrop area is part of a single large thrust sheet with the floor thrust surfacing to the west (shown on Fig. 5 as the thrust on the eastern margin of the Alberni outcrop area). Cathyl-Bickford (1992a) documented northwest-trending and younger northeast-trending sets of minor faults in the central part of the Comox area, suggesting both normal and dextral strike-slip senses of movement on the northwest set and sinistral strike-slip displacements on the northeast set. Displacements on all faults in this area appear to be minor, tens to hundreds of metres (England, 1990; Cathyl-Bickford, 1992a). Northeast-trending faults in the Nanaimo outcrop area also show minor displacements and appear to be minor normal faults which postdate thrusting (Massey and Friday, 1988).

One major fault which appears to postdate development of the Nanaimo thrust system occurs east of the southern Gulf Islands (Fig. 5). This structure, observed from marine seismic lines (Machacek, 1971; White and Clowes, 1984), is a steep northeast dipping fault which displaces both Nanaimo Group and lower Tertiary strata several kilometres down to the north.

Timing of the major contraction is loosely constrained as post-Cretaceous, based on the involvement of the entire Nanaimo Group in the deformation, and probably pre-Middle to Late Eocene, based on fission track ages from Nanaimo Group apatite (England and Massey, unpub. data, cited in England and Calon, 1991) and thermal history modelling (England, 1990). Northwest-trending folds in the Paleogene Chuckanut Formation of northwest Washington State were interpreted by Johnson (1984), as part of a Late Eocene event suggesting a connection between this deformation and the Nanaimo Group fold and thrust system. Both Massey and Friday (1989) and England and Calon (1991) speculated that contractional deformation of the Nanaimo Group is connected to accretion of the Pacific Rim and Crescent terranes to southern Vancouver Island during the late Eocene (Clowes et al., 1987).

LITHOSTRATIGRAPHY AND DEPOSITIONAL SETTING OF THE NANAIMO GROUP

The major features of each of the eleven Nanaimo Group formations are summarized in a table with accompanying figure in a separate Appendix (Appendix A; Tables 1 to 11, Fig. 2 to 12). Each table summarizes the age range, contacts, major lithostratigraphic features, depositional environment, and provenance. The formation descriptions and interpretations are distilled from the listed references on each table, supplemented by the author's field studies in 1990 to 1992. Depositional environment and provenance sections reflect

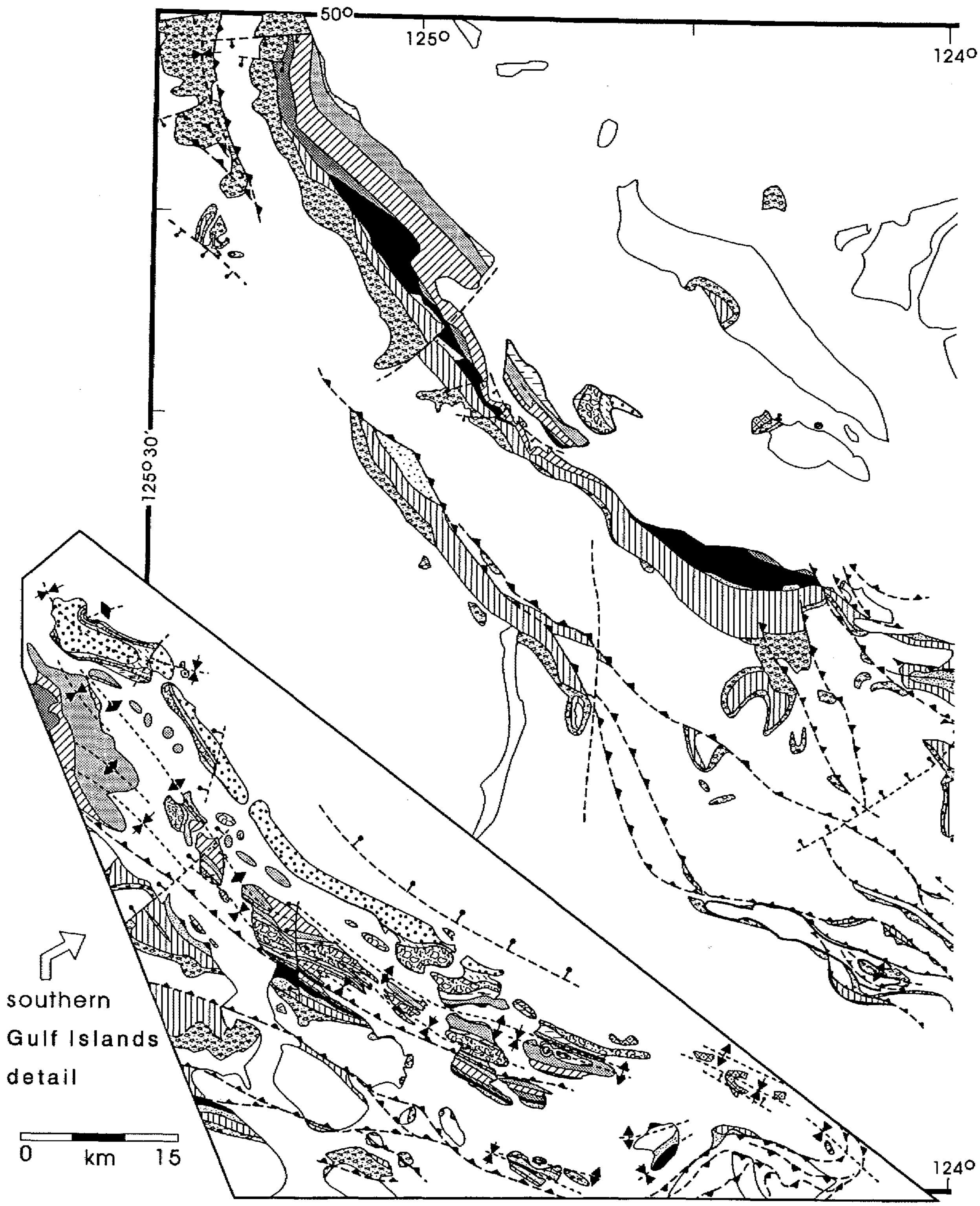
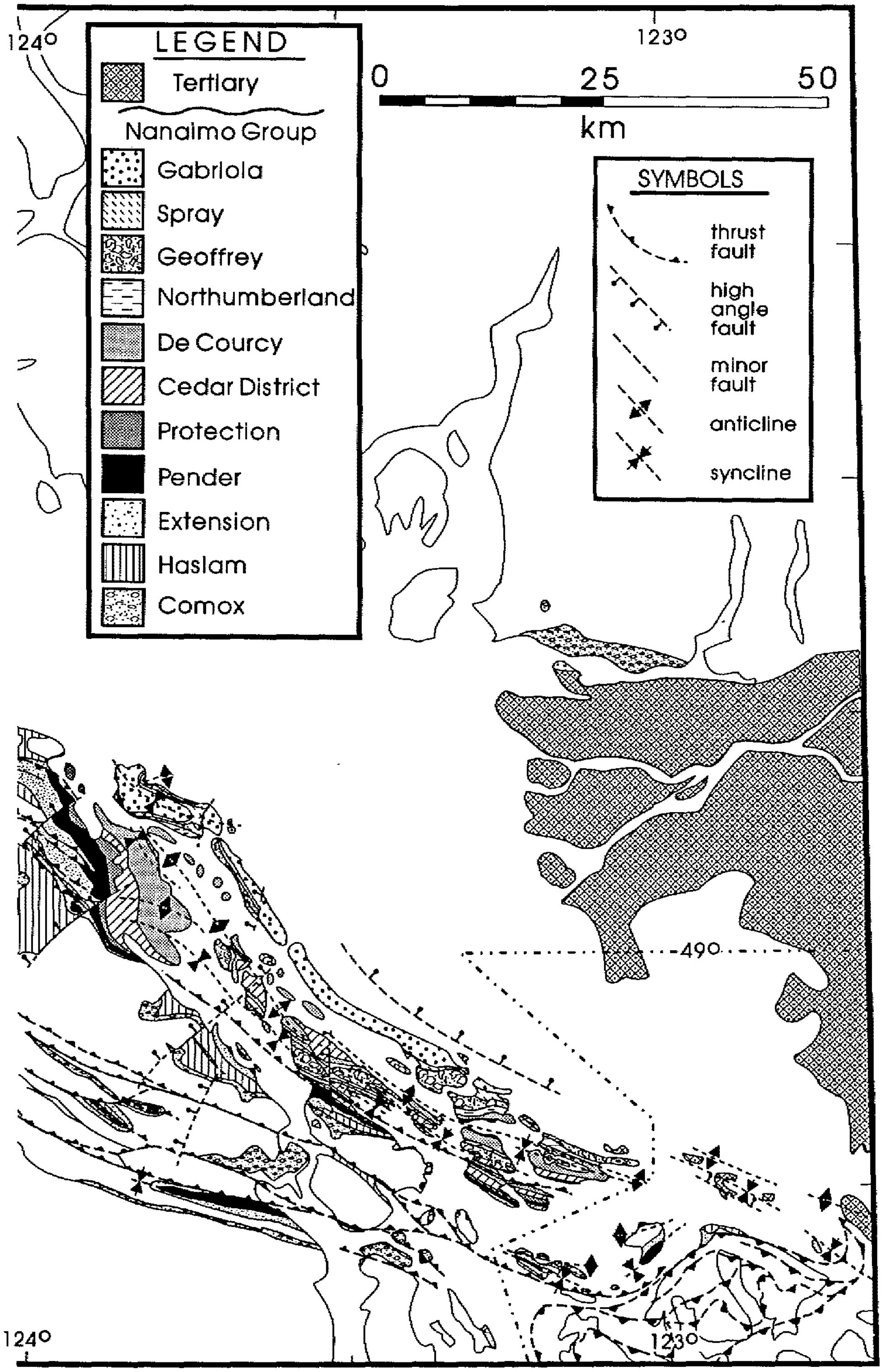


Figure 5. Simplified geological map of the Nanaimo Group. Strata generally show shallow to moderate dips to the northeast on Vancouver Island and the Gulf Islands, except for local reversals of dips across folds and steep dips in fault zones. Strata on Texada Island and on the mainland generally dip shallowly to the southwest or south. Main sources for compilation in addition to the author's mapping are Sutherland-Brown et al. (1986), Massey et al. (1987, 1988), Massey and Friday (1988, 1989), Bickford and Kenyon (1988), Brandon et al. (1988), Bickford (1989), England (1990), Cathyl-Bickford (1992a, b), Kenyon et al. (1992) and C.G. Cathyl-Bickford (unpub. data, pers. comm., 1992). For geographic details, see Figure 2.


124°
LEGEND
Tertiary
Nanaimo Group
Gabriola
Spray
Geoffrey
Northumberland
De Courcy
Cedar District
Protection
Pender
Extension
Haslam
Comox
123°
0
25
50
km
SYMBOLS
thrust fault
high angle fault
minor fault
anticline
syncline
49°
124°
123°

modern (generally post-1980) interpretations, again supported by the author's work. For each table an accompanying figure provides summary maps and ternary plots showing outcrop distribution, paleocurrent, and sandstone and conglomerate compositional data compiled from generally the same sources of information as for the table and from the author's own data (sandstone and conglomerate data sources are listed in Tables 1 and 3 of this review).

Not given in this review are examples of the many measured vertical profiles published for the Nanaimo Group. Major published examples of these profiles are: **Comox Formation** – Muller and Atchison (1970), Kenyon et al. (1992); **Haslam to De Courcy formations – Ward (1978a), Ward and Stanley (1982), Pacht (1984); Northumberland to Gabriola formations** – England and Hiscott (1992). Detailed section descriptions for specific areas (especially indi- vidual Gulf Islands) are also generally included in the appropriate graduate theses (see formation table references in Appendix A and compiled graduate thesis references in Appendix C).

Nanaimo Group depositional models

Muller and Jeletzky (1970) proposed the first regional-scale depositional models for the Nanaimo Group. They interpreted the nine formations they recognized (shown in Fig. 3) as containing four complete and one partial unconformity-bound sequences (Comox-Haslam; Extension/Protection-Cedar District; De Courcy-Northumberland; Geoffrey-Spray; and Gabriola as the basal part of an incomplete fifth sequence). They interpreted these sequences as stacked transgressive cycles with basal nonmarine to marginal marine fluvial-deltaic deposits (the sandstone and conglomerate-dominated formation), that were transgressively overlain by generally shallow marine prodeltaic deposits (e.g., delta bottomsets and delta slope generated turbidites) represented by the capping mudstone and fine-grained sandstone formation of each cycle.

The numerous (>15) detailed graduate studies completed between 1968 and 1980 in general retained the basic fluvial-deltaic and shallow marine interpretations for the roughly stacked, coarse-and fine-grained formations (e.g., Hanson, 1976 and other Oregon State theses, compiled in Appendix C). However, most of these researchers concluded that the unconformities proposed between cycles were not evident and documented formation contacts as laterally intertonguing and in many places vertically gradational. In some cases they reinterpreted many of the cycles as coarsening-upward, with the coarser formation representing progradation of a major deltaic and fluvial succession over relatively distal deltaic and prodelta deposits.

Post-1980 regional facies interpretations and depositional models for the Nanaimo Group are published in Ward and Stanley (1982), Pacht (1984), and England and Hiscott (1992) with more detailed analysis presented in the theses of Pacht (1980) and England (1990). Although differing in their interpretations of overall basin type (i.e. one vs. two depositional basins and tectonic setting) and in the specifics of some facies interpretations, all the above authors concluded that much of the Nanaimo Group was deposited in relatively deep marine environment (lower neritic to bathyal depths generally, meaning roughly sub-wavebase shelf to 2000 m water depths) with sediment gravity flow as the major depositional process, and submarine fans as the main depositional system. These authors retained mostly nonmarine to shallow marine interpretations for the entire Comox Formation and for most lower formations (Extension, Pender, and Protection formations, but not Haslam Formation) in the coal-bearing area centred around Nanaimo.

The change in models from fluvial-deltaic to submarine fan is due to two factors. The first was the explosion of literature on submarine fan deposition in the 1970s and early 1980s which documented sand and conglomerate-rich facies similar to most Nanaimo Group lithofacies (historical reviews are given in Walker (1992) and Pickering et al. (1989)). The second was paleobiological evidence for generally deep marine deposition both in the mudstone-rich formations and the thick sandstone and conglomerate successions above the Comox Formation (except for the coal-bearing formations in the Nanaimo area). Much of this paleobathymetric evidence came from regional studies of foraminifera, most notably those of Sliter (1973), Scott (1974a, b), McGugan (1979, 1981, 1982, 1990) and unpublished reports by Cameron (1988a, b). McGugan (1979, 1981) discusses the difficulties in interpreting paleodepth from foraminifera, a problem apparent from the variations in paleodepths suggested by the different studies. Sliter, using detailed statistical analysis of foraminifera diversity and types, concluded that the mudstone-rich units were generally deposited in deep marine areas (lower neritic to abyssal, about 150 m to 2000 m depths). Scott, and McGugan, also using detailed statistical analysis, but of a much greater number of samples and over a wider area, interpreted deposition at generally upper to lower neritic with only rare upper bathyal depths (about 50 to 1000 m depths). Cameron considered foraminifera types and diversity (with no statistical treatment, however) in combination with the characteristics of the sampled lithofacies to interpret depth ranges for each sample, generally suggesting middle neritic to mid-bathyal paleodepths. Significantly, Cameron also generally interpreted similar paleodepths for samples from mudstone interbeds within the sandstone/conglomerate formations. Although conflicting in exact depth determinations, these studies show agreement that both coarse-and fine-grained formations of most of the Nanaimo Group were deposited in sublittoral, generally outer neritic or deeper environments, a conclusion supported by the lithofacies present and the presence of marine macro- and trace fossils in both fine-and coarse-grained formations.

Trace fossils are present in most formations and abundant in some areas, but no specific studies of trace fossil types, ichnofacies, or paleoenvironmental implications have been published. England (1990), provides the most comprehensive listing of identified traces and briefly discusses possible paleodepth significance. In terms of commonly recognized ichnofacies, most traces are marine types, with representatives of *Cruziana*, *Skolithos*, and *Zoophycos* ichnofacies present. Pemberton and Frey (1992) provide a current review of common ichnofacies and the difficulties in assigning specific environmental and paleodepth interpretations without detailed and comprehensive study. Pending such studies, the most significant conclusion derived from the trace fossils are their compatibility with the neritic to bathyal paleodepths suggested by foraminifera and lithofacies evidence, and their presence in most sandstone units previously considered fluvial, adding to the evidence for marine deposition (both also concluded by England, 1990).

The following summary of depositional facies is condensed from the modern studies cited above (especially Pacht, 1980 and England 1990), supplemented with observations and interpretations by the author. Comprehensive references for the formations are given in the summary tables of Appendix A; only references relevant to specific depositional interpretations are cited below.

Nonmarine facies associations

Figure 6 presents a schematic representation of both the early basin setting within the region extending from west of Vancouver Island to the eastern Coast Mountains (Fig. 6A) and more detailed examples of the interpreted nonmarine to shallow marine depositional environments represented by lower units of the Nanaimo Group (Fig. 6B, C, D). Nonmarine and marginal marine facies associations make up most of the Comox Formation and occur extensively in the coal-bearing parts of the Extension, Pender, and Protection formations in the immediate Nanaimo area. Parts of the Extension Formation at the southern end of the basin are also nonmarine.

The most widespread nonmarine deposits occur in the basal unit of the Nanaimo Group, the Comox Formation (Appendix A, Table 1, Fig. 1), which is extensively exposed in the main outcrop areas and present both in the eastern outliers at Vancouver and Lang Bay and also in the western, Alberni, outlier. Thick bedded, massive conglomerates (Benson Member) overlie an irregular unconformity surface, and vary greatly in thickness from a maximum of about 300 m in paleolows, to thin or absent over paleohighs. Many conglomerates are thick, massive, poorly sorted composites of generally locally derived clasts (mostly volcanic and felsic intrusive) as shown in Figure 7A. These are interbedded with graded-stratified conglomerate and coarse sandstone (some crossbedded) and in general change upward or laterally to better sorted stratified pebble conglomerate and pebbly sandstone displaying fining upward trends, crossbedding, erosive bases, and complexly overlapping lateral geometry (Fig. 7B).

The features of these conglomerates and sandstones suggest deposition occurred in alluvial fan and associated fan to braidplain, braided fluvial environments (Fig. 6B). The irregular paleotopography of the unconformity surface is reflected in the generally scattered pattern of paleocurrents for the Comox Formation. Locally, trends are well developed, however, with west to southwest trends in northern and eastern areas probably reflecting orientation of major paleovalley systems as suggested by Atchison (1968) and Kenyon et al. (1992). Paleocurrents in thick paleovalley fills in southern areas (e.g., Saltspring Island and Duncan areas, possibly in part submarine) are more random, but in the northwestern San Juan Islands, conglomerates of the Comox Formation show a northwest radial trend suggesting alluvial fans and fan deltas derived from exposed San Juan terranes to the southeast.

At Nanaimo, conglomerates and pebbly, coarse-grained to granule sandstones make up a significant part of the Extension Formation (Millstream Member) and lower Protection Formation (Cassidy Member), and are associated with the major coal-bearing facies discussed below. Measuring several tens of metres thick, these successions consist of complexly overlapping lenticular beds that are massive to graded-stratified, with common crossbedded sandstone interbeds and rare carbonaceous to coaly mudstone interbeds (Fig. 7C, D). Thick successions of lenticular, erosive based, planar and less common trough crossbedded sandstone also occur. These successions in general display better organization and are better stratified than for the conglomerates discussed above, and are interpreted as deposits in braided fluvial deposits. Pacht (1980) suggested the conglomerate-rich successions resembled the Scott braided river facies type of Miall (1977) and the crossbedded sandstones were similar to the Platte facies type.

Thick successions of nonmarine sandstone-dominated facies also occur in the upper Comox Formation. A succession up to 500 m thick (generally d m) of the Comox Formation in the Nanaimo outcrop area (Saanich Member of England, 1989) and a thinner succession in the Comox outcrop area (Dunsmuir Member, up to 230 m thick) includes thick packages of medium- to coarse-grained sandstone displaying similar geometries and structures to those of the braided stream deposits described above (Platte type). This member also includes rare discontinuous carbonaceous mudstone and pebble conglomerates, and changes to a marginal to shallow marine interval which is transitional to the marine Haslam Formation. Coal beds in the Dunsmuir Member are described below.

A coal-bearing nonmarine to marginal marine facies association is present in northern outcrops of the Comox Formation and at Nanaimo in the Extension, Pender, and Protection formations (Appendix C, Tables 1, 3-5). Most deposition occurred in a coastal to marginal marine setting, which included back-barrier lagoons, interdeltaic swamps, with some coal also forming on sandy braid-deltas and less common coastal braidplain swamps (Fig. 6B). Detailed studies include those by Muller and Atchison (1970), Kenyon et al. (1992), and Cathyl-Bickford (1992b). The Comox Formation coals occur within the Cumberland Member, a succession of fine grained sandstone, siltstone, carbonaceous mudstone, and coal that overlie or intertongue with the alluvial conglomerates and coarse sandstones of the Benson Member. Coal also occurs in the Dunsmuir Member of the upper Comox Formation and is generally associated with marine mudstone, thick sandstone beds commonly containing marine trace fossils, and minor conglomeratic lenses. Mudstones containing marine fossils are less common, but also occur in the Cumberland Formation. In the Nanaimo area significant coal beds occur in the basal Extension Formation (Northfield Member), the Pender Formation, and the central part of the Protection Formation (Reserve Member). The coals are interbedded with fine-to medium-grained sandstones, rare

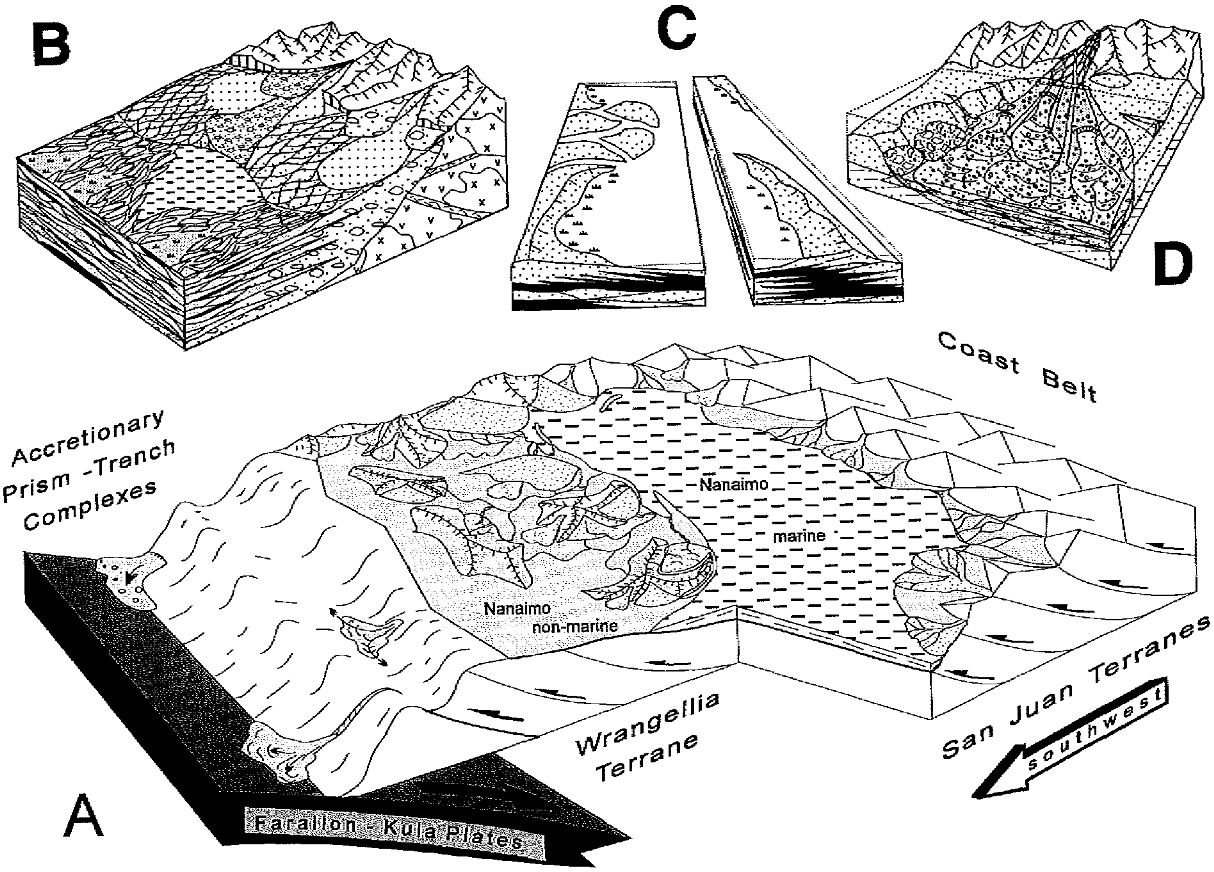
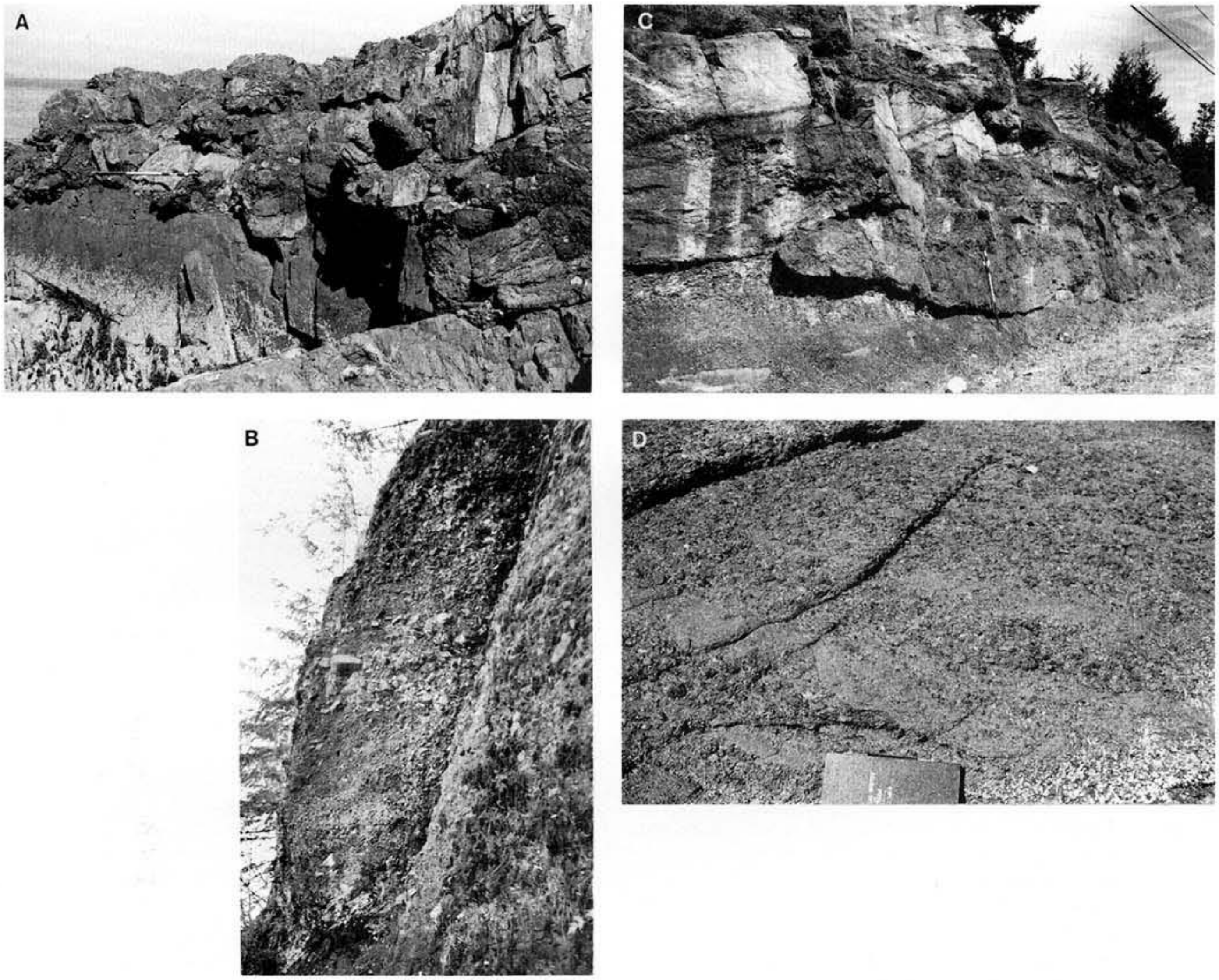


Figure 6. *A) Schematic representation of early Nanaimo Group depositional basin setting within southwest British Columbia. B) Example of a nonmarine alluvial fan and braidplain model, including possible abandoned braid belt swamps where some of the Comox Formation coals may have formed (modified from Wu Chonglong et al., 1992). C) Deltaic and back-barrier depositional environments envisioned for much of the coal- bearing Nanaimo Group (e.g., upper Comox Formation in Comox outcrop area; parts of Extension, Pender, and lower Protection formations in northern Nanaimo outcrop area) (modified from Nemec, 1992). D) Coarse grained fan delta deposystem showing subaerial to subaqueous transition without coastal distributary plain (modified from Nemec, 1992).*

conglomerates, and moderate to abundant mudstone. Most sandstone successions are stratified, overlapping and commonly crossbedded, some show fining-upward cycles typical of meandering channels, but others display the less organized overlapping geometry and crossbedding of braided fluvial deposits. Marine trace fossils and macrofossils are present in some interbeds within coal-bearing facies, and in places littoral bar or possible beach sandstones occur within the units or as capping deposits (e.g., upper Protection Formation). Flaser-bedding is apparent in some sandstone/mudstone successions, possibly as part of a marginal marine transition facies. These coal-bearing units change to marine facies over a short distance laterally in the Nanaimo area, a feature that also indicates coal deposition was coastal to marginal marine. This suggestion of marginal marine deposition for much of the coal facies association is supported by the high sulphur content of the coals and their generally thin, laterally discontinuous geometries.

Marginal to shallow marine facies associations

In addition to the marginal marine parts of the coal-bearing facies described above, parts of several formations contain features suggestive of deposition in littoral to shallow shelf depths. The Comox Formation in several areas contains a few metres to tens of metres of poorly sorted conglomerates whose matrix is rich in broken to articulated shell material, which in several places occurs immediately above the unconformity surface (Fig. 8A). The conglomerates are commonly overlain by coarse stratified fossiliferous sandstone with complex crossbedding, reactivation surfaces, and symmetric and asymmetric ripples. These strata probably represent talus or lag breccias deposited on the irregular unconformity surface, reworked in a coastal high energy marine setting, and buried by coastal shallow bar and beach deposits.



A) *Unconformity surface exposure of basal conglomerate in sharp but irregular contact with underlying Karmutsen Formation volcanic rocks. Unconformity surface is just below and parallel to 1.5 m measuring staff. Comox Formation conglomerate here comprises a poorly sorted mix of angular pebble-to boulder-sized clasts in a volcanic lithic wacke matrix. Most clasts compositions are identical to the subjacent Karmutsen Formation. GSC 1994-711A*

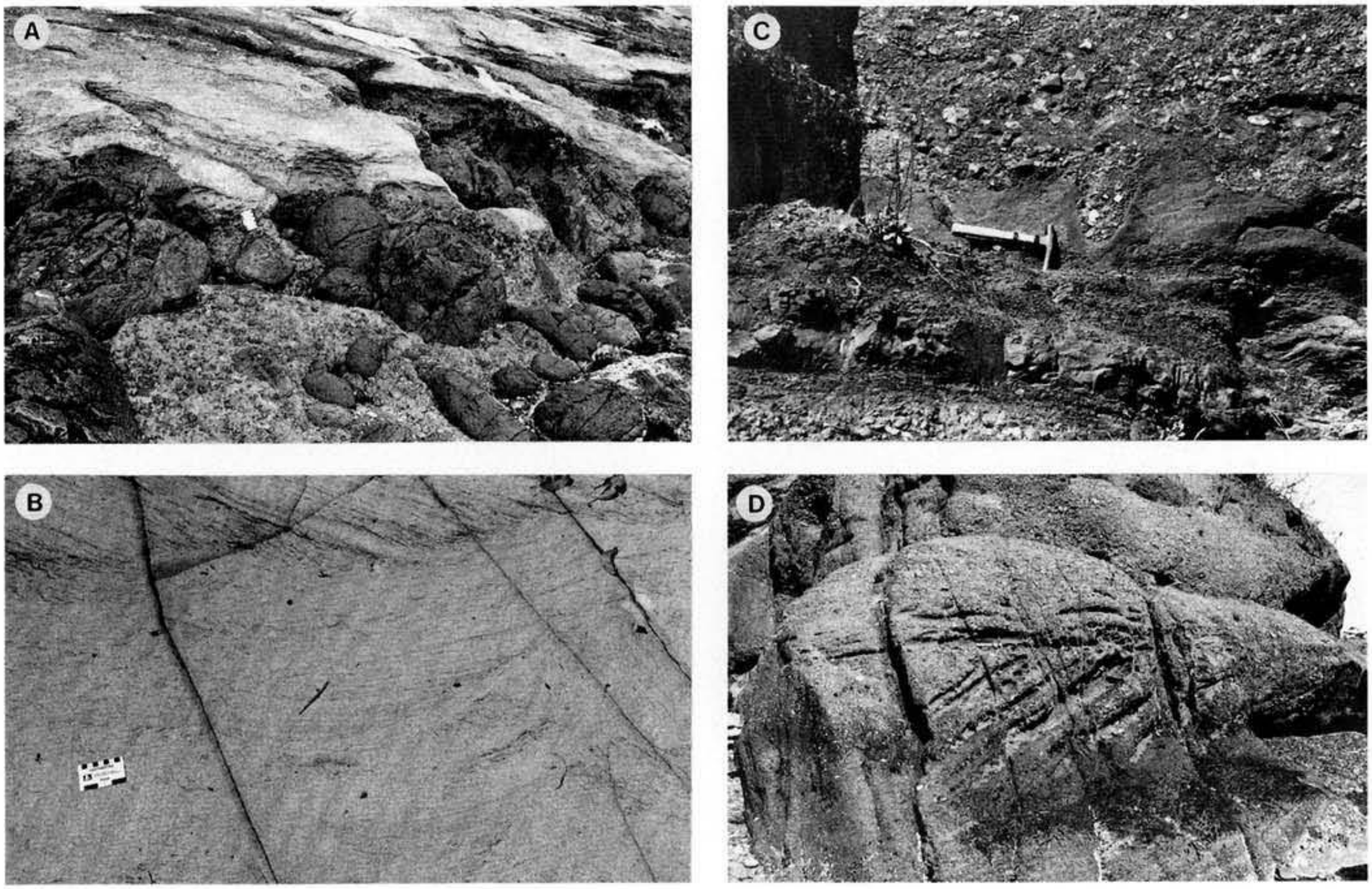
B) *Crudely stratified Comox Formation conglomerate in cliff exposure on north bank of Nanaimo River. This 30 m thick section of clast- and matrix-supported conglomerate consists of a poorly sorted mix of pebble-to boulder-sized clasts crudely organized into thick beds. These are interpreted as alluvial fan debris flow and interbedded streamflow deposits. GSC 1994-711B*

C) *Thick, channelized pebble conglomerate and coarse grained sandstone of the basal Protection Formation in a cliff exposure south of Nanaimo. The overlapping fining-upward lenticels have curved bases erosive into sandstone and mudstone of underlying cycles and are interpreted as deposits in transitional braided to meandering fluvial systems. Measuring staff is 1.5 m long. GSC 1994-711C*

D) *Complexly interbedded and overlapping pebble conglomerate and coarse grained sandstone from the Comox Formation within the Nanaimo townsite. Pebble conglomerate show slight normal grading into overlying sandstone, have curved, erosive bases and are laterally discontinuous over a few metres to tens of metres. These are interpreted as gravel bar deposits in a braided fluvial system. Notebook is 20 cm on longest edge. GSC 1994-711D*

Figure 7. *Typical features of the Nanaimo Group nonmarine facies associations.*

Thicker sequences (generally 10 to d m) of fine- to coarse-grained sandstone with common interbedded lenticular conglomerate also occur in several formations, including the Extension Formation in the northern San Juan Islands, the Comox Formation on Texada Island, the Protection Formation at Parksville and Nanaimo (Protection and Newcastle islands), and the Northumberland to Geoffrey Formation transition near the mouth of the Oyster River. In addition to common shallow marine macrofossils (including coquinas) and trace fossils, these successions contain wavy bedding with common symmetric and asymmetric ripples, complexly overlapping trough crossbed (Fig. 8B) and planar crossbed



A) Basal Comox Formation at Departure Bay, Nanaimo, showing volcanic boulders which unconformably overlie Karmutsen Formation volcanics of identical composition. The white matrix between boulders is granule carbonate sandstone composed mainly of broken shell material, suggesting deposition occurred along a rugged coastline where boulder rubble on the unconformity surface was buried by high energy shallow marine carbonate clastics. GSC 1994-711E

B) Protection Formation sandstone on Newcastle Island (Departure Bay, Nanaimo) displaying complexly overlapping, medium scale trough cross-stratification with planar cross-stratification at top. Presence of trace fossils and rare marine bivalves in this unit supports an offshore bar or megaripple origin in a sandy shallow shelf setting for most of the sandstone in this area. GSC 1994-711F

C) Clast supported pebble cobble conglomerate in lower Extension Formation on South Pender Island. This conglomerate is interbedded with marine trace-fossil-bearing mudstone and sandstone and in this picture includes large burrow structures (right of hammer head), all indicating shallow marine deposition. Features of this thick conglomerate unit on South Pender Island and on islands to the south suggest fan delta deposition in this area. GSC 1994-711G

D) Large scale planar crossbeds in lower Extension Formation fan delta deposits of South Pender Island. These crossbeds indicate west to northwest paleotransport directions and overlie mudstone (light grey rocks in bottom right corner of photo) which contains abundant Skolithos and other shallow marine trace fossils. Hammer in centre of photo is about 32 cm long. GSC 1994-711H

Figure 8. *Typical features of the Nanaimo Group marginal marine and shallow marine facies associations.*

sets (some bidirectional), reactivation surfaces, and large channel scours (especially associated with pebble conglomerates). These probably represent nearshore bar, beach, and barrier island deposits with some component of fan delta input, especially where conglomerates are abundant and large-scale planar crossbeds occur (e.g., parts of Protection Formation in the Nanaimo area).

Thicker facies of conglomerate and coarse-grained sandstone occur in the Extension Formation on the southern edge of the preserved basin, reaching about 500 m thickness at Pender Island. The generally massive to crudely stratified and rarely crossbedded conglomerate intertongues with deep marine shelf mudstone of both the underlying Haslam and overlying Pender formations, and on Pender Island it contains marine trace fossils in thin mudstone interbeds (Fig. 8C, D). On the San Juan Islands to the south the same facies includes extensive coarse sandstones with abundant broken shell material and complex crossbedding, suggesting deposition in shallow marine bars. The association of shallow and relatively deep marine features with the coarse, generally disorganized conglomerates in this area is best explained as alluvial deposition directly into a relatively deep submarine environment, a type of alluvial fan-delta or debris fan system similar to that described by Nemec (1990) and illustrated in Figure 6D. The great thickness, strong westerly paleocurrent trend and chert-rich nature of the conglomerates (Appendix A, Fig. 3) all suggest derivation from the San Juan terranes to the southeast for the Extension Formation in this area.

Successions 10 m to >100 m thick and dominated by horizontally bedded, fine-grained sandstone, siltstone, and mudstone occur in some units (e.g., upper Comox Formation at Chemainus, parts of Protection and Cedar District formations between Nanaimo and Duncan). These beds contain calcareous concretions, starved ripples (symmetric and asymmetric) and rare lenticular beds of graded sandstone or pebble conglomerate. Shell material and bioturbation are common. The successions lack evidence of tidal or high energy shallow marine deposition, but contain relatively shallow water foraminifera and are not obviously part of the submarine fan systems described below. This facies probably represents low energy shelf deposits in which graded sandstone and conglomerate were deposited as prodelta turbidites and in offshore bars.

Massive packages of silty mudstone with only rare sandstone interbeds also occur in several units, most importantly in the Haslam and Pender formations. This facies is generally associated with sandstone-mudstone couplets and is probably a component of submarine fan complexes, as described below. However, in some areas, 10 m to rare 100 m thick sections of massive to poorly and discontinuously laminated silty mudstone occur with articulated macrofossils which do not appear to have been transported. These thick massive sections are in places pervasively bioturbated, and contain microfossil assemblages suggestive of neritic deposition. This facies is most common in parts of the Pender and Haslam formations and, to a lesser extent, in the Northumberland Formation in the Comox outcrop area, but also occurs in the Haslam, Pender, and Cedar District formations south of Nanaimo. Other features of this facies include rare to common carbonate concretions (sideritic in some units), discontinuous starved ripple fine grained sandstones, and rare fine- to medium-grained sandstone interbeds, commonly graded or massive.

The rare wave ripples, articulated fossils and neritic microfossil assemblages suggest these muddy facies were deposited in quiet water outer shelf areas not influenced by major currents or sandy deposystems. The rare sandy turbidites were probably derived from storm-generated or distal delta slope slumping events. A similar muddy facies, described below, occurs with submarine fan facies associations and is interpreted generally as the deposit of interchannel fans. These massive facies are recessive, generally poorly exposed and thus not well-documented. Probably there is a gradation from the non-fan muddy facies to the fan facies, including slope deposits of interfan areas. However, the major slump structures and debris flows commonly present in slope facies have not been observed in these units.

Submarine fan and related facies associations

Most Nanaimo Group facies contain features suggestive of deposition directly or closely related to submarine fan complexes. The first regional study to suggest this was published by Ward and Stanley (1982), who provided a detailed re-interpretation of the Haslam Formation in its southern outcrop areas in terms of a classic submarine fan complex deposystem, and suggested similar models for most units above the Haslam Formation, except for coal-bearing formations in the Nanaimo area. Pacht (1984) summarized his detailed study of the southern areas (Pacht, 1980), and also suggested submarine fan models for most of these units. England (1990) expanded this interpretation to include most strata in northern (Comox) outcrop areas and provided micro- and trace fossil evidence to confirm the deep water setting (generally lower neritic to bathyal) for most deposition. England and Hiscott (1992) provided a submarine fan facies interpretation for the Northumberland and higher formations in the outer southern Gulf Islands.

A wide range of facies models for deep sea fan systems has been published, with many aspects of these systems remaining poorly understood. Recent reviews include those of Walker (1992), Pickering et al. (1989), and Shanmugam and Moiola (1988). Pacht (1980, 1984) considered paleocurrent trends, petrographic analysis, and the pattern of stratigraphic successions in his study to reflect a nonradial elongate fan system, controlled by a basement topography of faulted horsts, with several sediment sources, and with most transport parallel to the axis of the basin. However, other and more recent studies (Ward and Stanley, 1982; England, 1990; England and Hiscott, 1992) are more compatible with the normal inner-middle-outer fan zonation of most radial fan models. This type of model is also suggested by the overall radial regional paleocurrent patterns demonstrated in this study. For these reasons the major features of the submarine fan facies associations are summarized in those terms below and are illustrated in Figure 9A-C.

Inner fan conglomerate-sandstone facies associations

Inner fan facies associations are dominated by conglomerates and pebbly sandstones which occur in overlapping lenticular thick beds with channelized or irregular bases (Fig. 9C). Conglomerates are typically clast-rich (framework-supported more common than matrix-supported) and moderately to poorly sorted with pebbles, cobbles, and rare boulders in a coarse grained, arkosic matrix. Internally, conglomerates range from massive and nonstratified to graded-stratified with pebbly sandstone tops; normal grading is more common than inverse grading, and basal clast imbrication is common in stratified conglomerates (Fig. 10B, C). Most beds appear laterally discontinuous over a few tens to hundreds of metres, and complexly overlap with curved bases which commonly have eroded into underlying beds. Conglomerate clasts are generally subangular to subround and in some places sedimentary clasts are common, including both contorted ripups of semi-lithified mudstone and lithified blocks of older Nanaimo Group strata. Pebbly sandstone interbeds are, in general, more common in upper parts of major successions of this association, occurring as lenticular graded-stratified beds. Mudstone and fine grained sandstone occur as rhythmically interbedded turbidite packages, generally <1 m thick, rarely >10 m.

Thick successions of this facies association make up most of the Geoffrey Formation and parts of the Gabriola Formation (Hornby Island), the Extension Formation (Saltspring Island), and the De Courcy Formation (Denman Island). Most deposition occurred as high concentration turbulent flows in stacked and overlapping channels, or lenticular sheets contained within larger channel complexes (Fig. 10A, D), all features typical of major inner fan channels (e.g., Nelson and Nilsen, 1984; illustrated in Fig. 9C). For example, on Hornby Island the Geoffrey Formation comprises 400 m of conglomerate and pebbly sandstone contained within a major channel form displaying about 200 m of

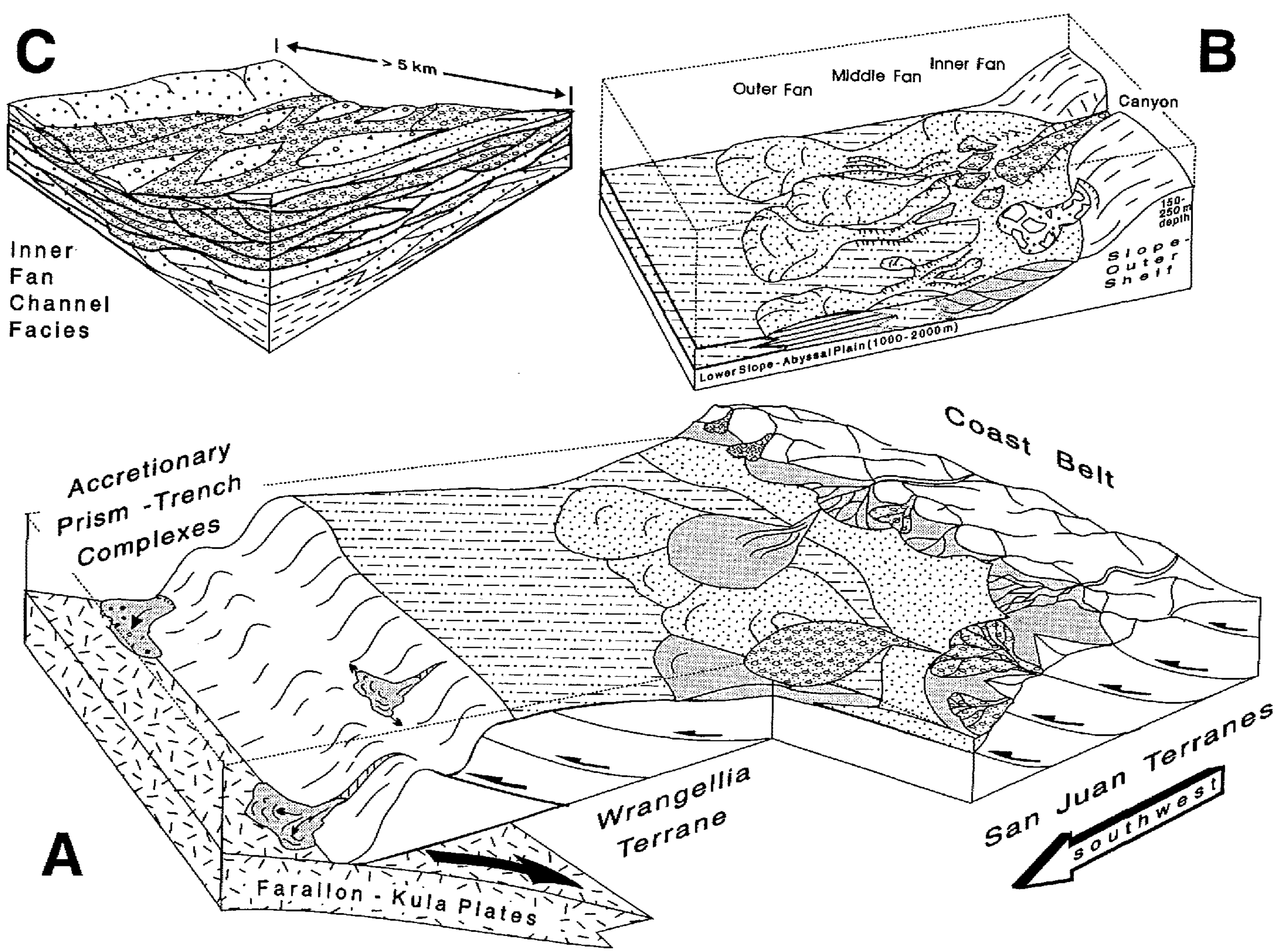
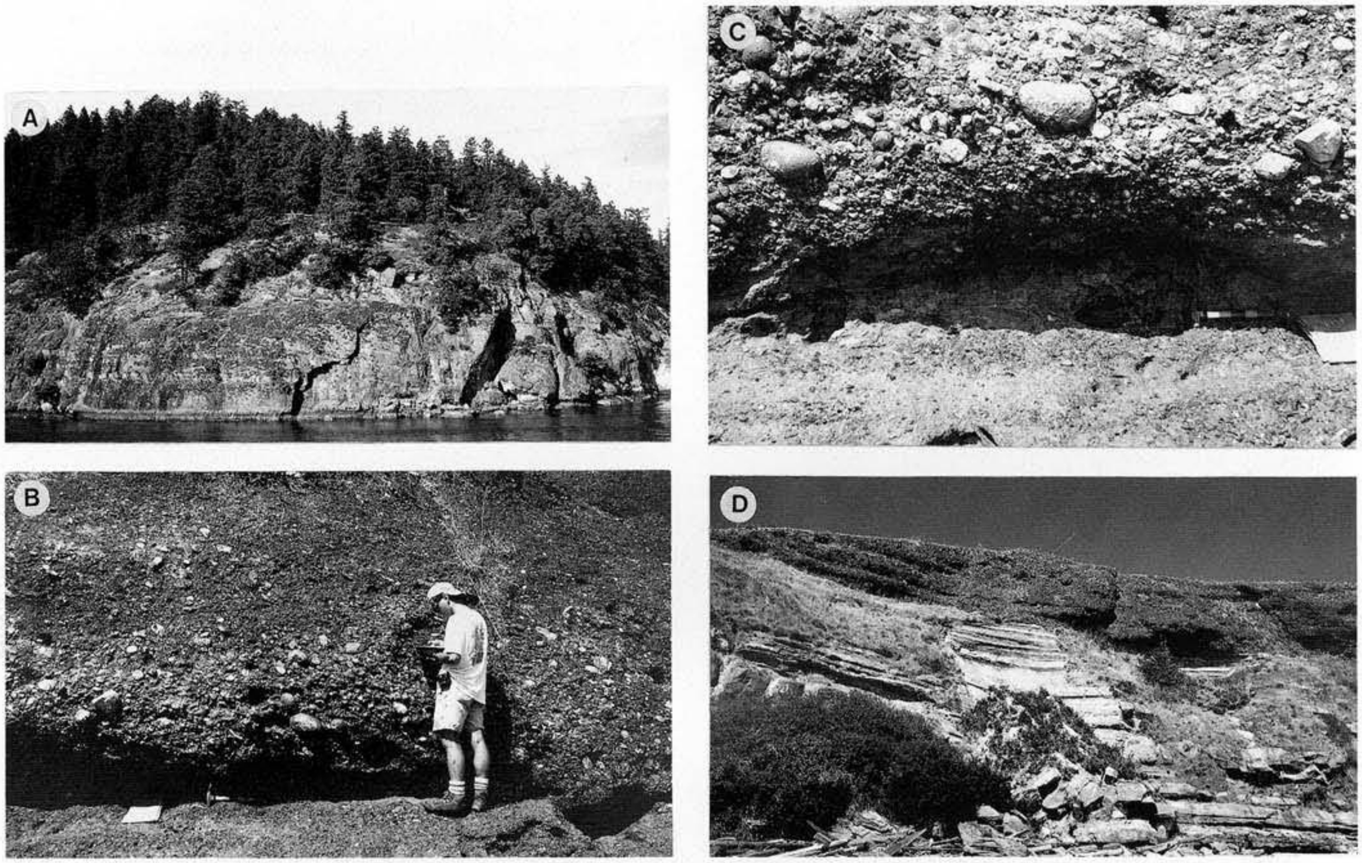


Figure 9. A) Schematic representation of Nanaimo Group basin during marine dominated deposition. B) Example of radial submarine fan sedimentation model envisioned for Nanaimo Group (modified from Shanmugam and Moiola, 1988). C) Example of inner fan channel facies environment (modified from Pickering et al., 1989).

erosion into the underlying Northumberland Formation over a lateral distance of more than 8 km (Fiske, 1977; England, 1990). England and Hiscott (1992) show that the Geoffrey Formation (their Galiano Formation) on the outer Gulf Islands includes a channel complex with about 350 m relief over about 12 km laterally. Smaller channel complexes include the Gabriola Formation at Hornby Island, where 300 m of conglomerates and pebbly sandstones occur within a channel form showing 50 m of erosion into underlying sandstones over several kilometres laterally.

The inner fan facies associations described above are in some ways similar to submarine canyon fills described from other areas (e.g., Morris and Busby-Spera, 1988). However the Nanaimo Group examples are associated with thick-bedded sandstones typical of middle fan channels and supra-fan lobes, and in some places laterally intertongue with interchannel fines and channel-levee deposits (described below). Some laterally discontinuous channel complexes similar to those described above, but contained within finer grained, probably deep shelf mudstone, occur as part of the



A) *Pebble-cobble conglomerate and coarse grained sandstone thick beds organized in overlapping discordant channel systems in inner fan channel facies of the Geoffrey Formation on the south coast of Galiano Island. Progressive overlap of cut and fill beds suggests lateral migration of central channel to the right (east). Paleoflow indicators from several sites on Galiano Island suggest flow directions to the west or northwest. Cliff height is about 50 m. GSC 1994-711I*

B) *Thick cobble-pebble conglomerate bed deposited by sediment gravity flow (gravelly turbidity flow) in the Geoffrey Formation, Galiano Island, displaying overall normal-grading to a crudely-stratified granule sandstone top, but with a reverse-graded basal 20 cm section. GSC 1994-711J*

C) *Reverse- to normal-graded lower part of gravel turbidity flow deposit in Geoffrey Formation on Galiano Island. Contact with underlying graded-stratified pebbly sandstone is at hammer. Note imbrication of tabular clasts, which, in this bed, indicates westerly paleoflow. GSC 1994-711K*

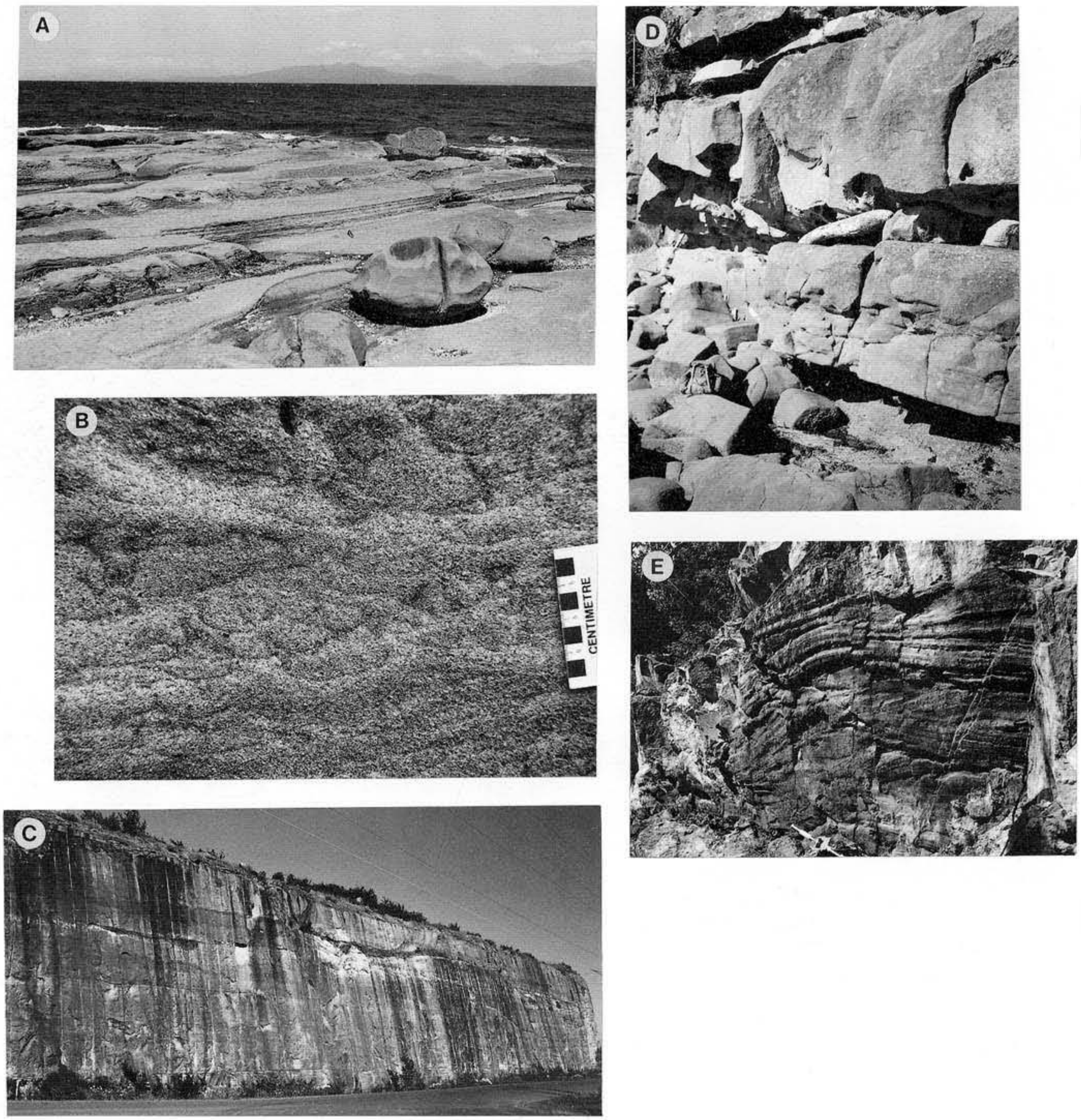
D) *Thick-bedded pebble-cobble conglomerate of the Gabriola Formation, Hornby Island, which occupies a broad channel form eroded into underlying sandstone beds. Truncation of sandstone beds to the east (right in photo) indicates this gravel-filled submarine channel was 50 m deep in this area and 1 km wide. GSC 1994-711L Conglomerate clast imbrication from several sites in this channel complex consistently indicates paleoflow towards the west or northwest.*

Figure 10. *Typical features of the Nanaimo Group inner submarine fan facies associations.*

Extension Formation in the Comox outcrop area west of Parksville and could be true submarine canyon fills (also suggested by England, 1990; Cathyl-Bickford and Hoffman, 1991). In addition, the thick paleovalley fill of Comox Formation conglomerates preserved on Saltspring Island, partly nonmarine, may represent a drowned valley (Hanson, 1976; England, 1990). The Extension Formation conglomerate on Pender Island, interpreted in a previous section as an coarse fan-delta, also probably provided input directly to submarine fan or canyon systems.

Middle and outer fan sandstone-rich facies associations

This broad division includes several facies generalized here into two major associations. The first comprises thick beds of pebbly to medium grained sandstone which are interbedded with thin bedded to massive silty mudstone and rare conglomerate (Fig. 11A), and in many places are organized into crude upward-fining and-thinning successions a few



metres to tens of metres thick, or less commonly upward coarsening/thickening. The sandstones are generally massive, nongraded to normal- or rarely inverse-graded, many with thin upper planar to convolute laminated parts (Bouma TA, TAB, TABC, TBC types where definable; see Pickering et al. (1989) for a review of the Bouma sequence of turbidite sedimentary structures and nomenclature used here). Beds appear laterally continuous for hundreds of metres, but pinchouts are relatively common, suggesting many beds are overlapping lenticels rather than regionally persistent. Load casts and dewatering pipe, sheet, and dish structures are common (Fig. 11A, B); flutes and grooves are rarely present. The thick beds are commonly separated by, or change up to, thin bedded, fine grained sandstone intercalated with laminated to massive silty mudstone (generally organized in Bouma TCDE, TBE, and TDE types). Pebble-cobble conglomerate occurs in discontinuous thick beds, commonly clast-supported and normal graded to nongraded, rarely graded-stratified with basal imbrication and upper crossbeds.

A second major facies association consists of thick successions of laterally continuous, thick massive sandstone beds with only rare silty mudstone interbeds, in some places forming amalgamated massive sets several tens of metres thick (Fig. 11C, D). These sandstones tend to be coarse- to medium-grained, rarely with pebbly bases, internally massive and nongraded to rarely crudely normal-graded, with an upper thin bedded to laminated top. Dewatering structures are common and, where rare silty mudstone interbeds are present, load and flame structures are common and rare flutes and grooves occur on sandstone bases (Fig. 11E). Upward thickening and coarsening successions on 10 to 50 m scales are present in a few places apparent, but generally the beds appear randomly stacked. Rare lenticular sandstones with slightly curved, erosive bases also occur, some with trough or planar internal crossbeds and fining up to a mudstone cap.

The features of these sandstone-dominated facies suggest high concentration, sand-rich, turbidity current deposition in mid-fan to outer fan regions. Broad, shallowly channelized flows account for the fining-upward successions, and nonchannelized flows account for the bulk of the thick sandstone beds, in both suprafan lobes (where they form parts of upward coarsening and thickening successions) and in nonchannelized sheets. They probably reflect rapid deposition in outer fan lobes or as nonorganized sheet systems. Interbedded sandstone-mudstone "classical" turbidites (used in the sense of ideal sand-mud turbidity current deposition, reviewed in Pickering et al., 1989 and Walker, 1992) within the thick sandstone successions probably represent channel fill and interchannel deposition in the middle fan area. Some contain abundant small slumps, typical of channel levee deposits. Similar "classical" turbidite successions commonly occur in transitional contacts at the base or top of the major sandstone formations, and in these areas likely represent major migration of middle and lower fan lobe and channel systems.

These sandstone-dominated facies associations make up the bulk of nonconglomeratic parts of the resistant formations (the Protection, De Courcy, Geoffrey, and Gabriola formations) in the north and south Gulf Islands. Both the broadly channelized sandstone association and the nonchannelized sandstone association occur in all of these units and are complexly interrelated in formation scale successions generally 100 to 400 m thick which commonly intertongue laterally and change gradationally upward to the inner fan facies association, where present, or to the mudstone-sandstone facies described below.

A) Thick, medium-grained sandstone beds interbedded with thin beds of silty mudstone and fine grained sandstone. Deposition was primarily by turbidity current processes with abundant deformation of thin, water-rich beds and mudstone by rapid deposition of overlying thick sandstone beds, especially evident by abundant load and flame structures in lower left and central layers in photo. Weathered out concretion in front right is about one metre in longest dimension. GSC 1994-711M

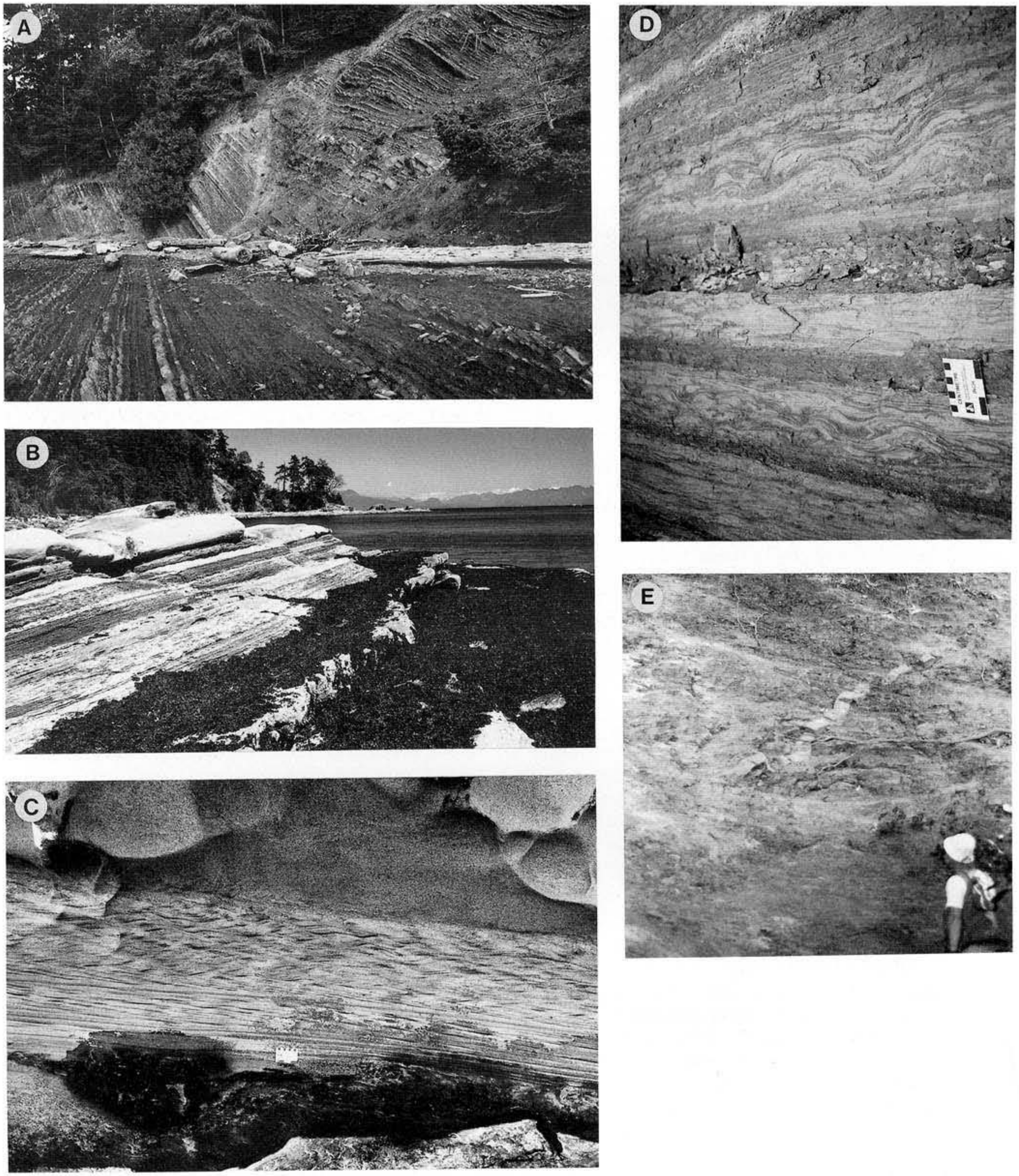
B) Upper part of thick sandstone bed displaying massive fabric and abundant dewatering deformation (dish structures). GSC 1994-711N

C) Amalgamated set of massive medium- to coarse-grained thick sandstone beds in 40 m high cliff exposure of the De Courcy Formation. Bedding is approximately horizontal here (vertical marks are from drilling during road construction) and beds are generally laterally continuous with slightly irregular but nonerosive contacts. GSC 1994-711O

D) Individual thick sandstone beds separated by recessive laminated to massive mudstone from basal Gabriola Formation on Gabriola Island. Sandstone beds are internally massive to very vaguely horizontally stratified and typically display slight normal grading in the upper few centimetres. Backpack in lower left centre is about 45 cm in height. GSC 1994-711P

E) Strongly grooved lower surface of a thick sandstone bed, from the transitional contact between the Pender and Protection formations on South Pender Island. Grooves are oriented about northwest-southeast. Paleocurrent measurements from nearby beds suggests paleoflow was towards the northwest. Divisions on measuring staff (lower centre) are each 10 cm long. GSC 1994-711Q

Figure 11. *Typical features of the Nanaimo Group submarine fan, middle and outer fan sandstone-rich facies associations.*


A
B
C
D
E

Middle and outer fan mudstone-sandstone facies associations

The mudstone-rich units of the Nanaimo Group are dominated by silty mudstone, generally with common sandstone interbeds forming thick (hundred of metres) units of rhythmically interbedded sandstone-mudstone couplets (Fig. 12A, B). However, thick successions, (tens of metres to rarely 100 m) of massive mudstone with only rare sandstone interbeds are also present in some places.

Within the thick successions of sandstone-mudstone couplets, mudstones tend to be silty, massive to faintly laminated, and generally range from a few centimetres to tens of centimetres in thickness. Sandstones are generally thin-bedded and very fine- to fine-grained, but range to thick bedded and medium grained. Sandstones display a range of features including common normal grading, crosslaminated ripples to climbing ripples, convolute laminae, common loading and relatively rare flute casts, and small-scale dewatering structures (Fig. 12C, D; Bouma types TBCDE, TCDE, TABCE, and TBE are most common). Most beds display continuous and even thickness for more than hundreds of metres laterally, but some occur as broad, thin lenses with curved, erosive lower contacts (especially common for thicker and coarser beds). Clastic dykes, synsedimentary slumping and large soft-sediment deformation features are generally rare, but well-preserved and common in some areas (Fig. 12D, E).

Upward fining-thinning successions, 1 to < 10 m thick, are evident in places where medium bedded sandstone (rarely pebbly) occurs at the base, and commonly show broad scour into underlying units. Other apparent fining-thinning or coarsening-thickening trends on these scales have not been statistically tested and can not be defined unambiguously. Walker (1992) suggests this is typical for these types of successions. On a formation scale (hundreds of metres) an overall coarsening and thickening upward of sandstone interbeds is common in the upper parts of formations, as part of the gradational change to the overlying sandstone-rich unit. Fining and thinning upward from the lower sandstone unit is also common. However, intertonguing of these units in many places complicates these trends.

The sandstone mudstone couplets are typical "classical" turbidites which were deposited by waning low-concentration turbidity currents in areas away from the main fan channel and lobe deposition on these sand-rich systems. Thinner successions occurring as intertongues or at contact intervals with the sandstone-dominated formations probably formed by overbank sedimentation as interchannel deposits on middle to lower fan regions. Some contain abundant small slump features and common rippling, typical of channel levee deposits. The thicker successions of couplets probably represent lower fan and fan fringe deposition in areas "distal" to main suprafan lobe deposition sites. The occurrence of these thicker and more laterally continuous mudstone-sandstone

A) *Thick section (about 50 m in photo, but 100 m in outcrop) of laminated to massive mudstone and fine grained sandstone-mudstone couplets in the upper Cedar District Formation on North Pender Island. Beds and laminae are generally laterally continuous across the extent of the exposure (50 m), but some pinch out laterally and occupy broad, low relief channels. Overall the section shows an upward (to right in photo) increase in the abundance and thickness of sandstone beds, as part of a gradational contact with the overlying De Courcy Formation. Deposition is interpreted as by turbidity current processes and the overall upward coarsening and thickening trend may reflect migration of middle submarine fan sand lobes over this lower fan facies. GSC 1994-711R*

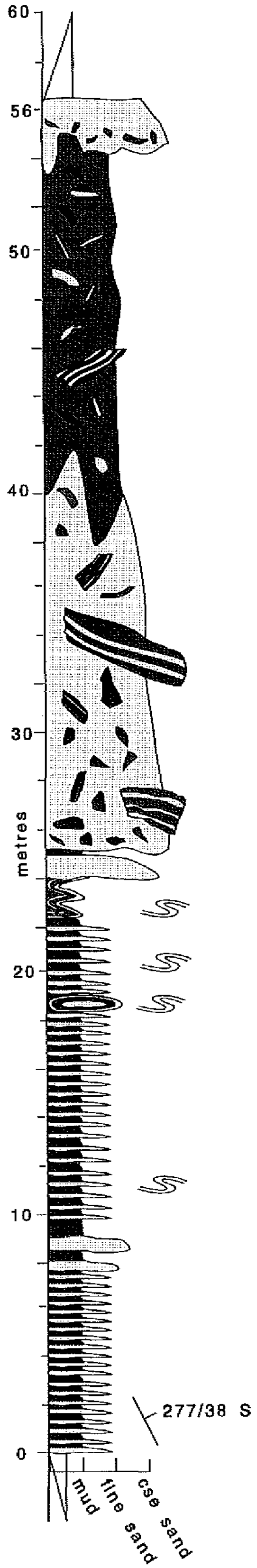
B) *Sandstone-mudstone couplets in the Spray Formation on Gabriola Island. Most thin sandstone beds are rippled at top, displaying Bouma TBCE turbidite features. Massive sandstone at top left is about 40 cm thick. GSC 1994-711S*

C) *Bouma TABC turbidite sandstone bed in the Gabriola Formation on Hornby Island. The bed sequence changes from a massive, medium grained sandstone at base (below scale card) to horizontally laminated fine grained sandstone at the scale card to a capping climbing ripple sandstone and siltstone, all overlain by coarse grained sandstone of the next thick bed. Climbing ripple drift (and thus paleoflow) in this bed is towards the northwest (left in photo). GSC 1994-711T*

D) *Bouma TCDE and rare TBCDE turbidity flow beds in the Gabriola Formation on Hornby Island. Basal planar laminated (B) or convolutely deformed (C) intervals change upward to a laminated siltstone (D layer, best seen left of lower scale card), all capped by laminated to massive silty mudstone (E layer, darker grey in photo). Deposition is interpreted to have occurred on lower fan or inter channel mid-fan areas. GSC 1994-711U*

E) *Clastic dyke of fine grained sandstone cutting massive to vaguely laminated mudstone of the Northumberland Formation on Gabriola Island. GSC 1994-711V*

Figure 12. *Typical features of the Nanaimo Group middle and outer submarine fan mudstone-sandstone facies associations.*



units probably reflects major autocyclic controls (slow lobe and major fan channel migration with time) and allocyclic (eustatic and/or tectonic controls on fan deposition).

Successions of massive to laminated silty mudstone with relatively rare sandstone interbeds that occur with these sandstone-mudstone couplets are probably also deposits of interchannel and outer fan fringe areas, but far enough from channel or lobe deposits that channel levee or crevasse splay deposition is not significant and only the finest overbank material is deposited. Such muddy facies are typical of interchannel areas on modern fans (e.g., reviews in Pickering et al., 1989). Only the close association of these massive successions with the sandstone-mudstone couplets distinguishes them from massive mudstone described in the shallow marine succession, and the two facies may thus be related. They could both include slope deposits, but slump structures and major debris flows are generally not present in this facies.

Chaotic deposits and synsedimentary deformation features

Thick, matrix-supported conglomerates, some containing only brecciated sedimentary clasts, occur as rare isolated beds in most submarine units. Matrix-supported conglomerates, relatively common in the conglomeratic inner fan facies described above, generally contain extrabasinal clasts and a sand-rich wacke matrix. These are interpreted as cohesive debris flows which have been transported along submarine canyons to the inner fan areas, a common feature of inner fan facies. Similar extrabasinal clast debris flows rarely occur in both the sandstone-dominated and mudstone-sandstone facies, some several metres thick and also containing abundant rip-ups of contorted sedimentary blocks (generally concentrated near the base of the bed). These are also interpreted

Figure 13. *Measured section through sedimentary breccia exposed in the central to upper De Courcy Formation on the northwest coast of North Pender Island. This 30 m thick sedimentary float-breccia comprises abundant angular sedimentary blocks in a matrix which changes upward from sand-rich to mud-rich. The upper contact with the overlying sandstone thick bed is extremely irregular, suggesting the mud matrix was water-rich when the overlying sandflow was deposited. Many of the blocks within the thick debris flow are internally deformed, indicating they were not completely indurated when incorporated. A shelf slope or canyon wall collapse origin is proposed for this feature. The section of mudstone-sandstone thin beds below the breccia unit contains several examples of synsedimentary folds (including spectacular sheath folds) and small clastic dykes. The sense of deformation on these folds suggests a northwest-dipping paleoslope for this area.*

as debris flows which have continued from inner fan areas and are sufficiently turbulent at the base to erode and incorporate some underlying material.

Intrabasinal blocks of sedimentary strata occur in the inner fan facies as parts of debris flows and are incorporated into the other conglomerate types. These probably represent failures on submarine canyon slopes, incorporated or reworked into normal flows and transported to inner fan areas. More spectacular sedimentary breccias very rarely occur in other facies. The best exposed and thickest example occurs on northwest Pender Island, in a middle to outer fan sandstone facies of the De Courcy Formation (Fig. 13). Here a single bed about 30 m thick comprises abundant contorted sedimentary blocks up to 5 m diameter floating in a massive, poorly sorted, sand-mud matrix, with an irregular upper surface and common dewatering features. This is interpreted as a slump-generated debris flow, one of the few thick examples preserved in the Nanaimo Group.

Contorted to semi-coherent layers of folded and contorted strata are common in the interbedded mudstone-sandstone turbidite facies, especially in the Cedar District and Northumberland formations on the Gulf Islands. Generally <1 m thick, these contorted layers comprise complexly folded strata contained within unfolded upper and lower strata, indicating a synsedimentary origin. Vergence directions of sheath and other synsedimentary folds generally indicate downslope deformation in directions ranging from west to northeast, but these directions have not been documented in detail. Most common where the mudstone-sandstone units intertongue with sandstone dominated facies, or in the upper gradational contact of those facies, the contorted layers probably represent depositional overloading of the water-saturated mud and sand by thicker sand flows, or oversteepening of slopes in channel margin and channel levee systems. No thick and regionally persistent deformed layers are reported from the Nanaimo Group, suggesting these features probably are not the result of major slope failure or deformation events.

Sandstone dykes and sills and other injection structures occur in some mudstone-sandstone facies (most abundant in Pender, Cedar District, Northumberland, and Spray formations on the Gulf Islands). They are generally thin (a few tens of centimetres) and traceable for tens of metres at most. As with the contorted bedding, these injection features are most common in the intertonguing or upper transition to the sandstone facies or below major conglomerate facies and are also interpreted as failure events of water saturated (overpressured) sands, primarily due to sudden loading of thick sand or gravel flows.

Evidence for provenance, and basin evolution

The major provenance trends for the Nanaimo Group are indicated by regional changes in paleocurrent patterns, by variations in sediment thickness and type, by compositions of sandstone and conglomerate clasts, and from U-Pb dates of detrital zircons.

Most early workers (e.g., Muller and Jeletzky, 1970 and most pre-1980 graduate theses) interpreted Wrangellia terrane as the major source of Nanaimo Group sediments, with a smaller component of detritus coming from the north Cascades and Coast Belt, mostly in the upper formations. More recent studies (e.g., Ward and Stanley, 1982; Pacht, 1984), which are based largely on sandstone compositions and paleocurrent trends, suggest that detritus from Wrangellia terrane is the predominant component only in the basal Comox Formation. The major source areas for all but the Comox Formation appears to have been the northern Cascade terranes of the San Juan Islands and mainland Washington State to the southeast, and the Coast Belt to the east, although in the Nanaimo area, a Wrangellia terrane source has generally been postulated for the coal-bearing Extension, Pender, and Protection formations.

The compilation of all known paleocurrent, sandstone and conglomerate clast data herein provides a much more complete dataset for evaluating the provenance of Nanaimo Group sediments than previously available. For example, Pacht (1984), in his extensive study, used 566 paleocurrent measurements from about 30 sites and point counts from 152 sandstone samples, most from De Courcy or older formations, and including a substantial number from Sucia Island sandstones now known to be of Paleocene age. The present study includes more than 2800 paleocurrent measurements from about 130 sites, 463 sandstone sample point counts (which excludes samples from the Paleocene rocks of Sucia Island), and 78 conglomerate clast composition counts (generally 100 counts per site). These compilations are presented in several ways. For individual formations, paleocurrents and conglomerate clast counts are shown in correct geographic position on the summary maps of Appendix A. Sandstone detrital compositions are also shown on these figures, using ternary diagrams with the detrital framework compositions subdivided into the grain populations recommended by Dickinson and Suczek (1979), and also divided into four arbitrarily defined geographic divisions from southeast to northwest (shown in Appendix A, Fig. 1A). Ternary plot apices and provenance divisions are defined in Figure 14. Inner dashed lines represent inferred provenance type divisions of Dickinson and Suczek, as modified by Dickinson et al. (1983). Changes in sandstone major component abundances between formations are also shown graphically in Figure 15. Sandstone and conglomerate composition summaries are shown for major time divisions, and for the Nanaimo Group as a whole, in Figure 16 (sandstone compositions for the entire Nanaimo Group are also shown in Fig. 14). Tables 1 and 2 provide numerical summaries and sources of information for the sandstone detrital composition compilations. Table 3 provides the same information for the conglomerate clast compositions.

The data summarized in these figures and tables demonstrate a wide range of compositions for Nanaimo Group detritus. Most sandstones are immature or submature, moderately-sorted feldspathic arenite. However both chert-rich and lithic-rich sandstones are abundant locally. The matrix is generally abundant, but less than 15%, which is the arenite/wacke boundary used by most classification schemes, e.g., Pettijohn et al. (1973); (others use a 10% cutoff, e.g., Dott, 1964), although true wacke is present in most formations. A carbonate cement is typical, with less common silica and rare chloritic to clay cements also reported. Diagenetic alteration products are common, including zeolite group minerals which decrease original porosity to less than five per cent in most places, and the alteration of original grains into pseudomatrix (Page, 1972, provides a detailed discussion of this and other diagenetic effects).

Changes in sandstone composition between formations is shown graphically in Figure 15 (but note the large standard deviations of these values in Table 1). In general the arkose is plagioclase-rich with plagioclase content increasing markedly in upper formations (generally Protection Formation and above). Polycrystalline quartz (most as chert) is a significant component of most sandstones, but is noticeably abundant in the lower formations, especially the Haslam to Pender formations. Sedimentary rock fragments (mostly carbonate and argillaceous fragments) show a marked decrease in abundance in upper formations. Volcanic rock fragments (generally mafic) are a minor but important component of all formations and show no strong trends in abundance. Pacht (1980, 1984) noted similar trends in his study, and in addition subdivided the sandstone of the Nanaimo area and south into five petrofacies he related to specific changes in source areas, both with time and in different parts of the area studied.

The trends apparent from this compilation generally support the interpretations of Pacht (1980, 1984) and Ward and Stanley (1982) for major source areas, but with some notable differences. The Comox Formation shows clear evidence of essentially local derivation. The scattered paleocurrent patterns, nonmarine to shallow marine facies, and control on thickness and conglomerate abundance by the rugged paleotopography all support this interpretation. Both the sandstone

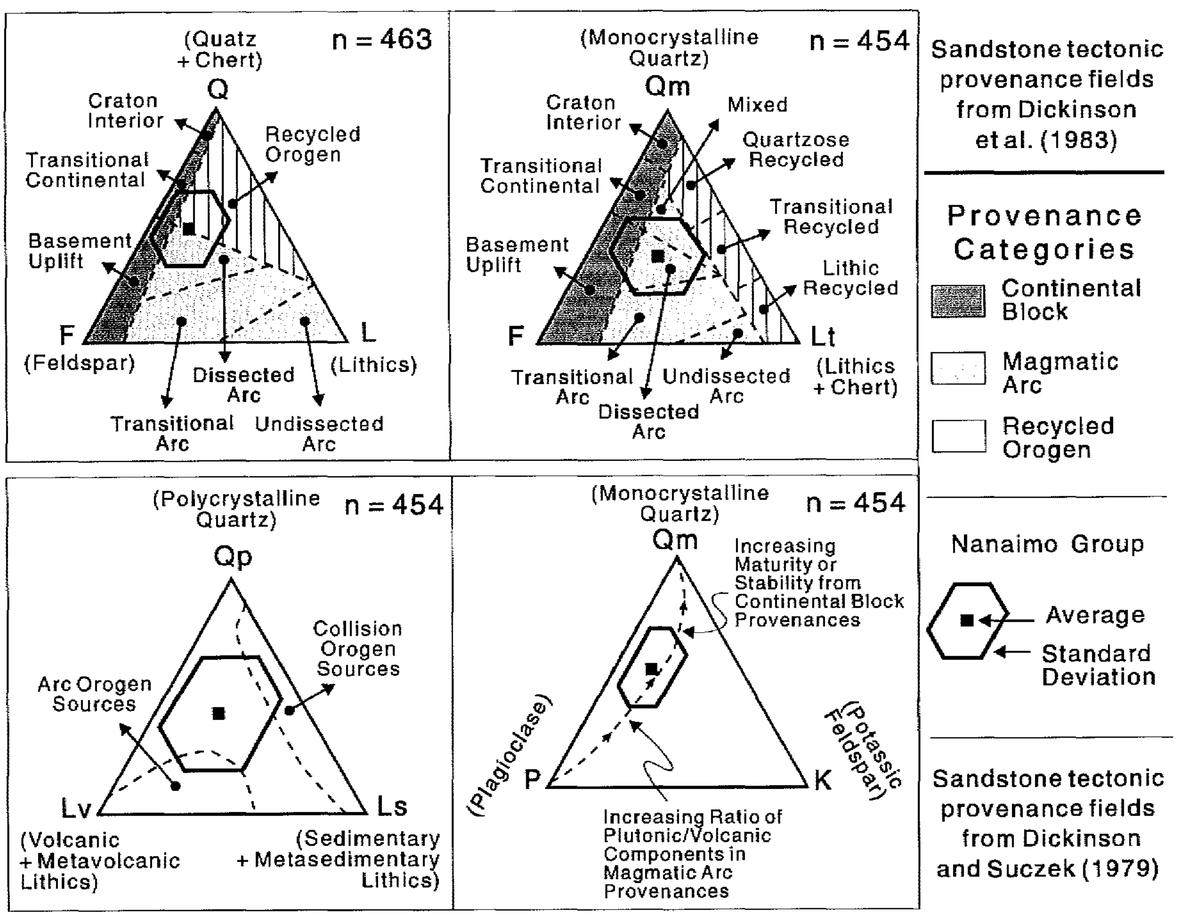


Figure 14. Sandstone detrital framework compositions subdivided into grain populations and plotted on ternary diagrams showing the tectonic provenance fields of Dickinson et al. (1983, QFL and QmFLt plots) and Dickinson and Suczek (1979, QpLvLs and QmPK plots). The mean value of all Nanaimo Group sandstone samples is shown (black square) with a polygon representing the standard deviation from the mean. Sources of information are summarized in Table 1. Note that for nine samples grain types were not identified in the original study to the level allowing plotting on any but the QFL plot, accounting for the change in total number of samples used between the QFL plot and the other plots.

and conglomerate clast compositions vary widely for different sites, and generally reflect the composition of local basement. For example, conglomerates of the Comox Formation in the Vancouver area contain 50% felsic intrusive clasts, and likely reflect the abundance of local Coast Belt sources. A notable increase in volcanic clasts in the sandstone versus sedimentary and chert clasts is also apparent between southern areas and the Nanaimo and northern areas (Fig. 6, QpLvLs plot). This probably reflects the increased importance of volcanic sources such as the Karmutsen Formation in the north, and an increased sedimentary component derived from the Sicker Group in the south. In the Comox outcrop area, Atchison (1968), observed abundant locally-derived volcanic and intrusive clasts in the Comox Formation conglomerates and documented marked thickening of the conglomerates towards paleohighs (note that specific clast compositional data of Atchison , 1968 were not reported in his original study and thus are not shown on Appendix A, Fig. 2 or used in the summary Fig. 16).

A dramatic increase in chert-rich sandstones and conglomerates is apparent for the uppermost Santonian to lower Campanian Haslam, Extension, and Pender formations. As noted by Ward and Stanley (1982) and Pacht (1984), this corresponds to a general southeast thickening of these formations, an increase in conglomerate abundance in the Extension Formation, and intertonguing of shallow marine conglomerates with the mostly deeper marine Haslam Formation in southeast areas. Paleocurrent patterns in the southern part of the basin also strongly support an overall east or southeast source. Although chert is known in Wrangellia rocks and is a major part of the Bridge River terrane of the southeastern Coast Belt (Monger and Journeay, 1994), it is abundant in some of San Juan terranes (e.g., Orcas Chert, Brandon et al., 1988). Pacht (1980) reported that radiolaria from chert clasts are identical to the radiolaria of the San Juan terranes. It seems probable that the San Juan terranes were the major source region for these formations in at least the southern part of the basin. Pacht (1984) suggested sandstones in the Nanaimo area are not chert-rich and contain a noticeably higher abundance of basaltic clasts than elsewhere, reflecting a local Wrangellia terrane source from the west or northwest. The more complete dataset of this study does not support this interpretation. Chert-rich sandstones are still abundant in the Nanaimo area, and conglomerates of the Extension formation also contain abundant chert. The amount of volcanic material in these sandstone is not noticeably more common. However, paleocurrents in this area do show more complex patterns than to the south, reflecting the coal-bearing shallow marine to non-marine facies preserved in this area, which suggests a more complex provenance than a simple Cascades source. As discussed below, detrital zircons from the Extension Formation in this area suggest that the major source was the western Coast Belt and perhaps the northwest Cascades.

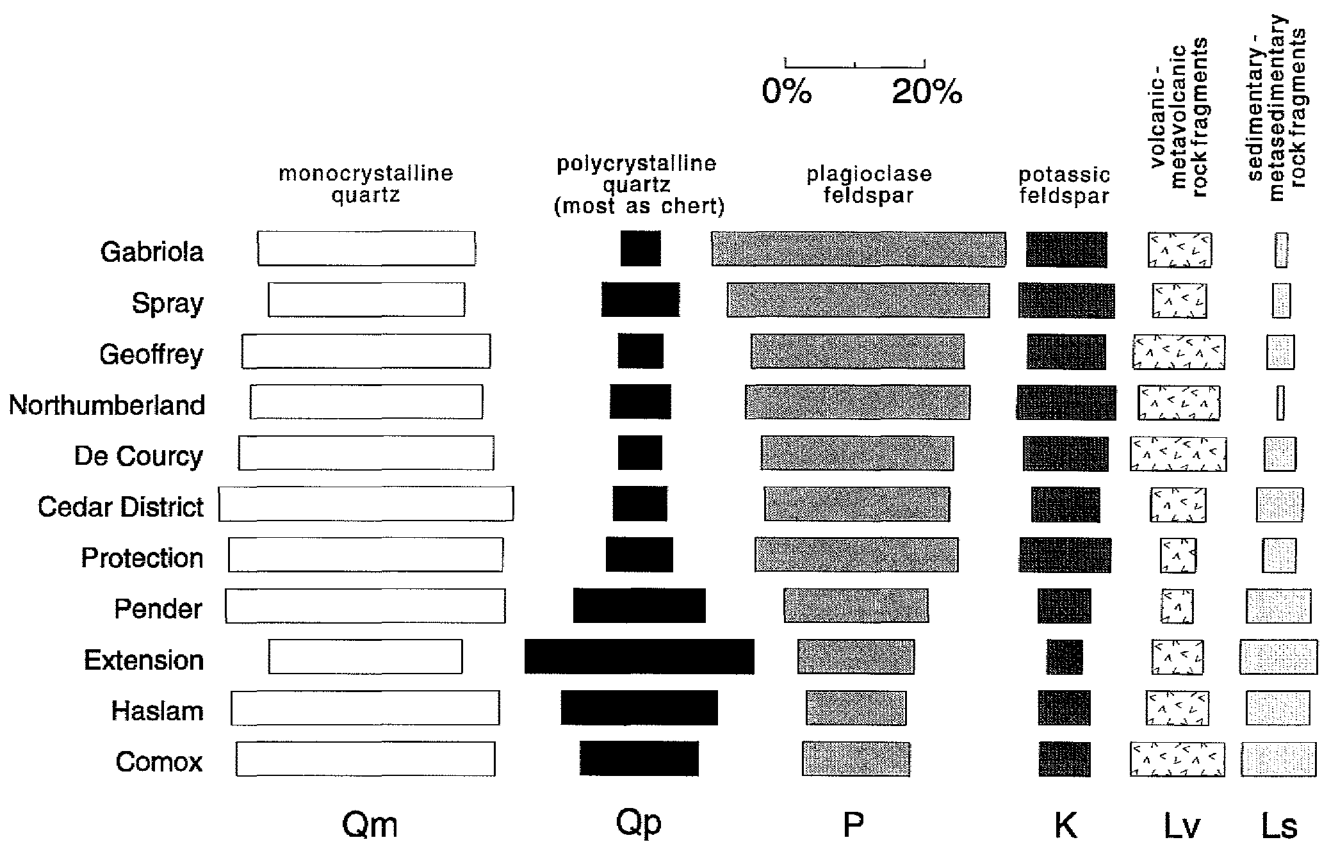


Figure 15. Mean values of major sandstone constituents subdivided by formation and shown graphically in correct relative stratigraphic position. Bar width corresponds to mean per cent. Mean values, standard deviations sources of information and numbers of samples for each formation are given in Table 1.

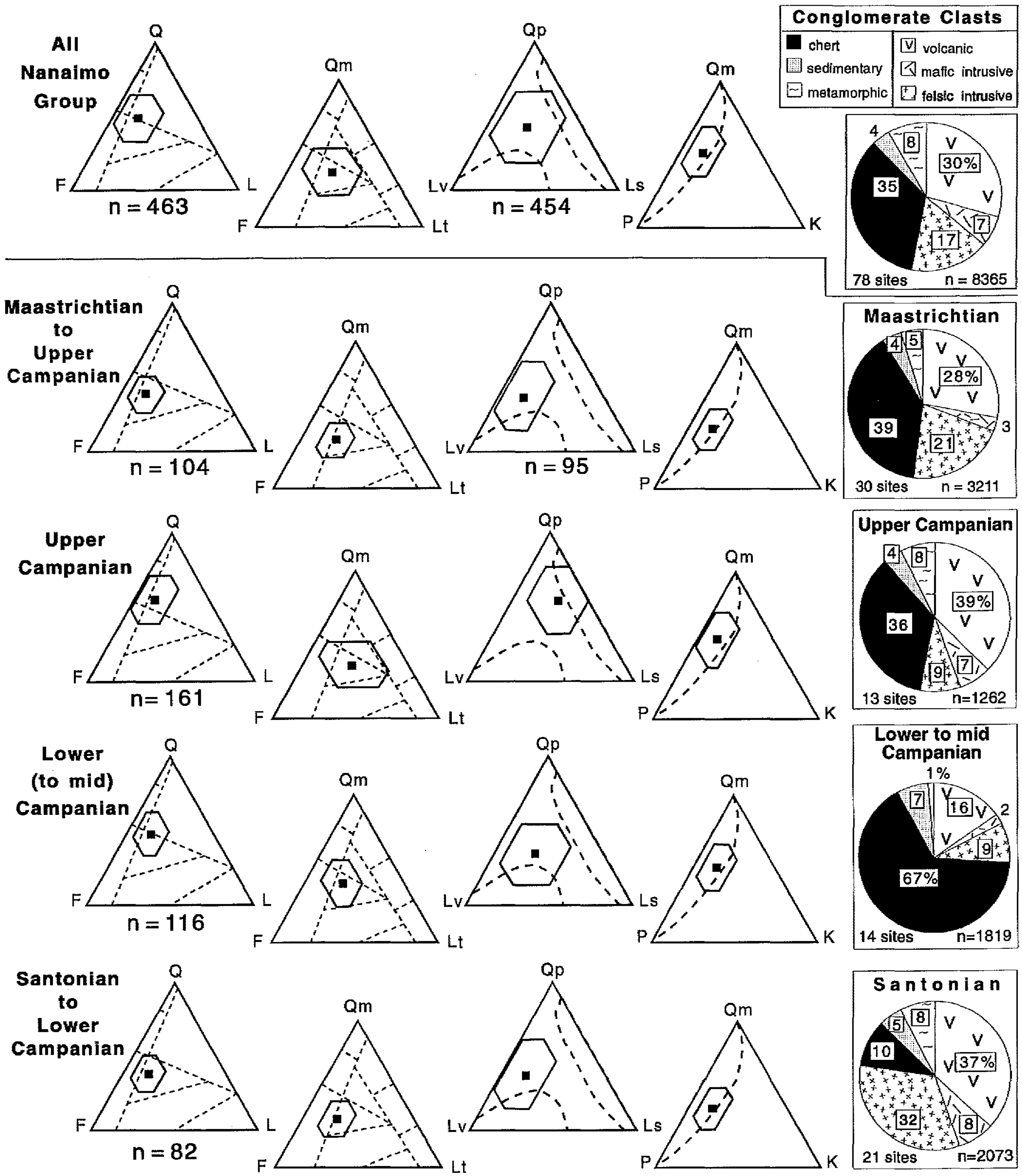


Figure 16. Variations of Nanaimo Group sandstone and conglomerate clast compositions with time. Sandstone detrital framework mode subdivisions and tectonic provenance divisions of plots are defined in Figure 17B and are subdivided using the format of Dickinson and Suczek (1979). The mean value for each time slice sample population is shown (black square) with a polygon representing the standard deviation from the mean. Subdivision of sandstone samples into time populations is imprecise for samples from some sites where the formation age is not tightly constrained, resulting in some overlap of ages for the populations (most relevant to lower Campanian sample sites).

Protection Formation and younger strata in general are dominated by plagioclase-rich arkose, although chert and other components are present and often abundant. Paleocurrents show a general radial trend toward the west, but with considerable variability. Facies are generally deeper marine, although shallow marine and rare marginal marine facies are present (especially for the Protection Formation in the Nanaimo area). Conglomerate clast types vary widely, with chert, intermediate to mafic volcanic, and felsic to intermediate plutonic types most abundant. In general, southern areas are slightly to abundantly richer in chert clasts, while more northerly areas are more volcanic-clast rich (especially notable for De Courcy Formation conglomerates, Appendix A, Fig. 8). Overall, an upward increase in plagioclase and decrease in chert is evident in the sandstones (Fig. 15). The general conclusions of Ward and Stanley (1982) and Pacht (1984) that the Coast Belt and to some extent the Cascades were the source area for the upper units of the Nanaimo Group is compatible with these observations.

Uranium-lead ages of twenty-two detrital zircons from three formations of the Nanaimo Group provide new evidence of changing source areas with time for Nanaimo Group deposition. The results are summarized here; a detailed discussion of the results and implications is presented separately (Mustard et al., 1994).

Table 1. Mean framework modes of Nanaimo Group sandstones subdivided in terms of individual formations. Sources of information numbers at right correspond to references listed at bottom. In cases where samples in the original study were identified as part of a different formation then is currently believed correct, the data has been plotted with the formation currently thought correct. Data was not used in cases where more recent work has shown the original strata sampled are not part of the Nanaimo Group (e.g., Pacht, 1980 included data from Sucia Island and nearby islands, then believed part of the De Courcy Formation, but now correlated with the lower Tertiary Chuckanut Formation).

Formation		F	Q	L	Qm	Qp	P	K	Lv	Ls	Matrix	Sources of Information
Gabriola	Mean (%)	52	34	14	31	6	42	11	9	2	14	2, 4, 12, 13, 17, 19
N = 29	Std. Dev. (%)	9	13	11	11	4	8	5	8	1	7	
Spray	Mean (%)	51	39	10	28	11	37	14	8	3	9	2, 4, 7, 11, 17,
N = 15	Std. Dev. (%)	14	14	9	11	8	11	5	8	3	6	18, 19
Geoffrey	Mean (%)	42	41	18	35	6	30	11	13	4	12	2, 4, 5, 6, 7, 11, 12,
N = 60	Std. Dev. (%)	11	13	11	14	4	12	6	10	4	6	13, 16, 17, 18, 19
Northumberland	Mean (%)	46	42	12	33	9	32	14	12	1	10	1, 2, 4, 11, 16, 17, 18
N = 41	Std. Dev. (%)	18	14	12	12	5	13	7	12	1	7	
De Courcy	Mean (%)	39	42	18	36	6	27	12	14	4	10	1, 4, 5, 6, 7, 11,
N = 60	Std. Dev. (%)	10	12	10	14	5	11	8	10	4	5	13, 16, 17, 18, 19
Cedar District	Mean (%)	36	50	14	42	8	26	10	8	6	11	1, 3, 6, 7, 11, 14,
N = 87	Std. Dev. (%)	17	12	11	13	7	15	6	8	6	5	16, 17, 18, 19
Protection	Mean (%)	42	48	10	39	9	29	13	5	5	7	1, 3, 7, 11, 15,
N = 54	Std. Dev. (%)	15	12	9	11	8	13	6	8	5	6	18, 19
Pender	Mean (%)	28	59	13	40	19	20	7	4	9	8	3, 6, 7, 10, 11,
N = 17	Std. Dev. (%)	23	20	12	17	21	19	9	7	9	5	15, 19
Extension	Mean (%)	22	60	18	28	33	17	5	7	11	6	3, 6, 7, 10, 11,
N = 37	Std. Dev. (%)	14	37	12	11	39	11	8	9	8	4	15, 19, 20
Haslam	Mean (%)	22	61	18	38	22	14	7	9	9	12	3, 6, 9, 10, 11, 20
N = 28	Std. Dev. (%)	9	16	14	15	11	6	7	11	6	10	
Comox	Mean (%)	22	54	24	37	17	15	7	13	10	7	3, 6, 8, 9, 10, 13,
N = 54	Std. Dev. (%)	8	15	13	17	11	8	7	16	12	6	15, 19, 20

References

1. Allmaras, 1979	8. Johnson, 1978	16. Simmons, 1973
2. Carter, 1977	9. Kachelmeyer, 1978	17. Stickney, 1976
3. Fahlstrom, 1982	10. Mercier, 1977	18. Sturdavant, 1975
4. Fiske, 1977	12. Packard, 1972	19. Thom, 1983
5. Grieve, 1974	13. Page, 1972	(cited in England, 1990)
6. Hanson, 1976	14. Rahmani, 1968	20. Ward and Stanley, 1982
7. Hudson, 1974	15. Ruddiman, 1980	21. Mustard, this study

Table 2. Mean framework modes of Nanaimo Group sandstones subdivided into major time divisions. Subdivision of sandstone samples into time populations is imprecise for samples from some sites where the formation age is not tightly constrained, resulting in some overlap of ages for the populations (this is most relevant to lower Campanian sample sites).

		F	Q	L	F	Qm	Lt	Lv	Qp	Ls	P	Qm	K
Combined	Mean (%)	35	49	16	35	37	27	33	43	24	35	52	14
Nanaimo Grp	Std Dev	16	16	12	16	16	18	26	24	19	16	17	10
		N=463			N=454			N=454			N=454		
Maastrichtian	Mean (%)	47	39	15	47	33	21	48	37	15	44	41	15
Geoffrey	Std Dev	12	12	9	13	12	10	23	24	14	15	14	8
to Gabriola formations		N=104			N=95			N=95			N=95		
Upper	Mean (%)	37	48	15	37	40	22	41	34	25	35	51	14
Campanian	Std Dev	12	15	10	12	17	13	24	21	18	15	16	9
Cedar District to Northumberland fm					N=161 for all								
Lower	Mean (%)	32	55	13	32	37	31	17	55	27	33	55	12
Campanian	Std Dev	18	16	11	18	15	23	19	23	18	16	17	10
Extension to Protection fm		N=116											
Santonian	Mean (%)	22	56	22	22	38	40	24	48	28	26	61	13
(to low Campani	Std Dev	9	16	16	9	17	18	25	19	23	11	17	12
Comox + Haslam fm		N=82 for all											

Table 3. Summary of clast compositional data for formations containing significant conglomerates. Clast composition subdivisions are shown at top, with the main rock types in each subdivision listed below the general name. In cases where samples in the original study were identified as part of a different formation than is currently believed correct, the data has been plotted with the formation currently thought correct. Sources of information numbers at right refer to the listed references at the base of Table 1.

DETAILED CLAST TYPES (%)

Formation		Mafic Volcanic mafic volcanic metavolc	Mafic Intrusive mafic dyke diorite gabbroic	Felsic Intrusive felsic dyke granitic granodiorite	Chert chert quartzite	Sedi-mentary sandstone siltstone limestone	Meta-morphic schist gneiss phyllite	Sources of Information
Gabriola	MEAN	43	7	18	24	6	2	2, 4, 21
n = 846	STD DEV	18	5	8	17	5	4	
Geoffrey	MEAN	21	3	21	49	4	3	2, 4, 6, 12
n = 2365	STD DEV	14	3	11	12	5	4	16, 17, 21
De Courcy	MEAN	39	7	9	34	4	8	1, 6, 7, 16
n = 1262	STD DEV	21	10	7	28	7	6	17, 18
Extension	MEAN	16	1	9	67	7	1	3, 6, 7, 15
n = 1819	STD DEV	21	2	7	28	8	1	
Comox	MEAN	37	8	32	10	5	8	6, 8, 9, 21
n = 2073	STD DEV	18	8	22	9	4	14	

Three samples were collected, all from the Nanaimo area. Sandstone of the Extension Formation and the Protection Formation from the southern Nanaimo townsite were sampled to test interpretations that they were derived from a local Wrangellia terrane paleohigh to the west or northwest (Pacht, 1984; England, 1990). A sample from the base of the Gabriola Formation on Gabriola Island was also sampled to test the Coast Belt source generally suggested for this submarine fan facies (e.g., England and Hiscott, 1992).

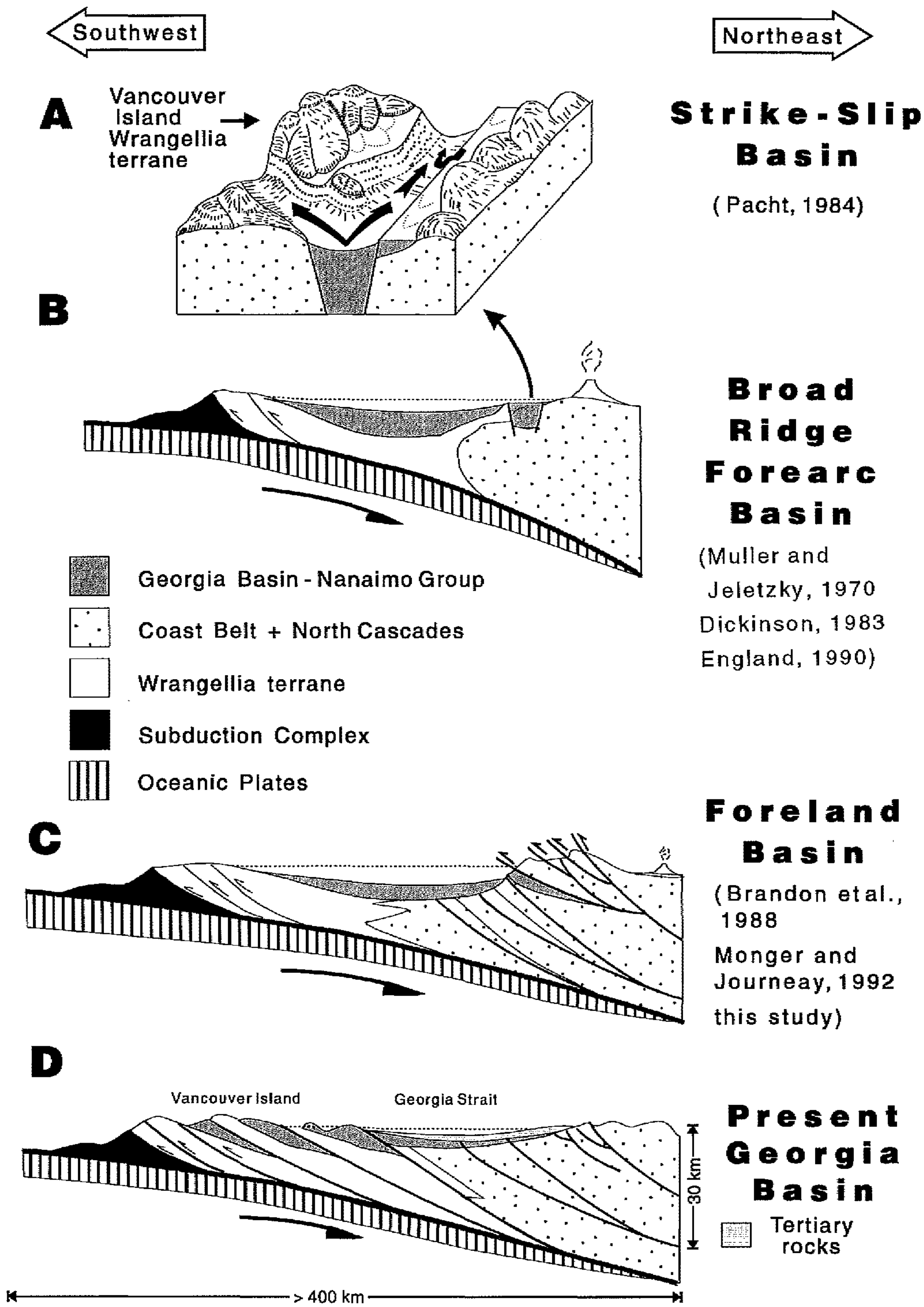
The detrital zircon ages strongly indicate that the Coast Belt and northwest Cascades were the dominant source even in the Nanaimo area by at least earliest Campanian time and that Wrangellia terrane probably was not a major source of detritus. Most zircons from the lower Campanian Extension and Protection formations are Early Cretaceous (ca. 110 Ma, 120-125 Ma) or Late Jurassic (mostly ca. 145 to 155 Ma, but including 159 and 167 Ma zircons). These ages are all younger than possible Wrangellian sources, but are typical of the western Coast Belt. Two zircons with ages of ca. 220 and 320 Ma are most likely from the San Juan terranes, which contain igneous rocks of both Late Triassic (Haro Formation and Deadman Bay volcanics) and Late Mississippian-Early Pennsylvanian (East Sound Group) age (Brandon et al., 1988), or from other parts of the northwest Cascades (e.g., Marblemount Intrusives with 220 Ma U-Pb age, Mattinson, 1972). Paleozoic igneous rocks in Wrangellia terrane are Devonian or older (e.g., Parrish and McNicoll, 1992), precluding this terrane as a source for the 320 Ma grain. As neither the basaltic volcanics of the Triassic Karmutsen Formation of Wrangellia terrane, nor associated mafic intrusives contain significant zircon, it is unlikely that the 220 Ma grain came from a Wrangelli source.

The submarine fan Gabriola Formation sandstone (Maastrichtian age) contains abundant zircon with ages as follows: Precambrian zircons, zircons of late Mesozoic age with Precambrian inheritance, 87 Ma, and a predominant 72-73 Ma population. These surprising results indicate derivation from varied sources which may include the eastern Coast Belt (87 Ma), Paleozoic or late Precambrian sedimentary rocks (recycled Precambrian zircons), and eastern Cordillera (Idaho Batholith and Omineca Belt) plutons (72-73 Ma grains and grains with inheritance). Thus, the source areas for upper Nanaimo Group sediments appear to be considerably more widespread then previously believed, suggesting that major fluvial drainage systems bisected the western Cordillera during the last stages of Nanaimo Group deposition.

A common interpretation in most previous provenance studies of the Nanaimo Group has been that abundant intermediate to mafic volcanic detritus in the sandstones and conglomerates reflects a source from Wrangellia terrane rocks exposed to the west to northwest of the strata (e.g., Muller and Jeletzky, 1970; Pacht, 1984 for Nanaimo area; England, 1990 for Nanaimo area). However, Wrangellia terrane strata are known from the eastern side of Georgia Strait as septa within the dominant plutonic bodies of the western Coast Belt (Roddick and Woodsworth, 1979; Friedman and Armstrong, 1990; Monger, 1991a; Monger and Journeay, 1994). In addition, thick successions of arc volcanics and associated clastics of the Early Cretaceous Gambier Group are preserved as pendants in the Coast Belt (e.g., Lynch, 1991) and probably formed an extensive cover over Coast Belt plutons in mid- to Late Cretaceous time. Thus abundant volcanic material was present east of the present Georgia Strait during Nanaimo Group sedimentation. Examination of isolated sedimentary outliers of the basal Nanaimo Group on the eastern side of Georgia Basin by Mustard and Rouse (1991, 1994) and in the Vancouver area by Rouse et al. (1975) shows that these areas were nonmarine to shallow marine environments with sediment derived from local sources. For these reasons the volcanic-rich material of most Nanaimo Group strata is believed compatible with a western Coast Belt source, as suggested by the detrital zircon evidence.

A different method of examining Nanaimo Group provenance uses the tectonic provenance discrimination plots defined by Dickinson and Suczek (1979) and since modified by several workers (most notably Dickinson et al., 1983). The premise of these plots is that the detrital framework modes of sandstone suites can be related to the tectonic setting of basins of deposition and the associated provenance, although sandstone composition also reflects factors such as diagenesis, types of transport, etc. (reviewed in Suttner, 1974). Figure 14 illustrates the tectonic provenance fields and defines the framework mode apices for the different plots. The mean value, outlined by a standard deviation envelope, is shown for the entire Nanaimo Group (Fig. 14 and top of Fig. 16) and subdivided suites based on four major time slices (Fig. 16). Conglomerate clast compositions are also shown for the time slices and the Nanaimo Group as a whole on Figure 16. Table 2 provides a numerical summary of the sandstone data.

The three main provenance categories distinguished by Dickinson and Suczek (1979) are continental blocks, magmatic arcs, and recycled orogens. Each of these major classifications can be subdivided, depending on features such as the degree of uplift and dissection of arc terranes, compositional variation in recycled orogen terranes, and the importance of stable shield versus sedimentary cover sequences in continental block terranes (see Dickinson et al., 1983, and Dickinson and Suczek, 1979 for detailed discussion of these factors and definition of individual fields). Within these broad classifications, the traditional source areas for the Nanaimo Group can be best described as two different dissected arcs (Wrangellia terrane and the Coast Belt), and a complex recycled orogen (northwest Cascade/San Juan terranes). The Wrangellia terrane is distinctive for its high component of mafic to intermediate volcanic cover rocks (e.g., Bonanza Group and Karmutsen formations), significant bodies of intermediate to mafic intrusives (Island Intrusives) and mixed sedimentary packages of the Sicker Group (see Monger et al., 1991 for a recent overview). The Coast Belt is dominated by granitic intrusives, but as discussed above, contains several pendants of volcanic successions and in mid- to Late Cretaceous time probably had an extensive volcanic cover (see Woodsworth and Monger, 1991 for a recent overview). The northwest Cascades comprises a complex series of small and large accreted terranes dominated by oceanic



A) Strike-slip basin setting as proposed by Pacht (1980, 1984) which he suggested formed as a small basin in a larger forearc setting (shown with arrow from Fig. B) or as a separate pull-apart in a transpressive plate margin setting. Diagram is modified slightly from Pacht (1980).

B) Forearc basin setting proposed by several authors.

C) Foreland basin setting proposed here.

D) Schematic of present Georgia Basin illustrating preservation of remnants of the originally more extensive Nanaimo Group strata as thrust slices in a west-vergent fold and thrust belt (probably formed during Eocene transpression).

***Figure 17.** Schematic illustrations of different tectonic settings proposed for the Nanaimo Group.*

sedimentary and metasedimentary rocks (most argillaceous to cherty), but with significant volcanic, ophiolitic, and intrusive slices, all deformed in mid- to Late Cretaceous contractional and transpressional events (see Tabor et al., 1989 for a recent overview).

Plotting of all Nanaimo Group sandstones onto these tectonic provenance diagrams does show the influence of these major source areas, although it is difficult to discriminate between the importance of any specific source, especially Coast Belt versus Wrangellia terrane (Fig. 14, top of Fig. 16). The entire Nanaimo Group on the QFL diagram overlaps both the dissected arc and recycled orogen fields, reflecting the mixed source areas for the entire basin, but plots in the dissected arc field for the QmFLt diagram, probably reflecting the abundance of volcanic as well as chert lithic material in the recycled orogen source area of the northwest Cascades. This is also shown on the QpLvLs diagram where the lithic content is widely variable, again suggesting a mixed source.

The time slice plots of Figure 16 show trends compatible with the provenance evidence discussed previously. The Santonian to Early Campanian slice reflects an early volcanic-rich component probably derived from early dissection of the Coast Belt volcanic cover and the volcanic-rich Wrangellia terrane. Both the Early Campanian and Late Campanian time slices show the major influence of chert-rich recycled orogen material from the northwest Cascades and possibly southeastern Coast Belt (also shown in the conglomerate clast plots). This is most noticeable in the QFL and QpLvLs plots for these slices where the increased chert content dramatically shifts the values towards the recycled orogen field. An upward (relatively increased Qm) shift in the values in the QmPK plots from Santonian to Late Campanian time might represent continued dissection of the Coast Belt arc, with the source changing from predominantly the volcanic cover to originally deeply buried plutonic bodies (increasing the abundance of monocrystalline quartz). The youngest Nanaimo Group strata (Late Campanian to Maastrichtian) show a decrease in influence of the recycled orogen source and appear to reflect the dominance of dissected magmatic arc material, possibly the dominance of dissected eastern Coast Belt sources versus Cascade sources during this time period. However, the detrital zircon results discussed previously indicate that major river systems were transporting material from the eastern Cordillera to the Nanaimo basin at least in Maastrichtian time, casting doubt on any simple interpretation of how the source areas have changed for this time slice. Perhaps the overlap of the sandstone data into the basement uplift fields for this time period reflects this new eastern source area.

TECTONIC SETTING

The ultimate control on the Late Cretaceous development of Georgia Basin was the interaction of the North American with the Kula and Farallon oceanic plates. Engebretson et al. (1985) provide a summary of the known plate positions and convergence vectors for this time. They show a change in Late Cretaceous time (at about 80 Ma) at the latitude of present Vancouver Island from a strongly convergent margin with the Farallon plate being subducted at a high angle beneath the North American plate, to an increasing transpressive regime with major dextral strike-slip faults as the Kula plate subducted obliquely to the north and northeast in latest Cretaceous to early Tertiary time. However, there is a large margin of uncertainty concerning the position of the Farallon-Kula plate boundary along western North America during this time period, and thus the change from largely convergent to transpressive regimes in the Late Cretaceous is not well constrained for the specific time of interest. For example Duncan and Kulm (1989) calculate both a high rate and high angle of Farallon plate convergence with respect to the North American plate at 70 Ma for the Oregon coast area. Major dextral strike-slip faults in southwest British Columbia and northwest Washington State (e.g., Straight Creek-Fraser, Ross Lake-Yalakom fault systems) show evidence of significant offset (100 km) in the Paleogene (mostly Eocene), although the Ross Lake-Yalakom fault systems could have been active in latest Cretaceous time (constraints on timing and amounts of displacement are reviewed in Monger and Journeay, 1992). Other dextral strike-slip faults which probably were active during the Late Cretaceous occur in the southwest Coast Belt and northwest Cascades (e.g., Harrison Lake Fault (Journeay and Csontos, 1989) and faults in the North Cascades documented by Brown, 1987)), but probably did not provide major lateral displacements on the order of the Straight Creek-Fraser and related fault systems.

Most researchers have suggested the Nanaimo Group was deposited in a forearc basin (Muller and Jeletzky, 1970; Dickinson, 1976; Muller, 1977a; Ward and Stanley, 1982; England, 1990) based on the setting of the basin east of the underthrusting zone of the Farallon/Kula oceanic plates beneath the North American plate and west of the Cretaceous and early Tertiary magmatic arc of the Coast Mountains. A strike-slip basin model was suggested by Pacht (1984). A foreland basin model was proposed, but not discussed in detail by Brandon et al. (1988) on the basis of mid- to Late Cretaceous thrusting he recognized immediately south of the Nanaimo Group (San Juan thrust system). Recent syntheses of the regional Cretaceous evolution of western North America and the southwest Coast Belt also refer to the Nanaimo Basin as a foredeep which developed in front of widespread mid- (to Late) Cretaceous west vergent thrusting now recognized in much of the northwest Cordillera (Rubin et al., 1990; Monger and Journeay, 1992), but as with the original Brandon et al. (1988) suggestion, there is no discussion of the direct evidence for this model. The merits and problems of these proposed basin models are discussed below.

Strike-slip basin model

Pacht (1980, 1984) proposed that the Nanaimo Group was deposited in a restricted pull-apart basin formed by extension and transcurrent faulting, with local horsts and grabens controlling sedimentation (shown schematically in Fig. 17A). He speculated that this pull-apart basin might be unrelated to a forearc setting on the western North America margin, or, alternatively, might have developed within an overall forearc setting on a margin dominated by oblique subduction, as a type of small intramassif forearc basin described by Dickinson and Seely (1979) for forearc regions in transpressive regimes, and illustrated at the arrow origin on Figure 17B.

New evidence concerning both the regional geology of the western Cordillera and the characteristics of strike-slip basins available since the study of Pacht suggest a strike-slip basin model is not appropriate for the Nanaimo Group. Faults interpreted by Pacht (1980, or Muller, 1977b, cited in Pacht) as synsedimentary and strike-slip on Vancouver Island have since been shown to be early Tertiary thrusts (England and Calon, 1991) and seismic investigations do not suggest any major transcurrent faults are present beneath the present Georgia Strait (White and Clowes, 1984). Pacht also did not consider the Alberni or Comox outcrop areas or the eastern outliers of the Nanaimo Group in his reconstructions of the Nanaimo Basin. Including these areas, the Nanaimo Basin was much more extensive then Pacht envisioned. As discussed in the regional setting summary, minimal estimates of the original basin size are 230 km northeast-southwest (roughly along the basin axis) and 120-130 km northeast-southwest. These figures are certainly low, reflecting considerable Tertiary uplift and erosion both to the east and west. This basin size is not compatible with most well documented active margin strike-slip basins, which tend to be narrow, a few tens of kilometres wide at most, and elongate, with a stacked apparent basin thickness approaching the value of the basin width (e.g., Nilsen and McLaughlin, 1985). Independent of the lack of evidence of synsedimentary bounding faults, the internal features of the Nanaimo Group are not typical of strike-slip basin fills. The provenance evidence, paleocurrents, facies types, and overall stratigraphy of the basin do not support an interpretation that local horsts and grabens confined basin sedimentation, or that Wrangellia terrane on Vancouver Island was exposed and controlling sedimentation on the northwestern margin during any but the earliest stages of basin fill. Depositional and stratigraphic features typical of strike-slip basins are summarized in Christie-Blick and Biddle (1985). Typical of strike-slip basins, but not the Nanaimo Basin, are: geological mismatches between clast types of the sediment fill and adjacent basement; extremely rapid rates of subsidence and abnormally thick basin fills (the 4 to 5 km thick Nanaimo fill, accumulated over about 20 to 25 million years, is not unusual); and evidence that synsedimentary oblique fault movements control facies distribution and paleocurrent trends (the latter characteristically nonradial and parallel to the basin axis except at the basin margins in pull-apart basins).

Forearc basin model

There is little doubt that the Nanaimo Basin formed between a zone of Late Cretaceous igneous activity in the eastern Coast Belt and a basically oceanic plate boundary to the west, thus fitting the broadest definition of a forearc basin as used by most authors (e.g., Dickinson and Seely, 1979; Ingersoll, 1988). More specifically, the Nanaimo strata were deposited on a complex basement composed of early Mesozoic and Paleozoic Wrangellia terrane and Middle Jurassic to Early Cretaceous arc massif rocks of the western Coast Belt, and probably overlapped the northwest Cascade terranes to the southeast. The basin was separated from the active subduction zone and accretionary complex presumably located well to the west, by an unknown width of Wrangellia terrane. This type of setting could be termed a broad ridged forearc basin in the Dickinson and Seely (1979) classification, as illustrated in Figure 17B, although with an unusually thick and extensive block of transitional crust (Wrangellia terrane) providing part of the forearc basin basement in this setting. Although basin size and major characteristics of the basin fill are broadly compatible with forearc basin depositional models, there are some problems. Forearc basins typically (but not ubiquitously) derive material from the active arc volcanic region of the arc-trench system and, in later stages, from the exposed roots of this system, providing both a strong magmatic arc provenance signature and evidence of coeval volcanism (pyroclastic layers or synsedimentary volcanic beds), as reviewed in Dickinson and Seely (1979) and shown for the classic Mesozoic Great Valley forearc basin of northern California by Ingersoll (1983). The Nanaimo Group does show a strong magmatic arc provenance, though mixed with the signature of the Cascades recycled orogen source. Significantly, however, the detrital zircon and other provenance evidence demonstrate that the magmatic arc sources, mainly the western Coast Belt, but also eastern Wrangellia in earliest stages, are of older arc provenance. The Wrangellia magmatic arc sources are part of an exotic terrane accreted to the western margin prior to Nanaimo sedimentation and unrelated to subduction or basin formation in the Late Cretaceous. The western Coast Belt magmatic rocks are part of an arc massif formed in place, but prior to Nanaimo Basin formation and acting as a major source area not because of uplift or effusive buildup directly related to coeval magmatic activity as in typical forearc basins. In addition, there is little evidence of synsedimentary volcanism directly related to Nanaimo Group sedimentation. A single ca. 83 Ma tuffaceous layer present in the Comox Formation in the Quinsam area (Kenyon et al., 1992) and a detrital zircon age of ca. 87 Ma from Gabriola Formation sandstone (Mustard et al., in press) are the sole indications of any possible link to the narrow zone of Late Cretaceous magmatic arc activity present in the eastern Coast Belt and Cascade Ranges during Nanaimo sedimentation. However, as discussed in the provenance section, the presence of recycled Precambrian, discordant Mesozoic, and abundant 72-73 Ma detrital zircons in the Gabriola Formation probably indicates major river systems were transecting the Coast Belt and northern Cascades and providing detritus from the eastern Cordillera Omineca Belt at least during Maastrichtian time. This second source of a possible magmatic arc signature, but from an area not directly part of the western margin arc-trench system, also suggests that the traditional forearc basin model is not required from the provenance evidence.

Forearc basins as defined by Dickinson and Seely (1979) form between an active arc massif and a trench subduction complex. Basin subsidence is dominantly a result of uplift in the arc massif due to pluton emplacement, and formation of a structural high on the edge of the subduction complex, the latter resulting from underthrusting of oceanic crust and accretionary complex material. However, as discussed above, the Late Cretaceous Georgia Basin received sediment from older arc massif, transitional crust and recycled orogenic terranes, most to the east, and some considerably removed from the basin. It is difficult to envision how the uplift of these source areas is directly linked to either the subduction complex present some distance to the west of the basin or the narrow zone of arc magmatism present to the east.

Foreland basin model

A foreland basin setting for Nanaimo Group deposition was first proposed by Brandon et al. (1988) on the basis of mid- to Late Cretaceous thrusting (San Juan thrust system) they recognized immediately south of the Nanaimo Group. A problem with this model at the time was the poor control on the age of the thrusting, which Brandon et al. (1988) could demonstrate was younger than about 100 Ma, but could not show overlapped with the time of Nanaimo Group sedimentation. In addition, there was little evidence for coeval thrusting in the Coast Belt, which several studies had suggested was a major source for much of the Nanaimo Group detritus (e.g., Pacht, 1984).

New support for this foreland basin model comes from several recent studies, most importantly recognition that Late Cretaceous thrust systems in the southern Coast Belt and northwest Cascades were active both during initial Nanaimo Group basin formation and the major period of sedimentation. A complex history of Late Cretaceous shortening has been demonstrated for a series of northwest-trending contractional (and in late stages strike-slip) fault systems preserved in the central and eastern Coast Belt. The structural history of these systems is summarized in Journeay et al. (1992) and Monger and Journeay (1994) and detailed along with results from related geochronologic studies in Journeay and Friedman (1993). A major period of southwest-directed thrusting is evident during the Cenomanian and Turonian (about 98 to 90 Ma), best preserved in the central Coast Belt (Lillooet River Fault System of Journeay and Friedman, 1993). This may include southwest-directed thrusting in the western Coast Belt documented by Lynch (1991), although age constraints in that region are poor (Lynch suggests thermal events at about 93 and 83 Ma are caused by the thrusting). Both northeast- and southwest-vergent thrusting and associated (but probably late stage) dextral strike-slip faulting can be demonstrated in the eastern Coast Belt for latest Cretaceous time, with southwest-directed thrusting active (along with other faulting) from about 86 to 68 Ma, corresponding to the main period of Nanaimo Group sedimentation.

Broadly west-directed thrusting of mid- to Late Cretaceous age is well established in the northwest Cascades (see Brandon et al., 1988 for San Juan thrust system, McGroder, 1991 for mainland Washington State Cascade thrusts). Miller and Paterson (1992) have shown that at least some thrusts in the northwest Cascades can be constrained between 96 and 85 Ma. Garver (1988) recognized the clastic Obstruction Formation of probable Cenomanian-Turonian age in the San Juan Islands, which contains clasts derived from San Juan terrane sources and also has been deformed into northwest-verging folds. Garver suggests this sandstone and conglomerate unit represents northward prograding submarine fan deposits derived from active thrust sheets, and that the strata were then involved in later stages of thrusting. Significantly, Turonian age strata have recently been confirmed at the base of the Nanaimo Group sequence immediately west of the San Juan Islands (Haggart, 1991, 1994) and possible Cenomanian strata are known from drilling beneath the Fraser Delta in the Vancouver area (Mustard and Rouse, 1991). If the interpretation of Garver is correct, the proximity of synthrust detritus in the San Juan thrust system to coeval strata at the base of the Nanaimo Group strongly supports a thrust system control on initial Nanaimo Group sedimentation.

The evidence cited above and a re-examination of the Nanaimo Group basin fill led Mustard and Monger (1991) to suggest a foreland basin model is appropriate for the Nanaimo Group (Fig. 17C). The thick Wrangellia composite terrane of Paleozoic and Jurassic volcanic arcs, oceanic sedimentary piles, and major intrusive bodies provided a semi-rigid foreland basement, which included western elements of the Coast Belt, and was loaded and flexurally deformed in front of the westerly propagating thrust stacks of Coast Belt and Cascade composite terranes to form the initial Nanaimo Basin. Continued uplift, at least partly due to thrusting in the source regions (and possibly also transpressional effects), provided periodically rejuvenated sediment supplies to the Nanaimo basin throughout its history. Direct linkage of thrusting and sedimentation is hampered by the several kilometres of Late Tertiary uplift of the western Coast Belt (Parrish, 1983; Monger, 1991b) which has created a gap between the best preserved thrust systems and the foreland fill, and caused erosion of the original eastern part of the Nanaimo Basin, so that only deeper basin marine strata preserved for much of the upper Nanaimo Group. In addition, early Tertiary compression thrust the Nanaimo Group southwest, and formed in a thrust contact between the southern Nanaimo Group and the San Juan terranes (England and Calon, 1991, also shown on the basis of vitrinite reflectance data for the Nanaimo Group sediments by England, 1990). This obscured the original southeastern margin of the Nanaimo Group, in the place where it would have been incorporated into, or overthrust by, the leading edge of the late Cretaceous thrust system, with the exception of the minor thrust-deformed sedimentary packages preserved on the San Juan Islands (Garver, 1988).

The evolution of Nanaimo Group source areas with time indirectly fits a foreland basin hypothesis with: 1) early loading and forebulge migration away (west) from the thrust systems, with early thrust-related uplift on the east side providing a provenance signature of the western Coast Belt and eastern Wrangellia terrane (plus minor forebulge erosion on the west side providing an initial Wrangellia terrane signature); 2) continued forebulge migration and basin deepening, submerging the foreland area and forming a basin axis parallel northwest-southeast to the main trend of the thrust systems, with the main influx of sediment from the uplifted and eroding thrust stacks, in this case the San Juan terranes and the Coast Belt. An evolution in the importance of the Coast Belt versus the San Juan thrust systems as source areas is also predicted by the current knowledge of timing of thrusting in these areas. If the evidence of Garver (1988) is correct, the San Juan thrust system was most active immediately prior to the earliest stages of Nanaimo basin fill. Thrusting in the Coast Belt (and perhaps the northwest Cascades of mainland Washington) also occurred during later (Campanian to Maastrichtian) deposition of the Nanaimo Group. This fits well with the observation that detritus from the San Juan thrust stacks is most prevalent in lower Nanaimo Group strata and less common in upper strata, where a dominance of Coast Belt sources is apparent both in the provenance and the facies trends. The thrust events in the latest Cretaceous appear restricted to the eastern Coast Belt and this corresponds to the only provenance signature of eastern Coast Belt sources (the 87 Ma detrital zircon in the Gabriola Formation). In contrast, western Coast Belt sources are common in lower units of the Nanaimo Group, mixed with detritus from San Juan terranes. The final stage in this foreland basin evolution is a cessation of major uplift in the Coast Belt during the Maastrichtian as transform faults became important (e.g., the Ross Lake-Yalakom Fault systems) and a change in major drainage patterns to include large river systems bisecting the Coast Belt and northern Cascades and providing an influx of sediment from eastern Cordillera sources.

NANAIMO GROUP: ECONOMIC RESOURCES

Historically, the principal exploited resource of the Nanaimo Group has been extensive coal deposits in the Extension, Pender, and Protection formations in the Nanaimo area and in the Comox Formation in the Cumberland and Quinsam areas (located on Fig. 2). Major mining activity in these areas continued from the late 1800s into the 1950s with over 55 million tonnes of coal mined. Production declined drastically, although reduced mining efforts continued into the mid-1980s and, at Quinsam, to the present (reviewed in Bickford et al., 1990 and Kenyon et al., 1992). Coals of the Nanaimo Group are generally high volatile A to B bituminous types with moderate sulphur contents, although variations are present (Bickford and Kenyon, 1988; Kenyon and Bickford, 1990; coal geochemistry also provided in Van der Flier-Keller and Goodarzi, 1992).

Nanaimo Group sandstones have provided an important source of building stone, used in several buildings on Vancouver Island, the Vancouver area, and as far away as San Francisco. As reviewed in White (1988), quarries were located on several Gulf Islands and in Nanaimo townsite, in the Protection, De Courcy, and Gabriola formations. Most stone was quarried in the mid- to late 1800s; the last commercial quarrying occurred in the 1920s.

The most recent exploration activity in the Nanaimo Group has focused on kaolin deposits in the Lang Bay sedimentary outlier located on the northeast side of Georgia Strait (Fig. 2). Kaolin occurs in high concentrations at the unconformity surface in the residual regolith of the deeply weathered granodiorite basement, and in dilute concentrations, but higher volumes, within fluvial silty mudstone facies in the overlying Comox Formation (White, 1986; Hora, 1988; Mustard and Rouse, 1991). The Lang Bay deposit was extensively drilled in the late 1980s and is currently in advanced stages of evaluation. A recent study by Read (1994) of the potential for similar kaolin deposits at the sub-Nanaimo Group unconformity surface concluded that the potential was generally low, due to the lack of intermediate plutonic basement for most strata, and the lack of the special conditions of formation and preservation required for these deposits, but that there was potential for similar deposits in the Quinsam and southern Comox outcrop areas.

The potential for significant traditional oil and gas-reserves in the Nanaimo Group is generally considered poor. Gordy (1988) provided an evaluation of the hydrocarbon potential of the entire Georgia Basin, including a review of past exploration activity. He concluded that although extensive sandstone bodies and appropriate caprocks and structural traps are present within the Nanaimo Group, the sandstones contain insufficient porosity and permeability to act as significant reservoirs and that there is a lack of high quality hydrocarbon source rocks. This evaluation is supported by more recent studies. Thermal maturity values based on vitrinite reflectance data show that most of the Nanaimo Group is mature, generally within the oil window, but in areas near Tertiary intrusives or where overthrust, rocks are overmature and gas prone. Values range from about 0.5% Ro to 5% R_o with most values % R_o (Kenyon and Bickford, 1990; Bustin, 1990; England, 1990; Bustin and England, 1990). However, source rock potential is generally poor, with organic matter principally kerogen types II and III and the extensive marine shales containing low values of total organic carbon (%) and moderate to low hydrogen indices (Bustin, 1990; Bustin and England, 1990). In addition, England et al. (1989) and England (1990, 1991) suggest that any major period of hydrocarbon generation in the basin resulting from burial thermal metamorphism predated by several million years the main period of structural trap formation in the middle to late Eocene (the fold and thrust event documented by England and Calon, 1991).

There remains some potential for secondary hydrocarbon generation due to thrusting which caused tectonic burial of upper units, or where thick Eocene to Miocene sediments continued to bury the older strata. The few exploration wells which have tested Nanaimo Group strata tend to support the pessimistic evaluations cited above (reviewed in Gordy, 1988). Two shallow wells drilled in the mid-1980s south of Nanaimo (BP Laurel Harmec and BP Yellow Point) and a single well drilled on Saturna Island in 1957-58 (Charter Saturna No. 1) did not discover significant shows or reservoir development. Two dry wells drilled in the southwest Fraser Delta in the early 1960s (Richfield Point Roberts and Richfield Sunnyside) penetrated several hundred metres into Nanaimo Group strata preserved beneath about 3000 m of Tertiary strata (Hopkins, 1966; Bustin, 1990; Mustard and Rouse, 1991). The most recent exploration test of Nanaimo Group strata was conducted in 1988 by American Hunter Ltd. and partners about 10 km south of the U.S.-Canadian border in northwest Washington State (AHEL Birch Bay No. 1); probable Nanaimo Group sediments were encountered below about 1000 m in the 2800 m hole. Hurst (1991) reports these Cretaceous strata include impermeable but fractured sandstones with Type III kerogen and low total organic carbon values.

Although traditional oil and gas prospects are not favourable for the Nanaimo Group, a potential for coalbed methane reserves on eastern Vancouver Island has been suggested. Kenyon (1991) provides a preliminary evaluation of the coalbed methane potential for these strata. The combination of major deposits of high volatile A to B coals showing vitrinite reflectance values suggesting thermal maturity into the window of coalbed methane generation, with large scale fracture and joint systems in the coalfields to provide enhanced permeability, suggests that possible large reserves of methane may be associated with the major coal-bearing areas. Kenyon (1991) quotes test well data indicating high quality and encouraging volumes of methane from some coals in the Nanaimo area, and provides a preliminary resource estimate of about one trillion cubic feet of methane for the eastern Vancouver Island coalfields. Exploration for this resource is in early stages, however, and significant future work is required to substantiate these early estimates.

SUMMARY

The Nanaimo Group, a part of the Cretaceous to Neogene Georgia Basin, comprises 4 km of sedimentary rocks of Turonian to Maastrichtian age. Nanaimo Group sediments originally extended 230 km along a northwest-southeast axis with a northeast-southwest breadth greater than 120 km and the basin axis centred over present Georgia Strait or eastern Vancouver Island. The basin was deformed by Eocene compression into a fold and thrust belt, and uplifted and eroded into the present outcrop pattern by Neogene uplift of the Coast Belt to the east, and tilting of the west side of Wrangellia terrane during subduction of the Juan de Fuca oceanic plate to the west.

Lithostratigraphic nomenclature has undergone a complex evolution; eleven formations are recognized. These consist of sedimentary successions, each hundreds of metres to >1000 m thick, with resistant sandstone- or conglomerate-rich formations separated by recessive mudstone- and fine grained sandstone-rich formations. Formation contacts are generally conformable and gradational with common lateral intertonguing relationships. The basal Comox Formation formed in alluvial and coastal marine environments on a rugged unconformity surface of several hundred metres paleorelief. Coal-bearing parts of the Comox Formation in the Comox outcrop area and the Extension to Protection formations in the Nanaimo area formed in coastal fluvial, lagoonal, and marginal marine back-barrier environments, associated with fluvial and shallow marine sandstone-mudstone facies. Strata above the Comox Formation and outside of the coal-bearing units of the Nanaimo area were deposited in marine, generally outer neritic to bathyal depths, mostly from sediment gravity flows organized in facies typical of submarine fan models. These include inner fan lenticular conglomerate and coarse sandstone channel deposits, middle to outer fan channelized and nonchannelized, thick sandstone beds, interchannel levee sandstone-mudstone turbidites, laterally extensive outer fan to fan fringe thin-bedded sandstone/mudstone turbidites, and thick successions of massive to poorly bedded silty mudstone. Slope or channel margin failure causing thick debris flows of intrabasinal sedimentary clasts were rare; common synsedimentary folds, clastic dykes, and other deformation structures are thin and not laterally persistent, probably representing minor interfan failures and overpressured sand loading effects.

The abundance of provenance evidence clearly indicates an evolution of source areas with time. Initial detritus was derived from local basement, indicated by the prevalence of Wrangellia terrane compositions in Comox Formation sandstone and conglomerate. Strata of lower formations above the Comox Formation contain detrital zircons and clast types supporting mixed derivation from the northwest Cascades and western Coast Belt, with little evidence of continued Wrangellian terrane detrital input. Higher formations are increasingly feldspar-rich and have a predominant Coast Belt signature, but the uppermost formation (Gabriola Formation) includes both Precambrian and Maastrichtian age detrital zircons probably derived from recycled sedimentary packages and Mesozoic intrusives of the eastern Cordillera, which indicates detrital input from large river systems in latest Cretaceous time.

A foreland basin model for Nanaimo Group deposition is preferred here, based on recent evidence that generally west-directed thrusting in the Coast Belt and northwest Cascades overlaps in time with initial basin sedimentation. This suggests that the direct cause of basin formation was lithosphere flexure in front of the active thrust sheets. The presence of sedimentary rocks coeval with initial Nanaimo Group strata and incorporated into San Juan thrusts, and evidence of continued thrusting in the Coast Belt and perhaps northwest Cascades during the main stages of Nanaimo Group sedimentation support the foreland basin model, as does the evolution of the provenance signatures of the detrital fill for the basin. A strike-slip basin model is not appropriate for the Nanaimo Group, as previously interpreted synsedimentary strike-slip master faults are now interpreted as postdepositional thrusts, and Nanaimo Group thickness, extent, and internal characteristics differ markedly from well documented strike-slip basins. Most researchers have suggested the Nanaimo Group was deposited in a broad ridged forearc basin, based on the setting of the basin east of the subducting Farallon or Kula oceanic plate and west of the magmatic arc of the Coast Belt. However, the Nanaimo basin was separated from the subduction zone to the west by the thick composite crust of Wrangellia terrane and from the narrow Late Cretaceous arc by eastern Wrangellia and western Coast Belt crust. The lack of evidence for significant detritus from an active arc, as opposed to Wrangellia, old arc massif crust, and source areas well east of the active arc-massif, suggest the arc-trench system did not directly control basin evolution.

The principal economic resource of the Nanaimo Group has been high volatile A to B bituminous coal deposits in the Nanaimo and Cumberland areas of the basin. The importance of coal mining has dwindled since the 1950s, with only one deposit currently being exploited. Other economic resources include building stone quarried at the turn or the century, and potential mining of kaolin-rich in situ or reworked regolith at or just above the unconformity in the north and northeast parts of the preserved outcrop areas. Oil and traditional gas potential is poor for the basin. Although potential sandstone reservoirs are abundant and thermal maturity values generally are sufficient for hydrocarbon generation, porosities are extremely low, organic-rich source rocks are not extensive, and the peak period of hydrocarbon generation may have preceded the time of structural trap formation in the mid- to late Eocene. However, the extensive coal deposits and good thermal maturity values suggest significant coalbed methane resources could be present in parts of the basin.

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APPENDIX A

Summary of Nanaimo Group formations

This appendix presents the major features of the eleven Nanaimo Group formations, each summarized in a table with accompanying figure (Tables A1 to A11, Fig. A2 to A12). Each table summarizes the age range, contacts, major lithostratigraphic features, depositional environment, and provenance. The formation descriptions and interpretations are distilled from the listed references on each table, supplemented by the author's field studies in 1990 to 1992. Depositional environment and provenance sections reflect modern (generally post-1980) interpretations, again supported by the author's work. A more complete discussion of these depositional models is provided in the report. For each table an accompanying figure provides summary maps showing outcrop distribution, extant paleocurrent data, conglomerate clast composition data, and ternary plots showing sandstone detrital framework compositions. Sources of data additional to the author are listed in Table 1 paper. The ternary plots in these figures provide sandstone detrital framework compositions for each sample subdivided both into the grain populations recommended by Dickinson and Suczek (1979) and also in terms of four geographic divisions splitting the Nanaimo Group roughly from southeast to northwest (shown on Fig. A1A, with formation data summaries and references given in Table 1 of the paper). Inner dashed lines represent inferred provenance type divisions of Dickinson and Suczek (1979), as modified by Dickinson et al. (1983). Ternary plot apices and provenance divisions are defined in Figure A1.

References

References cited for these tables are provided in the paper or as part of the separate listing of graduate student theses compiled in Appendix C.

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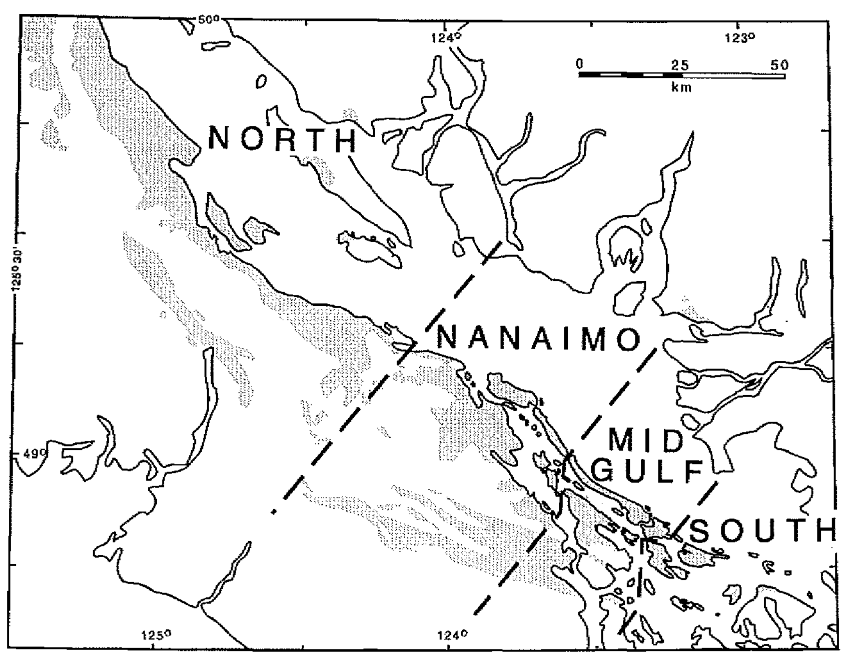

Figure A1A) Division of Nanaimo Group into four geographic areas for comparison of sandstone compositions on ternary plot diagrams shown on Figures A2 to A12 inclusive.

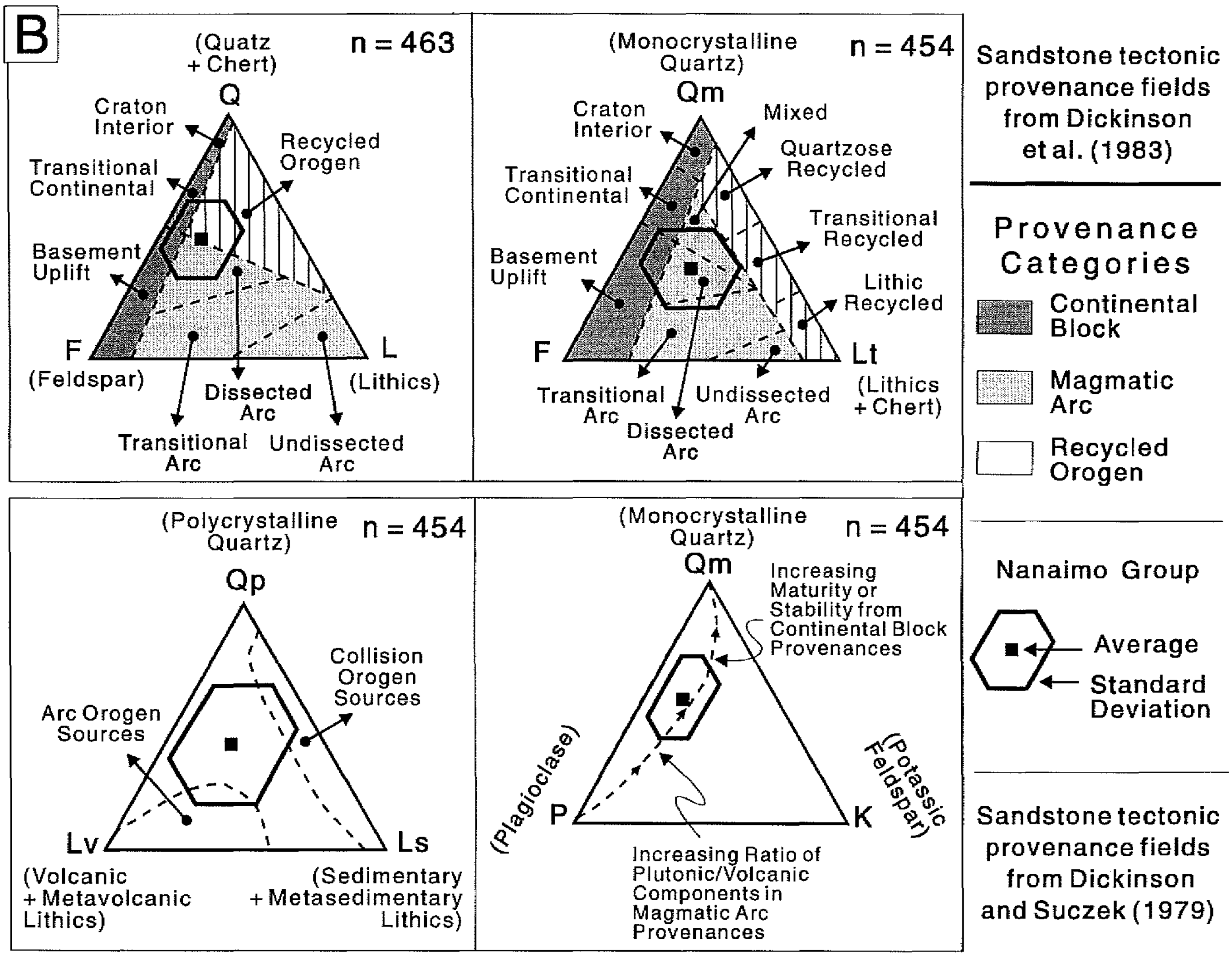

Figure A1B) Sandstone detrital framework compositions subdivided into grain populations and plotted on ternary diagrams showing the tectonic provenance fields of Dickinson et al. (1983, QFL and QmFLt plots) and Dickinson and Suczek (1979, QpLvLs and QmPK plots). The mean value of all Nanaimo Group sandstone samples is shown (black square) with a polygon representing the standard deviation from the mean. Sources of information are summarized in Table 1. Note that for nine samples grain types were not identified in the original study to the level allowing plotting on any but the QFL plot, accounting for the change in total number of samples used between the QFL plot and the other plots.

Table A1. Comox Formation.

AGE	Mid- to Late Santonian generally, but may range to earliest Campanian in northern areas (e.g., Quinsam) and eastern outliers (Vancouver, Lang Bay). Minor Turonian strata in southeast basin may be older and underlie Comox Fm.
UPPER CONTACT	Generally gradational, rarely sharp (but conformable) change to Haslam Formation; contact at top thick sst bedsets
COMOX FORMATION MAJOR LITHOFACIES	*——Nanaimo - Cowichan - Duncan outcrop areas ——* **Saanich member:** coarse grained, medium-to thick-bedded arkosic to lithic arenite; crossbedded to massive; less common interbeds of fine grained sandstone, siltstone, and mudstone, rare carbonaceous mudstone and discontinuous coaly beds; upper thick-bedded massive sandstone with rare marine fossils in thin silty mudstone interbeds thins and fines up to Haslam Formation turbidites *——Alberni - Quinsam - Comox outcrop areas ——* **Dunsmuir member**: medium grained, arkosic arenite with rare interbeds of carbonaceous mudstone, coal, and conglomerate lenses; probable equivalent of Saanich mbr. **Cumberland member:** interbedded siltstone, fine- to medium-grained lenticular sandstone, carbonaceous mudstone; laterally persistent coal beds to 7.5 m thick; sandstones complexly intertongue with finer strata **Benson Member:** basal conglomerate; generally poorly sorted, matrix- to clast-supported, heterogenous mix of pebbles to boulders in a medium- to coarse-grained arkosic to lithic arenite (rarely wacke) matrix. Generally thick-bedded and poorly stratified, but well stratified in places. Common interbedded coarse grained arkosic to lithic arenite, crossbedded to massive; rare siltstone and carbonaceous mudstone; very rare coarse grained bioclastic calcarenite and algal laminated limestone <3 m thick at unconformity; rare preserved basal regolith (e.g., Lang Bay, Lasqueti Island)
DISTRIBUTION AND THICKNESS	Present and persistent in all outcrop areas and eastern outliers. Total thickness up to 650 m, but typically <200 m thick; generally thickest filling paleolows and thins to absent over paleohighs; Benson Mbr. conglomerates up to 300 m thick in paleovalleys, but typically a few tens of metres and absent in some areas. Saanich mbr. up to 500 m at Saanich Peninsula, but thins north and generally 50-100 m; Cumberland mbr. generally 30-150 m; Dunsmuir mbr. generally 120-230 m and thickens northward slightly, but laterally persistent
LOWER CONTACT	sharp, angular unconformity with underlying Wrangellia terrane (Coast Belt intrusives at Vancouver and Lang Bay); unconformity surface rugged with high relief (to 400 m, in west-to southwest-trending paleovalleys)
DEPOSITIONAL ENVIRONMENT	Basal conglomerate-sandstone reflect local high paleorelief controls in alluvial fan and braided fluvial environments in most areas, but shallow marine high energy deposits are locally present, including submarine canyon fills (e.g., Lasqueti & Saltspring Islands; Saanich Peninsula); Upper finer strata generally coastal fluvial-floodplain and marginal to storm wavebase marine with significant back-barrier lagoonal to inter-deltaic coal-forming environments, especially in northern areas. Transition to Haslam Fm. reflects general deepening marine conditions during late stages of deposition.
PROVENANCE	Paleocurrents generally scattered, especially in basal conglomerates; reflecting control of local irregular topography, but a broadly radiating westerly trend is apparent in many areas, especially in upper members; Fluvial deposits in major paleovalleys indicate southwest flow in northern areas. Conglomerate clasts generally appear locally derived, but sandstone more arkosic then expected for local sources and may reflect early input from dissected Coast Belt to east.
REFERENCES (REGIONAL STUDIES IN BOLD)	**Atchison, 1968**; **Bickford and Kenyon, 1988**; Bickford et al., 1989, 1990; Buckham, 1947a,b; Carson, 1960; Cathyl-Bickford and Hoffman, 1991; Clapp, 1912a,b, 1914a, b; Clapp and Cooke, 1917; **England, 1990**; Fahlstrom, 1982; Haggart, 1991, 1994; Hanson, 1976; Johnson, 1978; Kachelmeyer, 1978; **Kenyon and Bickford, 1990**; Kenyon et al., 1992; Mackenzie, 1922, 1923; Massey and Friday, 1988, 1989; Mercier, 1977; **Muller and Jeletzky, 1967, 1970**; Muller and Carson, 1969; **Muller and Atchison, 1970**; Mustard and Rouse, 1991; **Pacht, 1980, 1984;** Page, 1972; Rouse et al., 1975; Ruddiman, 1980; **Usher, 1949, 1952; Ward, 1976a, 1978a; Williams, 1924;** Winsby, 1973.

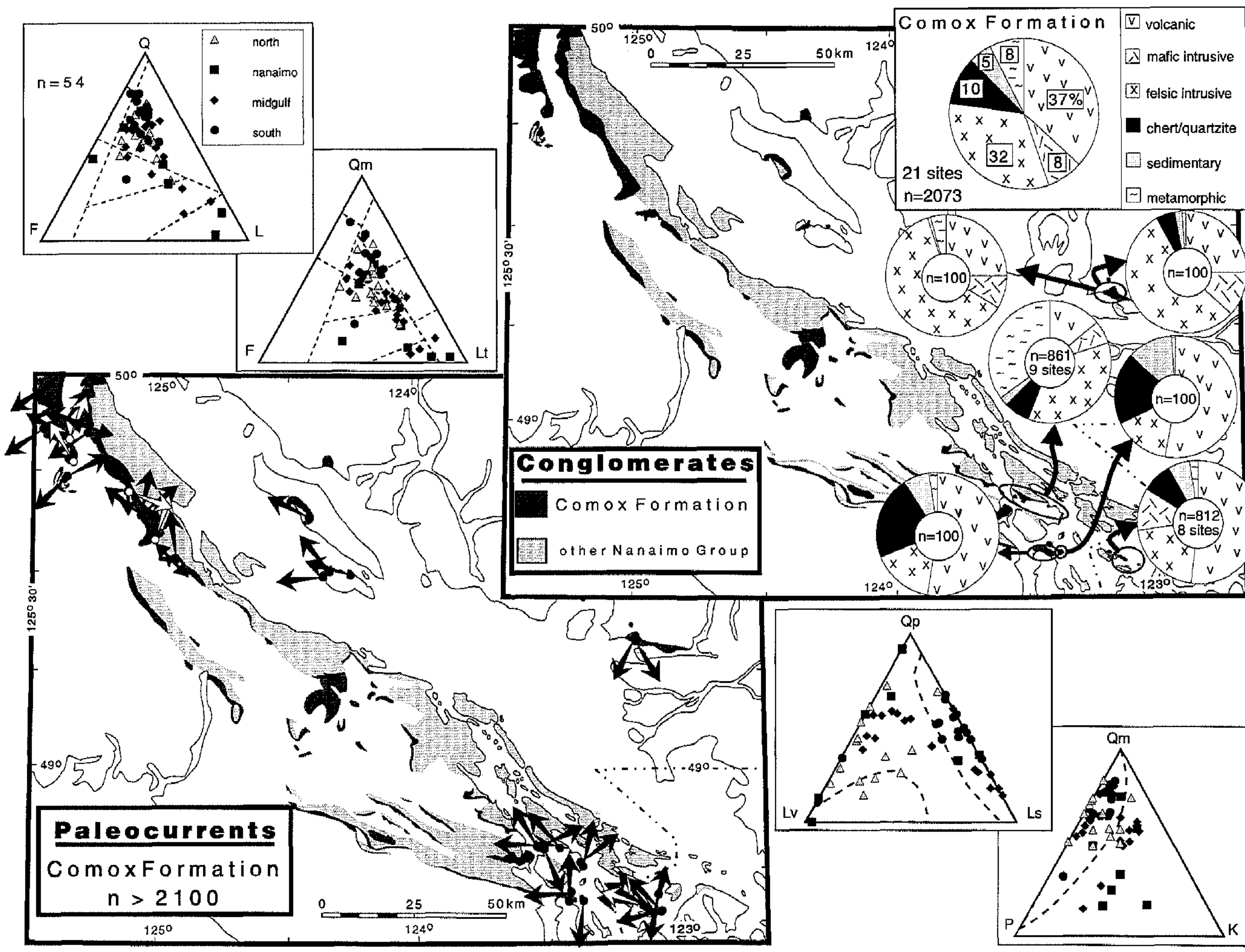

Figure A2. Comox Formation.

Table A2. Haslam Formation.

AGE	Late Santonian to Early Campanian (Elongatum and Schmidti biozones; fossils common)
UPPER CONTACT	Generally sharp, rarely gradational to Extension Fm. conglomerate/sandstone; erosive where at base of Extension Fm. conglomerate channels. Gradational to Pender Fm. mudstone-siltstone where Extension Fm. is not well developed (e.g., parts of Comox outcrop area). In south basin coarsens upward generally and intertongues with Extension Fm. sandstone and conglomerate.
HASLAM FORMATION MAJOR LITHOFACIES	Grey mudstone and siltstone with variable amounts of arkosic (rarely lithic) arenite thin beds. Mudstone and siltstone massive to faintly laminated or thin-bedded generally; in some places with few sandstone interbeds and forming monotonous massive sequences tens of metres thick. Sandstone interbeds vary from rare to common, thin to less common thick beds, generally fine grained, but rarely medium- or coarse-grained and pebbly. Bouma turbidite sequences common (T_{ABE}, T_{ACDE}, and T_{CDE} types most common but others present) with rare to common tool marks and internal ripples; clastic dykes and synsedimentary slumps rarely present. Both coarsening-thickening and fining-thinning upward sequences on metre and tens of metres scales can commonly be recognized. Overall coarsening-upward trend on formation scale to the overlying Extension Fm. is locally developed in some southern areas (e.g., Stuart Island), with rare conglomerate beds near the Extension Fm. contact and in some places occurring as thick, matrix-supported and poorly sorted conglomerates in mudstone-siltstone sequences. Common molluscan fossils and abundant foraminifera, especially in massive mud-siltstone sequences; common trace fossils and bioturbation in some areas
DISTRIBUTION & THICKNESS	Extensive and generally continuous in all major outcrop areas, generally 50-200 m thick, but probably >500 m thick in southern areas (and in places structurally thickened). General thinning to north and west, but not consistent;
LOWER CONTACT	Generally gradational from Comox Fm. sandstone thick beds, rarely sharp but conformable on Comox Fm.; very rarely angular unconformity on Wrangellia terrane where Comox Fm missing.
DEPOSITIONAL ENVIRONMENT	Marine outer shelf and slope environments suggested by trace fossil, foraminifera, and macrofossil types and diversity, but specific depth ranges are not well constrained; massive to poorly bedded mudstone-siltstone sequences represent low energy areas of deposition on outer shelf to slope. Laminated siltstone and fine grained sandstone include distal prodelta and outer submarine fan low-concentration turbidites, most common in northern and western areas and lower parts of formation, but present throughout. Sandstone-rich parts of formation represent overlapping submarine fan low to high-concentration turbidite deposition. Overlapping of fan lobes and migration of complexes caused both lateral and vertical variations in facies. Inner fan and feeder channel fills are preserved in south basin and in overlying Extension Fm. channels. Rare conglomerate thick beds in massive mudstone-siltstone represent single debris flow events into intrachannel or distal areas.
PROVENANCE	Paleocurrents are scattered but define ageneral radial pattern to the northwest, typical of submarine fan deposits. Sandstones are mixed arkosic and lithic types with scattered detrital modes indicating mixed source areas, generally from recycled orogen and chert-rich sources of San Juan terranes, but also with a high component of intrusive quartz and potassic feldspar, possibly representing Coast Belt contribution. General thinning and fining trends to northwest (and west?) also suggest San Juan terranes (and Coast Belt) as main source areas.
REFERENCES (REGIONAL STUDIES IN BOLD)	Bickford et al., 1990; Cathyl-Bickford and Hoffman, 1991; Clapp, 1912a,b, 1914a; Clapp and Cooke, 1917; **England, 1990;** Fahlstrom, 1982; Hanson, 1976; Hudson, 1974; Kachelmeyer, 1978; Mackenzie, 1923; Massey and Friday, 1988, 1989; **McGugan, 1962, 1964,** 1990; Mercier, 1977; **Muller and Jeletzky, 1967, 1970; Pacht, 1980, 1984;** Ruddiman, 1980; **Scott, 1974a, b; Sliter, 1973; Usher, 1949,1952; Ward,**1973, **1976a,1978a; Ward and Stanley, 1982; Williams, 1924;** Winsby, 1973.

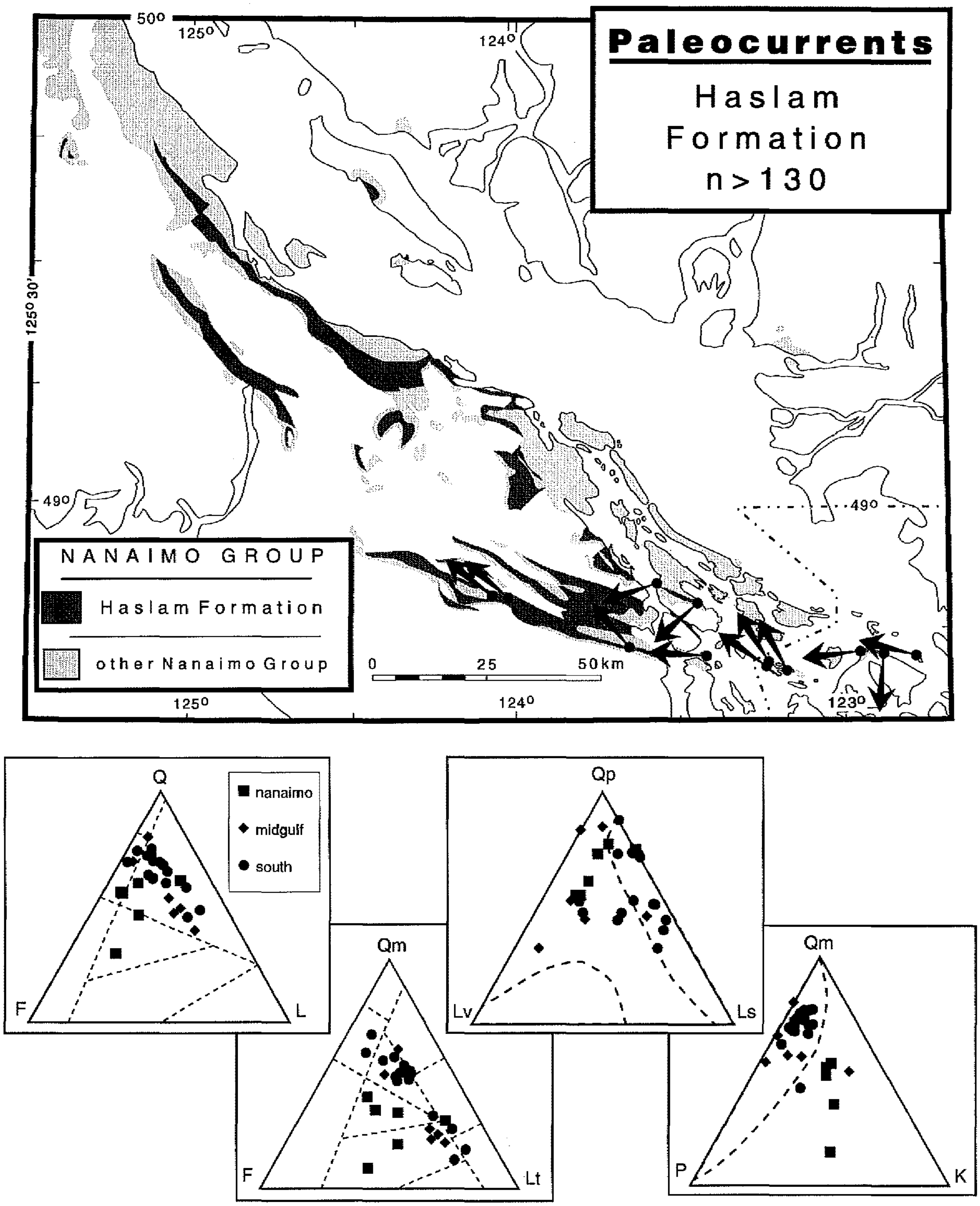

Figure A3. Haslam Formation.

Table A3. Extension Formation.

AGE	Early Campanian (Schmidti biozone but fossils are rare)
UPPER CONTACT	Generally gradational over a few metres to tens of metres in a fining- and thinning-upward transition to Pender Fm.
EXTENSION FORMATION **MAJOR LITHOFACIES**	Typically a thick sequence of pebble-cobble conglomerate with increasingly common sandstone interbeds in upper part. A thin, basal, coal-bearing facies is locally present at Nanaimo. Conglomerate generally clast-supported, moderately to poorly sorted with subround pebbles to boulders in a medium- to coarse-grained lithic (to arkosic) arenite matrix. Beds are generally massive, nongraded to poor normal or reverse-graded, but graded-stratified and crossbedded conglomerates are present. In southern areas (e.g., Saltspring - Pender - Orcas Islands) the Extension Fm gradually changes upward to thick- to medium-bedded medium-to very coarse-grained lithic arenite. Most sandstone beds are massive, poorly normal graded, rarely planar crossbedded, and in some places contain molluscan fossils, generally broken. In the Comox outcrop area, thick beds of matrix-supported pebble-cobble conglomerate with a mudstone matrix and common sedimentary ripups are locally present. More typically, sequences of massive, lenticular conglomerate overlap in overall channellized successions up to 150 m thick, which locally erode into underlying strata, but laterally interfinger with both the Haslam Fm. and overlying Pender Fm. Broken molluscan fossils are common in some places. In the local Nanaimo area the Extension Fm. can be subdivided into two units: a basal **Northfield member** with coal seams interbedded with 10-30 m of sandy siltstone and fine grained sandstone; and an upper conglomerate-rich **Millstream member** with minor interbeds of sandstone, which locally are channellized into, but in other areas gradationally interfinger with, sandstone of the Northfield member or Haslam Fm.
DISTRIBUTION AND THICKNESS	Widespread and generally continuous in southern areas, discontinuous lenses in northern areas. Ranges from 45 to 500 m, typically 100-200 m; thickest in southern Gulf Islands (e.g., Pender Island, but with some structural thickening) and thins northward to <200 m in the Nanaimo area and 45 to 140 m thick where present in Comox outcrop area
LOWER CONTACT	Generally sharp but conformable, locally erosive and channellized a few metres into Haslam Fm. and rare interbedded conglomerate at contact ; rarely gradational from coarsening/thickening upward Haslam Fm. sandstone sequences.
DEPOSITIONAL ENVIRONMENT	Generally inner submarine fan facies to Haslam Fm. turbidites and intertonguing Pender Fm.; Thick channels of conglomerates in some areas may be submarine canyon fills (e.g., southern Comox outcrop areas) with local conglomerate lens from large debris flows off fan deltas. Locally probably shallow marine high energy deposits related to fan-delta deposition of conglomerates and coarse sandstones (e.g., southern areas), in places reworked in offshore high energy barforms and storm deposits; probable coastal braided fluvial to alluvial fan deposits in Nanaimo area which grade into fan-delta deposits and interdeltaic to back-barrier lagoonal coal swamps.
PROVENANCE	Chert rich conglomerates and sandstones in southern areas and general westward paleocurrents reflect major source in San Juan terranes. Complex paleocurrents in Nanaimo area reflect mixed depositional environments and possible influence of remnant exposed paleohighs, but sandstone and conglomerate compositions and detrital zircon ages suggest major sources are San Juan terranes and Coast Belt areas. Northern conglomerates more rich in felsic intrusive clasts and show generally southwesterly paleocurrents, possibly reflecting a predominant Coast Belt source.
REFERENCES (REGIONAL STUDIES IN BOLD)	Bickford, 1989; Bickford et al., 1990; **Bickford and Kenyon, 1988;** Buckham, 1947a,b; Cathyl-Bickford, 1992b; Cathyl-Bickford and Hoffman, 1991; Clapp, 1912a,b, 1914a; Clapp and Cooke, 1917; **England, 1990**; Fahlstrom, 1982; Hanson, 1976; Hudson, 1974; Mackenzie,1923; Massey and Friday, 1988, 1989; Mercier, 1977; **Muller and Jeletzky, 1967, 1970**; **Pacht, 1980,1984;** Ruddiman, 1980; **Usher, 1949, 1952**; **Ward, 1976a, 1978a; Ward and Stanley, 1982;** Winsby, 1973.

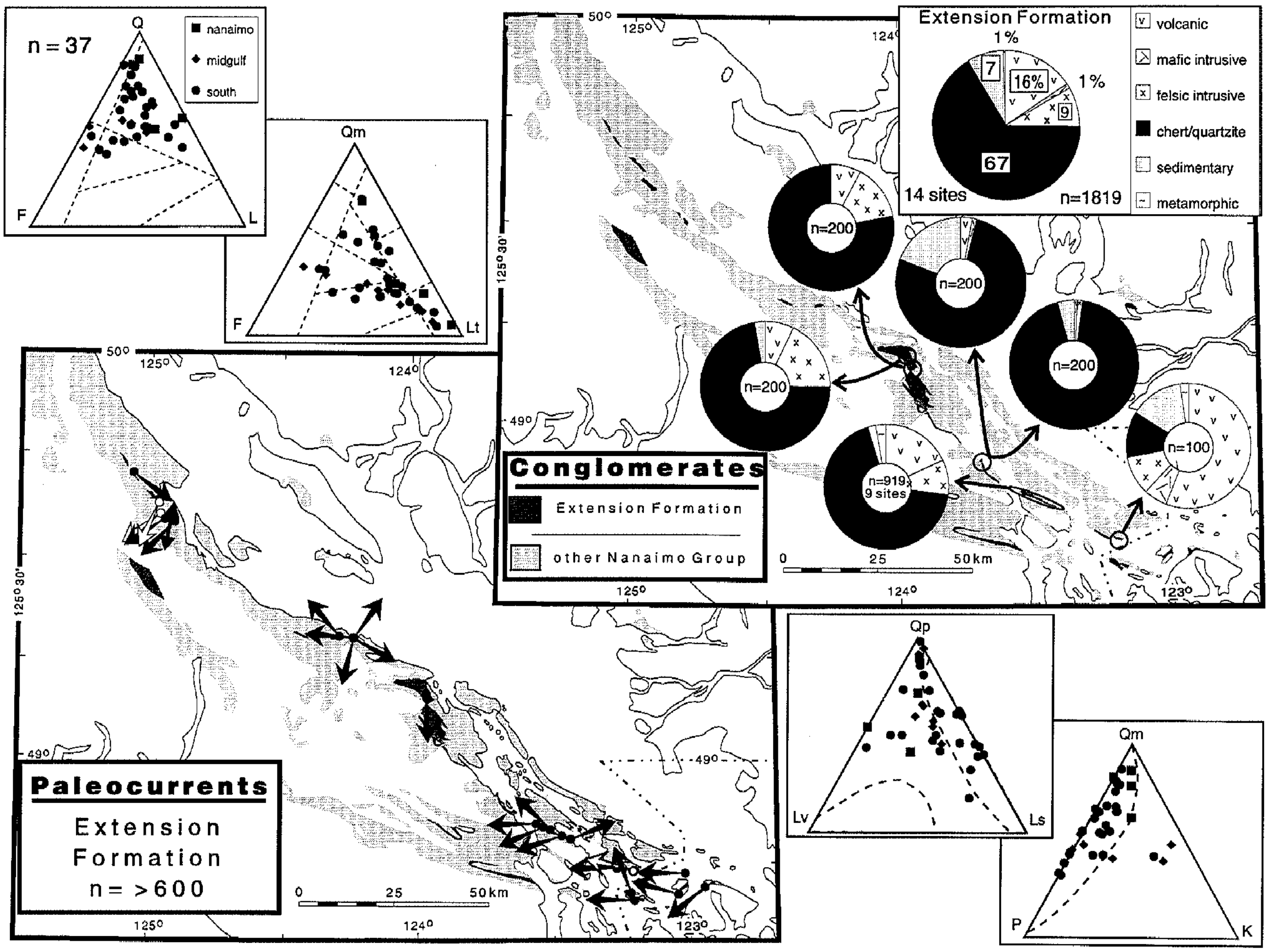

Figure A4. Extension Formation.

Table A4. Pender Formation.

AGE	Early Campanian (Chicoensis biozone; fossils common in some areas)
UPPER CONTACT	Generally gradational to Protection Fm.; in places sharp, erosive contact with Protection Fm. sandstone or conglomerate, but laterally intertonguing regionally. Gradational to Cedar District Fm. where Protection Fm. is not present.
PENDER FORMATION **MAJOR LITHOFACIES**	Outside of Nanaimo area, typified by thick sequences of massive to thin-bedded mudstone and siltstone with common interbeds of fine grained (rarely medium grained) sandstone. Sandstones are arkosic (rarely lithic) arenite (rarely wacke) and occur in thin to medium thickness beds which commonly display Bouma turbidite sequence features (T_{BDE} and T_{CDE} most common). Beds range from laterally persistent to discontinuous. Fining-and thinning-upward trends are apparent on several scales. In several places, a formation-scale thinning and fining upward trend is apparent from the underlying Extension Fm. into the lower Pender Fm. changing to an overall thickening and coarsening upward trend into the overlying De Courcy Fm. Molluscan macrofossils and foraminifera are common and locally abundant with rare discontinuous bivalve coquinas. At Nanaimo, two members are recognized: a basal **Cranberry member** comprising about 50-200 m of coarse grained lithic arenite and wacke interbedded with sandy mudstone and rare discontinuous coal beds; minor lenticular beds of chert-pebble conglomerate occur in the lower part of the unit. This member roughly fines upward in a gradation from the underlying Extension Fm. conglomerate-sandstone up to the overlying **Newcastle member**. The Newcastle mbr. generally thickens and fines southward and consists of 40 m to possibly >130 m of dark grey mudstone including several coal seams 1 to 18 m thick (generally <2 m thick), and rare discontinuous beds of granule to pebble conglomerate or coarse-grained lithic arenite. Molluscan macrofossils are locally common in siltstone-mudstone.
DISTRIBUTION & THICKNESS	Widespread and generally continuous in Comox and Nanaimo outcrop areas. Generally about 100 to 200 m thick in the Comox outcrop area and at Nanaimo, but thickens to south to >300 m at Pender and Waldron islands
LOWER CONTACT	Generally conformable and gradational from Extension Fm. sandstones and interbedded mudstone in fining-thinning upward transition; gradational from Haslam Fm. where Extension Fm. absent
DEPOSITIONAL ENVIRONMENT	Outside of Nanaimo area appears to be predominantly outer shelf to upper slope deposits. Generally hemipelagic mudstone plus low-concentration turbidites generated from fan-deltas in shallow marine or as outer submarine fan facies related to inner and middle fan facies of enclosing and interfingering Extension and Protection formations. Some shallow marine deposits in most southerly areas may reflect submarine fan and shallow marine fan-delta progradation from southeast. At Nanaimo shallow to marginal marine deposition occurred in generally low energy environments interfingering with fluvial-deltaic deposits and back-barrier coal swamps.
PROVENANCE	Sandstone compositions vary from quartz-rich to arkosic to lithic arenite and wackes, reflecting mixed source areas with no strong trends. Paleocurrent data is rare, but in combination with enclosing Protection and Extension formation data, a broadly westerly trend is indicated (varying greatly from generally SW in northern areas, mixed in the Nanaimo area and west to NW in southern areas). Paleocurrent in outer fan facies may be in part basin axial in a NW-SE overall basin orientation. Major source areas are interpreted as the Coast Belt to the northeast and San Juan terranes to the southeast as part of a continuous depositional history for the Extension, Pender, and Protection formations.
REFERENCES (REGIONAL STUDIES IN BOLD)	Bickford et al., 1990; Bickford and Kenyon, 1988; Buckham, 1947a,b; Cathyl-Bickford, 1992a; Clapp, 1912a,b, 1914a; Clapp and Cooke, 1917; **England, 1990**; Fahlstrom, 1982; Hudson, 1974; Mackenzie, 1923; Mercier, 1977; **Muller and Jeletzky, 1967, 1970**; **Pacht, 1980, 1984;** Ruddiman, 1980; **Usher, 1949, 1952**; **Ward,** 1973, **1976a, 1978a; Ward and Stanley, 1982;** Winsby, 1973.

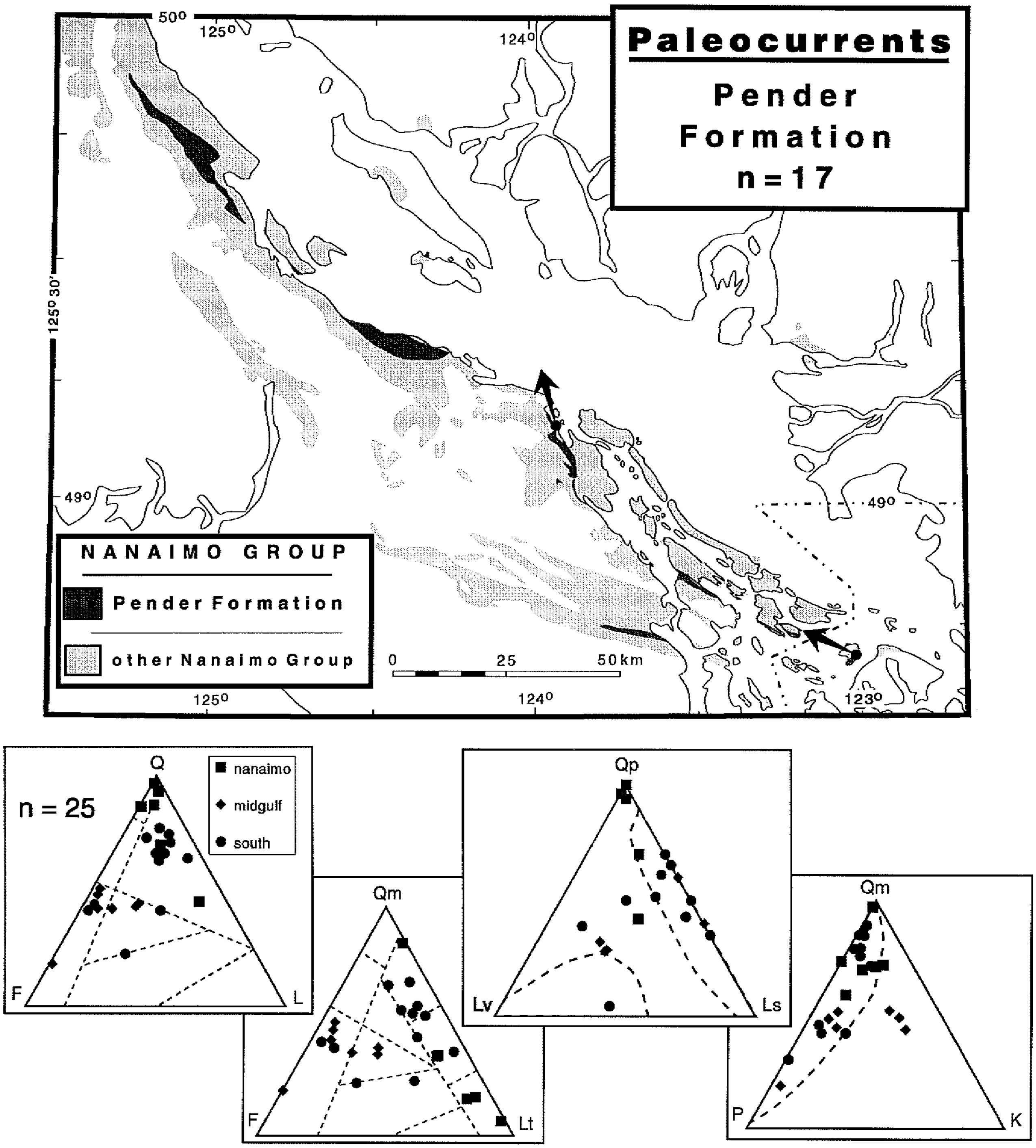

Figure A5. Pender Formation.

Table A5. Protection Formation.

AGE	Early to early Late Campanian (fossils rare and non diagnostic, but age is well constrained by age of adjacent formations).
UPPER CONTACT	Gradational over tens of metres or less and laterally intertonguing to Cedar District Fm. , rarely sharp but conformable.
PROTECTION FORMATION MAJOR LITHOFACIES	Generally resistant sequences of distinctive, light grey to white, medium-to thick-bedded sandstone, with subordinate conglomerate and siltstone. In Nanaimo area three mappable units are defined. The upper **McMillan member** comprises up to 90 m of medium-to coarse-grained arkosic arenite in medium to thick beds, contains trough and planar crossbeds, reactivation surfaces, flaser bedding, and mudstone ripups with intercalated horizontal beds of fine grained sandstone to silty mudstone. Trace and very rare molluscan fossils are present. A central unit (**Reserve member**) comprises 40 to 60 m of sandy siltstone and fine-to medium-grained lithic arenite to wacke, with abundant discontinuous beds of carbonaceous mudstone and thin dirty coals. A basal unit (**Cassidy member**) consists of 80 to 105 m of coarse grained to granule arkosic wacke and arenite, generally thick bedded, massive, and locally including pebble conglomerate. In the southern Gulf Islands the formation thickens and comprises medium to thick beds of arkosic arenite, nongraded to normal graded with rare pebbly bases; generally massive, rare dewatering structures, crossbedded to rippled at top (T_A, T_{ABC} and T_{ABC} Bouma types most common). Interbeds include normal graded pebbly conglomerate in lenticular beds scouring into sandstone; fine-to medium-grained sandstone thin beds grade to silty mudstone (T_{BCDE}; T_{CDE} and T_{ABCDE} Bouma types). Rare thick bedded conglomerate is matrix-supported, massive and poorly sorted. Fining and thinning upward sequences on metre to ten metre scales are common. Rare molluscan fossils and foraminifera are present. In the Comox outcrop area discontinuous sandstone-conglomerate lenses typically comprise medium-to coarse-grained lithic arenite in thin to thick beds. Beds are massive, rarely graded or crossbedded. Interbedded with, and in some places coarsen up to granular to pebble conglomerates, generally poorly sorted, some normally graded. Broken shell material abundant in places and bioturbation evident in minor fine grained sandstone and mudstone; rare synsedimentary slump folds.
DISTRIBUTION & THICKNESS	In Comox outcrop area occurs as 30 m to >60 m thick discontinuous lenses which are broadly channellized into the Pender Fm. Increasing to 180 to 250 m in the Nanaimo area and continuous and southeast-thickening to >400 m at Pender Island.
LOWER CONTACT	Generally gradational or sharp but conformable on underlying Pender Fm. In some places erosive, with Protection Fm. sandstone or rare conglomerate channellized several metres into Pender Fm.
DEPOSITIONAL ENVIRONMENT	General change in southern areas from minor marginal to shallow marine lower deposits to deep shelf submarine fan deposition for most of formation. Most southern sandstone-dominated areas consist of stacked middle to upper submarine fan density-modified grain flows, high to low concentration turbidites, plus less common debris flows and channellized traction carpet gravel-rich flows. Thick megasequence with common fining-and thinning-up sequences suggest rapid lateral migration of small fan lobes and complex overlapping of large fan complexes. The Nanaimo area generally comprises high energy shallow shelf to littoral deposits with tidal channels, storm deposits, and sandy barforms. Minor nonmarine strata reflect braided fluvial to braid delta deposition. Isolated conglomeratic-sandstone bodies in north areas represent submarine channel fills with some deposition in off channel, high energy shallow to littoral marine shelf areas.
PROVENANCE	Paleocurrents are generally westerly and sandstones show very high component of plutonic detritus, suggesting main source is deeply dissected Coast Belt as also suggested by ages of detrital zircons from the Nanaimo area.
REFERENCES (REGIONAL STUDIES IN BOLD)	Allmaras, 1979; **Bickford and Kenyon, 1988**; Buckham, 1947a,b; Cathyl-Bickford, 1992a; Cathyl-Bickford and Hoffman, 1991; Clapp, 1912a,b, 1914a; **England, 1989, 1990**; Fahlstrom, 1982; Hanson, 1976; Hudson, 1974; **Muller and Jeletzky, 1967, 1970**; **Pacht, 1980, 1984**; Ruddiman, 1980; **Usher, 1949, 1952**; **Ward, 1976a, 1978a**, Winsby, 1973.

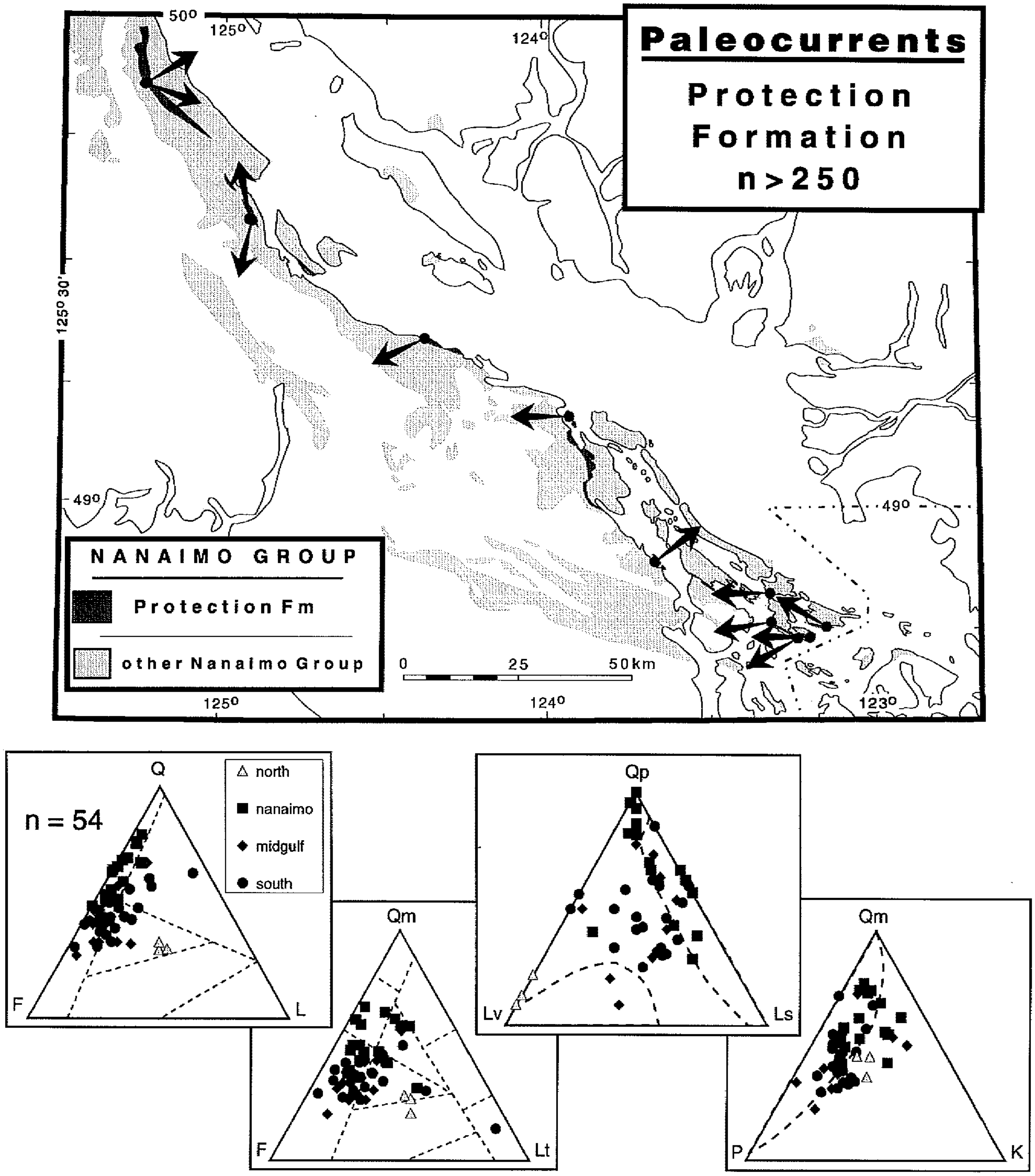

Figure A6. Protection Formation.

Table A6. Cedar District Formation.

AGE	mid to early Late Campanian (Vancouverense and cf. Pacificum biozones; fossils common in some areas)
UPPER CONTACT	Generally gradational with coarsening and thickening upward sandstone transition to De Courcy Fm. A lateral intertonguing relationships with the De Courcy Fm. is well developed in many areas which results in transitional and poorly defined contact
CEDAR DISTRICT FORMATION **MAJOR LITHOFACIES**	Interbedded mudstone, siltstone, and about 25-30% fine-to very fine-grained arkosic arenite to wacke. Mudstone/siltstone dominated generally in Nanaimo and northern areas with 10 m to >100 m of massive to faintly laminated strata which commonly contains carbonate concretions and in places is bioturbated; sandstone interbeds are rare (<10%) throughout; Coarsens and thickens upward in most areas to sandstone/mudstone couplets. Sandstones are generally thin-bedded and very fine-to fine-grained arkosic arenite, but range to thick-bedded and medium grained and include rare lithic wackes. Sole marks, clastic dykes, synsedimentary slumping, and large soft-sediment deformation features are generally rare, but well-preserved and common in some areas. Sandstones display a range of features including common normal grading, crosslaminated ripples, convolute laminae, common loading and relatively rare flute casts, and small-scale dewatering structures (Bouma types T_{BCDE}, T_{CDE}, T_{ABCE} and T_{BE} are common). Most beds are continuous and even thickness for >100s of metres laterally, but some occur as broad, thin lenses with curved, erosive lower contacts (especially common for thicker and coarser beds). Individual poorly sorted, matrix-supported pebble conglomerates are rarely present, most commonly in southern areas (e.g., Saturna, Sucia Islands); contain distinctive milky quartz and metamorphic clasts and abundant sedimentary ripups in >10 m thick beds. Macrofossils and trace fossils are common and locally abundant. Foraminifera are common and generally diverse.
DISTRIBUTION AND THICKNESS	Widespread and apparently continuous in Comox and Nanaimo areas and on most Gulf Islands. About 120 to 200 m thick in northern areas, thickening southward to about 300 m in Nanaimo area and 500 m in southeast. Complex interfingering with De Courcy Fm. causes local variations in thickness from general trend.
LOWER CONTACT	Generally gradational over tens of metres or less from Protection Fm. sandstone and laterally intertonguing; rarely sharp but conformable. Gradational from Pender Fm. in areas where Protection Fm. is not present.
DEPOSITIONAL ENVIRONMENT	Generally "classic" turbidites as part of large, lower and middle submarine fan complexes; complex overlapping of small lobes and migration of fan channels accounts for most variations in type, small scale fining-up sequences and channellized flows. Rare large debris flow conglomerates include resedimented shallow water and older detritus and fossils. Common slump structures and lack of sub-wavebase features suggest outer shelf and slope deposition generally, also suggested by macrofossil and microfossil types and diversity and by thick assemblages of massive and faintly laminated silty mudstone (deposited in areas between major fan channel complexes). General formation-scale coarsening and thickening upward trends and complex interfingering to De Courcy sandstones represent continued progradation of fan complexes in broadly western direction.
PROVENANCE	Radial westerly paleocurrent patterns, synsedimentary slump fold vergence directions and lateral thickening and coarsening trends all suggest overall western facing paleoslope and westerly fan depositional patterns with paleocurrents becoming basin axial in some places. Sandstone compositions suggest mixed source areas with strong components of deeply dissected Coast Belt plutonic detritus and possible input from northwest Cascades.
REFERENCES (REGIONAL STUDIES IN BOLD)	Allmaras, 1979; Breitsprecher, 1962; Cathyl-Bickford, 1992a; Clapp, 1912a,b, 1914a; **England, 1989, 1990**; Hanson, 1976; Hudson, 1974; Janbaz, 1972; **McGugan, 1962, 1964,** 1981; **Muller and Jeletzky, 1967, 1970**; **Pacht, 1980, 1984; Rahmani, 1968; Scott, 1974a, b**; Simmons, 1973; **Sliter, 1973;** Stickney, 1976; Sturdavant, 1975; Usher, 1949, 1952; **Ward,** 1973, **1976a, 1978a**, Winsby, 1973.

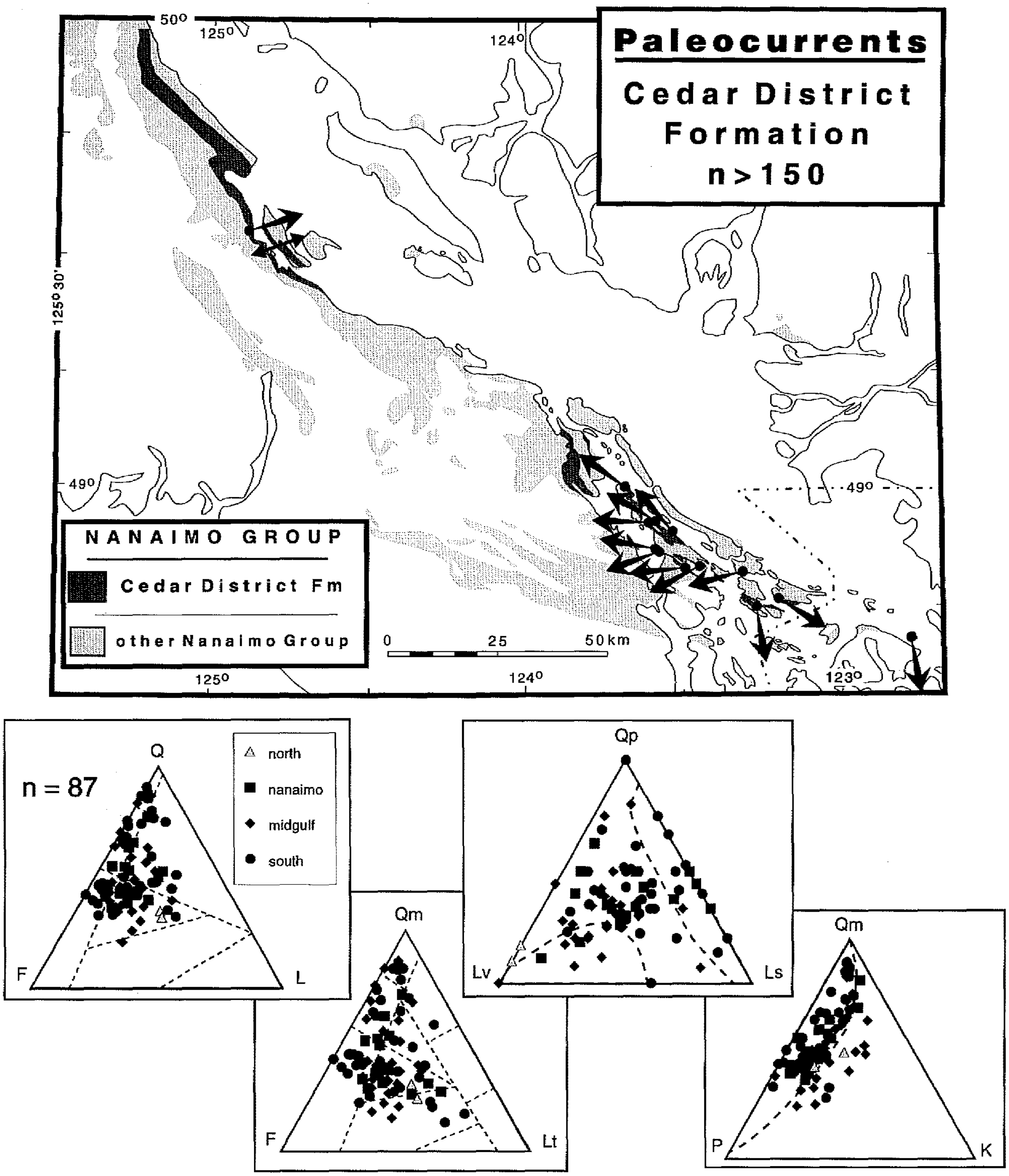

Figure A7. Cedar District Formation.

Table A7. De Courcy Formation.

AGE	Late Campanian (rare and nondiagnostic fossils, constrained by age of adjacent formations).
UPPER CONTACT	Thinning-fining up transition to mud-rich Northumberland Fm. with lateral intertonguing common.
DE COURCY FORMATION MAJOR LITHOFACIES	Thick-bedded, medium-to coarse-grained sandstone and rare to common pebble-cobble conglomerate thick beds and lenses; all interbedded with lessor amounts of fine grained sandstone, siltstone, and mudstone. Dominated by stacked, laterally continuous thick beds of arkosic arenite; in some places as amalgamated massive sets several 10s of m thick. Thick sandstones are generally massive, nongraded to normal or rarely inverse graded, many with thin upper planar to convolute laminated parts (Bouma T_A, T_{AB}, T_{ABC}, T_{BC} types). Pipe, sheet, and dish dewatering structures and load casts are common; flutes and grooves are less common. Thick beds are commonly separated by thin-bedded, fine grained arkosic arenite and wacke intercalated with laminated to massive siltstone to mudstone (generally organized in Bouma T_{CDE}, T_{BE}, and T_{DE} types). Fining-and thinning-up sequences of metres to roughly ten metre scales are common (but thick noncycle successions also occur). Several upward coarsening-thickening megasequences >10 m thick occur in lower parts (intertonguing transition from Cedar District Fm.). Pebble-cobble conglomerates occur in generally laterally discontinuous thick beds; commonly clast-supported and normal graded to nongraded, rarely graded-stratified with basal clast imbrication and rare trough or planer crossbeds. Conglomerates generally occur as individual beds in sandstone dominant successions, but in some places are predominant in sequences up to 80 m thick (e.g., Denman Island), with less common sandstone interbeds. Thick conglomerate-rich successions are broadly channelled into underlying sandstone and transitional laterally and vertically into overlying sandstone (defining fining and thinning upward megasequences >100 m thick). Rare matrix-supported, disorganized conglomerates occur as individual thick beds, generally containing abundant ripups of older strata. Synsedimentary deformation is common, including folding and small faults, clastic intrusions, and sedimentary float-breccias, some >30 m thick. Trace fossils are common in some places; foraminifera and molluscan macrofossils are rare.
DISTRIBUTION & THICKNESS	Well exposed and continuous in eastern Nanaimo area, central Gulf Islands and on Denman Island. About 200 to 250 m thick on Denman Island, thickening from about 300 m at Nanaimo area to 450 m in southern Gulf Islands.
LOWER CONTACT	Generally gradational with coarsening and thickening upward sandstone transition from Cedar District Fm. Lateral intertonguing relationships are well developed in many areas, resulting in transitional, poorly defined contacts.
DEPOSITIONAL ENVIRONMENT	Generally middle and upper submarine fan facies in major fan complexes on broadly west-northwest sloping shelf (subwavebase) and slope environment. Part of evolving submarine fan deposystem with both underlying/interfingering Cedar District Fm. and overlying/interfingering Northumberland Fm. as lower fan and off fan deposits. Thick sandstone beds from mid- to upper fan density modified grain flow and high concentration turbidity flow in broad overlapping fan lobes and channel/off-channel deposits (fining-and thinning-up sequences). Thick conglomerate-sandstone sequences are major fan channel deposits and upper fan high concentration turbulent flows, traction carpet deposits and clast-rich debris flows. Conglomerate-dominated megasequences (e.g., Denman Island) are submarine channel fills.
PROVENANCE	Paleocurrents show radial patterns typical of submarine fan complexes, ranging from SE to NW with a strong westerly main trend. Sandstones and conglomerate compositions suggest mixed sources - chert-rich in south reflecting San Juan terranes, arkosic sands and volcanic clast-rich conglomerates in north area are probably Coast Belt derived.
REFERENCES (REGIONAL STUDIES IN BOLD)	Allmaras, 1979; Buckham, 1947a,b; Cathyl-Bickford, 1992a; Clapp, 1912a,b, 1914a; **England, 1989, 1990**; Grieve, 1974; Hanson, 1976; Hudson, 1974; **Muller and Jeletzky, 1967, 1970**; **Pacht, 1980; Page, 1972;** Simmons, 1973; Stickney, 1976; Sturdavant, 1975; **Usher, 1949, 1952**; **Ward, 1976a**; Williams, 1924; Winsby, 1973.

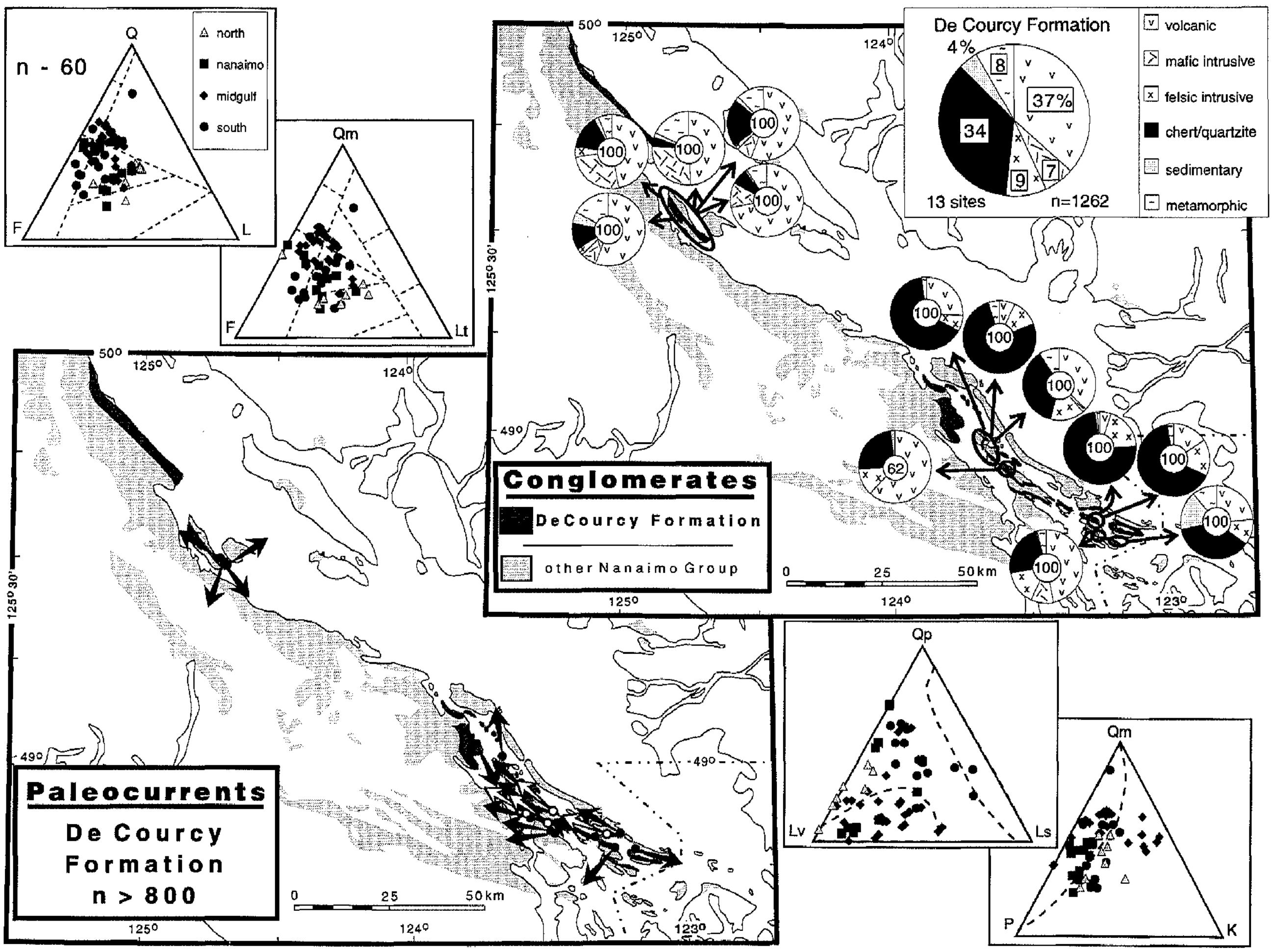

Figure A8. De Courcy Formation.

Table A8. Northumberland Formation.

AGE	Late Campanian to earliest Maastrichtian (Suciaensis biozone and age-diagnostic foraminifera); Campanian- Maastrichtian boundary defined by foraminifera on Hornby Island (McGugan, 1982).
UPPER CONTACT	commonly sharp where Geoffrey Fm. downcuts in channels, but gradational and slightly intertonguing in other areas
NORTH-UMBER-LAND FORMATION MAJOR LITHOFACIES	Typically recessive grey mudstone and siltstone with variable amounts of sandstone interbeds. Mudstone-siltstone occurs in thin to thick sequences, generally massive to discontinuously laminated, with rare to common calcareous and sideritic concretions. Sandstone as isolated thin beds or in rhythmically interbedded sandstone-mudstone sequences. Beds generally comprise very fine-to fine-grained arkosic arenite (rarely wacke), internally massive to laminated, rarely containing rippled or convolute upper laminae (Bouma types T_{DE}, T_{CDE}, T_{ADE} are most common). Beds are generally continuous laterally with planar to slightly loaded, rarely fluted bases. Medium-to coarse-grained arkosic arenite thin to thick beds and rare pebble conglomerates occur in some areas, most commonly as part of a basal transition (10s of m in some places) from the underlying De Courcy Fm. and rarely in a thin (< 10 m) coarsening-thickening upper transition to the Geoffrey Fm. (e.g., Saltspring, Denman Islands), or where lateral intertonguing is developed with the Geoffrey Fm. (e.g., east Mayne to west Saturna Islands, Oyster River coastline). Synsedimentary slump structures and sandstone dykes are common in some areas. Very slight formation-scale increase in abundance of sandstone upward, except in basal fining-thinning transition from De Courcy sandstone. Macrofossils are rare but present throughout with crab fossils abundant at coast near Oyster River; Diverse and abundant foraminifera have been recovered from most areas. Trace fossils are common to abundant in sandstone beds and in places original stratification is disturbed by strong bioturbation. Detrital plant material locally common.
DISTRIBUTION AND THICKNESS	Narrow and recessive generally but continuous on most north and south Gulf Islands and exposed at mouth of Oyster River. Varies between 100 to 400 m thick, generally 200 to 300 m, but locally deeply dissected by overlying Geoffrey Fm. erosive contact and lower interfingering with De Courcy Fm. also causes thickness variations.
LOWER CONTACT	gradational thinning-fining up transition from sandstones of De Courcy Fm.; lateral intertonguing common
DEPOSITIONAL ENVIRONMENT	Massive to laminated silty mudstone represent low energy deposition in areas "distal" to fan complexes. Sandstones from low concentration turbidites deposited as part of large lower and middle submarine fan complexes, both in areas "distal" to and between main fan channels. Rare slump structures and lack of subwavebase features suggest outer shelf and slope deposition generally, also suggested by macrofossil and microfossil types and diversity and by thick assemblages of massive and faintly laminated silty mudstone. Probably west-to northwest-facing paleoslope (based on evidence in enclosing coarse formations). Interfingering to De Courcy sandstones at base represents general fan system abandonment in early stages of deposition. Thin gradational coarsening-thickening upward transition and rare interfingering to Geoffrey Fm. plus deeply eroded Geoffrey Fm. channels in some places suggests rapid progradation of main fans and perhaps new or rapidly migrating major fan channels during late stages of Northumberland deposition.
PROVENANCE	Sandstone compositions suggest mixed sources, generally from uplifted and strongly dissected plutonic rocks of Coast Belt to east. Paleocurrent data sparse and flows directions may be basin axial or transverse to main fan slopes as levee or off-levee spillover deposits.
REFERENCES (REGIONAL STUDIES IN BOLD)	Allmaras, 1979; Carter, 1977; Clapp, 1912a,b, 1914a; **England, 1989, 1990; England and Hiscott, 1992**; Fiske, 1977; Grieve, 1974; Hanson, 1976; Hoen, 1958; Hudson, 1974; **McGugan, 1964, 1979, 1982;** Meding, 1964; **Muller and Jeletzky, 1967, 1970**; Packard, 1972; Richards, 1975; Simmons, 1973; **Sliter, 1973**; Stickney, 1976; Sturdavant, 1975; **Usher, 1949, 1952**; **Ward, 1976a,** Williams, 1924; Winsby, 1973.

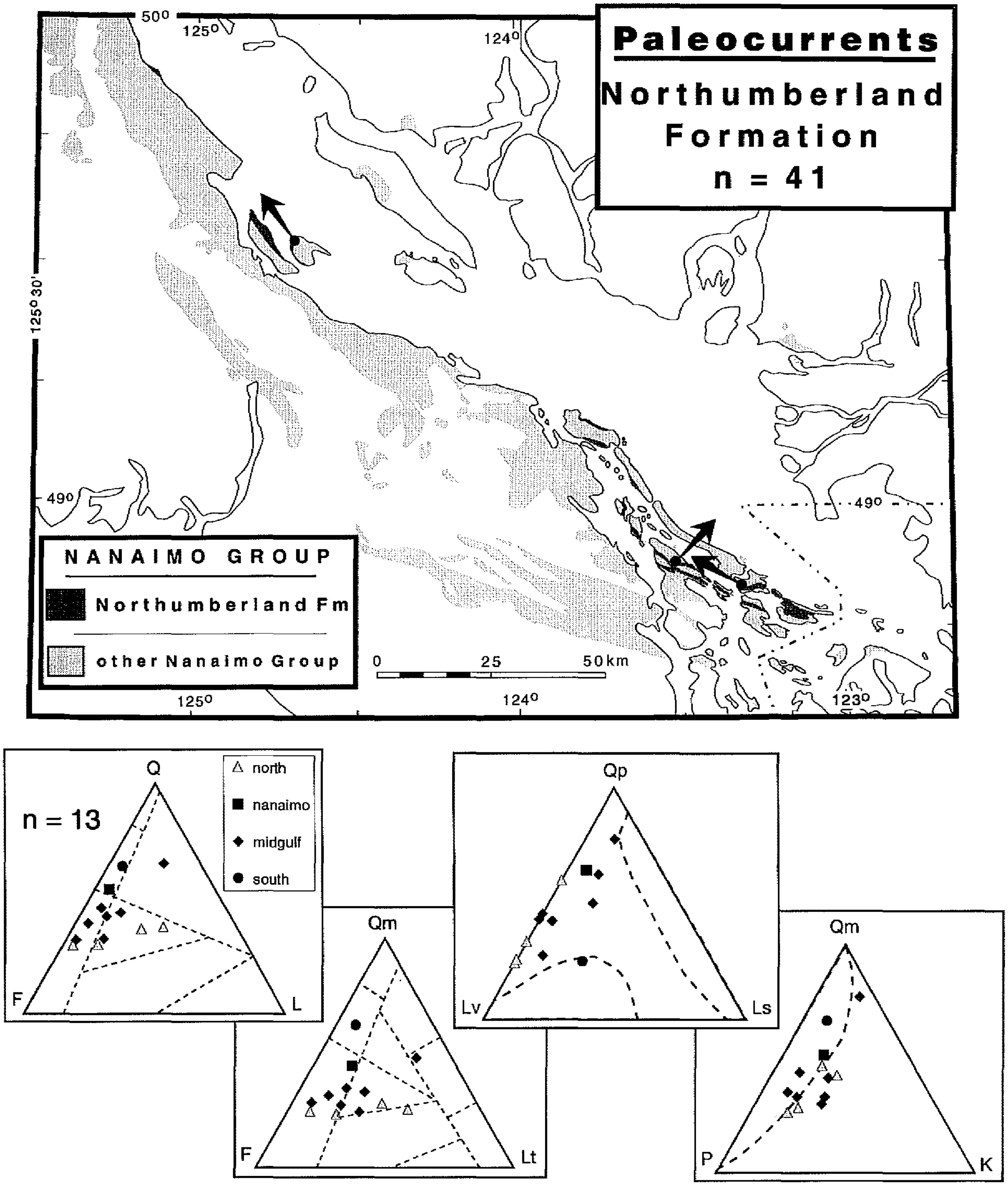

Figure A9. Northumberland Formation.

Table A9. Geoffrey Formation.

AGE	late Campanian - early Maastrichtian (rare and nondiagnostic fossils, well constrained by age of adjacent formations)
UPPER CONTACT	generally gradational in thinning-fining upward transition (several 10s of metres)
GEOFFREY FORMATION MAJOR LITHOFACIES	Resistant thick- to medium-bedded, coarse grained arkosic arenite and pebble to boulder conglomerate, with minor finer interbeds most common in sandstone-rich parts of the formation. In general a lower sandstone dominant unit can be recognized, plus a central conglomerate-rich unit which fines and thins upward to a capping sandstone-rich unit transitional to overlying mudstone-rich Mayne Fm. However, lateral diversity is common and in several places the lower sandstone unit is missing where the conglomerate-rich package occurs in an irregular erosive-based broad channel geometry with several hundred metres of downcutting present in packages traceable for 5-10 km laterally. In these areas conglomerate sharply overlie the mudstone-rich Northumberland Fm. with erosive contact. Sandstones throughout tend to be coarse-to less common medium-grained, normal graded with pebbly bases in some places, massive in lower part and changing to parallel laminated or cross-stratified and commonly with dewatering structures in upper parts. Bases are commonly loaded or scoured into underlying beds and rip-up clasts of mudstone are common. Conglomerates are typically clast-rich (framework-supported more common than matrix-supported) and moderately to poorly sorted with pebbles, cobbles, and rare boulders in a coarse grained arkosic matrix. Internally conglomerates range from massive and nonstratified to graded-stratified with pebbly sandstone tops; normal grading is more common than inverse grading, basal imbrication of tabular clasts is common in stratified conglomerates. Most beds appear laterally discontinuous, lenticular over a few tens to hundreds of metres, and complexly overlap with curved based erosive into underlying beds. Conglomerate clasts are generally subangular to subround and in some places sedimentary clasts are common (10-15%) including both contorted rip-ups of semilithified mudstone and lithified clasts apparently derived from erosion of older Nanaimo Group strata. Mudstone and fine grained sandstone occur as rhythmically interbedded packages, generally <1 m thick, rarely up to 10 m and increasing in thickness and abundance in the upper part of the formation. Thin-to medium-bedded sandstone in these sections display typical Bouma sequences (T_{BE}, T_{BCE}, and T_{BCDE} most common). Molluscan macrofossils are rare and trace fossils locally abundant.
DISTRIBUTION AND THICKNESS	Well exposed and continuous on outer Gulf Islands from Denman to Saturna on major inner islands (Saltspring, Pender). Varies from about 150 m to >500 m thick; thickest where broad conglomerate channels present (e.g., >400 m on Denman, >500 m on Galiano and Saturna islands)
LOWER CONTACT	sharp where major channel forms well developed (e.g., Denman, east Saturna islands), but gradational and intertonguing in other areas (e.g., Saltspring Island, east Mayne to west Saturna islands)
DEPOSITIONAL ENVIRONMENT	Middle and upper submarine fan facies as part of major fan complexes. Common channellized high concentration turbulent and nonturbulent flows account for many conglomerate and coarse sandstone deposits, but wide range of flow types including debris flow, high density debris flows, nonchannellized low concentration sandy "classic' turbidity flows, surging traction carpet deposits and modified grain flows. Probable outer shelf to slope depths suggest by lack of wavebase features and interfingering with enclosing mudstone of similar depths.
PROVENANCE	Radial paleocurrents typical of submarine fans suggests general west to northwest deposition of fan complexes; sandstones arkosic to lithic from mixed sources, most likely dissected Coast Belt and north Cascades.
REFERENCES (REGIONAL STUDIES IN BOLD)	Clapp, 1912a,b, 1914a; Carter, 1977; **England, 1989, 1990**; **England and Hiscott, 1992;** Fiske, 1977; Grieve, 1974; Hanson, 1976; Hoen, 1958; Hudson, 1974; **Muller and Jeletzky, 1967, 1970**; **Pacht, 1980, 1984;** Packard, 1972; **Page, 1972;** Simmons, 1973; Stickney, 1976; Sturdavant, 1975; **Usher, 1949, 1952**; Williams, 1924.

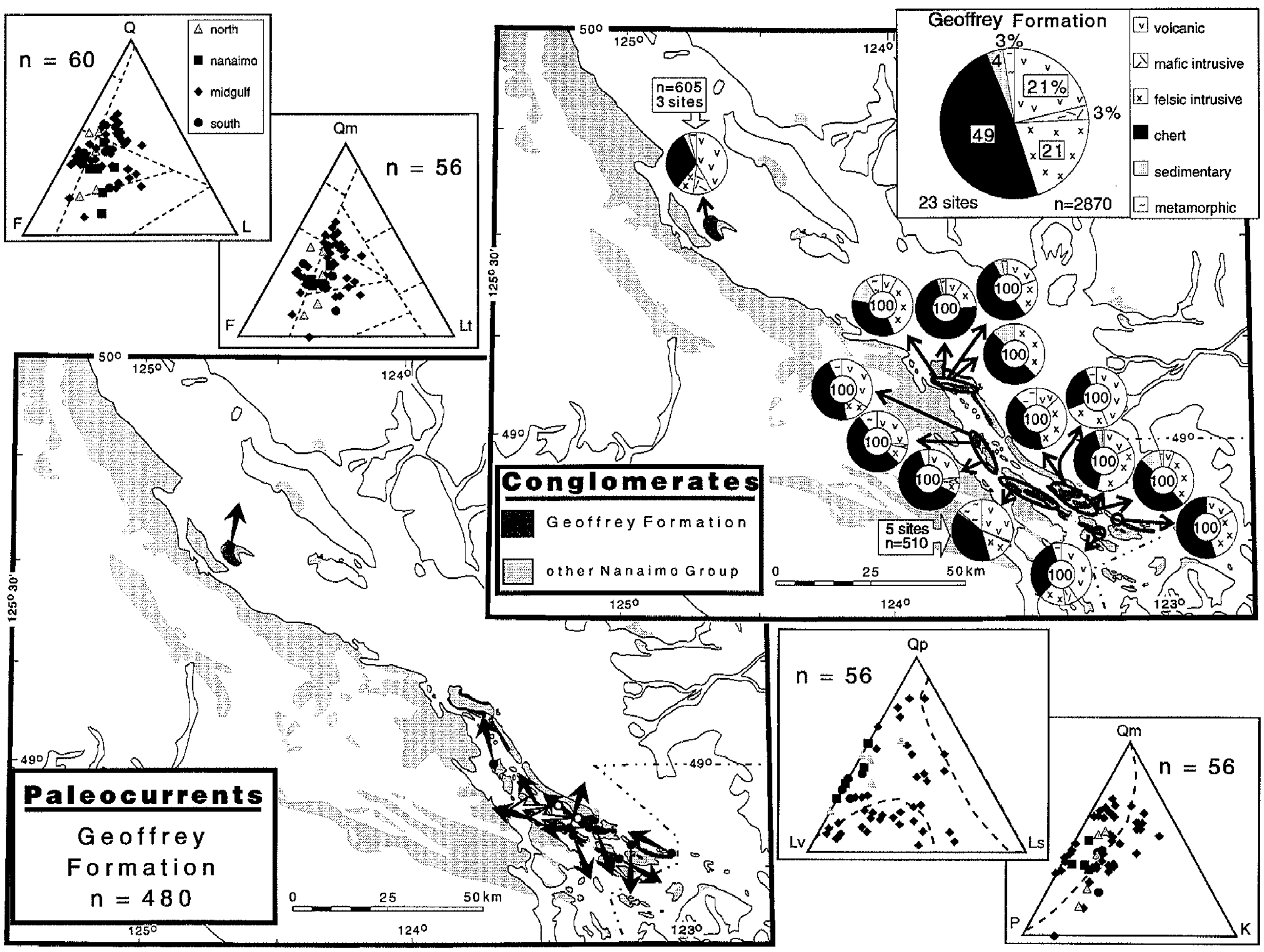

Figure A10. Geoffrey Formation.

Table A10. Spray Formation.

AGE	early Maastrichtian (Suciaensis biozone macrofossils and common Maastrichtian foraminifera)
UPPER CONTACT	generally gradational over <10 m in abrupt coarsening-thickening up transition to Gabriola Fm. sandstone thick beds, rarely sharp with thick-bedded sandstones directly overlying or slightly erosive into Spray Fm. mudstone-sandstone, but appears conformable
SPRAY FORMATION MAJOR LITHOFACIES	Typically recessive grey mudstone and siltstone with variable amounts of sandstone interbeds. Mudstone-siltstone occurs in thin to thick sequences, which are usually recessive, but where exposed are generally laminated to thin bedded or massive, with rare to common calcareous and sideritic concretions. Sandstone generally present as isolated thin beds or in rhythmically interbedded sandstone-mudstone sequences. Common in the lower transition from the Geoffrey Fm., the sandstones are arkosic arenite, generally fine grained thin beds which range from massive to laminated, some contain convolute lamina or ripple cross-laminae (Bouma sequence T_{DE}, T_{BCDE}, T_{CDE} are most common). Beds are generally laterally continuous and planar. Sole marks are very rare. Rare coarse- or medium-grained thick sandstone beds occur, most common in the lower part of the formation. These tend to be normal graded with laminated to rippled or crossbedded upper parts, have curved bases which erode slightly into underlying bases, contain rare to abundant basal mudstone rip-up clasts, and are laterally discontinuous over tens to a few hundred metres. Very rare matrix-supported pebble conglomerate thick beds, very poorly sorted and massive. Sandstone dykes and synsedimentary slumping structures (contorted and folded layers - generally verging to north or northwest) and disrupted layers are rare to common. Molluscan macrofossils are rarely present, foraminifera are common and diverse, trace fossils are common in some sandstone.
DISTRIBUTION & THICKNESS	Recessive but continuous on outer Gulf Islands from Hornby to Mayne and on Saltspring, Pender and Prevost islands. about 250 to 300 m thick on Hornby Island, varies from 100 to >300 m thick in southern Gulf Islands
LOWER CONTACT	generally gradational in >10 m fining-thinning upward transition from Geoffrey Fm; lateral intertonguing common with thick tongues of Spray Fm. in upper 100 m of Geoffrey Fm.
DEPOSITIONAL ENVIRONMENT	Sandstones from low concentration turbidites deposited as part of large lower and middle submarine fan complexes, both in areas "distal" to and between main fan channels. Rare slump structures and lack of subwavebase features suggest outer shelf and slope deposition generally, also suggested by macrofossil and microfossil types and diversity and by thick assemblages of massive and faintly laminated silty mudstone. Probably west-to northwest-facing paleoslope (based on evidence in enclosing coarse formations and synsedimentary fold vergence). Interfingering to Geoffrey Fm. sandstones at base represents general fan system abandonment in early stages of deposition. Thin gradational coarsening-thickening upward transition to Gabriola Fm. sandstones suggests very rapid progradation of main fans and perhaps new or rapidly migrating major fan channels during late stages of Spray Fm. deposition.
PROVENANCE	Paleocurrent data is sparse, but northwest to west flow generally indicated by paleoslope indicators and trends of enclosing formations. Sandstone compositions reflect uplifted and deeply dissected plutonic rocks of Coast Belt to east.
REFERENCES (REGIONAL STUDIES IN BOLD)	Clapp, 1912a,b, 1914a; Carter, 1977; **England, 1989, 1990**; **England and Hiscott, 1992;** Fiske, 1977; Hanson, 1976; Hoen, 1958; Hudson, 1974; **McGugan, 1979, 1982; Muller and Jeletzky, 1967, 1970**; **Pacht, 1980, 1984;** Packard, 1972; **Page, 1972;** Richards, 1975; **Sliter, 1973**; Stickney, 1976; Sturdavant, 1975; Usher, 1949, 1952; **Ward 1976a**; Williams, 1924.

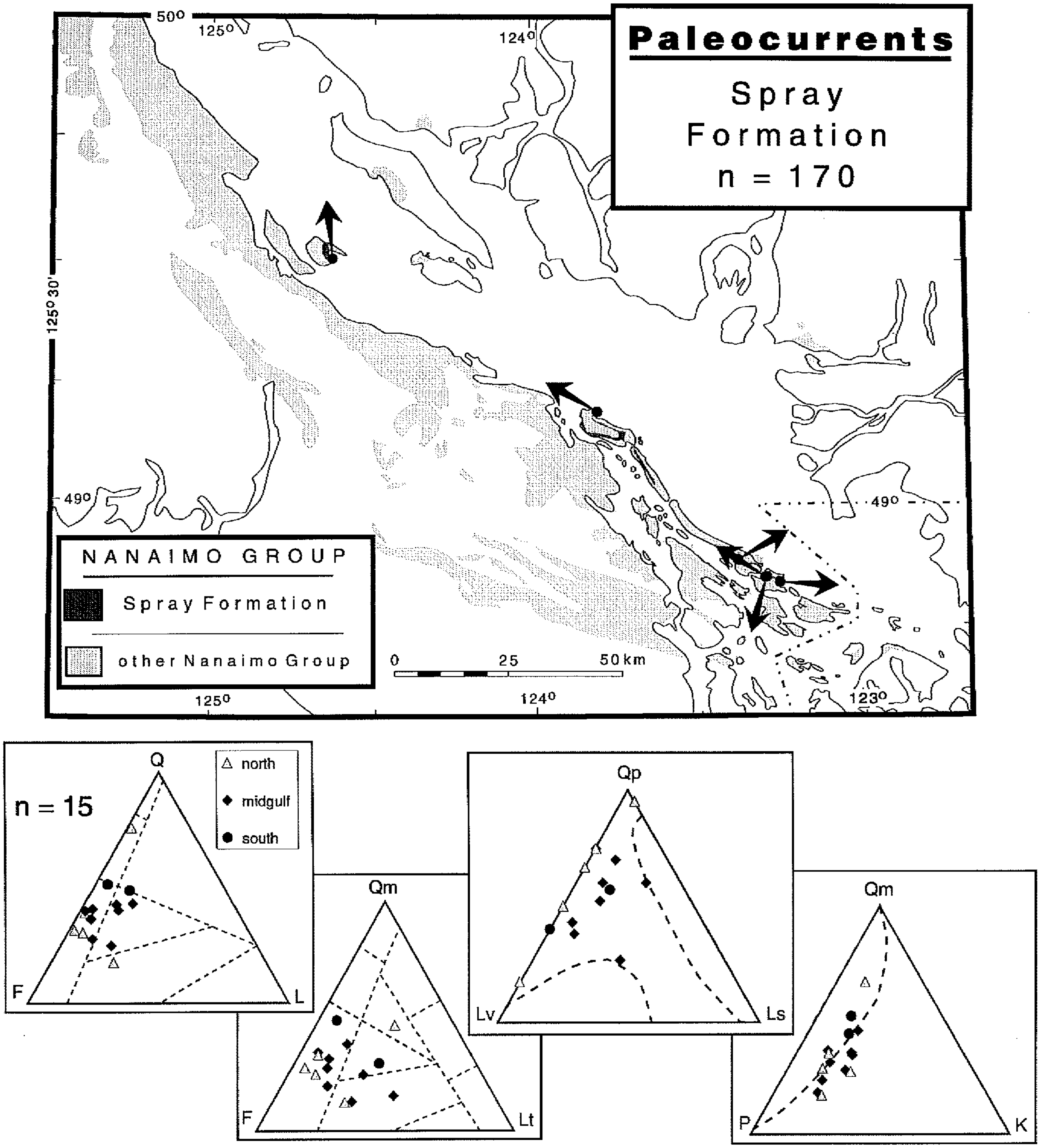

Figure A11. Spray Formation.

Table A11. Gabriola Formation.

AGE	Maastrichtian; poor age control; based on very rare macrofossils and gradational contact with underlying Spray Fm.
UPPER CONTACT	not exposed - probable cut out by unconformity at base of Paleocene clastics (closely superjacent to Nanaimo Group at Lasqueti, Tumbo and Sucia islands)
GABRIOLA FORMATION MAJOR LITHOFACIES	Thick-bedded coarse- to medium-grained arkosic arenite with an upper conglomerate-rich unit on Hornby Island. Greater than 90% sandstone except at Hornby Island, beds are laterally continous, massive to vaguely stratified in upper part and contain rare upper rippled or convolute thin beds (T_{ABC} or T_{AB} Bouma types where present), rarely normal graded with basal pebbles-granules or mudstone rip-ups. Laminated silty mudstone interbeds are rare to common, generally <20 cm thick and in many places deformed by loads and flame into overlying sandstone. Thicker packages (generally <10 m, rarely up to 50 m) of mudstone and thin-bedded, fine grained sandstones rarely occur and are similar to Spray Fm. lithotypes. Clastic intrusions and extensive synsedimentary disrupted layers are present in some places. Pebble conglomerate thick beds are rarely present in southern Gulf Islands and generally occur as graded-stratified bed which grades up to coarse grained sandstone. At Hornby Island a lower package >60 m of thick-bedded sandstones as above are sharply and erosively overlain by >250 m of pebble-cobble conglomerate thick beds with lesser amount of coarse-sandstone interbeds. The conglomerate unit has a curved lower contact, defining a steep-sided channel fill with >30 m relief into underlying thick sandstone beds. Lower conglomerates are clast-supported and massive in beds laterally continuous or overlapping lenticular over 100s of metres with <10% coarse grained massive to cross-stratified sandstone interbeds. Overall fining-thinning upward trend is apparent in top 100 m with increasingly abundant grade-stratified pebble-conglomerates and interbedded pebbly sandstone to coarse-grained sandstone. Fining-thinning up sequences a few metres or thinner are present, rarely capped by discontinuous silty mudstone beds up to 10 cm thick. Planar and trough crossbedding is present in some places in this upper part of the preserved formation.
DISTRIBUTION & THICKNESS	Continuously exposed on outer Gulf Islands from Hornby Island in north to Mayne Island in south. Upper contact not seen with exposed thickness varying from about 350 m (Hornby, Gabriola, Mayne islands) to >500 m at Galiano Island
LOWER CONTACT	generally gradational in <10 m coarsening-thickening upward transition from Spray Fm. mudstone-sandstone, rarely sharp and slightly erosive, but apparently conformable
DEPOSITIONAL ENVIRONMENT	Middle and upper submarine fan facies as part of major fan complexes derived roughly from east. High concentration nonchannellized turbulent and nonturbulent flows account for most coarse sandstone, with interbedded finer sandstone and mudstone from nonchannellized low concentration "classic' turbidity flows. Thick-bedded conglomerates at Hornby Island are major submarine channel fills, deposited from gravelly high concentration flows, including surging traction carpet and density modified grain flows. Higher graded-stratified pebble-conglomerate and sandstone are from turbulent high-concentration flows in smaller overlapping channels in upper fan areas. Probable outer shelf to slope depths suggest by lack of wavebase features and interfingering with subjacent mudstone of similar depths.
PROVENANCE	Paleocurrents show radial pattern typical of submarine fans with overall westerly to northwesterly major flow. Sandstone show mixed sources with high component from uplifted and deeply dissected Coast Belt to east. Abundant detrital zircons generally Cretaceous indicating derivation from Coast Belt intrusives (including ages as young as 72 Ma), but also early Proterozoic and Archean zircons probably derived from eastern Cordillera, suggesting major river systems adding sediments from distant sources.
REFERENCES REGIONAL IN BOLD	Clapp, 1912a,b, 1914a,b; Carter, 1977; **England, 1989, 1990**; **England and Hiscott, 1992;** Fiske, 1977; **Muller and Jeletzky, 1967, 1970**; Packard, 1972; Page, 1972; Stickney, 1976; **Usher, 1949, 1952; Ward, 1976a**; Williams, 1924.

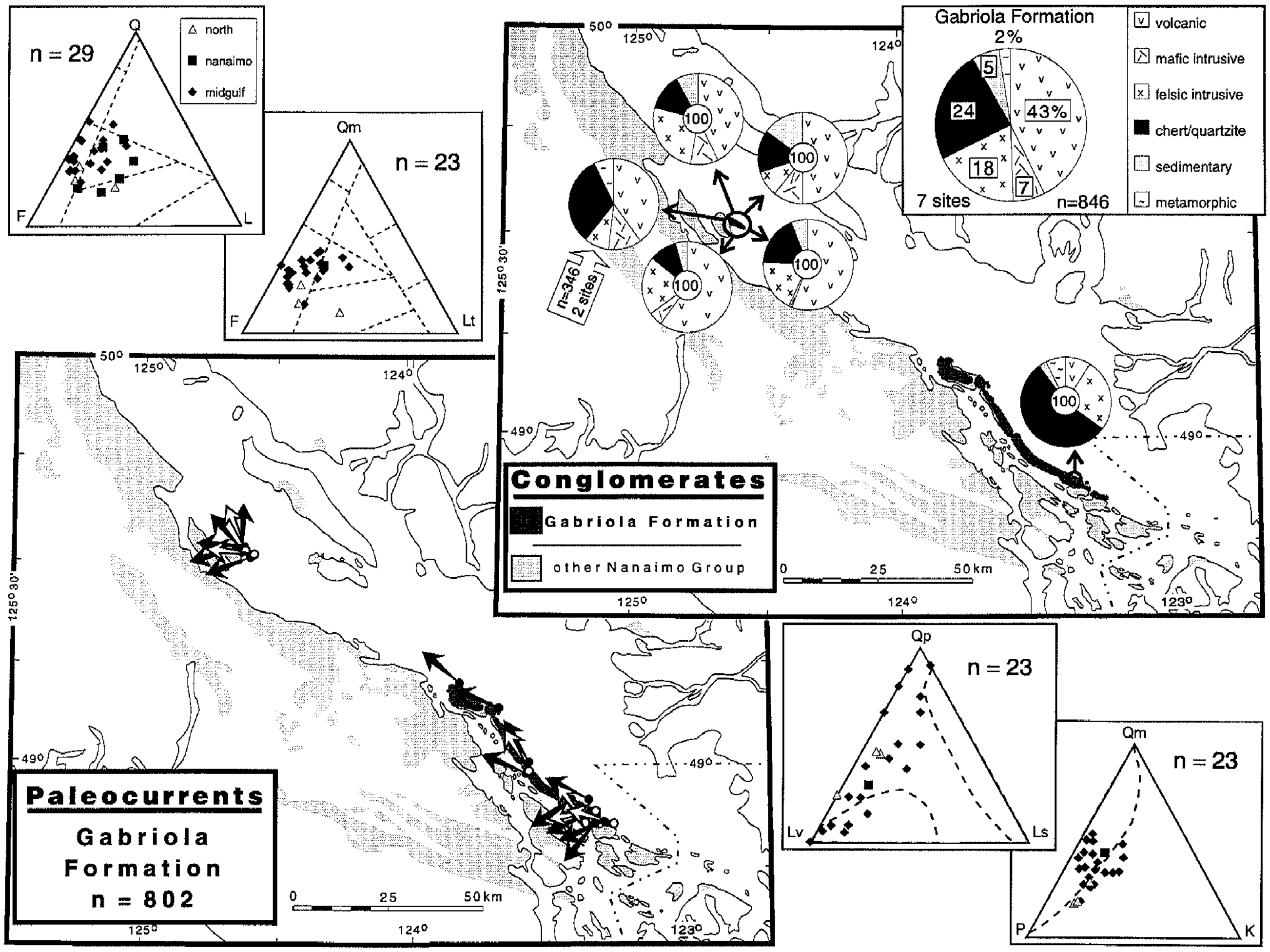

Figure A12. Gabriola Formation.

APPENDIX B

A review of the evolution of Nanaimo Group lithostratigraphic nomenclature

This review of the Nanaimo Group uses a unified formation nomenclature (Fig. 4) based on that of Muller and Jeletzky (1970), as modified by Ward, (1978a). However, several other formation divisions and names have been proposed, both as part of the natural evolution of basin studies in this region, and due to a fundamental disagreement on the original number of depositional basins during Nanaimo Group sedimentation. The evolution of the lithostratigraphy and the evidence concerning one versus two or more original depositional basins are presented in this Appendix, separate from the paper, because the involved arguments are not directly relevant to Nanaimo Group sedimentation and tectonic evolution.

Figure 3 shows the major formation schemes proposed in the last 100 years. Richardson (1872, 1873, 1878) and Clapp (1912a, b, 1914a) suggested deposition in discrete basins for the Comox, Nanaimo and Cowichan-Duncan areas (shown on Fig. 2). Buckham (1947a, b) considered that there were five separate basins of Nanaimo Group deposition. Usher (1952) recognized Comox and Nanaimo basins and proposed two separate sets of formation names for the two areas. Muller and Jeletzky (1970) retained the names Comox and Nanaimo for the separate geographic occurrences of these major outcrop areas, but suggested there was only one original depositional basin, and erected a unified scheme of nine formations for the entire Nanaimo Group, a scheme later expanded to eleven formations by Ward (1978a). The latter formation names have been widely accepted in the literature (e.g., Ward and Stanley, 1982; Pacht, 1984; Bickford and Kenyon, 1988; Gordy, 1988; Haggart, 1991). However, McGugan (1979, 1990) and England (1989), argued that the Comox and Nanaimo outcrop areas do represent discrete depositional basins and thus replacing formation names in one basin with those from another (as done by Muller and Jeletzky, 1970) was a contravention of stratigraphic practice as suggested in the North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature, 1983). This has resulted in several suggested revisions to the lithostratigraphy, most notably McGugan (1990), England (1989), and England and Hiscott (1992).

The objections to the unified formation nomenclature essentially rest on the perception that the Comox and Nanaimo outcrop areas reflect different depositional basins. If the areas are merely erosional remnants of a once continuous basin, the objection to replacing names in one basin with those in another is unfounded. There seems little reason to suggest there were originally two basins. The lithostratigraphy of the two areas is comparable, although lateral facies changes occur as would be expected over the 200 km of lateral extent. A significant test of this premise is the demonstrated ability of most researchers to successfully recognize and use the formation names of Muller and Jeletzky (1970) in both the Nanaimo and Comox and other outcrop areas (e.g., Ward, 1978a; Bickford and Kenyon, 1988; Massey and Friday, 1988). The age of the formations are slightly diachronous laterally, both between Comox and Nanaimo outcrop areas and within the "basins". This is again an expected occurrence in any extensive sedimentary sequence and is irrelevant to the problem of lithostratigraphic terminology (as is clearly stated in the Stratigraphic Code, Article 22). The misperception of separate depositional basins appears to stem from the paleotopography of the sub-Nanaimo Group unconformity surface. Local paleotopographic relief of several hundred metres can be defined in several areas, such as on Saltspring Island and in the Duncan areas of the southern Nanaimo outcrop area and in the Tsable River area of the Comox outcrop area (Atchison, 1968; Hanson, 1976; Muller and Atchison, 1970). An area encompassing the Nanoose region also contains a basement high, termed the Nanoose Arch or Uplift by several workers (e.g., Sutherland-Brown and Yorath, 1985; Massey and Friday, 1989). This pre-late Cretaceous structural culmination, coincidentally preserved at the eastern coast of Vancouver Island, has resulted in a very slight geographic separation of Nanaimo Group outcrops, thus the misperception of originally sparate basins. Sedimentation patterns of early Nanaimo Group deposition have been influenced by this paleotopography and the formations adjacent to all the paleohighs mentioned above show some onlap relationships and local facies changes towards the paleohighs (Muller and Atchison, 1970; England, 1990). Significantly, however, the main formations of the Muller and Jeletzky and Ward classifications are mappable up to paleohighs both in the Comox and Nanaimo outcrop areas, taking into account the facies changes near the paleohighs and basin-wide lateral variations in compositions. Finally, detailed mapping (1:20 000 scale) of the Nanaimo Group in the Nanoose area demonstrates that equivalent formations and members can be readily recognized across the Nanoose "arch" with only minor geographic separation, and that there was no substantial barrier in this area during late Santonian-early Campanian time (Cathyl-Bickford and Hoffman, 1991; Cathyl-Bickford, pers. comm., 1993). This is also suggested by provenance evidence from detrital zircons collected in the Nanaimo area, which are of ages incompatible with a Wrangellia terrane source in formations previously thought derived from "Nanoose Arch" Wrangellia basement (Mustard et al., in press). Thus the distinction of two separate depositional basins is not justified and the use of the same formation names in all the major outcrop areas correctly follows the recommended procedures of the North American Stratigraphic Code for abandonment of obsolete names (Article 20). In recognition of the past confusion of nomenclature, it is recommended that the use of the terms Nanaimo Basin and Comox Basin be abandoned, although the historic precedence of the terms Nanaimo and Comox suggest the terms Nanaimo outcrop area and Comox outcrop area be used as shown in Figure 2 in this report along with those for other major outcrop areas.

References

References are listed either in the report provided or as part of the separate listing of graduate student theses compiled in Appendix C.

Appendix C

Bibliography of Georgia Basin student theses

This appendix provides a complete listing of student theses directly concerned with the sedimentary and volcanic rocks of the Georgia Basin. Copies of all theses (many including original photographs and maps) are available for viewing and photocopying at the research library of the Vancouver office of the Geological Survey of Canada. The following compilation is subdivided by age, with theses concerning the upper Cretaceous Nanaimo Group separate from these concerned with the Tertiary rocks of Georgia Basin.

Nanaimo Group theses

Allmaras, J.M.
1979: Stratigraphy and sedimentology of the Late Cretaceous Nanaimo Group, Denman Island, British Columbia; M.Sc. thesis, Oregon State University, Corvallis, Oregon, 178 p.

Atchison, M.E.
1968: Stratigraphy and depositional environments of the Comox Formation (upper Cretaceous), Vancouver Island, British Columbia; M.Sc. thesis, Northwestern University, De Kalb, Illinois, 139 p.

Breitsprecher, C.H.
1962: Correlation of foraminifera and megafossils from the Upper Cretaceous Sucia Island, Washington; M.Sc. thesis, University of Washington, Seattle, Washington, 83 p.

Carter, J.M.
1977: The stratigraphy, structure and sedimentology of the Cretaceous Nanaimo Group, Galiano Island, British Columbia; M.Sc. thesis, Oregon State University, Corvallis, Oregon, 202 p.

England, T.D.J.
1990: Late Cretaceous to Paleogene evolution of the Georgia Basin, southwestern British Columbia: Ph.D. thesis, Memorial University of Newfoundland, St. John's, Newfoundland, 481 p.

Fahlstrom, B.E.
1982: Stratigraphy and depositional history of the Cretaceous Nanaimo Group of the Chemainus area, British Columbia; M.Sc. thesis, Oregon State University, Corvallis, Oregon, 115 p.

Fiske, D.A.
1977: Stratigraphy, sedimentology, and structure of the Late Cretaceous Nanaimo Group, Hornby Island, British Columbia, Canada; M.Sc. thesis, Oregon State University, Corvallis, Oregon, 164 p.

Grieve, D.A.
1974: Stratigraphy of the Upper Cretaceous Nanaimo Group sediments of Prevost Island, British Columbia, with special emphasis on sandstone petrology; B.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 53 p.

Hanson, W.B.
1976: Stratigraphy and sedimentology of the Cretaceous Nanaimo Group, Saltspring Island, British Columbia; Ph.D. thesis, Oregon State University, Corvallis, Oregon, 339 p.

Hoen, E.W.B.
1958: The geology of Hornby Island; M.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 114 p.

Holm, V.A.
1968: Continuous seismic profiles over Burrard Inlet, British Columbia; B.A.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 23 p.

Hudson, J.P.
1974: Stratigraphy and paleoenvironments of the Cretaceous rocks, north and south Pender Islands, British Columbia; M.Sc. thesis, Oregon State University, Corvallis, Oregon, 139 p.

Janbaz, J.E.
1972: Petrology of the Upper Cretaceous strata of Sucia Island, San Juan County, Washington; M.Sc. thesis, University of Washington, Seattle, Washington, 104 p.

Johnson, S.Y.
1978: Sedimentology, petrology, and structure of Mesozoic strata in the northwestern San Juan Islands, Washington; M.Sc. thesis, University of Washington, Seattle, Washington, 105 p.

Kachelmeyer, J.M.
1978: Bedrock geology of the North Saanich-Cobble Hill area, B.C., Canada; M.Sc thesis, Oregon State University, Corvallis, Oregon, 153 p.

Langhus, B.G.
1968: Paleoecological and biostratigraphic zonation of Upper Cretaceous foraminifera, Vancouver Island, British Columbia; M.Sc. thesis, University of Calgary, Calgary, Alberta, 79 p.

Mainala, S.M.
1975: Petrology and structure of Beaver Point area southeastern Saltspring Island; B.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 49 p.

Meding, M.G.
1964: The geology of the Oyster River; B.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 29 p.

Mercier, J.M.
1977: Petrology of the Upper Cretaceous strata of Stuart Island, San Juan County, Washington; M.Sc. thesis, University of Washington, Seattle, Washington, 157 p.

Neufeldt, T.W.
1973: Description of the Brothers Creek Formation, Burrard Inlet, B.C.; B.Sc thesis, University of British Columbia, Vancouver, British Columbia, 28 p.

Oliver, E.M.
1979: Cretaceous faunal studies, Vancouver Island; M.Sc thesis, University of Calgary, Calgary, Alberta, 191 p.

Pacht, J.A.
1980: Sedimentology and petrology of the Late Cretaceous Nanaimo Group in the Nanaimo Basin, Washington and British Columbia: implications for Late Cretaceous tectonics; Ph.D. thesis, Ohio State University, Columbus, Ohio, 368 p.

Packard, J.A.
1972: Paleoenvironments of the Cretaceous rocks, Gabriola Island, British Columbia; M.Sc. thesis, Oregon State University, Corvallis, Oregon, 101 p.

Page, R.J.
1972: A preliminary petrographic examination of the Gabriola, Geoffrey, DeCourcy and Comox Formations, Nanaimo Group, Vancouver Island and Gulf Islands, B.C.; M.Sc. thesis, University of Washington, Seattle, Washington, 29 p.

Rahmani, R.A.
1968: Sedimentology and petrology of the Cedar District Formation, late Cretaceous, southwestern British Columbia; M.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 140 p.
Ronning, P.A.
1973: Petrography and provenance of the Nanaimo Group sandstones, Saltspring Island, British Columbia; B.Ap.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 39 p.
Ruddiman, W.
1980: The geology and stratigraphy of the lower Nanaimo Group, Nanaimo, British Columbia; M.Sc. thesis, Oregon State University, Corvallis, Oregon, 111 p.
Scott, J.A.B.
1974: Upper Cretaceous foraminifera of the Haslam, Qualicum, and Trent River Formations, Vancouver Island, British Columbia; M.Sc. thesis, University of Calgary, Calgary, Alberta, 167 p.
Simmons, M.L.
1973: The stratigraphy and paleoenvironment of Thetis, Kuper, and adjacent islands, British Columbia; M.Sc. thesis, Oregon State University, Corvallis, Oregon, 114 p.
Stickney, R.B.
1976: Sedimentology, stratigraphy, and structure of the late Cretaceous rocks of Mayne and Samuel Islands, British Columbia; M.Sc. thesis, Oregon State University, Corvallis, Oregon, 226 p.
Sturdavant, C.D.
1975: Sedimentary environments and structure of the Cretaceous rocks of Saturna and Tumbo Islands, British Columbia; M.Sc. thesis, Oregon State University, Corvallis, Oregon, 195 p.
Usher, J.L.
1949: The stratigraphy and paleontology of the Upper Cretaceous rocks of Vancouver Island, British Columbia; PhD thesis, McGill University, Montreal, Quebec, 196 p.
Ward, P.D.
1973: Stratigraphy of Upper Cretaceous rocks on Orcas, Waldron, and Sucia Islands; M.Sc. thesis, University of Washington, Seattle, Washington.
1976: Stratigraphy, paleoecology and functional morphology of heteromorph ammonites of the Upper Cretaceous Nanaimo Group, British Columbia and Washington; Ph.D. thesis, McMaster University, Hamilton, Ontario, 189 p.
Williams, T.B.
1924: The Comox coal basin; Ph.D. thesis, University of Wisconsin, Madison, Wisconsin, 143 p.
Winsby, J.A.
1973: Geology of the Upper Cretaceous Nanaimo Group, Saltspring Island, British Columbia; B.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 34 p.

Tertiary Georgia Basin student theses

Blanchet, P.H.
1943: The Tertiary basalt at Sentinel Hill; B.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 37 p.
Blunden, R.H.
1971: Vancouver's downtown (coal) peninsula urban geology; B.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 45 p.
Cruver, S.K.
1981: The geology and mineralogy of bentonites and associated rocks of the Chuckanut Formation, Mt. Higgins area, North Cascades, Washington; M.Sc. thesis, Western Washington University, Bellingham, Washington, 105 p.
Frizzel, V.A.
1979: Petrology and stratigraphy of Paleocene nonmarine sandstones, Cascade Range, Washington; Ph.D. thesis, Stanford University, Palo Alto, California, 151 p.
Hartwell, J.N.
1979: A paleocurrent analysis of a portion of the Chuckanut Depositional basin near Bellingham, Washington; M.Sc. thesis, Western Washington University, Bellingham, Washington, 85 p.
Holm, V.A.
1968: Continuous seismic profiles over Burrard Inlet, British Columbia; B.A.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 23 p.
Hopkins, W.S.
1966: Palynology of Tertiary rocks of the Whatcom basin, southwestern British Columbia and northwestern Washington; Ph.D. thesis, University of British Columbia, Vancouver, British Columbia, 184 p.
Horton, D.G.
1978: Clay mineralogy and origin of the Huntingdon fire clays on Canadian Sumas Mountain, southwest British Columbia; M.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 96 p.
Johnson, S.Y.
1982: Stratigraphy, sedimentology, and tectonic setting of the Eocene Chuckanut Formation, Northwest Washington; Ph.D. thesis, University of Washington, Seattle, Washington, 222 p.
Kelly, J.M.
1970: Mineralogy and petrography of the basal Chuckanut Formation in the vicinity of Lake Samish, Washington; M.Sc. thesis, Western Washington University, Bellingham, Washington, 63 p.
Kerr, S.A.
1942: The Tertiary sediments of Sumas Mountain; M.A. thesis, University of British Columbia, Vancouver, British Columbia, 48 p.
Pongsapich, W.
1970: A petrographic reconnaissance of the Swauk, Chuckanut and Roslyn Formations, Washington; M.Sc. thesis, University of Washington, Seattle, Washington, 63 p.
Reiswig, K.N.
1982: Palynological differences between the Chuckanut and Huntingdon Formations, northwestern Washington; M.Sc. thesis, Western Washington University, Bellingham, Washington, 61 p.
Robertson, C.A.
1981: Petrology, sedimentology, and structure of the Chuckanut Formation, Coal Mountain, Skagit County, Washington; M.Sc. thesis, University of Washington, Seattle, Washington, 41 p.
Thomson, R.V.
1958: Petrographic and sedimentary comparison of sandstones of the Kitsilano and Burrard Formation; B.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 21 p.
Wootton, A.E.
1959: The Tertiary eruptives at Vancouver, British Columbia, British Columbia; B.Sc. thesis, University of British Columbia, Vancouver, British Columbia, 36 p.

Stratigraphy and evolution of Tertiary Georgia Basin and subjacent Upper Cretaceous sedimentary rocks, southwestern British Columbia and northwestern Washington State

Peter S. Mustard[1] and Glenn E. Rouse[2]

Mustard, P.S. and Rouse, G.E., 1994: Stratigraphy and evolution of Tertiary Georgia Basin and subjacent Upper Cretaceous sedimentary rocks, southwestern British Columbia and northwestern Washington State; in Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia, (ed.) J.W.H. Monger; Geological Survey of Canada, Bulletin 481, p. 97-169.

Abstract: Georgia Basin encompasses Strait of Georgia, eastern Vancouver Island, Fraser Lowland, and northwest Washington State. At Vancouver, a succession of Late Cretaceous clastic rocks less than 500 m thick unconformably overlie Coast Belt intrusions or Albian-Cenomanian sedimentary rock. The Late Cretaceous rocks correlate with the lower Nanaimo Group and were deposited in alluvial fan and fluvial-floodplain environments. Tertiary Georgia Basin is dominated by Paleocene-Eocene clastic rocks 2.5 km thick in the Fraser Lowland, and thicken to nearly 6 km at Bellingham. Paleogene strata previously named Kitsilano and Burrard formations are included with the better exposed and coeval Huntingdon Formation, a northern continuation of the Chuckanut Formation, not a younger unit as previously interpreted. Deposition occurred in a terrestrial basin with proximal lower alluvial fan deposits changing to fluvial systems towards the basin axis. Paleocurrents and provenance suggest derivation from local sources. The Miocene Boundary Bay formation is less than 1.2 km thick and preserved in the Fraser delta subsurface. It comprises fluvial sandstone and mudstone similar to the underlying Paleogene rocks, although rare marine microfossils also suggest marginal marine environments. Oligocene rocks are limited to scattered exposures of igneous intrusions and rare subsurface dykes and sills. Tertiary Georgia Basin formed as an intracontinental strike-slip basin when dextral strike-slip faults were active in the western Cordillera. Late Eocene compression deformed the basin into northwest-trending and plunging folds. Traps and porous sandstone are present, but low thermal maturity and lack of marine kerogen in the sub-Miocene succession indicate gas as the most likely potential hydrocarbon.

Résumé : Le bassin de Georgia englobe le détroit de Georgia, l'est de l'île de Vancouver, les basses terres du Fraser et le nord-ouest de l'État de Washington. À Vancouver, une succession de roches clastiques du Crétacé tardif de moins de 500 m d'épaisseur recouvre en discordance des intrusions survenues dans la chaîne Côtière ou des roches sédimentaires d'âge albien-cénomanien. Les roches du Crétacé tardif sont mises en corrélation avec la partie inférieure du Groupe de Nanaimo et se sont déposées dans des milieux de cône alluvial et de cours d'eau-plaine d'inondation. Le bassin de Georgia (Tertiaire) contient principalement des roches clastiques paléocènes-éocènes qui ont 2,5 km d'épaisseur dans les basses terres du Fraser et qui s'épaississent jusqu'à presque 6 km à Bellingham. Les strates paléogènes antérieurement appelées formations de Kitsilano et de Burrard sont incluses dans la Formation de Huntingdon, qui est contemporaine et mieux exposée et qui constitue le prolongement vers le nord de la Formation de

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Chuckanut, et non une unité plus récente comme le voulait une interprétation antérieure. Les sédiments se sont accumulés dans un bassin terrestre, les dépôts inférieurs proximaux de cône alluvial faisant place à des dépôts fluviatiles vers l'axe du bassin. À en juger par les paléocourants et la provenance des matériaux, les sédiments ont vraisemblablement une origine locale. La formation miocène de Boundary Bay a moins de 1,2 km d'épaisseur et se rencontre dans la subsurface du delta du Fraser. Elle comprend du grès et du mudstone fluviatiles semblables aux roches paléogènes sous-jacentes, bien que la présence de rares microfossiles marins indiquerait aussi des milieux marins marginaux. Les seules roches oligocènes se rencontrent dans des affleurements dispersés d'intrusions ignées et de rares dykes et filons-couches de subsurface. Le bassin de Georgia (Tertiaire) s'est constitué sous forme d'un bassin de cisaillement intracontinental à un moment où il y avait mouvement le long de décrochements dextres dans la Cordillère occidentale. Une compression survenue à l'Éocène tardif a déformé le bassin, formant des plis à direction et à plongement nord-ouest. Il existe actuellement des pièges stratigraphiques et des grès poreux, mais la faible maturité thermique et l'absence de kérogène d'origine marine dans la succession sub-miocène indiquent que le gaz naturel serait en toute probabilité l'hydrocarbure présent.

INTRODUCTION

The Georgia Basin is a northwest-oriented structural and topographic depression encompassing Strait of Georgia, eastern Vancouver Island, the Fraser River lowlands of southwest British Columbia, and the northwest mainland of Washington State. Sedimentary rocks of the Georgia Basin comprise two main tectonostratigraphic packages: the Upper Cretaceous Nanaimo Group, well-exposed on eastern Vancouver Island and the Gulf Islands of Strait of Georgia, and a Tertiary basin mostly preserved in the Vancouver area and northwest Washington State (Fig. 1, 2). The Tertiary part of Georgia Basin, termed Bellingham or Whatcom Basin by some authors (e.g., Miller and Misch, 1963; Hopkins, 1968) has recently been the target for renewed hydrocarbon exploration, with three unsuccessful wells drilled in 1991 and two in 1993 (Fig. 3, Table 1). As part of the Geological Survey of Canada Georgia Basin project, the stratigraphy and extent of the Tertiary part of Georgia Basin have been re-examined. Field studies of Tertiary and underlying Late Cretaceous strata of the greater Vancouver area and of Tertiary outliers on the present margins of the Tertiary part of Georgia Basin have been combined with continuing palynological studies of these rocks and samples from recent exploration drilling in the basin. This paper summarizes these results, suggests revisions to the Late Cretaceous and Tertiary stratigraphic nomenclature of the Greater Vancouver area, linking the Canadian stratigraphy to the Tertiary Chuckanut Formation of northwest Washington State, and provides a summary of models for the evolution of the Tertiary part of Georgia Basin.

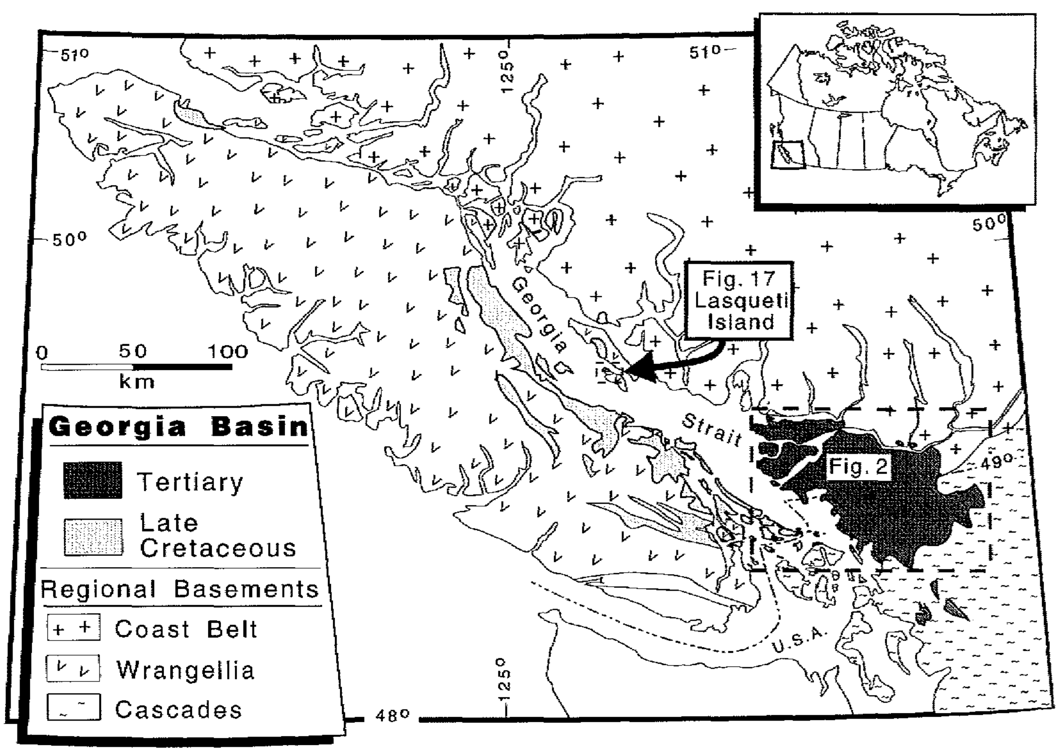


Figure 1. Regional setting of Georgia Basin with Tertiary Georgia Basin shown in dark grey.

Table 1. Major hydrocarbon exploration wells in Tertiary Georgia Basin. Well locations are shown in Figure 3.

Locality	Well	Year	T.D. (m)	elev (m)
1	Spartan No. 1	1918	610	40 (topo)
2	Spartan No. 2	1922	876	66 (topo)
3	Allenbee South Brazeau No. 1	1950	1433	90 (topo)
4	Boundary Bay No. 3	1921	1253	<3 (topo)
5	Conoco Dynamic Mud Bay	1991	1700	5 (KB)
6	Great Basin	1955	1842	63 (KB)
7	Richfield Pure Point Roberts	1962	4509	59 (KB)
8	Richfield Pure Sunnyside	1962	3321	114 (KB)
9	Royal Can-Van Tor Big Horn Kuhn No. 1	1955	1586	29 (KB)
10	Standard Ferndale Community	1945	1899	30 (topo)
11	AHEL Birch Bay	1988	2781	27 (KB)
12	AHEL Terrell No. 1	1991	1832	79 (KB)
13	AHEL Ferndale No. 1	1991	1348	33 (KB)
14	Pelican Dome No. 1	1941	1676	183 (topo)
15	El Paso Ross No. 1	1962	1435	214 (KB)
16	Squalicum Lake No. 1	1967	1689	167 (KB)
17	Hillebrecht No. 1	1948	1064	20 (topo)
18	Can-Am Stremler No. 1 (Lyndon No. 1)	1962	2245	40 (KB)
19	Kris Whatcom No. 1	1955	1740	38 (topo)
20	Hercon-Key Evans No. 1	1956	2396	42 (KB)
21	Richfield Pure Abbotsford	1961	956	70 (KB)
22	Fraser Valley Chilliwack	1959	1885	110 (KB)
23	Conoco Dynamic Stateside Campbell River	1993	2432	77 (topo)
24	Conoco Dynamic Murray Creek	1993	2635	75 (topo)

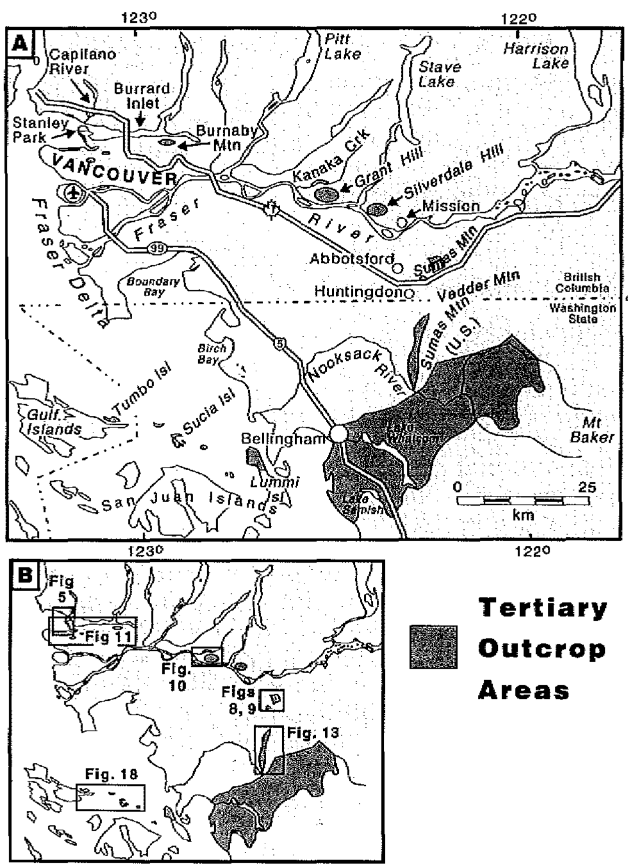


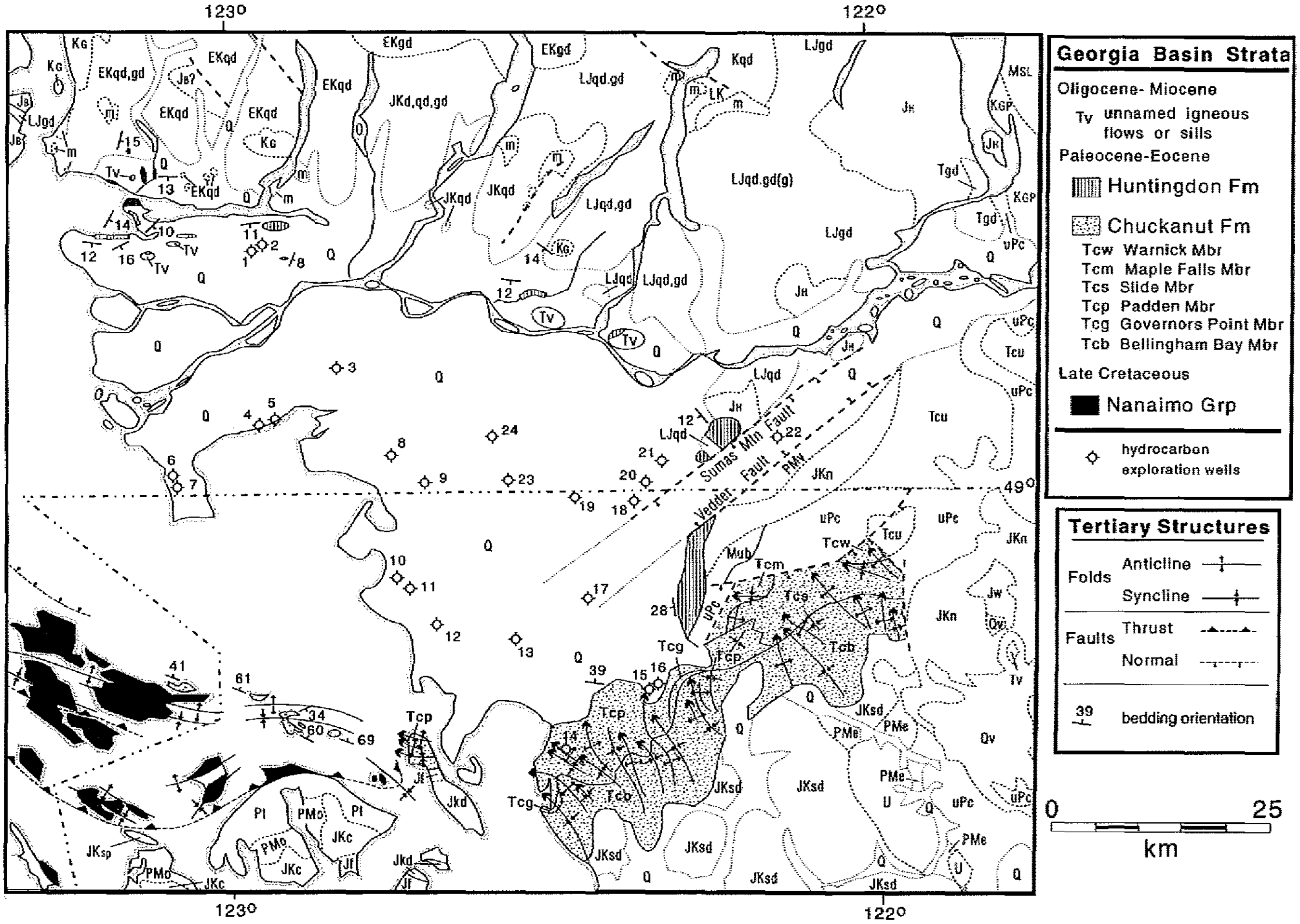
Figure 2. Major Tertiary outcrop areas of Georgia Basin (dark grey, excluding Lasqueti Island). Place names of towns (white dots), and other geographic features mentioned in text are also shown.

REGIONAL SETTING

As shown in Figure 3 and documented in this study, the Tertiary rocks of Georgia Basin unconformably overlie sedimentary rocks of the Late Cretaceous Nanaimo Group (Mustard, 1994) at the present west and northwest outcrop margins, at Vancouver, and in the subsurface beneath the Fraser Delta. In the eastern Fraser Valley the Tertiary rocks unconformably overlie Jurassic to Early Cretaceous Coast Belt granitic intrusives and remnants of associated volcanic/volcaniclastic cover sequences (Monger, 1990; Woodsworth and Monger, 1991). In northwest Washington State, the Tertiary strata unconformably overlie the northwest Cascades, a complex series of small and large accreted terranes dominated by oceanic sedimentary and metasedimentary rocks (mostly argillaceous to cherty successions), but with significant volcanic, ophiolitic, and intrusive slices (see Tabor et al., 1989 and McGroder, 1991 for recent reviews).

The main structural control on the sub-Georgia Basin rocks and to some extent Georgia Basin itself is southwest- to west-vergent thrusting that took place from the Cretaceous to the Eocene, a response to underthrusting of the Farallon or Kula oceanic plates beneath the North American plate (Yorath et al., 1985; Monger, 1991a, b). A mid- to Late Cretaceous, west-vergent thrust system is preserved at the southeast margin of the Georgia Basin (Brandon et al., 1988; McGroder, 1991) and in the eastern Coast Belt (Journeay et al., 1992; Journeay and Friedman, 1993) and includes thrust systems active during both basin formation and major periods of Nanaimo Group sedimentation. Dextral strike-slip faults influenced depositional patterns during the Tertiary stage of Georgia Basin fill (Johnson, 1984b, c; this study). The basin was also affected by early Tertiary compression, resulting in southwest directed thrusting that included the Nanaimo Group (England and Calon, 1991) and northwest-plunging and-trending folds in the Tertiary Chuckanut Formation (Johnson, 1982).

The extent of the original Tertiary basin can be estimated from the stratigraphic architecture, facies relationships, and provenance summarized in this paper and also discussed in Mustard and Rouse (1991, 1992). Source proximal alluvial fan and fluvial conglomerates are present in the northwest (Lasqueti Island), west (Tumbo and Sucia islands), east (U.S. and Canada Sumas Mountains), and southeast outcrop areas (much of the Chuckanut Formation). All pre-Miocene strata are terrestrial and both paleocurrents and clast types indicate derivation from their respective outer margins, suggesting the original basin margins were nearby and the basin was probably not significantly more extensive than implied by current outcrop patterns (with the exception that England and Calon, 1991 suggest about 20% shortening in a southwest direction in the underlying Nanaimo Group due to Eocene compression). There is no clear control on the south margin of the basin. Precise estimates of original basin extent are hampered on the east side by extensive late Neogene Coast Belt uplift (Parrish, 1983) which has caused erosion of most early Tertiary strata east or northeast of the present outcrop areas.



Legend for non-Georgia Basin rocks:

Coast Belt stratified rocks: **PMv** – Vedder Complex: amphibolitic gneiss; **TCU** – Cultus Formation: argillite and siltstone; **JB** – Bowen Island Group: mafic to felsic volcanics; **JH** – Harrison Lake Formation: intermediate to felsic volcanics; **JKm** – argillite-matrix mélange; **MSL** – Slollicum Schist: metavolcanics, metaseds; **KG** – Gambier Group: volcanic and sedimentary rocks; **KGP** – Peninsula Formation, conglomerate and sandstone; **Q** – Quaternary deposits; **m** – metamorphic rocks, uncertain protolith affinity.

Coast Belt plutonics: General composition: **qd** – quartz diorite; **gd** – granodiorite; **d** – diorite; **g** – granite. Age: **LJ** – Late Jurassic; **EK** – Early Cretaceous; **K** – Cretaceous; **LK** – Late Cretaceous; **JK** – Jura-Cretaceous (uncertain); **T** – Tertiary.

Northwest Cascades-San Juan Islands: **JKsp** – Speiden Formation; **Pt** – Paleozoic Turtleback terrane; **PMo** – late Paleozoic-early Mesozoic Orcas-Deadman Bay terrane; **JKc** – Constitution Formation; **Jf, Jkd** – Fidalgo Complex and Lummi Group of Decatur terrane; **Ksd, JKs** – Darrington Phyllite and Shuksan blueschist-greenschist of Shuksan metamorphic suite; **U** – ultramafic rocks of the Twin Sisters body; **PMe** – upper Paleozoic-lower Mesozoic Elbow Lake Formation; **uPc** – upper Paleozoic Chilliwack Group; **JKn** – upper Jurassic-Lower Cretaceous Nooksack Group; **Jw** – Jurassic Wells Creek Volcanics; **Mub** – Mesozoic ultramafic rocks; **Tv, Qv** – Tertiary and Quaternary volcanic rocks; **Q** – Quaternary deposits. Major sources for regional geology are Monger (1986, 1993), Brandon et al. (1988); McGroder (1991); and Monger and Journeay (1992).

Figure 3. Simplified geological map of southwest British Columbia and northwest Washington State showing the main Tertiary stratigraphic units and the major regional geological components of the northwest Cascades and southern Coast Belt. Selected historic hydrocarbon well sites are also located (see Table 1 for well information).

STRATIGRAPHY OF TERTIARY GEORGIA BASIN

With the exception of isolated occurrences of Paleocene rocks on Lasqueti, Tumbo, Sucia, and nearby islands (Rouse et al., 1990; Mustard and Rouse, 1991, 1992; this study), the Tertiary rocks of the Georgia Basin are mainly exposed in the lower Fraser Valley along the north and northeast margins of the present Fraser Delta, and in mainland northwestern Washington State (Fig. 2, 3). The main stratigraphic components of this area (summarized in Fig. 4) are: Paleocene-Eocene rocks of the Vancouver area, formerly termed the Kitsilano and upper Burrard formations, but here renamed Huntingdon Formation in recognition of their correlation to the better exposed Paleocene-Eocene Huntingdon Formation of Canadian Sumas Mountain (formerly believed to be upper Eocene to Oligocene age); the Paleocene-Eocene Chuckanut Formation of Washington State; minor Oligocene igneous rocks; and Miocene sedimentary rocks known from subsurface drilling.

Previous studies and stratigraphic terms

Early investigations of the Vancouver area geology are reviewed by Johnston (1923) and Rouse et al. (1975). A report of coal on the south shore of Burrard Inlet (Wood, 1859, pers. comm. to Captain G.H. Richards, H.M. Surveying Ship "Plumper" reproduced in Blunden, 1971) provided the first mention of the sedimentary rocks of the Vancouver area. Subsequent investigations included brief descriptions of the sedimentary rocks of the area, but no stratigraphic subdivision and early studies suggested a Cretaceous, and later studies a Tertiary age for the entire succession (e.g., Bauerman, 1885; Bowman, 1888; LeRoy, 1906; Bowen, 1913; Camsell, 1918). Daly (1912) defined the first formation, suggesting an Eocene age and the name Huntingdon Formation for the sandstones and conglomerates of Canadian Sumas Mountain (Fig. 2).

Johnston (1923) conducted the first comprehensive regional geological study of the Vancouver area. He subdivided the sedimentary rocks of Vancouver into a basal Burrard Formation and overlying Kitsilano Formation. He

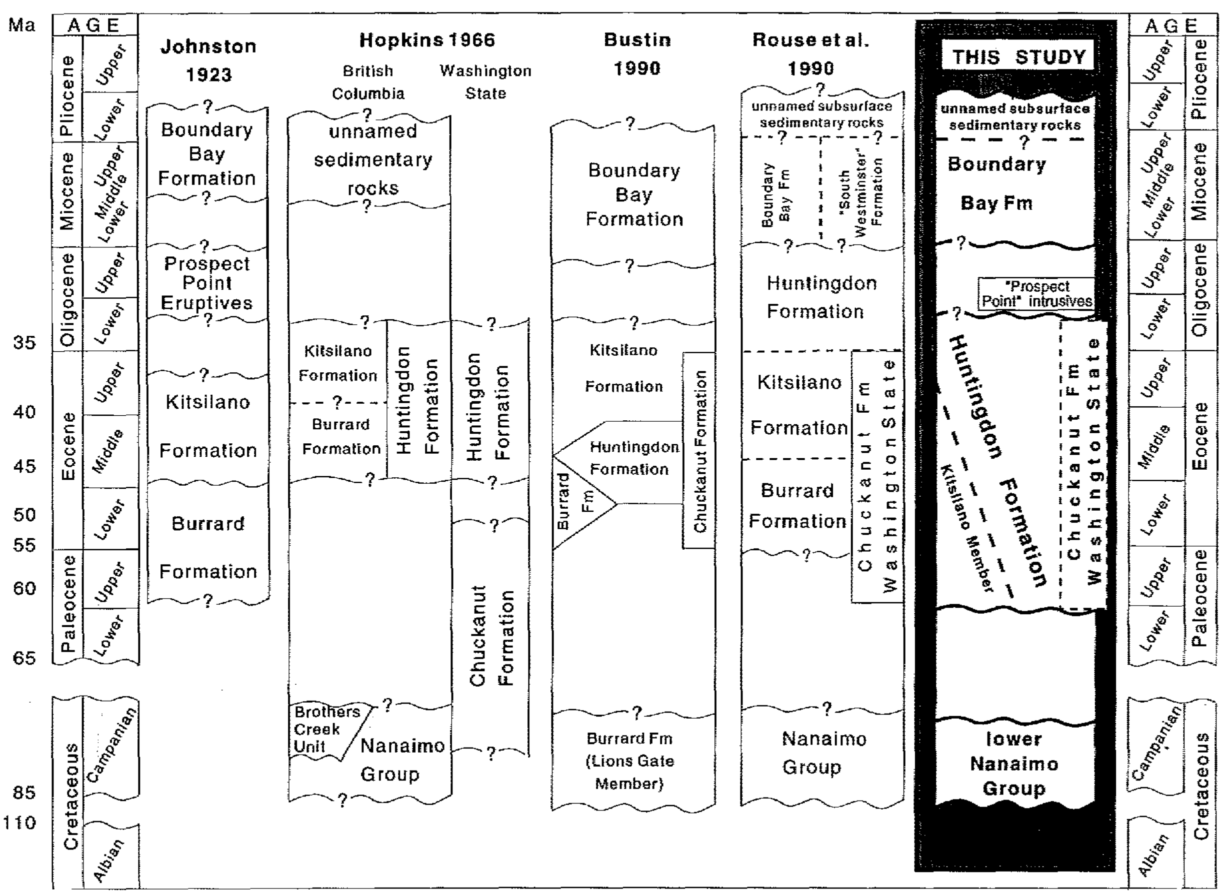


Figure 4. Major formation nomenclatures proposed for the Tertiary strata of Georgia Basin. Members proposed for several of the formations by various workers are discussed in the text. Epoch or Stage boundary ages are from the time scale of Harland et al. (1990).

suggested both formations were Eocene age, and described strata from a hydrocarbon exploration drillhole (wellsite 4 on Fig. 3), which he suggested were Pliocene to Miocene in age and named the Boundary Bay Formation. Subsequent studies of the Burrard Formation determined that the lower 600 m of the succession is Late Cretaceous in age (Rouse, 1962; Crickmay and Pocock, 1963), leading Rouse et al. (1975) to propose redefinition of this Upper Cretaceous succession as the Lions Gate Member of the Burrard Formation (discussed below). Hopkins (1968) confirmed the presence of a thick succession of Miocene strata in the subsurface of the Fraser Delta area.

Paleogene sedimentary rocks exposed in the Bellingham area and to the east in northwest Washington State are classified as the Chuckanut Formation, with the exception of a small outlier of sedimentary rocks exposed on a ridge named Sumas Mountain (termed "U.S. Sumas Mountain" in this study to distinguish from the "Canadian Sumas Mountain" of the Abbotsford area) and correlated with the Huntingdon Formation by Miller and Misch (1963). The Chuckanut Formation comprises up to 6000 m of nonmarine conglomerate, sandstone, mudstone, and minor coal (Fig. 3; see Fig. 16 below). The Chuckanut Formation was defined by McLellan (1927), Glover (1935), and Weaver (1937). Important early studies include Moen (1962) and Miller and Misch (1963). Johnson (1982, 1984a, b) defined seven members in the Chuckanut Formation; he later revised this to six members (Johnson, 1991; Fig. 3).

REVISED STRATIGRAPHY

Detailed examination and sampling of the major outcrops and drill cores has provided new data on the extent, age, and sedimentology of the Georgia Basin successions. Because of the dispersed nature of the main outcrop areas, separate descriptions and interpretations are presented below. A more detailed discussion of the provenance and sedimentology of the basin fill is provided in a separate section which follows the outcrop area description.

Upper Cretaceous lower Nanaimo Group (undivided) of the Vancouver area (formerly Lions Gate Member, Burrard Formation)

The basal sedimentary rocks of the Vancouver area are exposed in several localities at the margins and north of Burrard Inlet (Fig. 3, 5A-C). At the northern margin (best exposed in the Capilano River canyon immediately north of the Highway 1 bridge; shown on Fig. 6A-B) these rocks unconformably overlie Early Cretaceous quartz diorite intrusions of the western Coast Belt. The sedimentary rocks were originally defined as the lower part of the Burrard Formation and an Eocene age was attributed to them (Johnston, 1923). However, the lower 600 m of the succession (Fig. 5A-C, 6A-B) was recognized by Rouse (1962) and Crickmay and Pocock (1963) to be Late Cretaceous. Named the Brothers Creek Formation in a thesis by Neufeldt (1973), Rouse et al. (1975) redefined this Upper Cretaceous succession as the Lions Gate Member of the Burrard Formation. At Stanley Park this unit is disconformably overlain by a poorly exposed succession of similar looking, but less well-indurated, fluvial sandstone, mudstone, and conglomerate containing Late Paleocene and Early Eocene palynomorphs (the upper Burrard Formation of Johnston, 1923 and Rouse et al., 1975). A once-exposed conglomerate bed which may have marked the contact between upper and lower Burrard Formations is now covered or eroded (described in Rouse et al., 1975).

Lithofacies and sedimentology

The detailed measured section from Capilano Canyon is typical of the basal 50-70 m of the Upper Cretaceous strata (Fig. 6B, right side, this corresponds to unit 1 of Rouse et al., 1975). Less complete exposures of this interval occur on Brothers Creek immediately south of Highway 1 and as an isolated outlier on McDonald Creek at about 2500 feet elevation (Fig. 5A). At Capilano Canyon fractured and altered quartz diorite is unconformably overlain by brown-weathering, poorly sorted cobble-pebble conglomerate with 70-80% clasts in a medium- to coarse-grained arkosic wacke matrix. Clast-supported and clast-rich matrix-supported conglomerates are present. Clasts are subangular to subround, generally subspherical and randomly oriented, but rarely display poor tabular clast imbrication. Clast compositions are >50% diorite or granite, and the reminder mafic volcanic or intrusive, all types typical of the local Coast Belt basement. Bed contacts are indistinct. Conglomerate intervals several metres to more than 10 m thick contain poor normal or reverse grading in some parts, indicating amalgamation of thinner (1-3 m thick?) beds. Rare interbeds of coarse grained arkosic arenite up to 1 m thick are normally graded and capped by silty and slightly carbonaceous mudstone. These interbeds are cut out laterally over a few tens of metres by overlying conglomerate beds. The conglomerate-rich succession changes upward to >15 m of stacked, overlapping interbeds of pebble-cobble conglomerate and coarse- to medium-grained arkosic arenite. Conglomerate is generally clast supported, having crude normal grading and imbrication of tabular clasts, and is erosive into underlying sandstones in a geometry of overlapping irregular based sheets. Sandstones are crudely bedded to wavy and discontinuous and contain rare silicified logs up to 20 cm diameter and 1 m length. Drillhole data and intermittent exposures in Brothers Creek show that this basal conglomerate-rich interval is overlain by more than 50 m of brown-weathering, medium grey, arkosic arenite with mudstone interbeds increasingly common upwards (unit 2 of Rouse et al., 1975). The sandstone is medium- to coarse-grained, thin bedded, and occurs in discontinuous overlapping sheets with curved bases (some pebble-rich). Poorly exposed trough crossbeds are rarely present.

The covered interval between Brothers Creek and Prospect Point at Stanley Park can be characterized from extensive engineering test drillhole information from Burrard Inlet and the Capilano River delta. Originally described in Rouse et al. (1975) this interval is about 350 m thick, although minor faults may be present, and comprises a lower section about 150 m thick of interbedded sandstone and sandy mudstone which is probably a continuation of the gradational fining

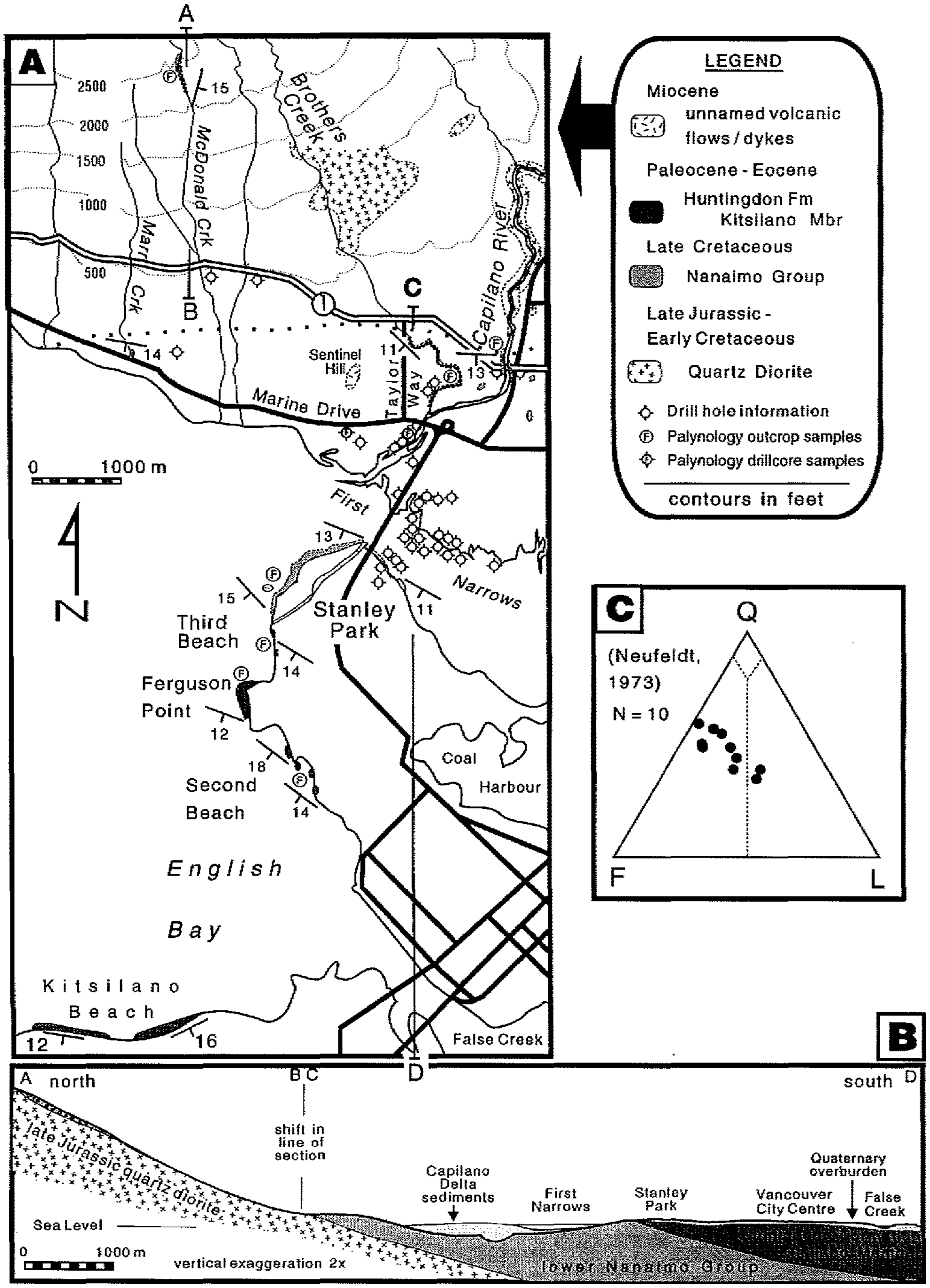


Figure 5. A) Main area of outcrop of Upper Cretaceous strata underlying Paleocene-Eocene rocks in west and north Vancouver (expanded and updated from Fig. 2 of Rouse et al., 1975). Drillhole information additional to Rouse et al. (1975) is from Belanger and Harrison (1976, re-released in Mustard and Roddick, 1992). Main diorite outcrops in upper Brothers Creek are now mostly obscured by subdivision development (these contacts from Rouse et al., 1975, all others re-mapped by the senior author). B) Simplified cross-section A-B-C-D, modified and extended from the section of Rouse et al. (1975). Line of section is located on Figure 5A. C) Ternary plot summary of sandstone detrital clast compositions (replotted from data of Neufeldt, 1973, sandstone classification boundaries of Dott, 1964).

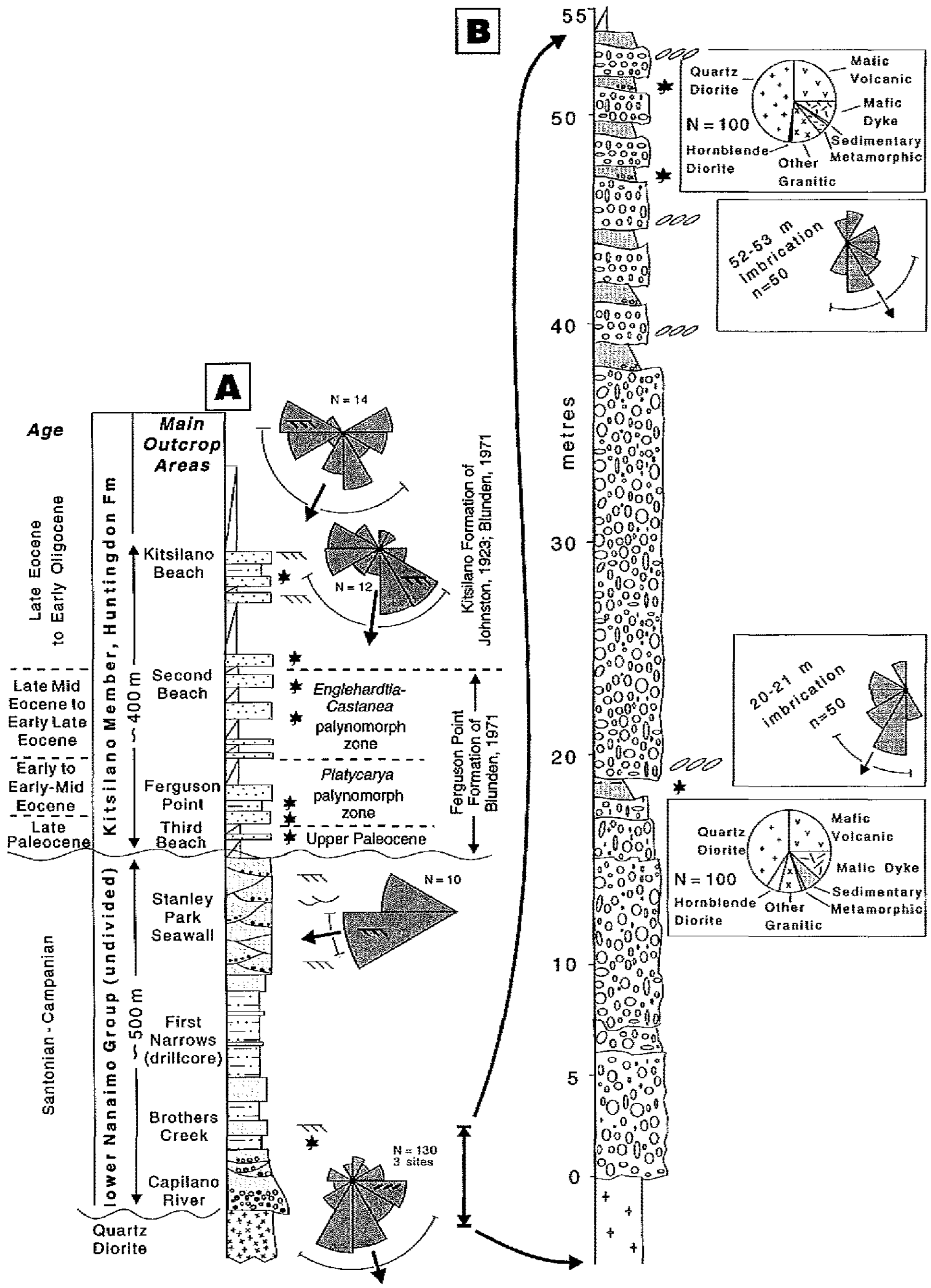


*Figure 6. A) General stratigraphic section through Upper Cretaceous Nanaimo Group strata (formerly termed Lions Gate Member of the Burrard Formation) and Paleocene-Eocene lower Huntingdon Formation (formerly termed upper Burrard and lower Kitsilano formations). Revised from the drillcore and outcrop based section in Figure 2 of Rouse et al. (1975) with added information from outcrop studies of the senior author and studies of Blunden (1971) and Holm (1968). **B)** Detailed columnar section illustrating basal Cretaceous rocks exposed in Capilano River at Highway 1 crossing (located on Fig. 5A).*

upward succession exposed in Brothers Creek. This section gradationally changes upward to about 150 m of mudstone, siltstone, and minor sandstone ("sandy shales" in the original engineering reports, unit 3 of Rouse et al., 1975). This mudstone-rich interval is overlain (either abruptly or with a thin gradational contact) by about 40 m of medium- to coarse-grained sandstone which is the basal part of the exposed sandstone unit at Stanley Park (unit 4 of Rouse et al., 1975). The Late Cretaceous sandstone which makes up the cliffs in Stanley Park north of Third Beach are about 200 m thick and comprise coarse- to medium-grained arkosic arenite organized into overlapping trough or channel forms each 10 m to more than 30 m in length and up to 5 m thick. Complex overlapping and erosion of lower by higher channels has obscured the original sequence, but a composite complete cycle typically has a pebble rich, curved base which changes upward to trough crossbedded or wavy planer bedded sandstone, capped by planer crossbedded or plane bedded sandstone and rarely by silty mudstone. Mudstone ripups are common in the sandstones, especially at channel bases. Coalified plant material (including large branches) are rare to common. No marine micro- or macrofossils have been recovered. Paleocurrents measured from planar crossbeds are consistently southwest directed (Fig. 6A-B).

The lithofacies described above suggest alluvial fan debris flows and lower-fan braided stream and sheetflow deposition for the lower conglomerate-rich succession; braidplain and perhaps floodplain fluvial deposition for the sandstone and mudstone of Brothers Creek and most of the Burrard Inlet covered interval (possibly including lacustrine deposition for the mudstone dominate part of this interval), and coarse-load sand-dominant meandering river deposition for the upper part of the succession. The locally derived clasts, lack of sorting or organized bedforms in the lower conglomerates, submature and coarse, arkosic nature of the sandstones, and generally south to southwestern paleocurrents all indicate provenance from nearby uplifted Coast Belt sources.

Subsurface Cretaceous strata

In addition to the exposed "Lions Gate Member", several exploration wells in the Fraser delta area have intersected Upper Cretaceous strata. Hopkins (1966) identified probable Late Cretaceous palynomorphs from the lower 1300 m of the Richfield Point Roberts well and the lower 600 m of the Richfield Sunnyside well (wells 7 and 8 on Fig. 3, also shown on Fig. 15, below). This was confirmed for the Sunnyside well by more recent palynological studies (Mustard and Rouse, 1991) which also identified Cenomanian and Albian sedimentary rocks in the lower 20 m of the well. A recent exploration well in northwest Washington State (AHEL Birch Bay, well 11 in Fig. 3; also shown on Fig. 15, below) intersected about 1000 m of strata interpreted as Late Cretaceous (to possibly Paleocene) at the base of the 2781 m deep well (Hurst, 1991). Resampling of the Birch Bay well cuttings and continued work on the Sunnyside well samples for this study has allowed a more precise definition of the Cretaceous strata in the subsurface of these areas. In addition, re-examination of the well cutting from the Point Roberts well for this study showed that this well drilled completely through the Cretaceous-Tertiary section (not reported in original drill logs), intersecting quartz diorite in the lower 30 m of the hole, and providing the only complete penetration of the Georgia Basin succession in this part of the basin.

Palynology

The Upper Cretaceous rocks flanking English Bay and from wells in Boundary and Birch bays have yielded a large and well preserved palynoassemblage. Some of the more characteristic spores and pollen are illustrated in Plate 1. Others have been reported earlier by Rouse (1962, 1977), Crickmay and Pocock (1963), Hopkins (1966), Rouse et al. (1971), and Mustard and Rouse (1991). The overall assemblage is dominated by fern spores (Pl. 1, fig. 1-6, 9, 10), and the angiosperm pollen *Proteaciditics thalmanni* (Pl. 1, fig. 11, 12) and *Tricolpites divergens* (Pl. 1, fig. 14-16), together with pollen representing the tree genera *Fraxinus* (ash) and *Quercus* (oak). The other pollen represent herbaceous genera that comprised most of the ground cover. Taken together, the assemblage represents most closely a fen-type of vegetation colonizing relatively flat lowland areas between estuarine or deltaic depositional sites. Based on the affiliation of the spores and pollen with modern taxa, together with the leaves identified by Bell (1957) from the Nanaimo Group on Vancouver Island, the paleoclimate is interpreted as very warm temperate to subtropical, probably similar to the present-day climate of the Gulf Coast of North America.

The English Bay palynoassemblage has been equated most closely to palynomorphs of the Protection Formation of Vancouver Island, described as mid-Campanian by Muller and Jeletzky (1970). However, some of the key palynomorphs such as *Proteacidites thalmanni*, *Cupaneidites* spp., *Cicatricosisporites* spp., and *Quercoidites* spp. also occur in other Nanaimo Group formations (e.g., Extension, Comox, and Spray). Although many of these have been reported from other rock units in the western region of North America, e.g., the Maastrichtian-Danian of California (Drugg, 1967), and the Maastrichtian of Wyoming (Farabee and Canright, 1986), the assignment of the English Bay assemblages to a Santonian-Campanian range would seem to be most reasonable until additional detailed analyses are undertaken.

Correlation to Nanaimo Group of Vancouver Island

The relationship of the Upper Cretaceous stratigraphy on the east side of the Georgia Basin to the Nanaimo Group is not well defined. The scant paleocurrent data from the lower Burrard Formation suggests southwesterly and westerly paleoflow directions towards the main area of Nanaimo Group sedimentation (Fig. 6A, B; also Johnston, 1923; Rouse et al., 1975; Mustard, 1994). The subsurface strata tentatively identified as Cretaceous appear to thicken to the west and are continuous across the basin (Machacek, 1971; White and Clowes, 1984; Gordy, 1988). The palynoassemblages are most similar to those from the Campanian Protection Formation on Vancouver Island, although several key types occur in other Nanaimo Group formations. A lithostratigraphic comparison to Vancouver Island Nanaimo Group rocks suggests a slightly different correlation, with the poorly sorted and locally derived basal conglomerates of Vancouver strongly resembling the lower part of the Comox Formation, the basal unit of the Nanaimo Group elsewhere in the basin (generally Santonian to lower Campanian in age). In this lower unit, the fining upward trend into interbedded sandstone and mudstone, and subsequent coarsening up into pebbly sandstone at the top of the Vancouver succession is similar to the Comox-Haslam-Extension formation successions on parts of Vancouver Island.

We suggest that this poorly exposed succession is best termed lower Nanaimo Group and does not warrant formal formation status. Previous practise referring to this unit as the lower Burrard Formation (Lions Gate Member) promotes confusion with Paleocene-Eocene age rocks previously assigned to the upper Burrard Formation, which we suggest below should be amalgamated with rocks previously assigned to the Kitsilano Formation and termed the Kitsilano Member of the Huntingdon Formation. In addition, the scattered and generally poorly exposed nature of the outcrops in the Burrard Inlet-English Bay area prohibit definition of a type section or even meaningful reference sections, a prerequisite of formal formation definition as used in the North American Stratigraphic Code (1983). For these reasons, we suggest the terms Burrard Formation, and thus Lions Gate Member, should be abandoned.

Paleocene-Eocene Huntingdon Formation (redefined)

The Huntingdon Formation was originally described by Daly (1912), based on exposures at Canadian Sumas Mountain (Fig. 2, 3, 7A-C), and named for the town of Huntingdon (Fig. 2). The age of the Huntingdon Formation based on palynology was considered Middle to Late Eocene (Hopkins, 1966), or Late Eocene to Early Oligocene (Reiswig, 1982; Rouse et al., 1990). Detailed examination of new exposures at Sumas Mountain has provided evidence that the Huntingdon Formation includes Late Paleocene to Early Eocene strata at the base and thus is coeval with strata previously termed upper Burrard and Kitsilano formations at Vancouver (documented below). The open pits and cliffs at Canadian Sumas Mountain provide vertical and lateral exposure of this succession vastly superior to exposures elsewhere in Fraser Valley or Greater Vancouver area. For these reasons, we here formally redefine Huntingdon Formation to encompass the entire Late Paleocene and Eocene stratigraphy of the lower mainland, superseding the terms upper Burrard Formation and Kitsilano Formation. Kitsilano Formation we suggest be redefined as the Kitsilano Member to denote the Paleocene-Eocene succession in the Vancouver-Burnaby area. Hopkins (1966) also concluded that upper Burrard and Kitsilano formations were part of a continuous and indistinguishable succession and suggested (informally) that a common name should be used for these strata and that the Huntingdon Formation is of the same age as the upper Burrard and Kitsilano formations. The original description of the Huntingdon Formation by Daly (1912) did not include a type section. Figure 8 illustrates a proposed composite type section and reference sections (located on Fig. 7A), all from Canadian Sumas Mountain.

Huntingdon Formation: Canadian Sumas Mountain

The Huntingdon Formation at Canadian Sumas Mountain, British Columbia (Fig. 2) is a Late Paleocene and Eocene to Early Oligocene succession of terrestrial clastic rocks greater than 425 m thick (top not exposed). Strata unconformably overlie intensely weathered and altered Jurassic volcanic rocks of the Harrison Lake Formation (Fig. 7A). The weathered material (kaolinitic saprolite) on the unconformity is incorporated into mudstone of the lower Huntingdon Formation and includes high quality refractory clays (fireclays) which have been mined since the 1920s (Kerr, 1942; Cummings and McCammon, 1952; Horton, 1978).

Lithofacies and sedimentology

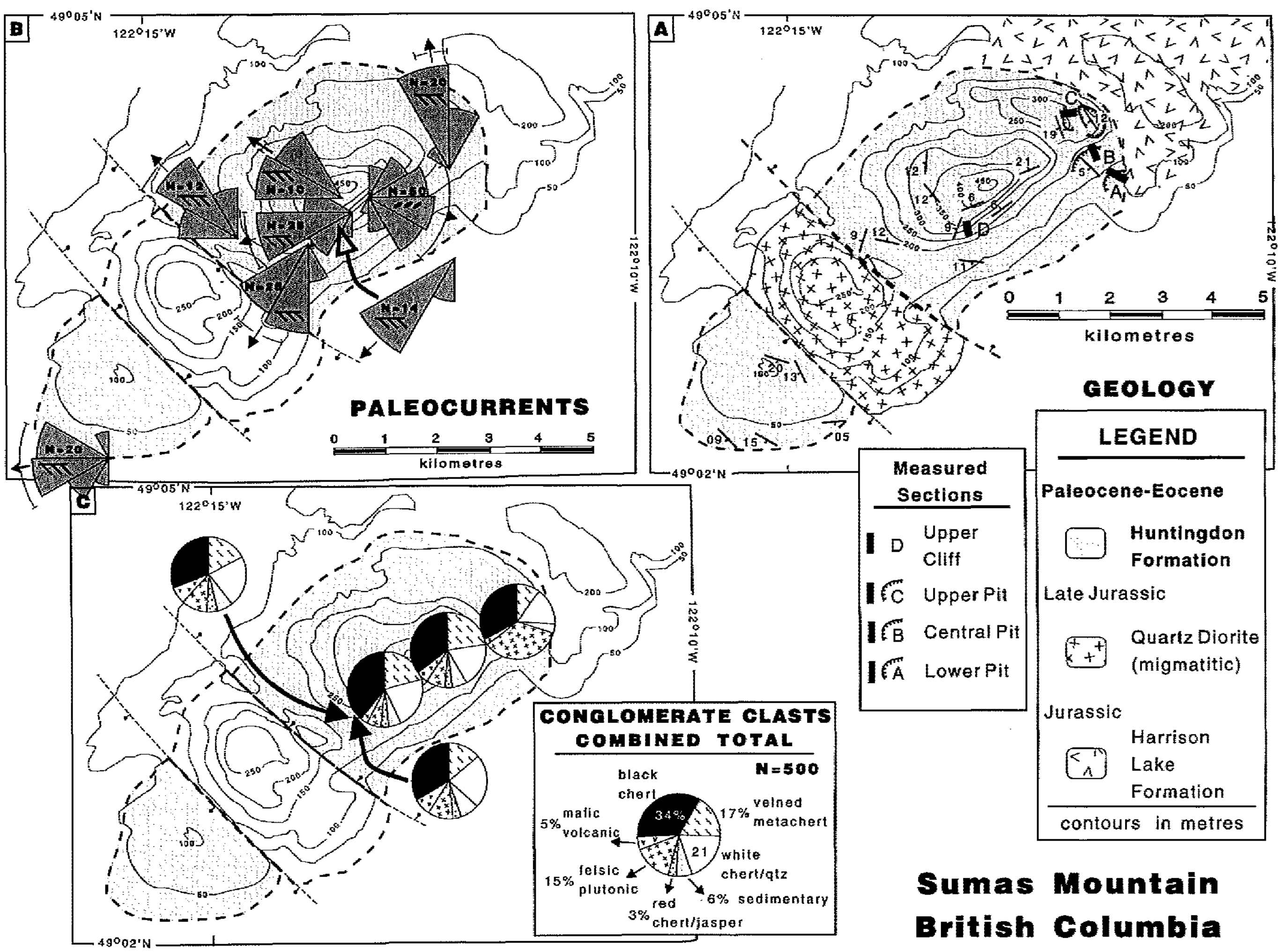
The succession coarsens upward (Fig. 8) from a mudstone- and fine grained sandstone-dominated lower section (Fig. 9A) of about 100 m thickness to an increasingly conglomerate-rich upper part which is roughly organized in repeated coarsening-upward cycles (50-100 m scale). Sandstone is fine- to very coarse-grained, chert-rich lithic arenite, and in the lower part of the formation occur in large channels several 100 m in width and up to 10 m thick (Fig. 9B, C). These are interpreted as major fluvial channels in a floodplain environment. Previous interpretations of depositional environments have focused on the origin of the fireclays, which Horton (1978) concluded were either preserved regolith or slightly transported and deposited as floodplain detrital muds in small ponds or lakes. The upper conglomerate-rich succession comprises abundant normal graded pebble-cobble conglomerate interbedded with coarse sandstone (Fig. 9D). These occur as complexly stacked and overlapping channelized beds at the tops of the coarsening-upward intervals (Fig. 9E). They are interpreted as lower alluvial fan facies and braided stream deposits which prograded over the floodplain facies. Crossbedding in sandstone and conglomerate clast imbrication indicates paleoflow ranged from southwest to north (Fig. 7B). Conglomerate clast compositions are dominated by dark grey to black chert (Fig. 7C), probably derived from the Vedder Complex or Cascade sources to the south or east, where such chert types are common.

Huntingdon Formation: Kanaka Creek

Johnston (1923) briefly described a succession of sandstone and mudstone at Kanaka Creek on the northeast side of the Fraser River west of Mission (Fig. 2, 10A-D) which he tentatively correlated with the Kitsilano Formation. A borehole described by Johnston from near the mouth of Kanaka Creek contained 380 m of interbedded medium- to coarse-grained sandstone, blue and grey shale, and minor conglomerate and lignite. The basal contact with nonsedimentary rocks was not reached in this well. The outcrop at Kanaka Creek is about 65 to 75 m thick and probably represents a lower part of the drillhole described by Johnston (1923). Projection of the lowest sedimentary beds northward towards exposed Coast Belt intrusions suggests the exposed sedimentary succession is within 100 m stratigraphically of the underlying granitic rocks.

Lithofacies and sedimentology

Figure 10A-B illustrates the main outcrop area, and provides a simple cross-section through the creek showing the dip of the succession, the location of palynology samples (Fig. 10C), and a summary paleocurrent rose (Fig. 10D). The Kanaka Creek exposures are dominated by coarse- to medium-grained lithic arenite organized into crudely defined to well defined, stacked fining-upward cycles which vary from 1 m to 5 m thick (rarely amalgamated into >20 m thick units as at Kanaka Falls). The basal sandstone of each fining-up cycle is thin- to medium-bedded and commonly occur in broad overlapping crossbedded sets, each less than 1 m thick. Upper sandstone in these cycles tend to be wavy to planar crossbedded and capped by laterally discontinuous dark grey mudstone up to 1 m thick. Cycle bases are curved to irregular and erode (tens of centimetres) into underlying mudstone or sandstone. Mudstone units are commonly cut out laterally



A) Geology and location of detailed measured sections (shown in Fig. 9).

B) Paleocurrent roses from sandstone crossbedding and conglomerate clast imbrication (all statistically significant and rotated to account for bedding tilt). See Figure 18D legend for explanation of rose construction. Centre of rose is located at measurement site where possible, some have been offset to avoid overlap of roses (large unfilled arrow indicates measurement site).

C) Conglomerate clast composition pie diagrams. Centre of pie is located at measurement site where possible. Closed arrows point to site for offset pies. Clast compositions are based on identification of 100 randomly picked clasts/site.

Figure 7. *Canadian Sumas Mountain, summary maps.*

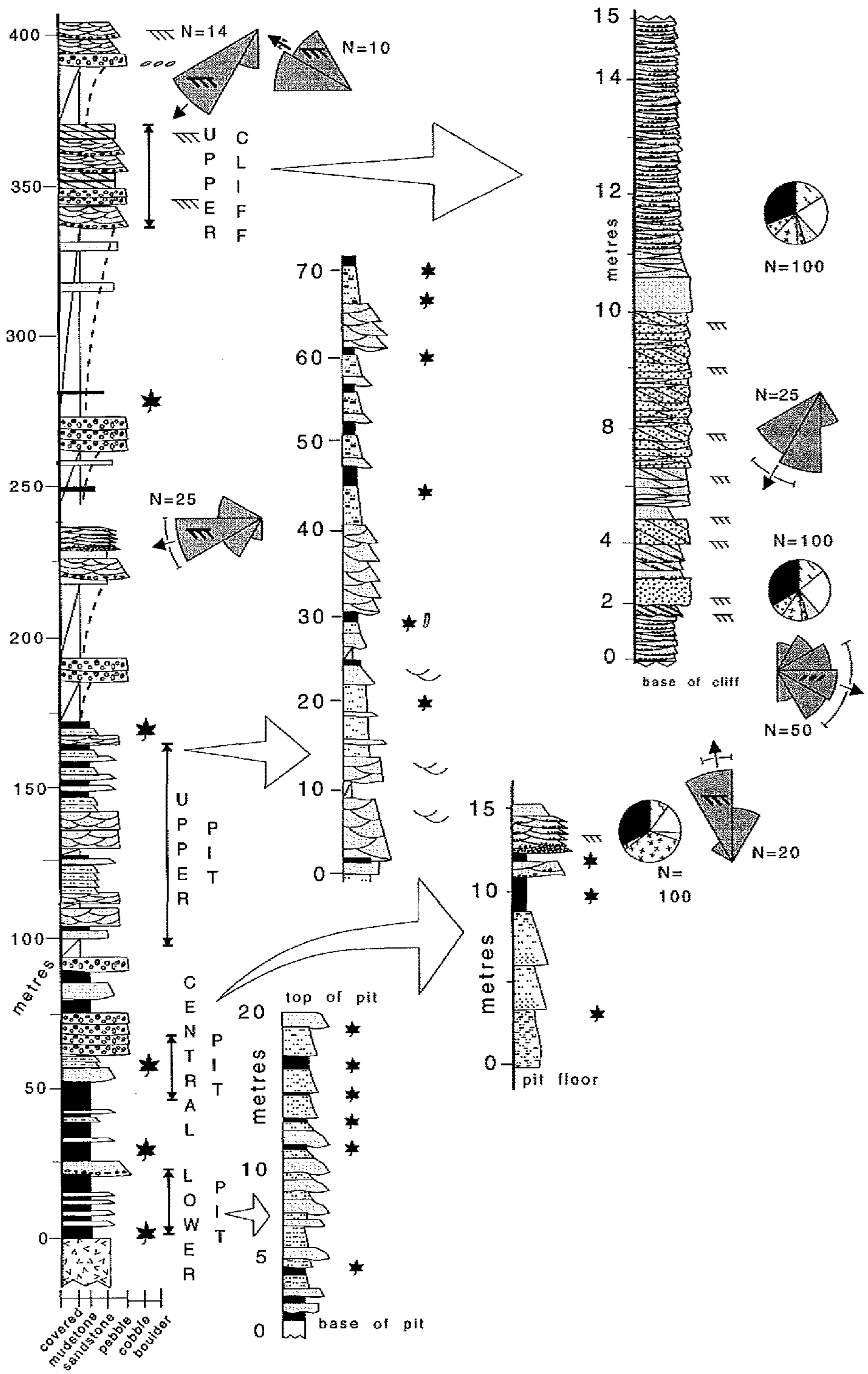


Figure 8. *Measured stratigraphic sections from Canadian Sumas Mountain. Main section at left is a composite drawn from the detailed sections at right (measured in mining pits and cliff exposures located on Fig. 8), rough section measurements by the senior author in creeks on the south side of Sumas Mountain, drillhole data from mine exploration summarized in Cummings and McCammon (1952) and section measurements by Kerr (1942). Conglomerate clast composition pies use same symbols as pies of Figure 7.*

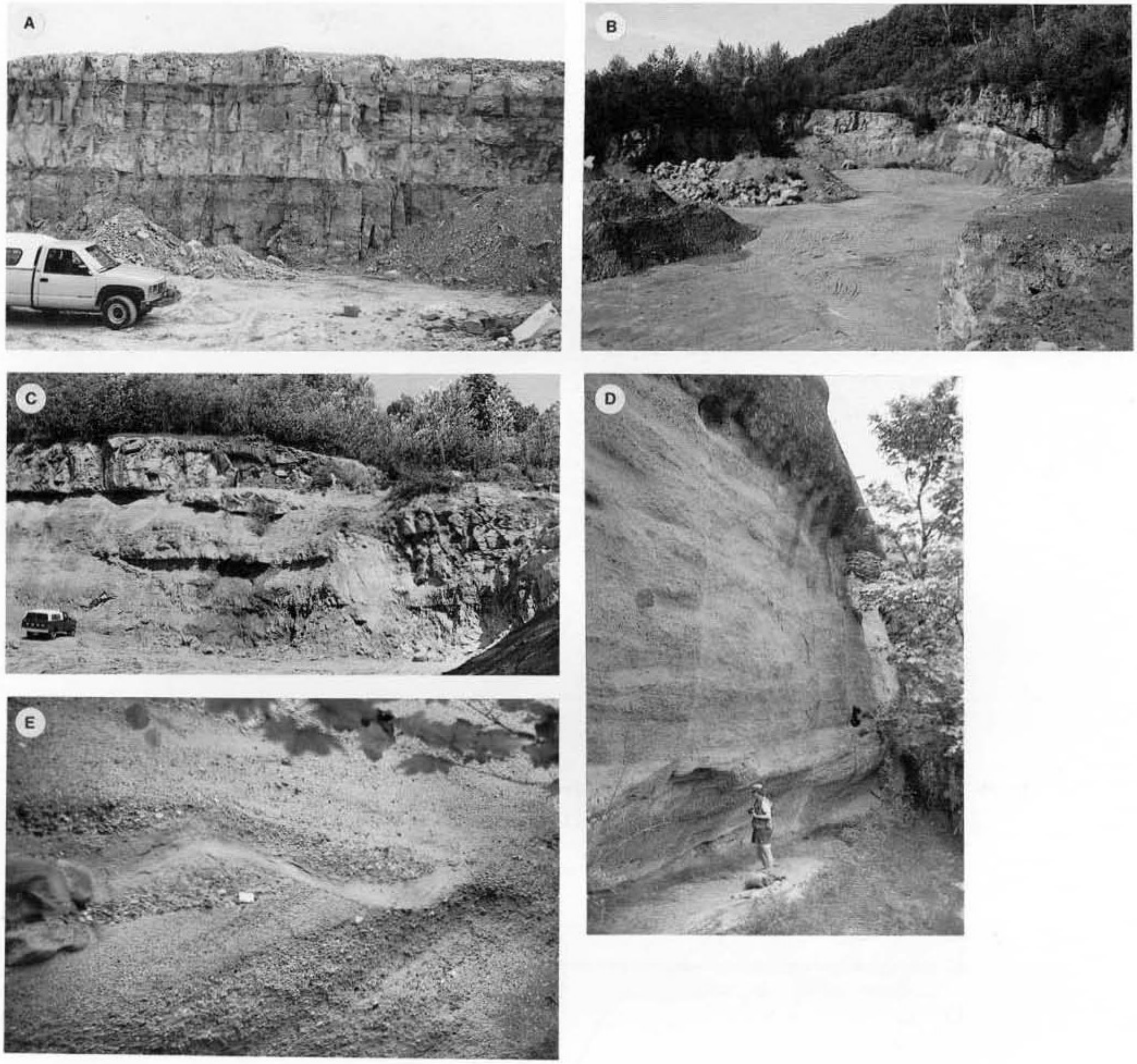
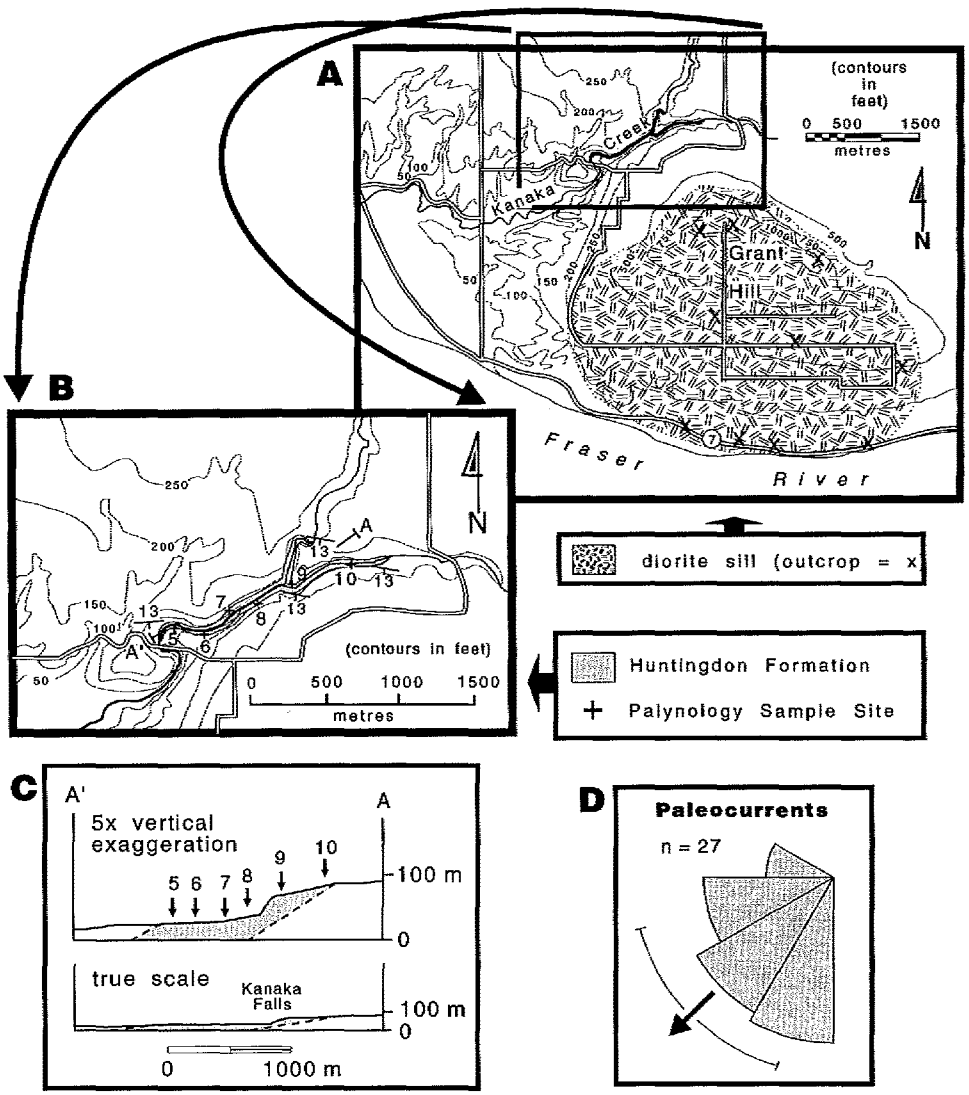


Figure 9. *Canadian Sumas Mountain outcrop photos.* ***A)*** *Pit wall exposure of lower Huntingdon Formation mudstone (darker units in photo are red-brown in outcrop) which cap fine grained sandstone and siltstone (light to medium grey). Units appear laterally continuous on scale of open pits (about 100 to 200 m) and are massive to faintly laminated and fine upward slightly. Sandstone-siltstone bases are planar to broadly convex up on underlying mudstone. Interpreted as floodplain and small lacustrine deposits. GSC 1994-712A* ***B)*** *Part of central pit at Sumas Mountain showing parts of overlapping pebbly sandstone channel forms (main wall on right). These major sandstone units have convex upward bases which are erosive on underlying fine grained sandstone and siltstone. Thickest exposed pit wall is about 35 m high. GSC 1994-712B* ***C)*** *Part of central pit displaying small and laterally discontinuous medium- to coarse-grained sandstone units (with thick more continuous pebbly sandstone at top of pit) contained within fine grained sandstone and planar laminated siltstone to silty mudstone. Interpreted as small fluvial channels and crevasse splay deposits in a floodplain environment. GSC 1994-712C* ***D)*** *One of several upper cliffs of Canadian Sumas Mountain. Cliffs consist of complexly overlapping and laterally discontinuous normal graded pebble conglomerate beds and very coarse grained sandstone (rarely cobble-rich at bases of beds). Beds are generally internally stratified with wavy planar bedding and small trough or planar crossbeds (commonly at tops of beds). GSC 1994-712D* ***E)*** *Upper cliff exposure showing complexly overlapping trough crossbedded and discontinuous pebble conglomerate and coarse grained sandstone. Interpreted as deposits of braided stream barforms. Scale card is 9 cm on longest edge. GSC 1994-712E*

with mudstone ripups abundant at the base of many sandstone crossbeds. Plant debris, including large branches and well-preserved leaves, are common in some mudstone and basal sandstone. In laterally extensive exposures the sandstone-mudstone cycles display a geometry of broad overlapping channels each several tens of metres wide. Paleocurrent indicators (most planar crossbeds) consistently demonstrate southwest paleoflow (Fig. 10D).

The fining-upward cycles dominated by coarse- to medium-grained sandstone, overlapping curved erosive bases, abundance of mudstone ripups and plant debris, and moderate sorting of these sandstone units is similar to both the lower part of the Canadian Sumas Mountain exposures and lower Kitsilano Member facies described below. These cycles are interpreted as sand-dominated meandering river deposits. The discontinuous interbeds of plant-rich mudstone represent areas of channel bank or interchannel deposition in floodplains and small lacustrine bodies of water.



A) Regional map (outlined on Fig. 2).

B) Detailed map of extent of exposed Huntingdon Formation in Kanaka Creek. Outcrop is nearly continuous in the creek bed and cliff sections in the shaded area.

C) Cross-section (located on B) showing general dip of Kanaka Creek exposures.

D) Paleocurrents, all from planar crossbeds, combined from the entire exposed interval (corrected for bed-tilt).

Figure 10. *Huntingdon Formation outcrop at Kanaka Creek.*

Huntingdon Formation, Kitsilano Member of the Vancouver area (formerly Upper Burrard and Kitsilano formations)

The Kitsilano Formation was defined by Johnston (1923) to include a thick package of conglomerate, sandstone, and minor mudstone which he suggested disconformably overlies the Burrard Formation in the Vancouver area, both dipping 10-15° to the south. As discussed above, the lower Burrard Formation contains Late Cretaceous palynomorphs and is correlated with the lower Nanaimo Group. Johnston (1923) suggested an unconformity is present on the south shore of Burrard Inlet (in outcrops no longer exposed) in which a succession (up to 300 m thick) of conglomerate unconformably overlies a sandstone-dominated unit. He interpreted the conglomerate as basal Kitsilano Formation and the underlying sandstone as upper Burrard Formation. Examination of temporary exposures of this conglomerate unit during road construction in this area (shown on the Fig. 11 map by an X south of Second Narrows bridge) and of the original photograph of the interpreted unconformity (Johnston, 1923, his Plate IIIB) suggests this contact is one of several erosive channel bases within the conglomerate-rich facies of this lower Tertiary unit (as in Fig. 12A), and does not represent a significant time gap or formation boundary, as confirmed by the new palynology information presented below (Hopkins, 1966 reached a similar conclusion based on the general aspect of the two successions and his palynology study). Thus we suggest that these strata be treated as a single entity and propose the term Kitsilano Member in recognition of the previously widely published term Kitsilano Formation. The latter we propose to replace with the regional name Huntingdon Formation, which also includes strata previously termed upper Burrard Formation.

Lithofacies and sedimentology

Hopkins (1966) measured 750 m of the Kitsilano Member south of Kitsilano beach during construction of a major sewer tunnel in this area. To this minimum thickness, about 300 m can be added from the Paleocene-Lower Eocene strata of Stanley Park, formerly termed upper Burrard Formation, to give a total minimum thickness of more than 1000 m for the Kitsilano Member, but with no top defined. The projected thickness calculated from the map pattern and assuming no major faults is about 1200 to 1300 m. This is slightly thinner than the 1500 to 2400 m thickness of Paleocene-Eocene strata present in the few subsurface wells on which detailed palynological studies have been completed (wells 5, 7, 9, 11, and 12 in Fig. 3; detailed in a separate section below), and significantly thinner than the estimated 6 km thickness of Paleocene-Eocene Chuckanut Formation in the Bellingham area as calculated by Johnson (1982, 1984a, b).

The main lithofacies of the Kitsilano Member are known from scattered outcrops in Stanley Park, Kitsilano Beach, Burnaby Mountain, Brunette Creek, the Grandview railway cut (Broadway and Victoria streets), and from a 1991 drill core near Capital Hill (Fig. 11). In addition, temporary exposures are common at construction sites in downtown Vancouver and areas immediately to the east and south and several were examined for this study. Johnston (1923) also summarized logs of drill core from two hydrocarbon exploration wells drilled about 1918-1920 southwest of Burnaby Mountain (wells 1 and 2 in Fig. 3). These wells intersected 610 m and 876 m of sedimentary strata (about 575 m and 850 m true thickness), all probably Kitsilano Member, but possibly including an unknown amount of Upper Cretaceous rocks. From the relative stratigraphic position of these outcrops and the associated wells the Kitsilano Member appears to consist of three units: a lower unit (about 250-300 m thick) of interbedded sandstone and minor mudstone, a middle unit (about 200-300 m thick) in which conglomerate is common, and an upper unit (>300 m thick) of sandstone and mudstone similar to the lower unit. However the conglomerate-rich facies is not laterally persistent and conglomerates are less common to the southwest and south of the Burnaby Mountain-Second Narrows areas.

A recent (1991) drill core from the Chevron Refinery north of Capital Hill is illustrated in Figure 11 and is typical of the sandstone-mudstone lower and upper units of the Kitsilano Member. Map relationships suggest the drill core section represents basal Kitsilano Member and the lower part of the hole could include Upper Cretaceous strata, although palynology samples from this core proved barren. The best outcrop exposures of this facies occur on the southwestern part of Stanley Park (e.g., Ferguson Point), at Kitsilano Beach, in the Grandview railway cut (Fig. 12B), and at Brunette River east of Burnaby Lake, the latter part of the upper unit of the Kitsilano Member (all located in the Fig. 11 map). The lower succession comprises repeated 1-5 m thick fining-upward cycles of medium- to coarse-grained lithic to arkosic arenite which grade upward to fine grained sandstone, siltstone, and dark grey mudstone, commonly carbonaceous (see Thomson, 1958 for detailed petrography of the sandstone). The bases of the fining-upward cycles commonly have erosive contacts, and pebbles and mudstone ripups are in many places present in the lower 10 cm of the sandstone beds. Sandstone is massive to planar crossbedded or wavy and trough-crossbedded in lower, coarser sections and wavy bedded or planer crossbedded in upper, finer sections. Plant debris and coaly lenses are common in some places. Mudstone and siltstone sections are generally less than 50 cm thick, rarely 5 m thick, and faintly laminated to massive where thin, but consist of 1-2 cm thick normal graded siltstone-mudstone rhythmites in thicker sections.

Conglomerate in the middle unit of the Kitsilano Member is rarely exposed. Figure 12A illustrates an unusually well-exposed example of the conglomerate-rich facies at the base of Burnaby Mountain (now covered). Conglomerate occurs as moderately sorted stacked discontinuous sheets or lenses which overlap laterally and have slightly curved bases erosive into underlying strata. The conglomerate comprises about 80-90% pebble and small cobbles in a coarse grained to granular arkosic matrix. Clasts are generally subrounded, but tabular clasts are present and a poor to fair clast imbrication can be defined in some places. Conglomerate beds are generally 0.5 to 1 m thick and grade upward into coarse pebbly sandstone (Fig. 12C) and in rare complete sections are capped

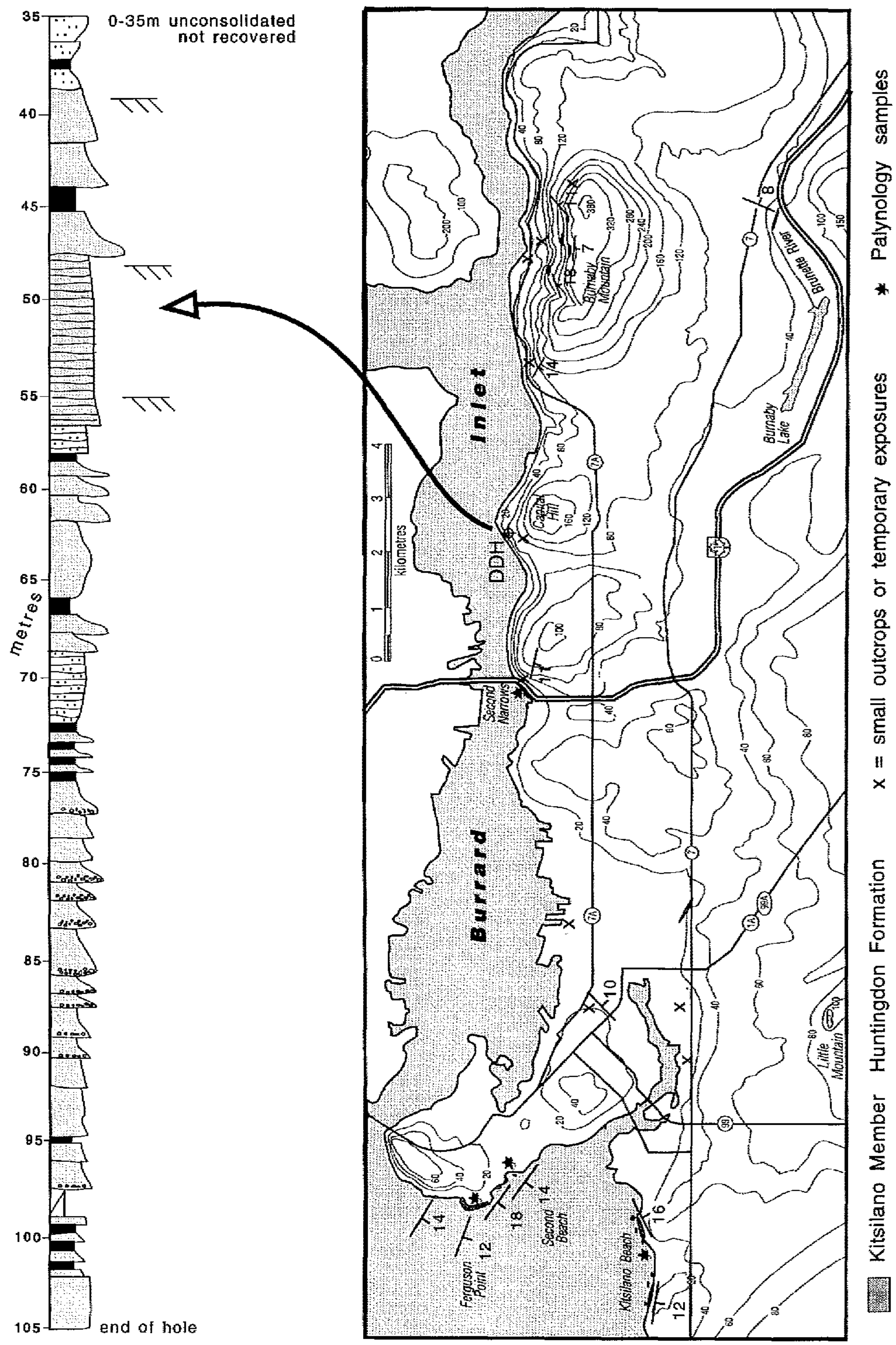
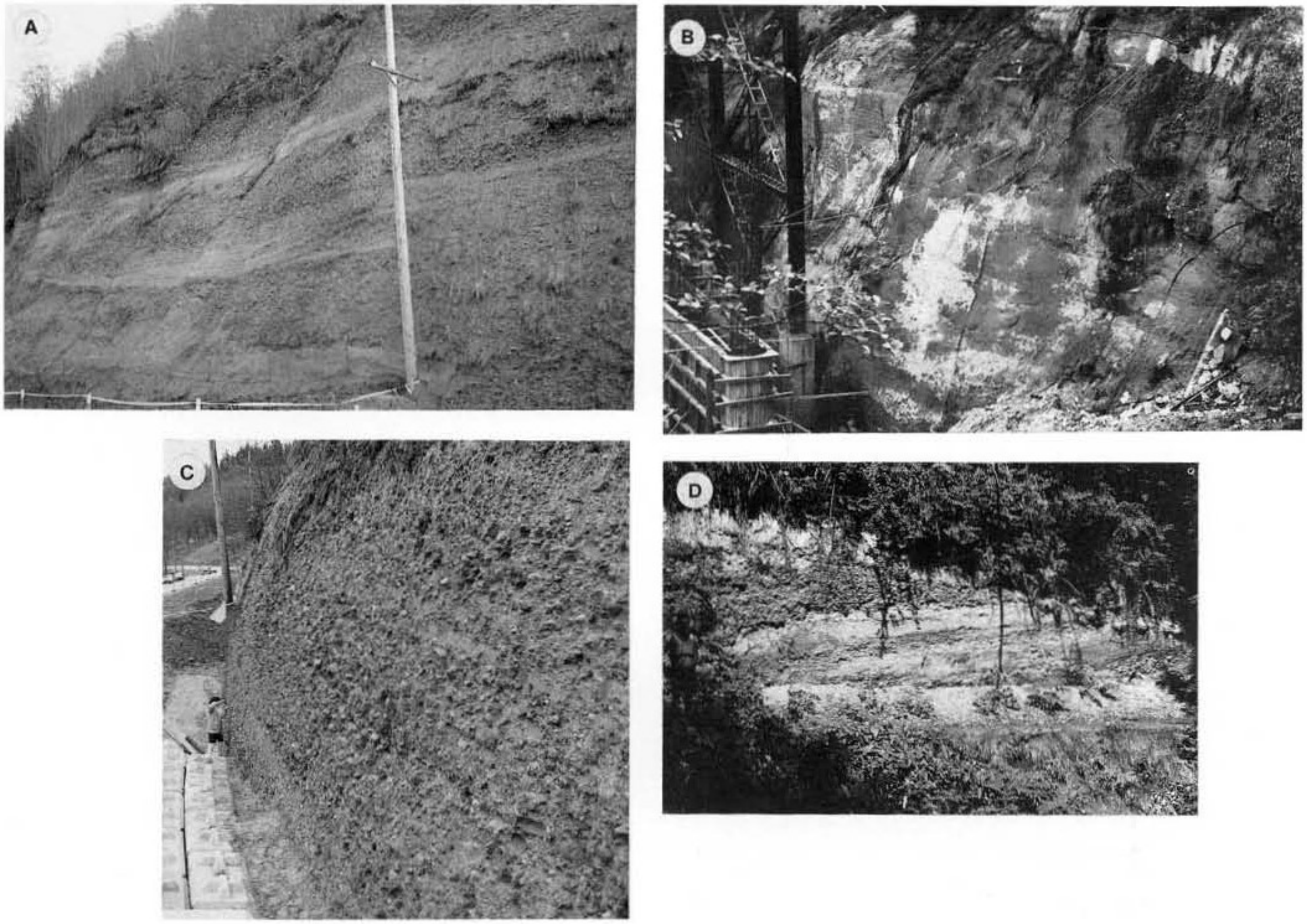


Figure 11. Main outcrop and temporary exposures of Kitsilano Member, Huntingdon Formation. Graphic log of 1991 diamond drill hole core from Chevron refinery site (located on Section). No palynomorphs were recovered from this core, but its position suggests it is representative of the lower Kitsilano Member in this area.



A) *Temporary vertical exposure (1993) on Barnett Highway at southwest end of Burnaby Mountain, illustrating typical conglomerate facies of central Kitsilano Member in this area. Pebble-cobble conglomerates grade in upper 10-20 cm to coarse- to medium-grained sandstone which are planar to slightly wavy bedded (rarely crossbedded) and rarely capped by carbonaceous silty mudstone. Shallowly left-dipping beds in centre of photo represent lateral migration of conglomerate channels and sand barforms with complex intertonguing of facies laterally. Total height of outcrop is about 30 m. GSC 1994-712F*

B) *North-facing cliff at Grandview railway cut (about 50 m vertical in 1992 photo, now partly covered) illustrating slightly irregular stacked sheets of coarse grained sandstone beds (light grey in photo) with discontinuous silty mudstone caps (medium grey in photo) which are cut out by slightly erosive and irregular base of overlying sandstone. GSC 1994-712G*

C) *Same outcrop area as in A, showing stacked and laterally overlapping bedded pebble-cobble conglomerates with rare discontinuous capping coarse grained sandstone. Bedding is defined by rare sandstone and by slight normal graded layers at tops of some conglomerates. Larger cobbles are also generally concentrated at the bases of beds. GSC 1994-712H*

D) *Outcrop on Joe's Trail, Burnaby Mountain showing lenticular pebble-cobble conglomerate which fines in upper part and intertongues laterally with coarse- to medium-grained sandstone. Outcrop is about 5 m maximum vertical exposure. GSC 1994-712I*

Figure 12. *Kitsilano Member outcrop photos.*

by a few tens of centimetres of brown to medium grey siltstone and silty mudstone, in total defining fining-upward cycles about 2-3 m thick (Fig. 12A). The mudstone and siltstone are typically carbonaceous with coaly lens and plant debris common. More commonly, mudstone caps are cut out laterally by overlying conglomerate and sections consist of complexly overlapping discontinuous beds of conglomerate and sandstone (Fig. 12D). Conglomerate clast imbrication and sandstone planar crossbedding measured from these units and from the Kitsilano Beach area define a generally southerly, but radial, paleoflow pattern with individual sites ranging from east to southwest directed (all shown on the Fig. 11 map).

The facies of the Kitsilano Member are similar to those described above from Canadian Sumas Mountain and Kanaka Creek and a similar alluvial interpretation is suggested. There is no indication of any marine component from either the lithofacies types or the recovered microfossil assemblages. The fining-upward cycles and mudstone interbeds of the lower Kitsilano Member are similar to those of the lower Huntingdon Formation at Canadian Sumas Mountain and a floodplain, meandering stream environment is also suggested. The well defined, coarsening upward sequences of Sumas Mountain are not apparent in the Kitsilano Member. However, the overall upward change to the central conglomerate-rich facies in the Kitsilano Member is interpreted as a general progradation of braided stream conglomerates over the floodplain fluvial succession. Exposed conglomerate of this facies is channelized and moderately sorted and interpreted as braided fluvial deposits. However, Johnston (1923) described and illustrated several poorly sorted and disorganized conglomerate beds from this facies (no longer exposed) which appear to be debris flow deposits, suggesting a lower alluvial fan setting for at least part of this facies. The southerly radial paleocurrent patterns measured in this member suggest a northern source area.

Palynology

Palynoassemblges from the Huntingdon Formation generally have been interpreted as Middle to Upper Eocene (Hopkins, 1966). Hopkins (1966) considered the Huntingdon Formation to be equivalent to the Kitsilano Formation. However, Reiswig (1982) concluded that the Huntingdon Formation possibly includes Oligocene palynoassemblages, and thus strata slightly younger than the Kitsilano Formation as originally defined. He also questioned the presence of an unconformity between the Huntingdon and Chuckanut formations, and suggested on the basis of lithological similarity and palynology that the Huntingdon Formation be considered the upper part of the Chuckanut Formation. Rouse et al. (1990, their Fig. 1) show the Huntingdon Formation as entirely Oligocene in age, separate from both the Kitsilano and Chuckanut formations, however new evidence from this study demands revision of this interpretation.

Outcrops at Third Beach on the Stanley Park seawall (Fig. 6A), those from the stratigraphically lowest outcrops on Canadian Sumas Mountain, and from Kanaka Creek (Fig. 10A) all contain correlative palynomorph assemblages. The palynoassemblage, illustrated in Plates 2-6, is dominated by conifer pollen and fungal spores, with fewer fern spores and angiosperm pollen. Significantly, there is a clear correlation of the main palynofacies with the principal lithofacies. The conifer pollen (Plate 2) and the fungal spores (Plates 4-6) predominate in the coarser sand-rich facies, wheras the fern-spores (Plate 2) and angiosperm pollen (Plate 3) are more common in the clay- and black mud-rich facies. This zonation is clearly shown at Third Beach, Stanley Park, where conifer pollen and fungal spores predominate in the lower sandier beds, whereas fern spores and angiosperm pollen are more common in the upper finer facies just north of Ferguson Point. At Kanaka Creek, the entire section is dominantly sandy and contains mainly conifer and fungal palynomorphs with few angiosperm pollen, suggesting correlation with the lower sandy facies at Third Beach.

The palynoassemblage from these strata correlates most closely with assemblages of the Upper Paleocene, particularly with assemblages reported by Krutzsch (1967) and Roche (1973) from western Europe, and by Leopold and MacGinitie (1972), Bihl (1973), Elsik (1974), Doerenkamp et al. (1976), Rouse (1977), Frederiksen (1979), and Mustard and Rouse (1991, 1992) in western North America. Particularly significant are the presence of *Pistillipollenites mcgregorii* (Pl. 3, fig. 15, 16) ranging from the late Paleocene to mid-Eocene (Rouse, 1977), and the so-called pre-Tilia pollen, viz. *Intratriporopollenites precrassipites* (Pl. 3, fig. 10, 11) and *I. prevescipites* (Pl. 3, fig. 12, 13). Other Late Paleocene indicators are *Triporopollenites mullensis* (Pl. 3, fig. 1, 2), *Quercoidites* sp. A (Pl. 3, fig. 8), *Tricolpites reticulatus* (Pl. 3, fig. 17), *Subtriporopollenites* sp. A. (Pl. 3, fig. 2), and *Tricolpites* sp. A (Pl. 3, fig. 22), all reported in the Late Paleocene zone P-4 by Rouse (1977). Similar Late Paleocene assemblages have been eported from several islands in Strait of Georgia and the northern San Juan Islands (Tumbo, Sucia, and Lasqueti islands, Mustard and Rouse, 1991, 1992).

There are also several fungal spores reported here for the first time that appear to be restricted to the Late Paleocene, viz. *Fusiformisporites paucistriatus* sp. nov. (Pl. 4, fig. 10, 11), *Multcellaesporites acuminatus* sp. nov. (Pl. 4, fig. 14, 15), and *M. bilobus* sp. nov. (Pl. 4, fig. 16).

The palynological setting suggests the presence of a floodplain through most of the Late Paleocene, with fluvial deposition in earlier periods and increasing marsh, pond, and/or lake deposition in later times. The paleoclimate appears to have been essentially wet and cool in earlier times, but warming and drying perceptibly in the latest Paleocene.

The unit lying immediately above the Late Paleocene interval at Third Beach consists of 120 m of arkosic, crossbedded sandstone which Blunden (1971) described and included in his "Point Ferguson Formation". We include this unit as part of the Kitsilano Member. In places, the sandstone contains thin, persistent fine silt and shale beds which yielded a small but well preserved and defined palynoassemblage.

The main palynomorphs include ferns, angiosperms, and fungal spores, the most characteristic of which are shown in Plate 7. The most definitive palynomorph is *Platycarya platycaryoides* (Pl. 7, fig. 4, 5), for which we name this the

Platycarya zone following Blunden (1971). This pollen is characterized by two creases or pseudocolpi; these occur on the two poles of the grain, and at different angles. The resulting configuration is often described as the "cross-swords" pattern, clearly observable in Plate 7, figures 4 and 5. Definitive work by Leopold and MacGinitie (1972) and Frederiksen and Christopher (1978) in North America, and several others in Europe has shown that *Platycarya* pollen ranges from the beginning of the Early Eocene, reaches a peak in the latter part of that time interval, and disappears by the early part of the Mid-Eocene. Together with the disappearance of *Pistillipollenites* in the middle of the Middle-Eocene, the dominance of *Platycara* pollen suggest the age of this *Platycara* zone is most likely Early Eocene, probably toward the latter part of that interval, approximately 54-49 Ma.

Samples of the *Platycara* zone have been recognized in the Canadian Hunter Birch Bay and Richfield-Pure Sunnyside wells (located on Fig. 3; see Fig. 15 below), and wells of the Strait of Georgia, and at Canadian Sumas Mountain. This zone was not encountered in the Conoco-Dynamic Mud Bay well (located on Fig. 3; see Fig. 15 below), which apparently stopped drilling immediately above the zone at 1700 m.

The sedimentary succession immediately overlying the *Platycarya* zone at Ferguson Point consists of about 95 m of interbedded massive sandstone, bentonitic sandstone, and carbonaceous mudstone and siltstone (Blunden, 1971; this study). These are best exposed at Second Beach, Stanley Park, especially at low tide (Fig. 11), and also occur in the Conoco-Dynamic Mud Bay #1 well in Boundary Bay (Fig. 15, below). This unit contains a moderately large and well-preserved palynoassemblage (Plates 8, 9) with the most representative and characteristic palynomorphs angiosperm pollen (Pl. 8, fig. 18-25; Pl. 9, fig. 1-14). The most diagnostic pollen are *Castanea/Castanopsis* (Pl. 8, fig. 4, 5), and *Engelhardtia* sp. (Pl. 8, fig. 7). These prompted the designation of this unit as the *Engelhardtia-Castanea* zone by Blunden (1971). Also characteristic are several fungal spores, viz. *Diporisporites* spp. (Pl. 8, fig. 22-25), *Diporicellaesporites* spp. (Pl. 9, fig. 1-2, 4-6), *Multicellaesporites desmodes* (Pl. 9, fig. 12, 13), and *Staphlosporonites conoideus* (Pl. 9, fig. 14). The age of this zone ranges from late Middle Eocene to early Late Eocene, constrained by the Early to early Middle Eocene age of the underlying *Platycara* zone, and the Late Eocene-Early Oligocene of the overlying unit.

The sedimentary rocks overlying the *Englehardtia-Castanea* zone at Second Beach, Stanley Park, previously termed Kitsilano Formation, but herein included as part of the Kitsilano Member of the Huntingdon Formation, contain Late Eocene-Early Oligocene palynoassemblages (Hopkins, 1966; Rouse et al., 1971; this study). Thus no significant time gap is apparent between these strata and the underlying Tertiary strata of Stanley Park, previously separated out as the upper Burrard Formation by Johnston (1923).

The part of the Kitsilano Member stratigraphically higher then the *Englehardtia-Castanea* zone at Second Beach contains a large and well preserved Late Eocene-Early Oligocene palynoassemblage originally described by Hopkins (1966, 1968), expanded in Blunden (1971), and shown here with added data from Canadian Sumas Mountain and the Conoco-Dynamic Mud Bay well in Plates 10-13. As with the previous Eocene assemblages, the two most prevalent groups are the angiosperm pollen and the fungal spores. Among the more diagnostic of the palynomorphs are:

The angiosperm pollen, viz.:

Nyssa kruschii Pl. 10, fig. 11
Nyssapollenites cf. *cruciatus* Pl. 10, fig. 15
Tetracolporopollenites lesquereuxianus Pl. 10, fig. 17, 18
Rhoipites latus Pl. 10, fig. 14; Pl. 11, fig. 4

Fungal spores:

Diporisporites - C, Pl. 11, fig. 9, 10
Diporicellaesporites bellulus Pl. 11, fig. 11-13
D. giganteus Pl. 12, fig. 1, 13
Striadiporites bistriatus Pl. 12, fig. 5, 6
Imprimospora sp. Pl. 12, fig. 8
I. tankensis Norris Pl. 12, fig. 9, 10
Fusiformisporites lineatus Pl. 13, fig. 1-3
Punctodiporites harrisii Pl. 13, fig. 4-7
Ctenosporites wolfei Pl. 13, Fig. 8, 9

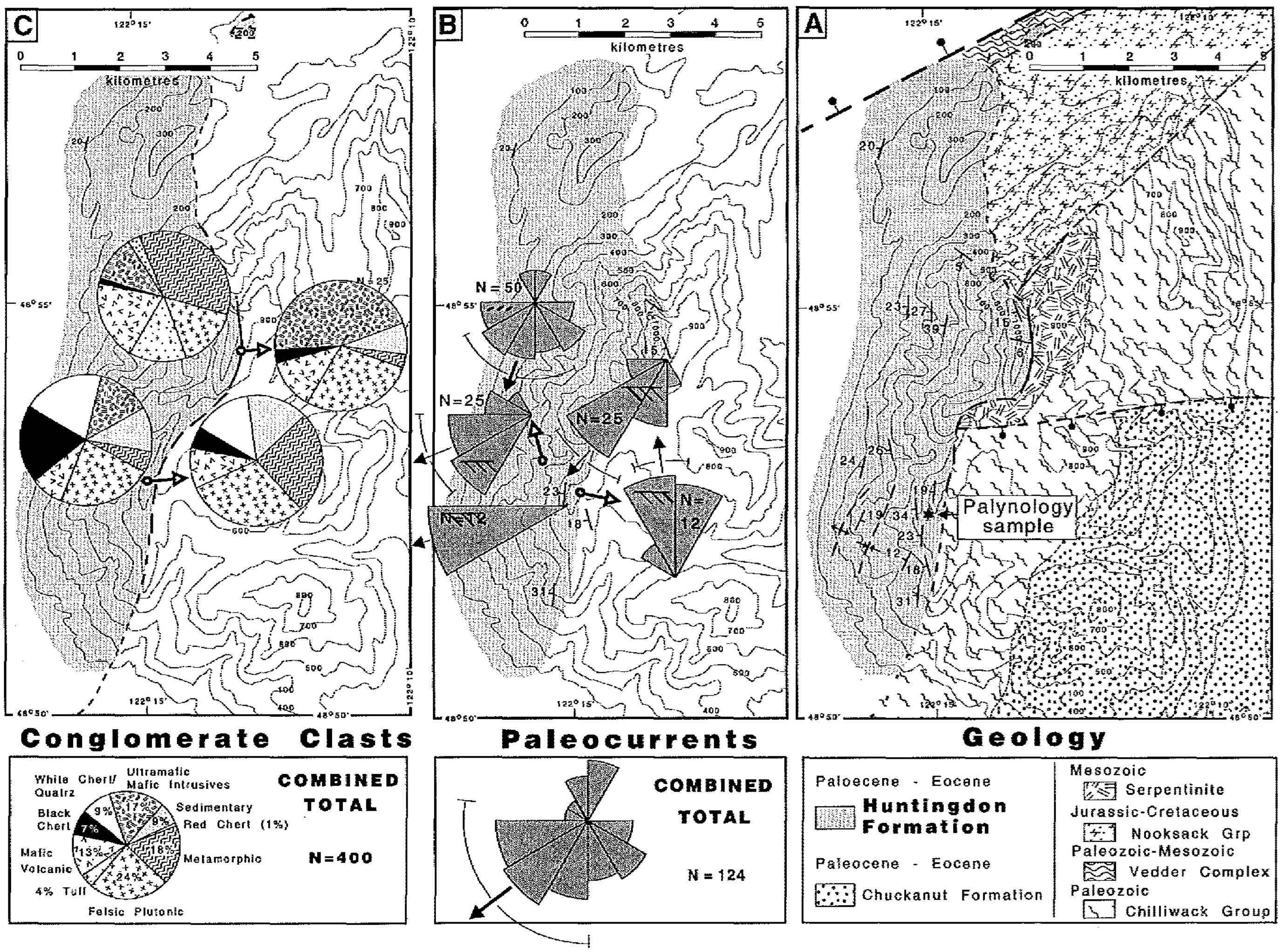
The overall palynoassemblage correlates most closely with assemblages described from the Jackson Group of the Gulf Coast by Frederiksen (1980), the British Columbia Interior and the Arctic (Rouse, 1977), and offshore Labrador (Williams, 1986), all assignable to a Late Eocene-Early Oligocene age. The paleoclimate appears to have been very warm and temperate.

Huntingdon Formation, U.S. Sumas Mountain

In northern Washington State, Moen (1962) and Miller and Misch (1963) mapped a discontinuous belt of sedimentary rocks which they correlated with the Huntingdon Formation of Canadian Sumas Mountain. This unit is poorly exposed in several discontinuous creek exposures and isolated outcrops on the western flank of Sumas Mountain, Washington State (Fig. 13A). The strata generally dip about 15 to 30° to the west, but are deformed at the southern end into northwest-plunging folds. An angular unconformity exposed on part of the main ridge of U.S. Sumas Mountain separates the Huntingdon Formation from subjacent metasedimentary and metavolcanic rocks of the Paleozoic Chilliwack Group or Mesozoic ultramafic intrusions. The thickness of this sedimentary unit at U.S. Sumas Mountain is estimated at 250 to 300 m (however Miller and Misch, 1963 estimated a thickness of about 500 m for the same area).

Lithofacies and sedimentology

As at Canadian Sumas Mountain, refractory clays are preserved as discontinuous lenses in the lower part of the U.S. Sumas Mountain succession, although not directly on the unconformity, nor in the abundance or quality of the Canadian area (the few mining attempts are reviewed in Moen, 1962). However the overall coarsening-upward trend of the Canadian exposures is not apparent in the Washington State exposures. Basal conglomerate at the unconformity is composed of blocks of the underlying lithologies (well exposed over laterally short intervals at the contact with the ultramafic body as shown in Fig. 13A). These are overlain by conglomerate interbedded with fine- to coarse-grained lithic arenite. Many of the conglomerate units in the lower part of the succession are cobble rich, with rare boulders and display vague bedding, poor sorting, no internal fabric, and high contents of poorly sorted mud to granule matrix (both matrix-supported and matrix-rich clast-supported types are present). However, most conglomerate is moderately-sorted and generally occurs as overlapping lenses of normal-graded pebbles-cobbles with slightly channelized bases erosive into underlying beds (Fig. 14A). Both planar and trough cross-bedding is present in some conglomerates (Fig. 14B). These conglomerate beds are interbedded with medium- to very coarse-grained lithic arenite, commonly with 10-20% small



A) Simplified geological map of U.S. Sumas Mountain. The basal Huntingdon contact is as remapped by the senior author. Other contacts are compiled from Moen (1962), Miller and Misch (1963), Johnson (1982), and Monger (1993).

B) Paleocurrent roses from sandstone crossbedding and conglomerate clast imbrication (all statistically significant and rotated to account for bedding tilt). See Figure 18D inset for explanation of rose construction. Centre of rose is located at measurement site where possible, one has been offset to avoid overlap (large unfilled arrow begins at measurement site). Paleocurrent rose at base is for all data.

C) Conglomerate clast composition pie diagrams. Centre of pie is located at measurement site where possible. Arrow begins at measurement site for offset pies. Clast compositions are based on identification of 100 randomly picked clasts/site. Pie graph at base is for all data.

Figure 13. *U.S. Sumas Mountain summary maps.*

pebbles. Some sandstone beds either are gradational from underlying conglomerate, forming fining-upward bedsets 1-2 m thick, or sharply overlie conglomerate as separate pebbly sandstones. Both types are commonly horizontally thin bedded, with less common planar or trough crossbeds, and massive or normally graded. Crossbeds and clast imbrication indicate paleoflow directions ranging from north to southwest (Fig. 13B). The conglomerate contains high percentages of ultramafic, metavolcanic, and metasedimentary clasts (Fig. 13C), all types common in older rocks underlying and east of the formation, suggesting a local source.

The most likely depositional environment was a braided stream and sheetflood deposystem on the lower part of an alluvial fan or fan-proximal braidplain. The overlapping, moderately sorted, and shallowly channelized conglomerate sheets and interbedded coarse grained conglomerates are typical of the gravel and sandy barforms of braided fluvial systems (Miall, 1992). The gravel-rich and generally poorly cyclic to noncyclic succession is similar to the Scott type of braided fluvial facies model, although transitional in some places to a Donjek type system (both defined in Miall, 1977). The poorly sorted, disorganized conglomerate units of the lower succession are interpreted as debris flows and suggest a lower alluvial fan setting for at least the lower part of the succession. The abundance of locally derived clasts and the coarse, poorly organized textures strongly suggest deposition along one margin of the basin for this time period. As at Canadian Sumas Mountain, prolonged weathering at the unconformity surface is interpreted to have produced the kaolin-rich regolith, but most of the U.S. Sumas Mountain fireclays appear to have been transported and deposited in small ponds or lakes (with dilution by other clastic material accounting for the lower quality fireclays).

Palynology and other age controls

Miller and Misch (1963) correlated the sedimentary rocks on U.S. Sumas Mountain with the Huntingdon Formation of Canadian Sumas Mountain based on a perceived similarity of compositions, especially the presence of kaolin-rich clay

***Figure 14.** Huntingdon Formation at U.S. Sumas Mountain. **A)** Vertical cliff (Dale Creek) about 75 m high consisting of interbedded pebble conglomerate and coarse grained to pebbly lithic arenite. Conglomerates occur in 10 to 50 cm thick beds and are typically internally stratified, poorly normally graded at top and in some places show clast imbrication. Sandstones are laterally discontinuous over a few tens of metres (cut out by overlying conglomerate bed) and internally are planar thin bedded or planar crossbedded. GSC 1994 - 712J **B)** Lower Huntingdon Formation on south-central part of U.S. Sumas Mountain comprising irregular beds of pebble and rare cobble conglomerate overlying very coarse grained to granule planar tabular sandstone (contact at hammer). Many conglomerates are normally graded and capped by granule to very coarse grained sandstone (in some places planar crossbedded). Bases are slightly irregular and erosive into underlying beds. GSC 1994-712K*

layers in both units. They also suggested that on the northern side of Squalicum Mountain, (about 5 km east of Bellingham) an angular unconformity separates the unit they mapped as Huntingdon Formation from the underlying Chuckanut Formation. However, Reiswig (1982) questioned this interpretation and suggested the rocks mapped as Huntingdon Formation at Squalicum Mountain are an upper facies of the Chuckanut Formation, based on the contained palynomorphs and detailed structural measurements in this area. Examination of the Squalicum Mountain area for this study also suggests that, although there are changes in bed attitude between definite Chuckanut Formation strata and the supposed Huntingdon Formation strata, these changes are consistent with folding present in the Chuckanut Formation, and there is no clear evidence of an unconformity between these map units. In addition, new samples from the "Huntingdon Formation" at Squalicum Mountain (G.E. Rouse, unpub. data) contain palynomorphs which confirm the conclusions of Reiswig (1982). This is consistent with the evidence from the type Huntingdon Formation on Canadian Sumas Mountain which spans the Upper Paleocene and Eocene epochs and thus is correlative to the Chuckanut Formation and not a younger unit.

All but one sample collected for this study from U.S. Sumas Mountain lacked identifiable palynomorphs. The single productive sample was taken from an isolated outcrop in the lower part of the section exposed at U.S. Sumas Mountain (Fig. 13A), and is probably within 100 m of the unconformity. The recovered palynomorphs include angiosperms and fungal spores similar to those described above for the upper Huntingdon Formation in Canada and also suggest a Late Eocene to possibly Early Oligocene age (G.E. Rouse, unpub. report to the Geological Survey of Canada, 1992).

The correlation of the U.S. Sumas Mountain strata with either the Huntingdon Formation or an upper member of the Chuckanut Formation is thus semantic. We continue to refer the U.S. Sumas Mountain area to the Huntingdon Formation because of the similar presence of "fireclays" and with respect to the historic precedence of the Miller and Misch (1963) correlation, but suggest the Squalicum Mountain strata be termed Chuckanut Formation (Padden Member) as suggested by Reiswig (1982) and mapped by Johnson (1982, 1984a).

Subsurface Paleocene and Eocene sedimentary rocks of the lower Mainland and Whatcom County, Washington State

Recent drilling in greater Vancouver (Conoco-Dynamic wellsites 5, 23, 24 in Fig. 3) and in Whatcom County (AMEX Birch Bay, Terrel No. 1, and Ferndale No. 1, wellsites 11, 12, and 13 respectively in Fig. 3) have provided new information on the subsurface stratigraphy of this part of the basin. In addition, resampling and further examination of well cuttings from two of the deepest exploration holes drilled in the basin (Richfield-Pure Point Roberts and Richfield-Pure Sunnyside, Fig. 15; wellsites 7 and 8 in Fig. 3) have provided a more complete palynological database than previously available, allowing correlation of subsurface stratigraphy to the major surface units as shown in Figure 15.

The Paleocene-Eocene component of this subsurface stratigraphy appears to be slightly thicker than implied from northern outcrop exposures, with about 1500 m of section present in the Richfield Point Roberts and Richfield Sunnyside wells, increasing in thickness to about 2400 m at the American Hunter Birch Bay well.

Lithofacies and sedimentology

The main features of this subsurface unit and the overlying Miocene strata are determined from electric logs, core and cutting descriptions from the Conoco-Dynamic Mud Bay well (provided courtesy of Conoco Exploration Ltd., Calgary), and descriptions of the older wells (Johnston, 1923). The Eocene-Paleocene intervals in these wells are sandstone-dominated, with interbedded mudstone and less common pebble conglomerate and rare thin coal. Bell-shaped SP-Resistivity electric log profiles are common and typical of fining-upward sandstone-mudstone cycles. Coarsening-upward and blocky interbedded layers are less common. Sandstone is generally medium- to coarse-grained and moderately- to well-sorted arkosic to lithic arenite. Clay or siliceous cements are typical; clay or minor chlorite alteration of feldspars and rock fragments is common. Porosity varies from poor to good (generally <20%). Lithic fragments generally make up 10-30% of the sandstone framework with volcanic and plutonic rock fragments, chert, and mica all common. Mudstone is generally grey and variably carbonaceous with rare to abundant plant fragments. Coal occurs as wispy fragments and as thin lignitic to sub-bituminous, generally a few centimetres thick and rarely a few tens of centimetres thick.

The general character of the subsurface Paleocene-Eocene strata is very similar to that of the surface exposures. The entire succession is interpreted as a thick (more than 2 km in some wells) fluvial sequence with laterally migrating meandering channels as part of a sand-dominated fluvial floodplain. There is no evidence from the core lithofacies or from the contained microfossils of any marine component in this succession.

Subsurface Miocene-Pliocene sedimentary rocks

A 400 m thick succession of sandstone, mudstone, plus minor conglomerate and lignite intersected in an early exploration drillhole (Boundary Bay No. 3, Fig. 15; well site 4 on Fig. 3) was recognized by Johnston (1923) as being younger than his Eocene Kitsilano Formation. He inferred that these strata were Miocene or possibly Pliocene in age and proposed the name Boundary Bay Formation for the succession. Hopkins (1966) and Rouse et al. (1990) considered these strata to be correlative with Miocene strata intersected in the Richfield drillholes (Fig. 15 and well sites 7 and 8 on Fig. 3). Miocene strata were also intersected in the American Hunter Birch Bay and Conoco-Dynamic Mud Bay wells (Fig. 15 and wells 11 and 5 respectively on Fig. 3). The corrected true thickness of the Miocene section ranges from about 200 m at Birch Bay to about 1180 m at Point Roberts. All thicknesses are minimums as the top in all holes is the sub-Quaternary unconformity.

Lithofacies and sedimentology

The strata consist of poorly indurated, intercalated sandstone and mudstone with minor pebble conglomerate and coal. Electric logs show blocky and bell-shaped patterns typical of interbedded sandstone-mudstone interbeds and fining-upward sandstone-mudstone successions ranging from about 2 m to 30 m thick. Sandstones are generally medium- to coarse-grained, arkosic to lithic arenite, and moderately sorted, but with high clay matrix in some samples. Pebble conglomerate is rare, although pebble-rich bases are present in some thick sandstone units. Mudstone is generally grey (but green-grey and red muds are present) and commonly carbonaceous, with several thin sub-bituminous to lignitic coal parting or beds present at the top of fining-upward cycles. Johnston (1923) reported several thin ash beds in this succession, but ash beds are not reported in subsequent well descriptions. Distinctive horizons are not present in well sections or geophysical logs and thus correlation between the widely spaced wells is not attempted. In general, formation-scale vertical trends in grain size or composition are not apparent. An exception is the succession intersected in the Mud Bay well which shows a distinctive overall coarsening- and thickening-upward trend from a lower unit about 150 m thick which comprises mostly mudstone, siltstone, and thin, fine grained sandstone interbeds to an upper unit about 700 m thick in which thick sandstone beds are common and are organized in stacked fining-upward cycles.

The stacked fining-upward sandstone-mudstone cycles of this Miocene sequence are interpreted as fluvial deposits similar to those of the underlying Eocene sections. Thicker cycles with pebbly sandstone bases represent major channels. Thin cycles and interbedded sandstone-mudstone successions represent overbank, floodplain, and possibly crevasse splay deposits. The presence of dinocysts in some parts of

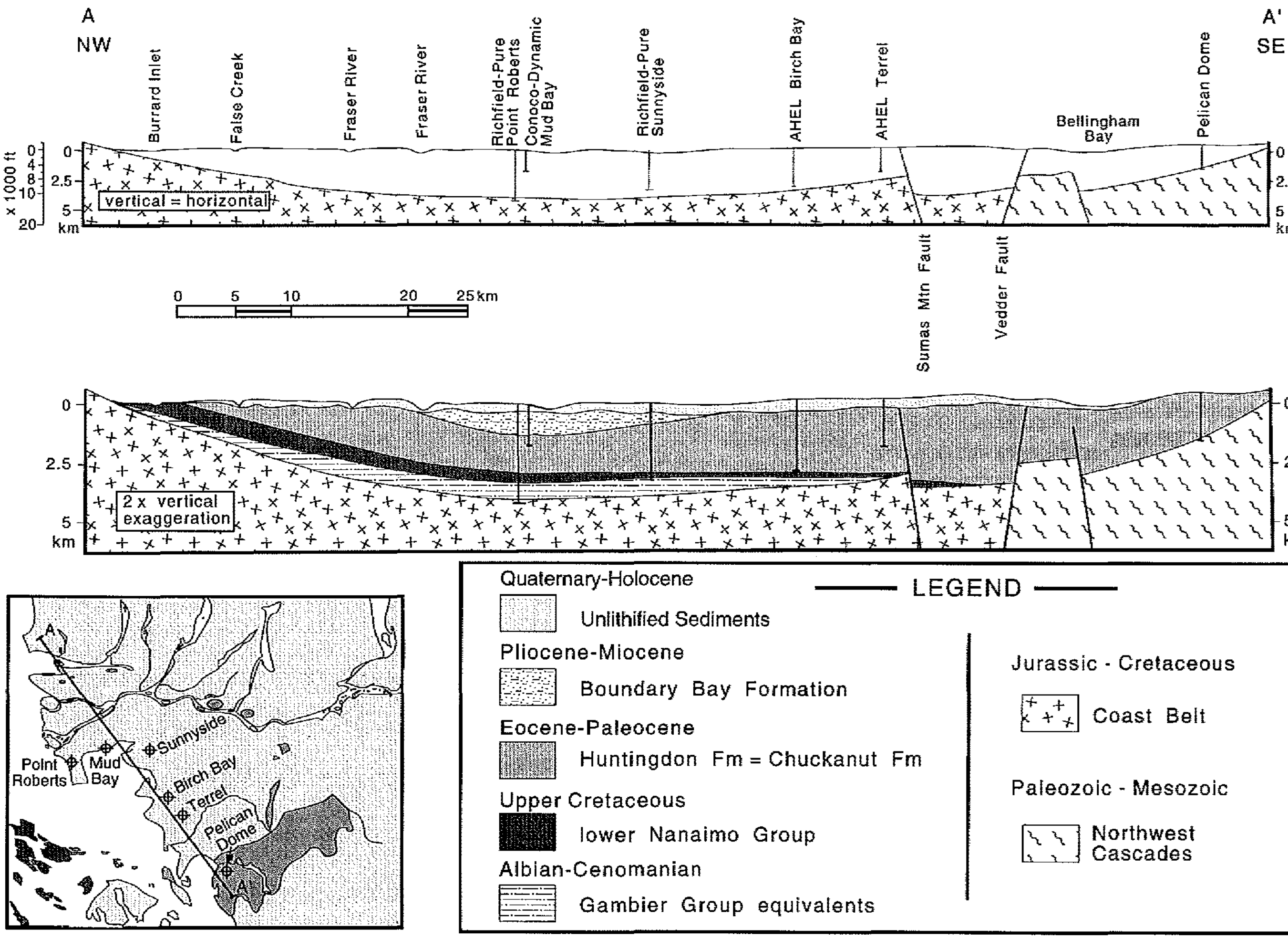


Figure 15. Stratigraphic cross-section from Burrard Inlet in North to Chuckanut Mountain in south showing main Tertiary and Late Cretaceous units of Georgia Basin in this area. Exploration wells in which palynological age controls are sufficient to delineate major unit boundaries are shown (except the Pelican Dome well which has no palynological control, but is shown because it intersects local basement on the line of section). Main faults are either projected from surface traces or known from petroleum exploration seismic lines in the southern part of the basin.

the sections provides the only evidence of marine influence in any of the Tertiary strata of the Georgia Basin. A marginal marine or estuarine setting is the most likely depositional environment for these marine intertongues.

Palynology

Hopkins (1966, 1968) studied the palynoassemblages from cuttings obtained in two drillholes (Fig. 15 and wells 7 and 8 of Fig. 3). He determined a Miocene age for the upper succession in these wells, estimating the thickness of the Miocene to be about 1200 m in both drillholes. A more recent palynological study of the well material by Rouse et al. (1990) also determined a Miocene age for these strata and for strata informally termed the "South Westminster formation" for a single poorly exposed outcrop south of the Fraser River in South Westminster.

Chuckanut Formation, northwest Washington State

Johnson (1982, 1984a, b, c, 1985, 1991) completed the most recent sedimentological and stratigraphical analysis of the Chuckanut Formation. Several M.Sc. graduate theses provide additional studies of local areas or specific aspects of Chuckanut Formation deposition (Kelly, 1970; Pongsapich, 1970; Hartwell, 1979; Robertson, 1981).

Lithofacies and sedimentology

Johnson (1984b) subdivided the 6 km thick Chuckanut Formation into seven members, since revised to six (Johnson, 1991; Fig. 3, 16A). To simplify this brief review, the Chuckanut Formation has been grouped into lower, middle, and upper subdivisions. The lower subdivision comprises the Bellingham Bay Member, a 3300 m thick succession of stacked and overlapping fining-upward cycles of conglomerate (minor), sandstone, and mudstone. Johnson (1984c) provides a detailed description of Bellingham Bay Member stratigraphy and sedimentation. The member is both thickest and coarsest in eastern exposures and paleocurrents demonstrate a radial but in general south to southwest pattern of paleoflow (Fig. 16B). Conglomerate clasts are predominantly chert, mafic volcanic, or sedimentary, but several other types are present (Fig. 16C). Bellingham Bay Member sandstone is arkosic (micaceous in places) and generally medium- to coarse-grained in lower parts of cycles, and fine grained in upper cycles. The stacked sandstone (conglomeratic in places) to mudstone cycles occur both as multi-story sheets interpreted as the deposits of fine-load meandering river systems and as ribbon sandstones within relatively mudstone rich units, interpreted as floodplain channel or crevasse splay deposits.

The middle Chuckanut Formation comprises the Governors Point Member which directly overlies the Bellingham Bay Member in western and central outcrop areas (Fig. 3, 16A), the Slide Member, present only in the eastern outcrop areas, and the Padden Member which is the western equivalent to the Slide Member (and probably equivalent to the upper Chuckanut Formation described below). The Governors Point Member thins eastward from a maximum of 375 m in western exposures and consists of amalgamated fining-upward couplets of trough crossbedded and planar bedded medium- to coarse-grained lithic arenite with a central succession of massive to crudely stratified conglomerate and pebbly sandstone. The Slide Member is about 2 km thick and consists of fining-upward sandstone-mudstone cycles with similar geometries and compositions to those of the underlying Bellingham Bay Member. The Padden Member thickness is not well-constrained, but a thickness of more than 3 km is suggested from map patterns in western outcrop areas. The upper part of the Padden Member is poorly exposed and dips beneath Quaternary cover at the northern limit of Chuckanut Formation exposures east of Bellingham (Fig. 3). A Paleogene succession more than 500 m thick on Sucia Island was included with the Padden Member by Johnson (1982, 1984b), but is discussed separately in this study. The Padden Member is dominated by medium- to coarse-grained lithic arenite complexly interbedded with massive to crudely stratified or crossbedded conglomerate, both rock types alternating with laminated mudstone and minor coal.

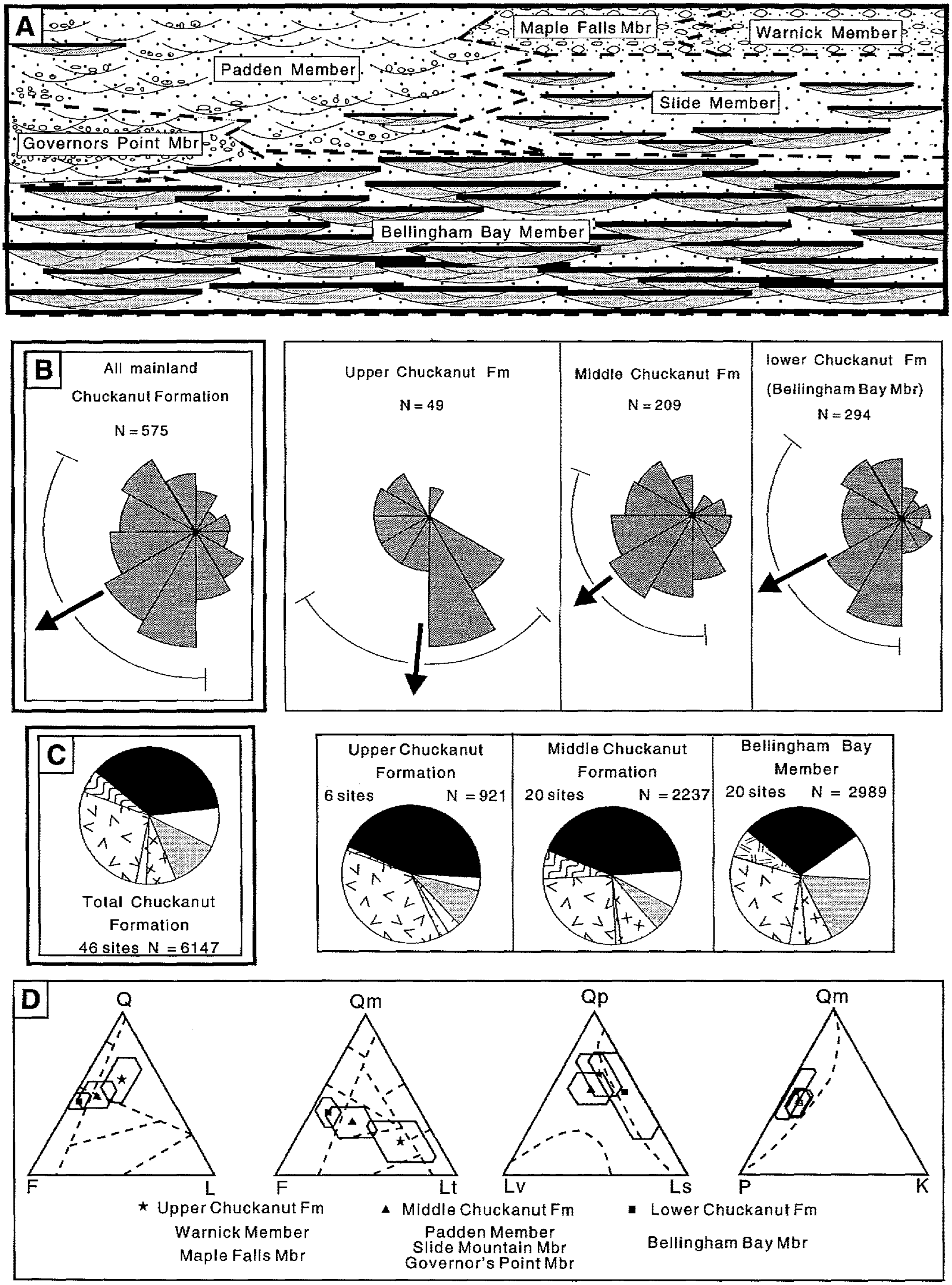
A) Schematic diagram of the six members defined by Johnson (1982, originally as seven members, modified in Johnson, 1991). Pattern fills reflect the major lithofacies of the member as summarized in the text.

B) Summary of paleocurrent data for mainland Chuckanut Formation occurrences (not including the Coal Mountain area studied by Robertson (1981), or the Sucia Island chain, reported on separately in this study). Data compiled from Johnson (1982; pers. comm., 1992) and Hartwell (1979). Lower, middle, and upper Chuckanut divisions are as defined in Figure 16D.

C) Conglomerate clast compositions from mainland Chuckanut Formation (not including Sucia Island chain, reported on separately in this study). All data from Johnson (1982). Clast composition patterns are the same as defined in Figure 22.

D) Summary of sandstone detrital clast data for mainland Chuckanut Formation using the framework mode subdivisions and tectonic provenance divisions of Dickinson and Suczek (1979) as modified by Dickinson et al. (1983 for QFL and QmFLt plots). The mean value for each stratigraphic subdivision is shown with the defined symbol with a polygon representing the standard deviation of the mean. Table 2 provides numerical summaries of the data and sources of information. Figure 23A provides an explanation of the tectonic provenance field divisions.

Figure 16. *Chuckanut Formation summary diagrams.*


A
Maple Falls Mbr
Warnick Member
Padden Member
Slide Member
Governors Point Mbr
Bellingham Bay Member
B
All mainland Chuckanut Formation
N = 575
Upper Chuckanut Fm
N = 49
Middle Chuckanut Fm
N = 209
lower Chuckanut Fm (Bellingham Bay Mbr)
N = 294
C
Total Chuckanut Formation
46 sites N = 6147
Upper Chuckanut Formation
6 sites N = 921
Middle Chuckanut Formation
20 sites N = 2237
Bellingham Bay Member
20 sites N = 2989
D
Q
F
L
Qm
F
Lt
Qp
Lv
Ls
Qm
P
K
★ Upper Chuckanut Fm
Warnick Member
Maple Falls Mbr
▲ Middle Chuckanut Fm
Padden Member
Slide Mountain Mbr
Governor's Point Mbr
■ Lower Chuckanut Fm
Bellingham Bay Mbr

Paleocurrents generally indicate northern to eastern source areas (Fig. 16B). Sandstone is arkosic to lithic with a distinctively higher component of chert and volcanic lithic fragments compared to the lower Chuckanut Formation (Fig. 16C). Conglomerate clast compositions also show an increased abundance of chert relative to the Bellingham Bay Member (Fig. 16D). All clast types are typical of local basement lithologies, suggesting local derivation of clasts from marginal, high relief areas. The Slide Member is interpreted as primarily the deposits of fine-load meandering fluvial systems, similar to the Bellingham Bay Member. The Governors Point Member is interpreted as braided fluvial deposits derived from local high relief areas, possibly related to syndepositional faulting Johnson, 1982, 1984b, inferred north of the deposit. The Padden Member is interpreted as coarse-load meandering and braided river deposits mostly derived from the north or northeast. Conglomerate clast types suggest a predominantly Coast Belt source, especially for Padden Member deposits.

The upper Chuckanut Formation consists of the Maple Falls and Warnick members, both restricted to eastern outcrop areas (Fig. 3, 16A). The Maple Falls Member is estimated to be about 800 m thick and appears to gradationally overlie and interfinger with the Padden and Slide members to the west and south, and to interfinger with the Warnick Member to the east, although original contact relationships are obscured by faulting and poor exposure. The Maple Falls Member consists of clast-supported conglomerate in thick (some to 12 m), massive boulder-rich beds and thinner crudely stratified beds with common clast imbrication. These conglomerates are interbedded with crossbedded to massive sandstone and massive to laminated mudstone. In addition, thicker interbeds are dominated by mudstone containing abundant lenticular beds of massive sandstone to pebbly sandstone. The Warnick Member is possibly up to 1000 m thick (based on map patterns) and contains conglomerate-rich and mudstone lithofacies similar to the Maple Falls Member. Both members are interpreted as interfingering alluvial fan and alluvial plain deposits. Paleocurrents are west- to southwest-directed in the Warnick Member and south directed in the Maple Falls Member. Conglomerate clasts are dominantly chert and mafic volcanic types and sandstones contain significantly more chert and volcanic lithic clasts than other members, although the Maple Falls Member is also arkosic in some places. Both conglomerate clast and sandstone compositions are compatible with local northern sources as implied by the paleocurrent data.

Palynology and other age controls

The age range of the Chuckanut Formation, based on early plant and palynological studies, was generally suggested to be Late Cretaceous to Eocene (Miller and Misch, 1963; Hopkins, 1966; Pabst, 1968; Griggs, 1971; Frizzel, 1979). Reiswig (1982) restudied the palynology and concluded that the Late Cretaceous palynoassemblages were reworked and that the non-reworked palynological evidence suggested a Middle Paleocene to Early Eocene age. The recent palynological study of Rouse et al. (1990) also suggested a Paleocene maximum age. Johnson (1982, 1984b) obtained a zircon fission-track age of 49.9 ± 1.2 Ma from a tuff bed situated about in the middle of the Chuckanut Formation succession (near the top of the Boundary Bay Member). He also estimated a maximum age for the Chuckanut Formation from fission-track ages of detrital zircons as about 55 Ma, that's Late Paleocene or Early Eocene, depending on the exact placement of the Paleocene-Eocene boundary (56.5 ± 2.5 Ma in Harland et al., 1990).

Oligocene igneous rocks

A series of igneous rocks occur as scattered separate dykes, sills, and possible flows along the northern edge of the preserved Tertiary basin (unit TV on Fig. 3). There were collectively termed the Prospect Point eruptives by Johnston (1923), named from the greater than 30 m thick dyke exposed at Prospect Point in Stanley Park (Fig. 5A). Other major igneous bodies include circular intrusions at Sentinel Hill (Fig. 5A) and Little Mountain, thin dykes at Kitsilano beach and in Brothers Creek and the Capilano Canyon, and a thick sill exposed at Grant Hill and Silverdale Hill west of Mission (Fig. 2). Other small igneous bodies, generally identified as sills or flows, have been reported from construction sites in the greater Vancouver area. A comprehensive study of these igneous bodies is contained in the unpublished theses of Blanchet (1943) and Wootton (1959) and by Hamilten and Dostal (1994). The igneaous bodies are fine to medium crystalline, have diabasic textures, are vesicular and augite-bearing, and of andesitic to basaltic compositions. Columnar jointing is common in the thickest bodies. Most are obviously intrusive into the containing sedimentary rocks. Wootton (1959) suggested some of the bodies could be either sills or flows, including the Sentinel Hill, upper Prospect Point, and Little Mountain bodies. Sketchley and Clowes (1976) conducted a gravity study of the Little Mountain body and suggested it was more likely a flow than a sill. Blanchet (1943) suggested that all occurrences originated as shallow intrusions, either dykes, sills, or small laccoliths. Potassium-argon dating of the Prospect Point and Little Mountain occurrences suggests they are Early Oligocene (32 ± 1, and 34 ± 1 Ma respectively, R.A. Armstrong, 1980, unpublished geochronological database, University of British Columbia). The Prospect Point body intrudes Late Cretaceous sandstone and the Little Mountain body occurs in the upper part of the Eocene Kitsilano Member; their nearly coeval dates suggest intrusive origins for both bodies. A sill interpretation is likely for the other low-dipping igneous bodies of this area such as the Grant Hill and Silverdale Hill examples, although neither has been studied in detail. The similar types of occurrences and roughly similar compositions suggests they are part of a minor early Oligocene igneous event. Subsurface sills or dykes cutting Eocene and Late Cretaceous strata are also reported from several exploration wells (e.g., 65 m thick diorite sill intruding Late Cretaceous strata at 2100 m depth in AHEL Birch Bay; well 11 on Fig. 3; described in Hurst, 1991). These are likely correlative with the igneous event described above, but have not been dated.

Tertiary rocks of Georgia Basin outside of mainland British Columbia and Washington State

Lasqueti Island

The northern end of Lasqueti Island contains an outlier of sedimentary rock shown on published maps as an unnamed Upper Cretaceous or Tertiary unit at the top of the Nanaimo Group (Fig. 17). Fieldwork in 1990 confirmed that the lower units are part of the Nanaimo Group and provided direct evidence for an unconformity separating these units from an overlying Paleocene sedimentary succession (Mustard and Rouse, 1991).

Lithofacies and sedimentology

The Paleocene unit is more than 70 m thick, comprising 15 m of interbedded fine- to medium-grained arkosic and micaceous arenite and dark grey mudstone (Fig. 17). The sandstone occurs in sets of slightly wavy, horizontal thin beds up to 3 m thick (most <50 cm). This sandstone-mudstone lithofacies is sharply overlain by more than 55 m of pebble-cobble conglomerate and interbedded coarse grained arkosic arenite. Conglomerate beds are generally clast-supported with a coarse grained matrix. Beds are 0.5-2 m thick and occur as overlapping lenses. Discordance of conglomerate bases indicate several metres of erosion into underlying sandstone in places. Clast imbrication indicates paleoflow ranging between west and southwest. About 40% of the clasts are

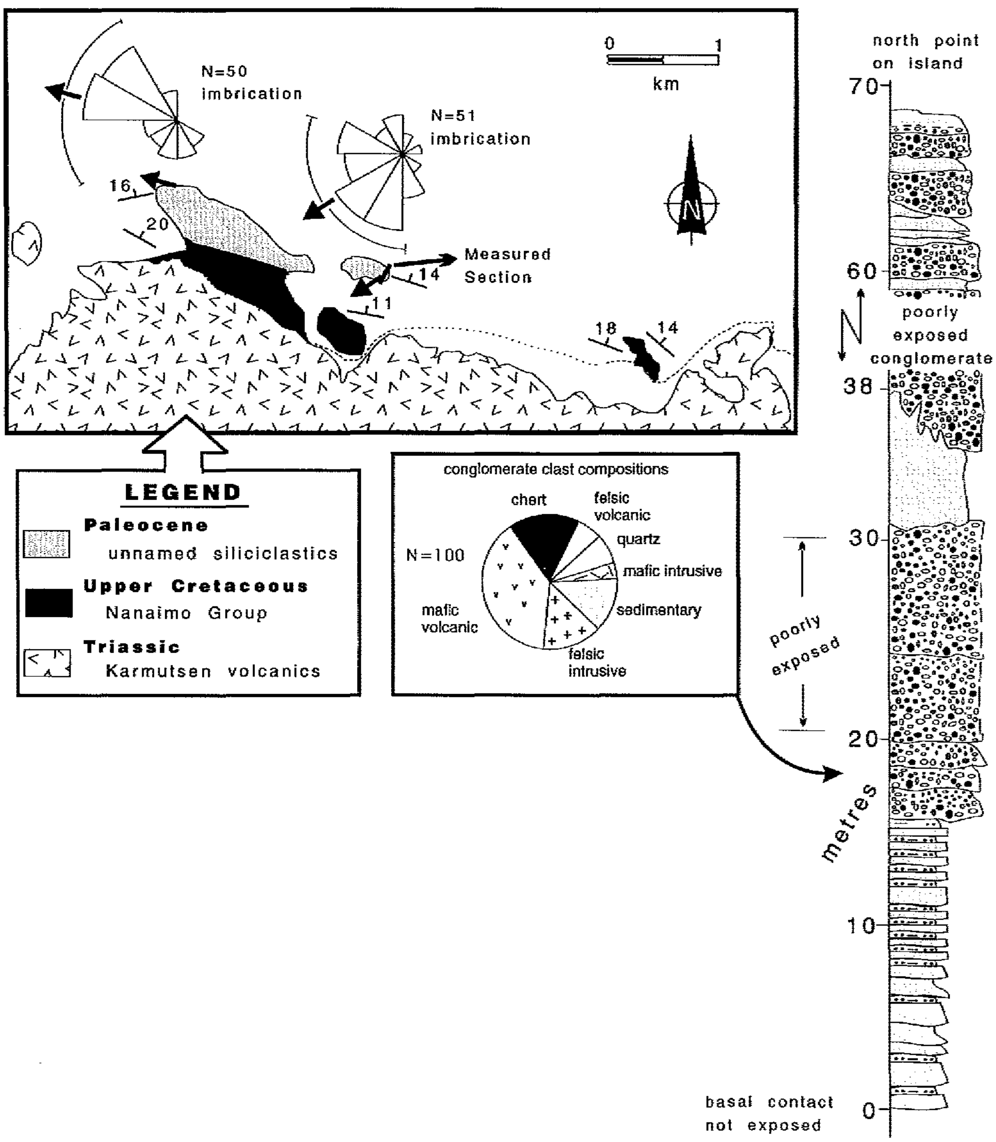


Figure 17. North end of Lasqueti Island (located on Fig. 1) showing geology of outlier of sedimentary rocks (left diagram) and on right a detailed measured section through Paleocene strata well-exposed on an unnamed island located on left diagram.

mafic volcanic and mafic intrusive compositions. Granitic, felsic volcanic, grey chert, and argillite make up remaining clast types, all compatible with the eastern provenance suggested by the paleocurrents.

The top contact of many conglomerate beds is slightly irregular with common projecting clasts. Rarely, a discontinuous, white, chalky carbonate layer a few centimetres thick is preserved at this contact, both draping and cementing pebbles. These layers are identical to subaerial caliche weathering horizons, common on alluvial fans. Most other features also suggest alluvial fan deposition, with the thick massive conglomerate deposited as debris flows and the thinner, better channelized conglomerate and sandstone deposited in braided channels. The underlying sandstone and mudstone beds are probably lower fan floodplain deposits.

Inclusion of this outlier as part of Tertiary Georgia Basin is speculative. It is also possible that the outlier represents deposition in an area not directly connected to the Vancouver-Bellingham area.

Palynology

Macrofossils are absent from this unit, but a reasonably large and well preserved palynoassemblage was obtained from the finer facies and indicate a Late Paleocene age (described in detail in Mustard and Rouse, 1991).

The most diagnostic palynomorphs are the angiosperms *Subtriporopollenites* -A (pre-*Tilia*), and *Pistillipollenites mcgregorii*. These occur together only in zone P-4, Late Paleocene, in the thick Eureka Sound Group succession on Ellesmere Island and correlative sections on other Arctic islands, and in the Alberta foothills, Mackenzie Delta, and Beaufort Sea regions (Rouse, 1977). This assemblage occurs just below early Eocene assemblages containing the earliest true *Tilia* pollen, e.g., *T. crassipites* and *T. vescipites*. The mixture of fungal spores and angiosperm pollen suggests lowland conditions, compatible with a floodplain environment suggested above. The general lack of algal cysts also supports a terrestrial interpretation.

Tumbo and Sucia islands

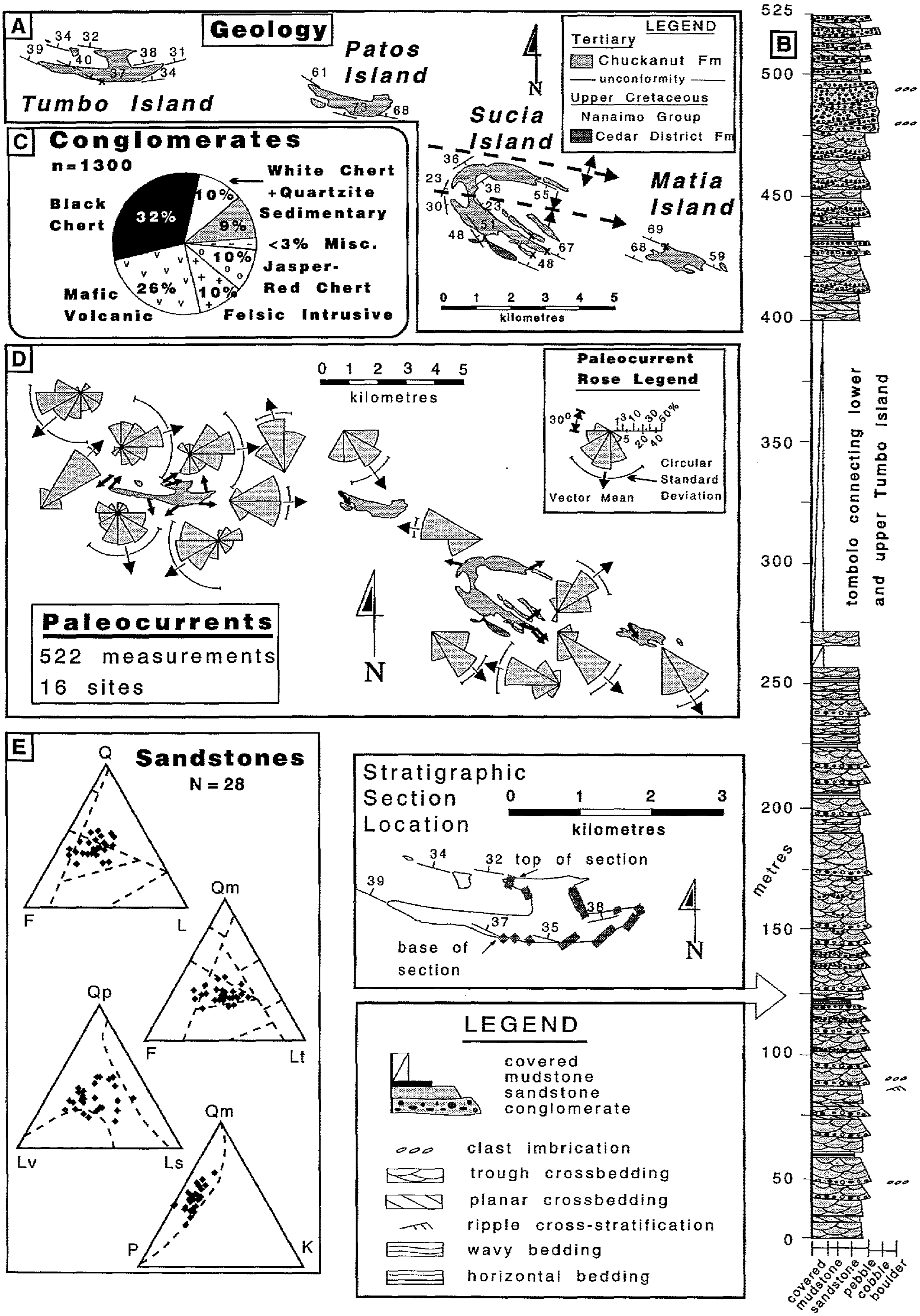
Tumbo Island rocks, shown on published maps as part of the Late Cretaceous Nanaimo Group, contain Paleocene palynomorphs (Mustard and Rouse, 1992). Most rocks of the Sucia Island chain (Fig. 18A) also contain Paleocene pollen, allowing correlation with Tumbo Island and with lower Chuckanut Formation on the mainland (Mustard and Rouse, 1992; suggested earlier by Vance, 1975; Johnson, 1982).

Lithofacies and sedimentology

The preserved section on Tumbo and Sucia islands is more than 500 m thick (Fig. 18B) comprising a lower sandstone-dominated succession of laterally overlapping and vertically stacked fining-upward sequences, each about 1-5 m thick. The sandstone is generally a medium- to coarse-grained feldspathic to lithic arenite (Fig. 18E), with abundant broken carbonaceous plants and rare carbonaceous and partly silicified logs. Trough crossbedding is abundant (Fig. 19A). Planar crossbedding and pebble imbrication is common (Fig. 19B, C). The paleocurrent indicators display a wide variance in paleoflow directions, but the predominant flow direction ranged from northeast to southeast (Fig. 18D). An overall coarsening-upward trend is evident with conglomerate much more abundant in the upper part of the Tumbo Island succession and slightly more so at Sucia Island (Fig. 19D). The most abundant conglomerate clast type is black chert, commonly containing deformed milky white quartz veinlets. Aphanitic green-grey volcanic clasts are slightly less common, followed by subequal amounts of white chert (rarely quartzite), red chert (including jasper), and felsic intrusive and sedimentary clasts (Fig. 18B, C), the last mostly massive, fine- to medium-grained arkosic sandstone.

Deposition occurred in a braided stream to lower alluvial fan system, flanked by marshes with abundant fungal and fern components, and low areas supporting angiosperms and conifers. The overall coarsening-upward trend suggests progradation of an upper braidplain or lower alluvial fan deposystem. The wide variance of paleocurrent indicators and overall radial pattern is typical of lower alluvial fan braided stream deposition. The abundant black and red chert and mafic volcanic

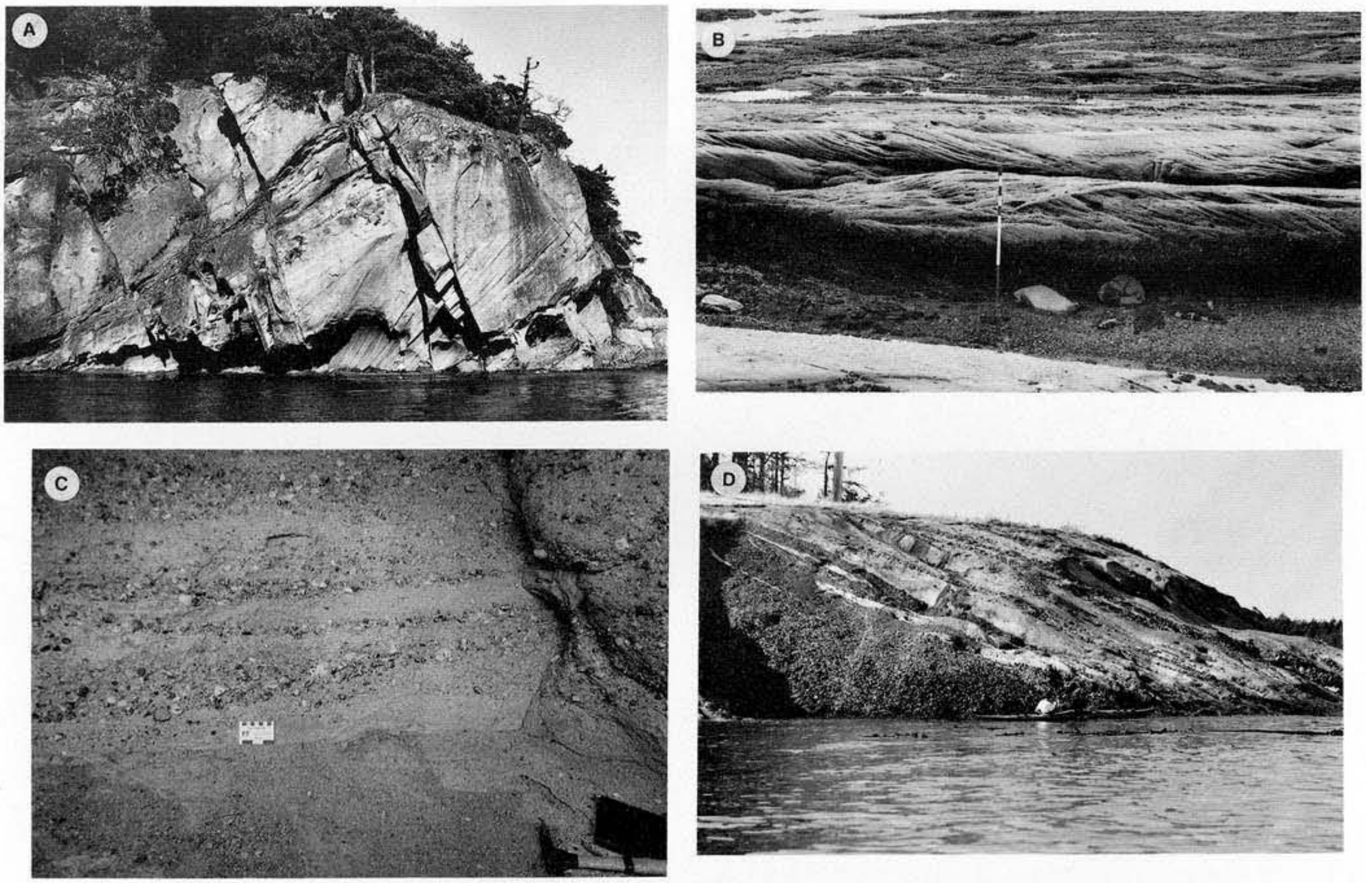
__Figure 18.__ __A)__ Composite measured stratigraphic section from Tumbo Island (located on inset map at lower left). __B)__ Geology of the Tertiary and associated rocks of Tumbo Island, British Columbia and the Sucia Island chain, northwest Washington State. Palynology sample sites are indicated by "X". __C)__ Conglomerate clast compositions combined from all thirteen sites from the four islands. See Figure 5 of Mustard and Rouse (1992) for compositions from individual sites. Clast counts were conducted by identifying all clasts encountered on a random line to a total of one hundred per site. __D)__ Summary diagram of paleocurrent data. Measurements have been corrected for fold plunge and bed-tilt. Paleocurrent measurement sites are located on each island with an arrow which is also the vector mean. Current roses for each site are shown near each locality and the same spatial grouping of the sites has been generally maintained. Current roses are plotted using a nonlinear scale as advocated by Nemec (1988). Only sites with a Rayleigh significance test value less than 0.05 were used for paleocurrent analysis (method of Curray, 1956). Circular standard deviation was calculated using the method of Krause and Geijer (1987). __E)__ Sandstone detrital compositions plotted on tectonic provenance ternary diagrams. Ternary plot fields and mode compositions are explained in Figure 23A. Numerical summaries of data and sources of information are given in Table 2.


A
Geology
Tumbo Island
Patos Island
Sucia Island
Matia Island
LEGEND
Tertiary
Chuckanut Fm
unconformity
Upper Cretaceous
Nanaimo Group
Cedar District Fm
0 1 2 3 4 5
kilometres
C
Conglomerates
n=1300
Black Chert 32%
White Chert +Quartzite 10%
Sedimentary 9%
<3% Misc.
Jasper-Red Chert 10%
Felsic Intrusive 10%
Mafic Volcanic 26%
D
0 1 2 3 4 5
kilometres
Paleocurrent Rose Legend
Vector Mean
Circular Standard Deviation
Paleocurrents
522 measurements
16 sites
N
E
Sandstones
N = 28
Q
F
L
Qm
F
Lt
Qp
Lv
Ls
Qm
P
K
Stratigraphic Section Location
0 1 2 3
kilometres
top of section
base of section
LEGEND
covered
mudstone
sandstone
conglomerate
clast imbrication
trough crossbedding
planar crossbedding
ripple cross-stratification
wavy bedding
horizontal bedding
B
metres
tombolo connecting lower and upper Tumbo Island
covered
mudstone
sandstone
pebble
cobble
boulder

clasts are all compositions common on the San Juan islands to the southwest. The sandstone clasts are identical to sandstone of the upper Nanaimo Group preserved in locations northwest of Tumbo and Sucia islands. The paleocurrent and provenance and the proximal sedimentation patterns suggest these deposits formed near the western margin of the early Tertiary basin.

Palynology

Reasonably well preserved and representative palynoassemblages were recovered from samples collected on the Sucia Island chain and Tumbo Island (Fig. 18A). The assemblages are dominated by fungal spores, with fewer angiosperm pollen and fern spores. Both are correlated with those from other rock units previously assigned to the Late Paleocene, particularly



A) Cliff on southwest Tumbo Island showing large overlapping trough crossbedded, medium grained sandstone of Lower Paleocene unit in this area. Kayak is about 5 m in length. GSC 1994-712L

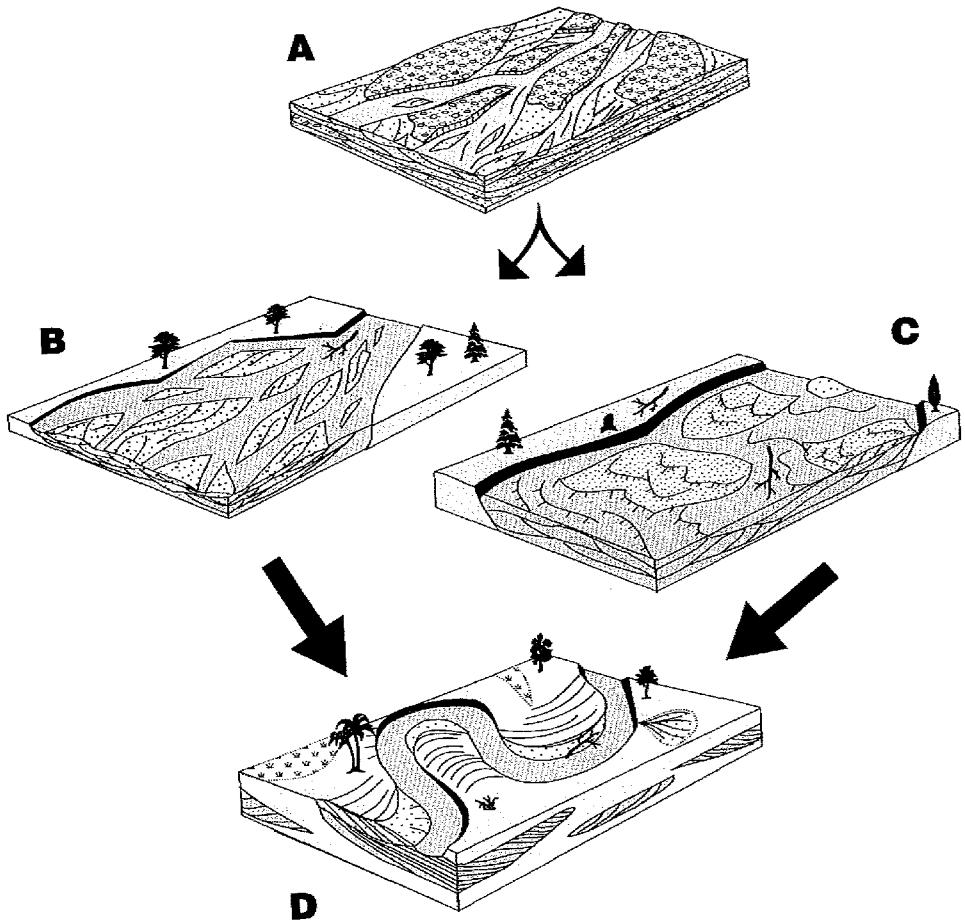
B) Overlapping irregular beds of planar crossbedded sandstone from central part of the Paleocene unit on Tumbo Island, probably formed as small bar complexes in this fluvial system. Measuring staff is 1.5 m long. GSC 1994-712M

C) Irregular to convex-up based pebble conglomerates grading up to and interbedded with coarse grained sandstone. Interpreted as braided fluvial deposits. Scale card is 9 cm on longest side. GSC 1994-712N

D) Exposure at Sucia Island from upper part of Paleocene section showing a 7-9 m thick succession with basal pebble-cobble conglomerate (base is at left edge of photo) which intertongues and grades at top into coarse grained sandstone. Sandstone is wavy to planar crossbedded and includes several discontinuous beds (barforms) of planar and trough crossbedded pebbly sandstone. Conglomerate tongue at right edge of photo thickens to right into basal conglomerate of overlying conglomerate-sandstone cycle. Kayak is about 5 m long. GSC 1994-7120

Figure 19. *Photos from Tumbo and Sucia islands.*

those from the Arctic and Rocky Mountain foothills regions (Rouse, 1977), and from Lasqueti Island (Mustard and Rouse, 1991). Diagnostic species include *Multicellasporites irregularis, M. "giganteus", Inapertisporites elongatus, Staphlosporonites allomorphus, Brachysporisporites cotalis, B. catinus, Callimothallus pertusus*, and *Dicellaesporites laevis* among the fungal spores (and reproductive units). Diagnostic angiosperm pollen are Intrabtriporopollenites-Spp(= pre-*Tilia*), *Rhoipites cryptoporus*, and *Paraalnipollenites alterniporus*. These species occur in assemblages immediately below the Eocene containing the earliest true *Tilia* pollen, i.e. *Tilia vescipites* and *Tilia crassipites*. The angiosperm pollen *Rhoipites cryptoporus*, first described by Srivastava (1972) from the Paleocene in Alabama, is a diagnostic middle and late Paleocene palynomorph from Arctic and western Canada (Rouse, 1977).



A) Proximal to lower alluvial fan facies with sediment gravity flow gravel and pebbly mudstone deposits interbedded with sheetflood pebbly sandstones, all partly reworked in broad braided fluvial channels.

B) Lower alluvial fan to proximal braidplain comprising low sinuosity shallow channels in which gravel and pebbly sand sheets, longitudinal and transverse bars are the major deposits.

C) Sandy braidplain deposition in large barforms with relatively deep channels. Deposits from this type of fluvial system include thick compound bars or sand shoals and complexly overlapping gravel-sand channel sequences.

D) Sandy mixed-load meandering river. Deposition occurs in well defined channels by migration of channels and lateral accretion. In a sand-dominated system such as envisioned for much of the Paleocene-Eocene Tertiary Georgia Basin well developed fining-upwards cycles with thick mudstone capped are less likely than overlapping sheet sand geometries showing complex channel fill sequences and poor preservation of the finer components of channel fills.

Figure 20. *Summary diagrams of main fluvial depositional types interpreted for the Paleocene-Eocene parts of Tertiary Georgia Basin examined for this study. Schematic diagrams are modified from Miall (1985, 1992).*

PROVENANCE AND SEDIMENTOLOGY SUMMARY

The lithofacies descriptions and interpretations of the previous sections indicate that the basin was dominated by alluvial depositional systems with alluvial fan debris flow; are both gravel-rich and sand-rich braided stream deposits common at the basin margins. Sand-rich braided and meandering fluvial deposition is the main process accounting for the bulk of the basin fill, with associated floodplain and possibly minor lacustrine facies also present. Figure 20A-D illustrates the main depositional processes interpreted for these environments. Sandstone occurs in predominantly overlapping irregular to sheet-like bodies where mudstone facies occur as minor interbeds or capping fining-upward channel cycles. The coarse sand and gravel facies are interpreted as the product of deposition in large sand- or gravel-bed braided rivers, more properly termed multiple-channel bedload rivers in recognition of the many variations and transitions between traditional braided and meandering channel models (Miall, 1985). In modern analogues, the overlapping sand or gravel sheets form by continued deposition as small and large barforms in migrating small channels (see reviews in Cant, 1982; Miall, 1985, 1992). Where mudstone interbeds are common, sandstone occurs both as sheets and ribbons, the latter probably representing small channel fills or crevasse splay deposits in major floodplain areas. In these environments, larger channels show evidence of lateral migration and both lateral and vertical accretion, possibly as point bars typical of classical meandering river systems (Smith, 1987). However, a wide variety of vertical and lateral sediment bodies that span the range from low sinuosity braided deposition to sinuous, single channel meandering deposition seems more characteristic of the Tertiary Georgia Basin.

The presence of lower alluvial fan conglomerate and sandstone facies in eastern (Canadian and U.S. Sumas Mountains), southeastern (Chuckanut Formation), northern (Kitsilano Member near Burrard Inlet), and western outcrop areas (Tumbo and Sucia islands) provides some control on the original extent of the basin. Paleocurrent patterns and conglomerate clast types all suggest local derivation from nearby margins present to the north and east for the mainland outcrop

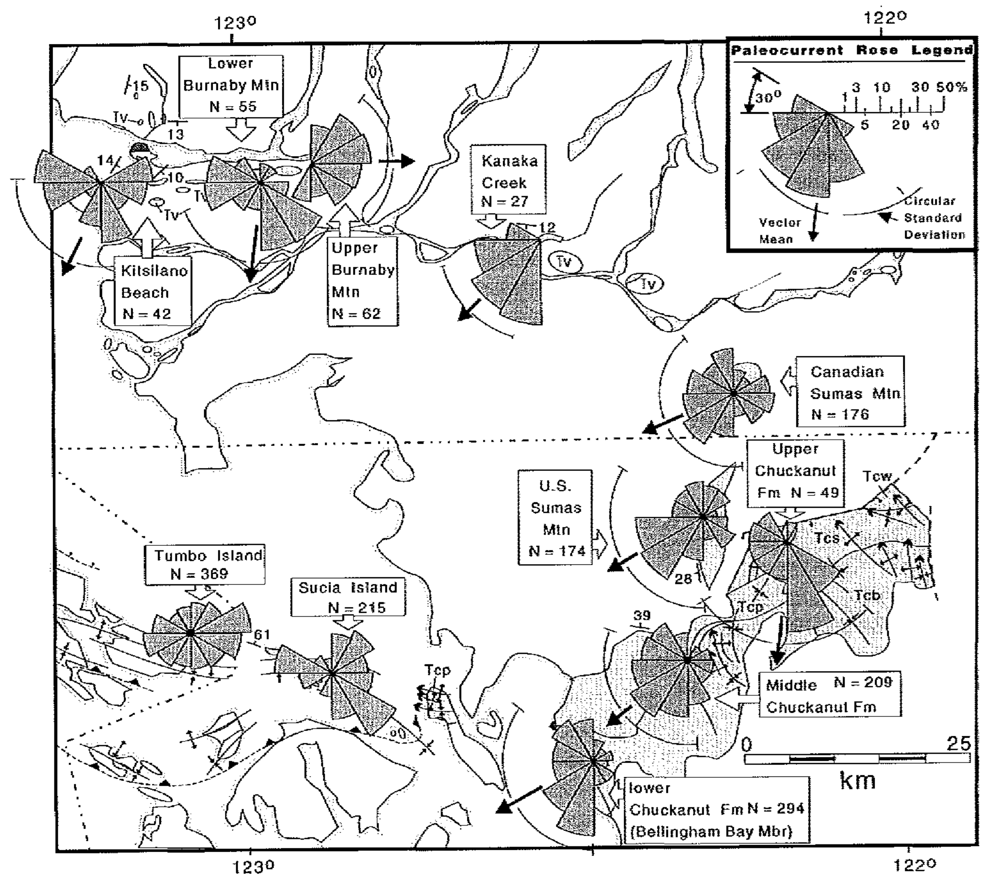


Figure 21. Summary paleocurrent diagram for Paleocene-Eocene part of Georgia Basin (see Fig. 17 for Lasqueti Island paleocurrents). Roses summarized all paleocurrents of the named area. Vector means and circular standard deviations are calculated using methods of Curray (1956) and Krause and Geijer (1987). Note that grouped paleocurrents of Tumbo and Sucia islands do not show a single statistically significant trend, probably reflect complex fluvial systems combined from over 500 m of stratigraphic section.

areas, and to the southwest through northwest for the Tumbo/Sucia Island areas (summarized in Fig. 21, 22). The lack of marine facies in any part of the study area, combined with the evidence for terrestrial deposition on most sides of the preserved basin suggests the basin was entirely intracontinental at least until the Miocene, when minor marine facies are present in the subsurface strata. This was first suggested by Johnson (1982) for the Chuckanut Formation. The Late Paleocene sandstone and conglomerate of Lasqueti Island are terrestrial and could have been deposited near the original northern margin of the basin. However, there is no direct evidence that the Paleocene outlier on Lasqueti Island was part of a single basin which included the other Paleogene deposits of this study. It is also possible that the Lasqueti Island strata are a remnant of Paleocene deposition not connected to the Vancouver-Bellingham area. Regardless of the inclusion or exclusion of the Lasqueti Island outlier, on at least three sides the Vancouver-Bellingham basin appears to have been intracontinental with separation of only a few tens of kilometres between these areas. There is no sedimentological or paleontological evidence for marine component to the basin before Miocene time. The main caveat to this conclusion is that there is no control on the southern margin to the Tertiary Georgia Basin.

Provenance evidence for Tertiary Georgia Basin is summarized in Figures 21, 22, 23A-B. Numerical summaries and sources of information for sandstone and conglomerate composition compilations are provided in Tables 2 and 3.

Detrital sandstone compositions for the basin are predominantly derived from studies of the Chuckanut Formation and the Tumbo/Sucia Island strata. These are shown on Figure 23A-B, as ternary diagrams where the detrital framework compositions have been subdivided into the grain populations recommended by Dickinson and Suczek (1979). The three main provenance categories distinguished by Dickinson and Suczek are continental blocks, magmatic arcs, and recycled orogens. These major groups are further subdivided depending on features such as the degree of uplift and dissection of arc terranes, compositional variation in recycled orogen

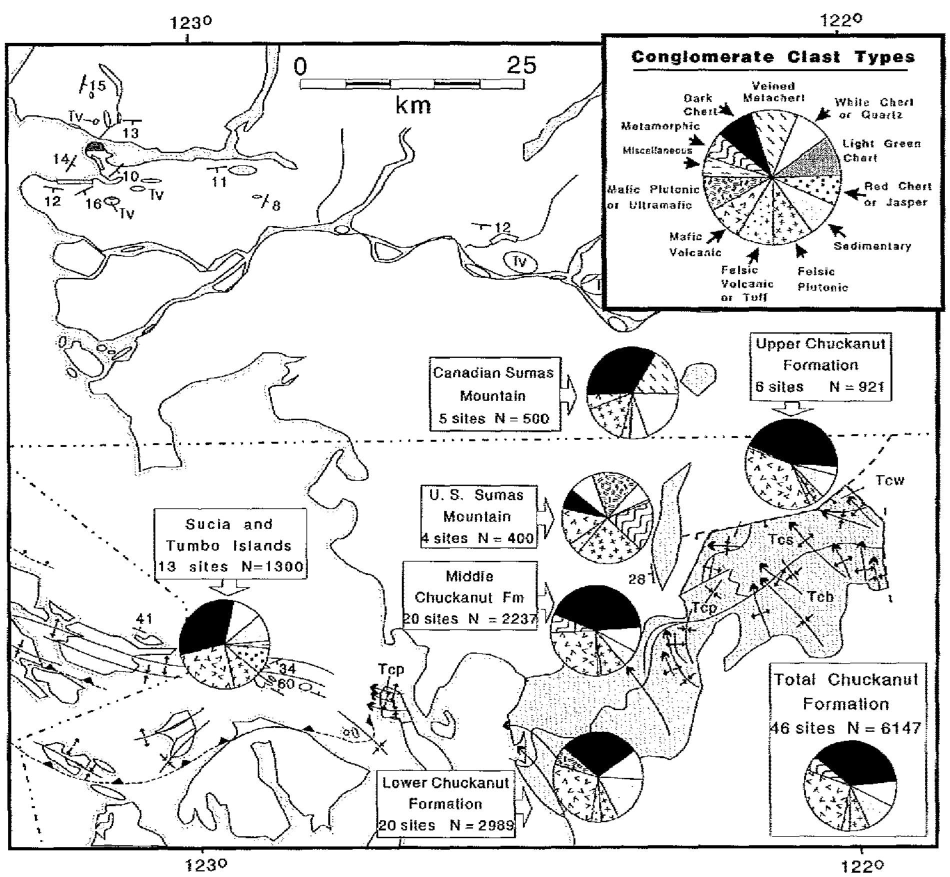


Figure 22. Summary conglomerate clast compositions for Paleocene-Eocene part of Georgia Basin (see Fig. 17 for Lasqueti Island conglomerate clast compositions). Pie diagrams summarize all conglomerate clast counts for a named area or part of formation with total Chuckanut Formation shown at lower right. Rock units are given in legend of Figure 3.

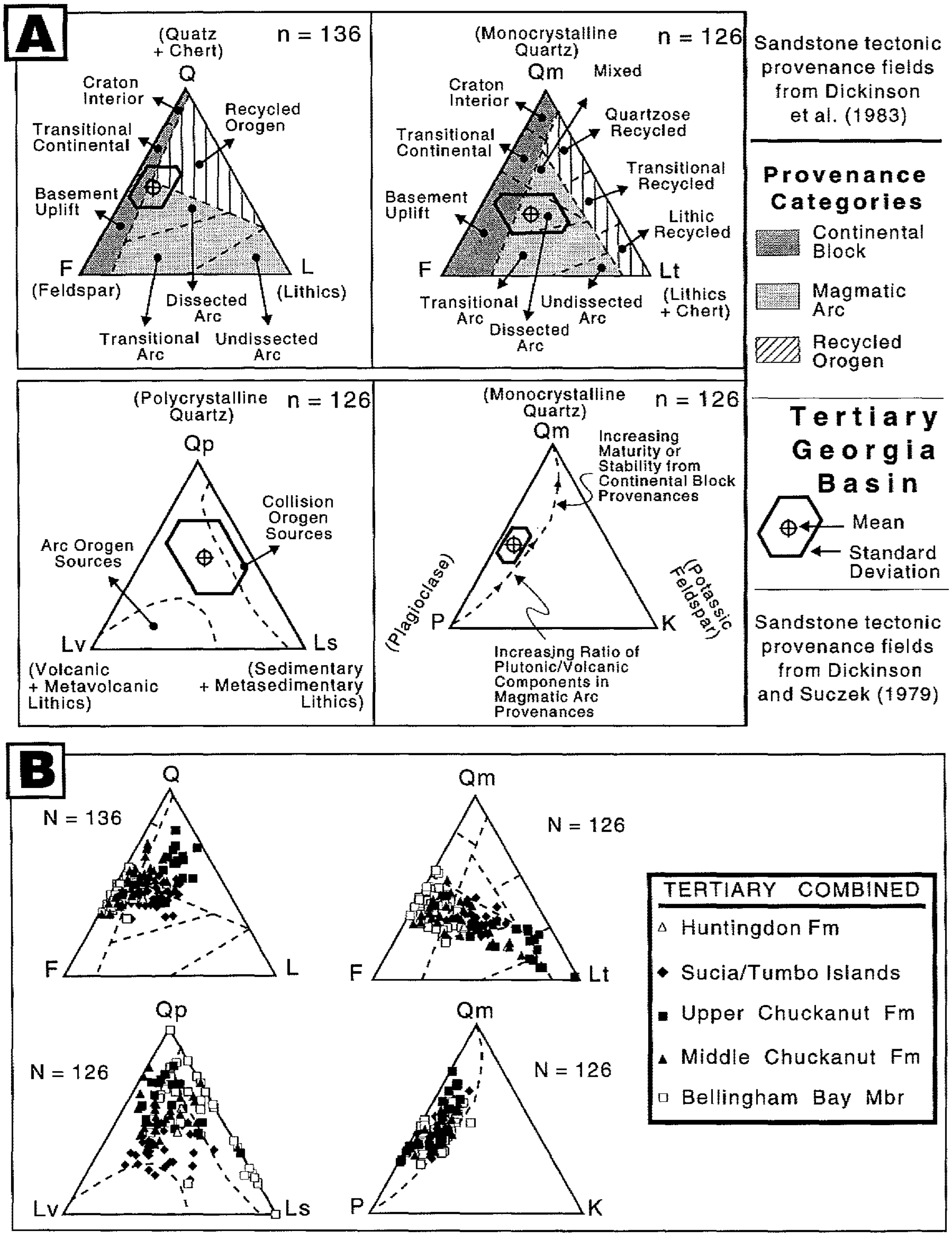


Figure 23. *A) Sandstone detrital framework compositions subdivided into grain populations and plotted on ternary diagrams showing the tectonic provenance fields of Dickinson et al. (1983, QFL and QmFLt plots) and Dickinson and Suczek (1979, QpLvLs and QmPK plots). The mean value of all Tertiary Georgia Basin is shown (crossed circle) with a polygon representing the standard deviation from the mean. Sources of information are summarized in Table 2. Note that for ten samples grain types were not identified in the original study to the level allowing plotting on any but the QFL plot, accounting for the change in total number of samples used between the QFL plot and the other plots.* ***B)*** *Individual point count values for entire Tertiary Georgia Basin shown split into major formation subdivisions (defined for Chuckanut Formation on Fig. 16D). Numerical summary and sources of information are provided in Table 2.*

terranes, and the importance of stable shield versus sedimentary cover sequences in continental block terranes (Dickinson and Suczek, 1979; Dickinson et al., 1983). Using these broad classifications, the subjacent and immediately adjacent source areas for the Tertiary Georgia Basin can be described as two different dissected arcs (Wrangellia terrane and the Coast Belt), and a complex recycled orogen (northwest Cascade/San Juan terranes). The Wrangellia terrane is distinctive for its high proportion of mafic to intermediate volcanic cover rocks (e.g., Bonanza Group and Karmutson formation), significant bodies of intermediate to mafic intrusions (Island Intrusives), and mixed sedimentary packages of the Sicker Group (see Monger, 1991a for a recent overview). The Coast Belt is dominated by granitic intrusions, but contains several pendants of volcanic successions suggesting that during Cretaceous and possibly early Tertiary time it had an extensive volcanic cover (Woodsworth and Monger, 1991). The northwest Cascades comprises a complex series of small and large accreted terranes dominated by oceanic sedimentary and metasedimentary rocks (mostly argillaceous to cherty successions), but with significant volcanic, ophiolitic, and intrusive slices, all deformed in mid- to Late Cretaceous contractional and transpressional events (Tabor et al., 1989).

The composition for the entire basin clearly reflects the variety of local basement lithologies, with the data spanning dissected arc, transitional continental, and collision orogen tectonic source fields (Fig. 23A-B, Q-F-L plot). An evolution is apparent for mainland Chuckanut Formation data where upper Chuckanut Formation sandstones are skewed towards the recycled orogen field. This probably reflects local uplifts on synsedimentary faults controlling late stages of sedimentation in these units, with local (lithic-rich) Cascade terrane sources providing the dominant detrital component (also suggested by Johnson, 1982). Lower members (especially Bellingham Bay and Slide members) of the Chuckanut Formation are considerably more arkosic than upper members. The traditionally interpreted sources for these lower units is

Table 2. Numerical summary of sandstone detrital clast compositions using modal classifications of Dickinson and Suczek (1979). Compiled from sources of information listed in right column.

		Tectonic Provenance Triangles Data												Sandstone Point Count Data							Sources of Information
		F	Q	L	F	Qm	Lt	Lv	Qp	Ls	P	Qm	K	Qm	Qp	P	K	Lv	Ls	Matrix	
Combined Tertiary	Mean (%)	41	47	12	41	33	26	21	49	29	46	45	9	33	14	34	7	7	6	4	all below
	Std Dev	16	11	9	14	10	21	16	20	22	10	9	5	11	15	13	5	7	4	6	
		N=136			N=126			N=126			N=126			N=126							
Upper Chuckanut Fm	Mean (%)	20	59	20	20	21	59	17	62	21	46	50	5	21	39	18	2	11	10	7	Johnson, 1982
	Std Dev	14	14	8	13	12	23	9	13	16	18	16	4	13	20	12	3	6	5	10	
		N=17 for all																			
Middle Chuckanut Fm	Mean (%)	39	49	12	41	32	26	26	53	21	47	44	9	32	14	34	7	8	5	5	Johnson, 1982 Kelly, 1970 Frizzel, 1979
	Std Dev	12	8	9	11	9	16	12	10	12	8	8	4	8	10	9	4	7	3	7	
		N=46			N=36			N=36			N=36			N=36							
Bellingham Bay Member Chuckanut Fm	Mean (%)	51	45	4	51	39	10	8	51	41	47	43	10	38	6	42	9	1	3	4	Johnson, 1982 Pongsapich, 1970 Frizzel, 1979
	Std Dev	8	6	4	6	9	8	11	26	29	9	8	5	7	6	9	5	2	2	4	
		N=45 for all												N=45 for all							
Sucia-Tumbo Islands	Mean (%)	37	43	21	36	31	33	36	35	29	43	47	10	32	11	30	7	12	9	2	Johnson, 1982 Pacht, 1980 Sturdavant, 1975
	Std Dev	14	10	6	9	5	9	12	8.8	11	8	8	4	8	4	11	3	5	4	2	
		N=26 for all												N=26 for all							

Table 3. Conglomerate clast composition summary statistics. All clast compositions measured by the senior author. Individual sites generally consist of 100 random clast identifications.

			Detailed Clast Types (%)											
Unit	Location		Dark Chert	veined meta chert	Red Chert/ Jasper	White Chert+ quartz	Lite Green Chert	Felsic Plutonic	Mafic Volcanic	Sedimentary	Dyke / Ultra-mafic	Felsic Volcanic	Meta-morphic	Tuff
Huntingdon Formation	Canadian Sumas Mtn	MEAN %	34	17	3	21		15	5	6				
		STD. DEV	3	5	1	3	0	12	2	2				
N = 500 5 sites														
Huntingdon Formation	Washington Sumas Mtn	MEAN %	7		1	9		24	13	9	17		18	4
		STD. DEV	8	0	2	10		7	5	6	19		15	8
N = 400 4 sites														
unnamed Paleocene	Lasqueti Isl	%	17			7		15	39	13	4	6		
N = 100 1 site														
unnamed Paleocene	Tumbo/ Sucia	MEAN %	26		8	5	11	8	21	7	7		3	4
		STD. DEV	13		4	3	5	6	6	5	5		3	2
N = 1300 13 sites														

high-grade metamorphic and plutonic rocks exposed to the east and northeast with some detritus from Coast Belt rocks to the north and northeast (Kelly, 1970; Hartwell, 1979; Johnson, 1982).

A recent isotopic provenance study by Heller et al. (1992) of Paleogene sandstone from several parts of western Washington State included samples from the Chuckanut Formation. Rubidium-strontium data from both whole rock and detrital white mica samples, and K-Ar, chemical, and stable isotope data from detrital white mica were analyzed. Two conclusions of Heller et al. (1992) are directly relevant to the provenance of Tertiary Georgia Basin sediments. First, there is a possible correlation of the continental Georgia Basin sandstone to marine sedimentary packages of the Olympic Mountains, suggesting both a much larger single basin than postulated in this study and a marine component for what we interpret as an intracontinental basin. Second, Heller et al. (1992) suggested a common source for the white mica from metamorphic and plutonic complexes of the southern Omineca Belt and that transport occurred in major rivers flowing from the east. Johnson (1982, 1985) proposed a similar source area for much of the Chuckanut Formation based on his detrital sandstone data and reconstructed depositional environment (particularly paleoflow patterns).

The early Tertiary transpressional regime and associated strike-slip faults of the Fraser-Straight Creek and other systems show a strong north to northwesterly structural trend (Fig. 24). The major river systems of Heller et al. (1992) would have to cut this trend almost at right angles, crossing several active strike-slip systems with associated components of dip-slip movement and northwest- to north-oriented elongate pull-apart basins. This drainage pattern is opposite to that typical of many strike-slip basins, where flow may be at high angles to master fault systems at the basin margins, but is reoriented to a direction roughly parallel to the main strike-slip trends (e.g., Nilsen and McLaughlin, 1985). Our study suggests that the Chuckanut Formation is part of the larger Tertiary Georgia Basin, which includes Paleogene strata of the sub-Fraser Delta area. Using this new information, the paleocurrent patterns for the entire basin display a southwest to south orientation on the east and north side of the outcrop areas (Fig. 21) rather than the west orientation suggested by Heller et al. (1992) or Johnson (1982, 1985). In addition, the nonmarine sedimentary rocks of the Tumbo and Sucia islands show complex paleocurrent patterns, which in association with the detrital sandstone and conglomerate clast compositions suggest derivation from local sources to the west and northwest. This finding is incompatible with either the eastern source interpretation of Heller et al. (1992) or linkage of Chuckanut Formation to Olympic Mountains marine strata west of the San Juan Islands. This requires a different source for the white mica, which is a distinctive component of some Chuckanut Formation sandstone units as well as of other west coast and Olympic Peninsula Paleogene sandstones. A source not considered by Heller et al. (1992) or Johnson (1982, 1985) is the Late Cretaceous Nanaimo Group where white mica is common, and in some places abundant. Dating of detrital zircons from sandstone in the top formation of the Nanaimo Group (Gabriola Formation), indicates derivation from the eastern Cordillera, including the southern Omineca Belt plutonic and metamorphic complexes (Mustard et al., 1994). Thus an abundant local source of white mica is present for the Tertiary succession, but these detrital micas should contain the isotopic signature of their original source in the eastern Cordillera. As documented in this study and Mustard (1994) the upper formations of the Nanaimo Group are missing at Vancouver and Lasqueti Island. In both places Late Paleocene sedimentary rocks unconformably overlie lower Nanaimo Group formations. It seems clear that a substantial thickness of white mica-bearing sandstone has been eroded from the top of the Nanaimo Group during the early Tertiary. The common presence of arkosic sandstone clasts identical to typical upper Nanaimo Group formations in the Paleocene conglomerates at Sucia and Tumbo islands provides strong evidence that the Nanaimo Group was cannibalized to contribute detritus to Tertiary Georgia Basin. We therefore suggest that this local source is more likely for the isotopically distinctive micas described by Heller et al. (1992), than a source requiring major river systems to cut laterally across the trend of active structures in central and western Washington State during the Paleogene. A logical test of this hypothesis would be a similar isotopic study of detrital mica from the sandstone of the upper Nanaimo Group.

TECTONIC SETTING

Engebretson et al. (1985) provided a summary of the known plate positions and convergence vectors for the Late Cretaceous and early Tertiary. They show a change in Late Cretaceous time (about 80 Ma) at the latitude of present Vancouver Island from a strongly convergent margin with the Farallon plate being subducted at a high angle beneath the North American plate to an increasing transpressive regime, including major dextral strike-slip faults, as the Kula plate was subducted obliquely to the north and northeast. Major Late Cretaceous to early Tertiary dextral strike-slip faults are common in southwest British Columbia and northwest Washington State (Fig. 24).

Several indirect lines of evidence suggest Tertiary Georgia Basin formation and deposition was controlled by strike-slip faulting (which also had major dip-slip offsets). The most complete detailed study of the Paleogene sedimentary rocks of Georgia Basin was that of Johnson (1982, 1984a, b, c, 1985, 1991), although restricted to the Washington State part of the basin (Chuckanut Formation). The estimated 6 km+ thickness of the Chuckanut Formation, which Johnson considered to have been deposited entirely in the Eocene, required abnormally high sedimentation rates and continuous uplift of marginal source rocks, both features typical of pull-apart basins (see reviews of Reading, 1980; Nilsen and McLaughlin, 1985). However, the more complete palynology dataset provided in this study indicates deposition spanned the Late Paleocene and Eocene, possibly into the Early Oligocene, perhaps 30 Ma instead of the 20 Ma estimated by Johnson (1982). In addition, the thickness of the Paleocene to Eocene succession in drillholes in the basin nowhere exceeds 2.5 km, considerably less than the 6 km+ accumulated total thickness suggested by Johnson (1982). In support of the thicker estimate in the southern area is the noted thickening of the succession to the south towards the Chuckanut Formation and the presence of several high angle faults mapped or shown in seismic data between the Chuckanut Formation outcrop area and the well sites used in this study. Depositional patterns and clast compositions documented for middle and upper members of the Chuckanut Formation by Johnson suggest that locally these faults were active during sedimentation and contributed significant volumes of sediment to the basin, and thus probably caused locally abnormally thick successions to be deposited. Johnson (1982, 1984a, b, c) considered these southwest-trending faults to form the northern boundary of the Chuckanut Formation and thus part of a basin separate from the early Tertiary sediments on the Canadian side of the international border. Our new palynological data, plus seismic

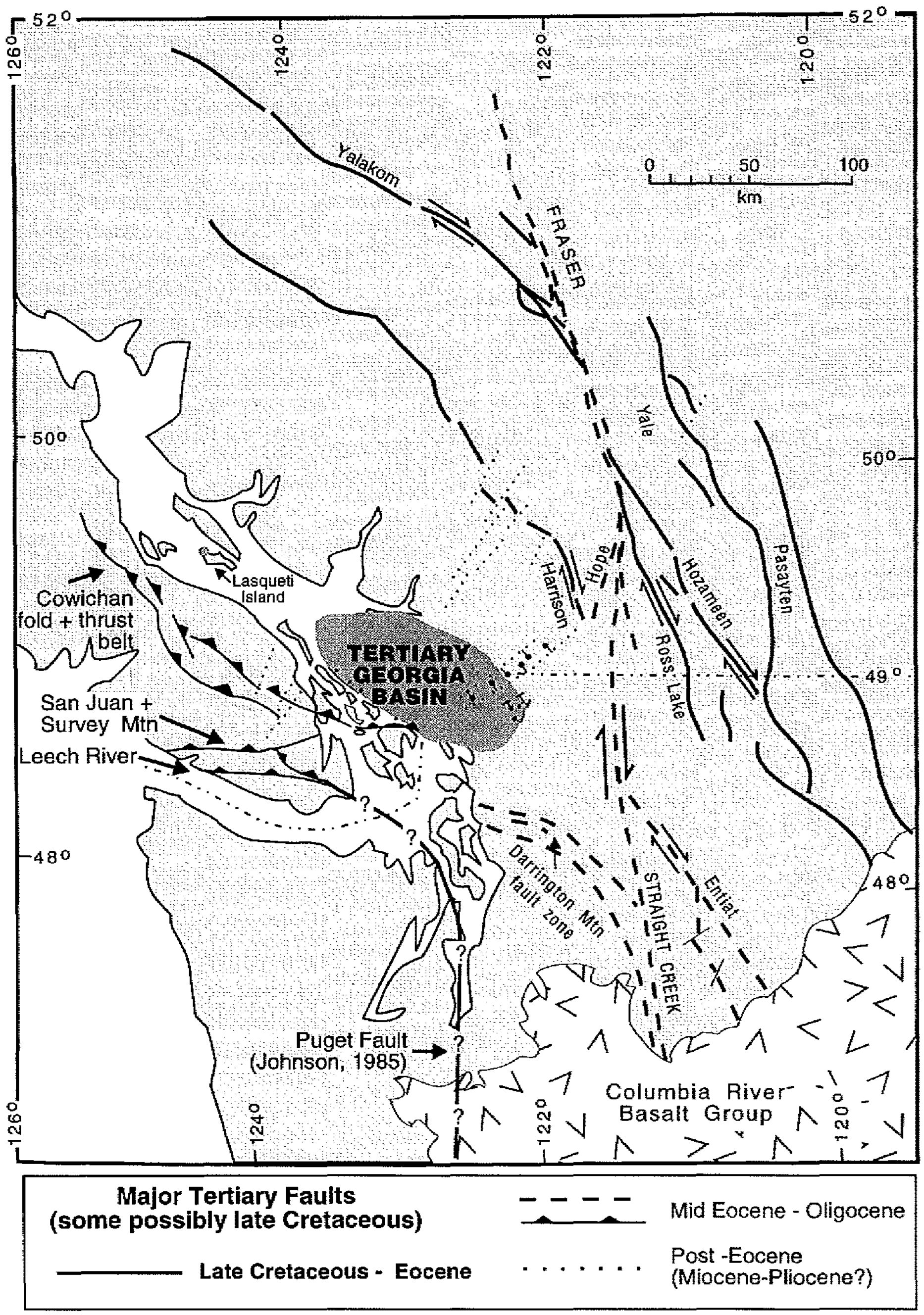


Figure 24. Major Tertiary faults of southwest British Columbia and northwest Washington State. Major sources of information are Johnson (1985), Tabor et al. (1989), England and Calon (1991), Wheeler and McFeely (1991), Monger and Journeay (1992), and unpublished seismic data used with permission of Conoco Exploration Ltd., Calgary, Alberta.

data and drilling subsequent to the study of Johnson (1982), demonstrate that the Huntingdon Formation in British Columbia is the same age as the Chuckanut Formation and thickens south beneath Quaternary and modern Fraser Delta-Nooksack Plain sediments to the Chuckanut Formation outcrop area. Therefore the Huntingdon Formation is laterally equivalent to the Chuckanut Formation rather than an unconformably overlying younger unit. Any interpretation of the basin depositional and tectonic setting must thus include both the Huntingdon and Chuckanut formations as part of one basin fill. We believe, however, that this is compatible with the pattern of basin fill and mapped faults along the northeast edge of the Chuckanut Formation and Johnson's interpretation that synsedimentary normal faults were active during some stages of the Paleogene sedimentation event. We suggest these high angle faults controlled minor subbasins during stages of the larger basin formation and fill, which encompassed the entire area of preserved Paleogene sediments (with the exception of the Lasqueti Island occurrence, possibly a separate deposit). Small grabens and half grabens at high angles to master faults are typical of pull apart basins (Christie-Blick and Biddle, 1985; Fig. 25B). This interpretation of a slightly larger early Tertiary basin does not conflict with the strike-slip basin model as proposed by Johnson and favoured here (Fig. 25A).

Johnson (1982, 1984a, b, c, 1985, 1991) proposed that the Fraser-Straight Creek fault system preserved in west central Washington State and British Columbia was the master dextral fault on the eastern side of the basin. These and other major early Tertiary fault systems of western Washington State and British Columbia are shown on Figure 24 (some possibly also active during the Late Cretaceous). Timing controls on dextral movements of the major faults are not entirely compatible with this system being related to the Paleocene-Eocene Georgia Basin formation or fill. As reviewed in Monger and Journeay (1992) movement on the Fraser fault system is determined to be post-47 Ma and pre-37 Ma by crosscutting plutons, suggesting at least this part of the system was not active during basin formation and the main period of deposition of the Tertiary Georgia Basin. However, indirect evidence of sedimentation related to Entiat and other

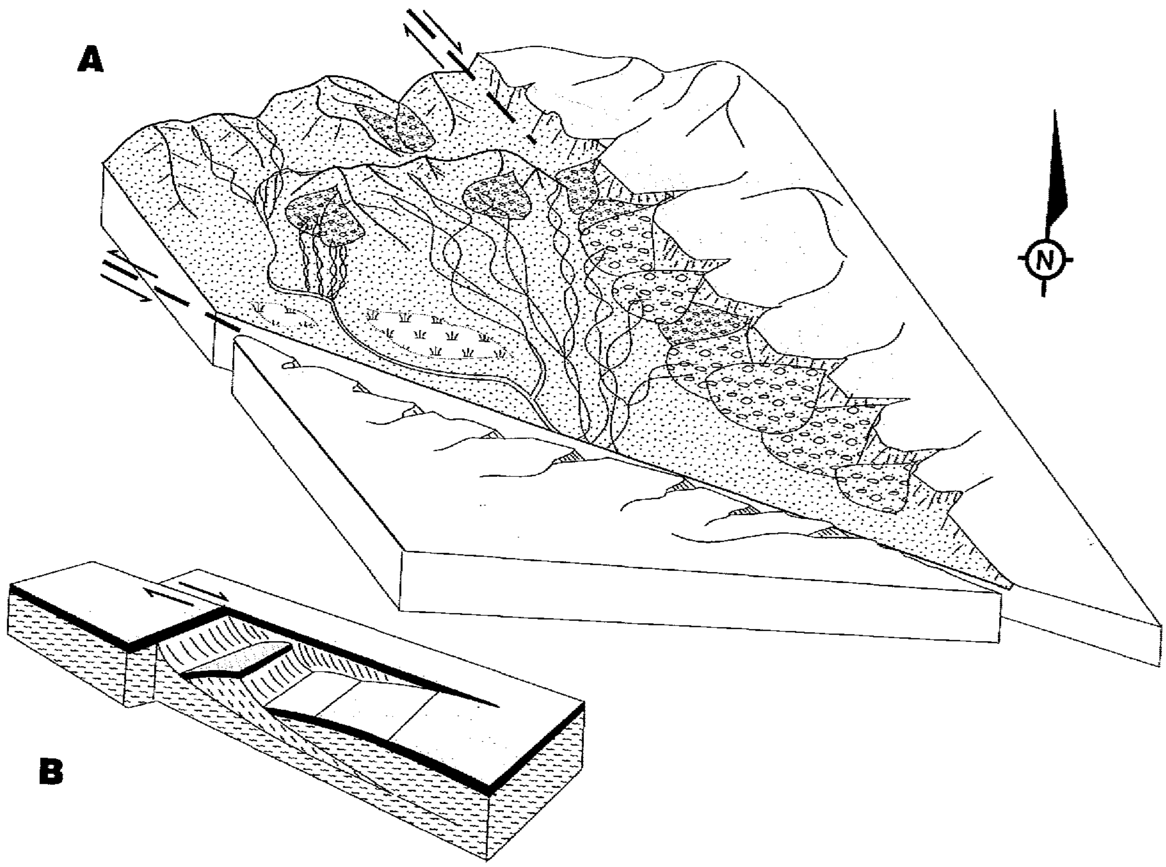


Figure 25. A) Schematic diagram of envisioned strike-slip basin setting for the main period of deposition of Tertiary Georgia Basin (Paleocene to Eocene or Early Oligocene). Dextral strike-slip master fault(s) on east side of basin show large dip-slip components of movement providing periodic rejuvenation uplifted eastern source for coarse clastic basin fill. Internal basin sources to north reflect synsedimentary minor normal faulting at high angles to main oblique-slip faults as shown in Figure 25B. Presence of major master faults on western boundary of basin is speculative; its western margin may have been bounded by minor fault zones showing mostly dip-slip senses of fault movement (a common feature of many large strike-slip basins, as discussed in text). ***B)*** *Schematic pull-apart basin model showing major dextral strike-slip fault and minor normal faults forming subbasins within the main pull-apart basin (modified from a similar diagram in Manspeizer, 1985).*

splays of the Straight Creek fault system in Washington State do suggest active strike-slip movement may have occurred locally in the early Eocene (reviewed in Johnson, 1985; Tabor et al., 1989). There are several northwest-trending dextral strike-slip faults cut by the Fraser-Straight Creek fault system which are more likely candidates for strike-slip faults directly related to the initiation and early filling of the Tertiary Georgia Basin. These include the Harrison, Yalakom, and Ross Lake faults and associated minor faults (Fig. 24). Monger (1986) reviewed evidence for dextral strike-slip movement of about 80-100 km on the Harrision fault and suggested timing of fault movement is Late Cretaceous to early Tertiary. Features of the Yalakom Fault are reviewed in Monger and Journeay (1992) with new evidence in Riddell et al. (1993) suggesting about 120 km of right lateral displacement during late Maastrichtian to middle Eocene time, (roughly 68 to 46 Ma). The complex Ross Lake fault zone of north-central Washington State and southwest British Columbia is generally considered an offset match across the younger Fraser-Straight Creek fault zone with the Yalakom system (e.g., Monger and Journeay, 1992). Tabor et al. (1989) provided a recent summary of the evidence for both dextral strike-slip (although amounts of displacement are unclear) and probable Paleocene to Middle Eocene timing of motion, although, as with the Yalakom fault system, timing is imprecise.

The apparent absence of obvious major strike-slip faults immediately west of the Tertiary Georgia Basin is not atypical of major strike-slip basins. As reviewed in Nilsen and McLaughlin (1985), well-documented strike-slip basins including the Miocene Ridge Basin of California and Devonian Hornelen Basin of Norway have a well-developed master fault system on one side of the basin, but the opposite side is bounded by minor faults, showing minor dip-slip offsets. This asymmetry of bounding structures is common for strike-slip basins (Reading, 1980). A major transcurrent fault system is thus not required on the western margin of the Tertiary Georgia Basin. Johnson (1985) inferred the presence of a major north-trending, dextral strike-slip structure buried below the Upper Eocene and younger deposits of the Puget lowland (Fig. 24), although evidence for such a structure is weak.

Deformation of the Chuckanut into north- to northwest-trending folds occurred during the Late Eocene. Miller and Misch (1963) and Johnson (1982, 1984b) suggested this compression event predated deposition of the sedimentary rocks of U.S. Sumas Mountain (correlated with the Huntingdon Formation by Miller and Misch, 1963). However, folds parallel to those of the Chuckanut Formation are present in the sedimentary rocks on U.S. Sumas Mountain and in light of the evidence that the Huntingdon Formation is laterally equivalent to the Chuckanut Formation, it seems likely that this deformation event affected both areas. England and Calon (1991) document compression effecting the Late Cretaceous Nanaimo Group on the Gulf Islands and eastern Vancouver Island. This compression caused both folding and southwest-directed thrusting of the Nanaimo Group strata with fault and fold trends generally northwest, parallel to the fold trends of the Chuckanut Formation. We suggest the major northwest-directed folds preserved in the Chuckanut Formation were also part of this event. Timing of the major contraction is probably Middle to Late Eocene or slightly younger, based on fission track ages from Nanaimo Group apatite (England and Massey, unpub. data, cited in England and Calon, 1991). Both Massey and Friday (1989) and England and Calon (1991) speculated that the contraction is connected to accretion of Pacific Rim and Crescent terranes to southern Vancouver Island during the Late Eocene as documented by Clowes et al. (1987).

HYDROCARBON POTENTIAL IN GEORGIA BASIN

Hydrocarbon exploration has occurred sporadically in the Georgia Basin since the first wildcat drilling in the 1890s. More than 125 exploration wells have been drilled in the basin, more than 75% in northwest Washington State (Whatcom County). The location of deeper wells are shown in Figure 3. None resulted in economically viable shows of oil or gas, but minor amounts of natural gas were reported in several. McFarland (1983) summarized the drilling activity in northwest Washington State prior to 1981, with early activity and potential reviewed in more detail by Glover (1935, 1936). Early drilling in the British Columbia Fraser Delta area is reviewed in Johnston (1923) and Hopkins (1966). Hopkins suggests that drill site selection prior to 1940 was generally based on little if any geological control (many appear to have been promotional ventures).

Gordy (1988) conducted a detailed examination of the hydrocarbon potential of the Georgia Basin. He provided summaries of previous well and seismic exploration in the basin, based both on geophysical and geological reports submitted to British Columbia and Washington State government agencies, and proprietary seismic and well information he was able to examine with the permission of private exploration firms. Most recent exploration activity has focused on the Tertiary strata of the Fraser Lowland and northwest Washington State (Whatcom County). Gordy (1988) considered potential to be very good for economically significant natural gas occurrences in the Tertiary strata. Porosity of Tertiary sandstones is generally good (estimated to average about 15%, but probably showing a wide range). Reservoir sandstones are abundant, and thick beds (up to 30 m) are present. However, the general lack of recognizable horizons in well and seismic data makes subsurface correlation of individual sandstone horizons or cycles extremely difficult. Source rocks include minor coal occurrences in the Chuckanut Formation. Both Gordy (1988) and Bustin (1990) concluded that organic matter in the basin are predominantly gas-generating types. This is confirmed by examination of both surface and subsurface samples and the regional sedimentological interpretations of this study.

The probability that Tertiary Georgia Basin was entirely continental during Paleocene and Eocene time has implications for hydrocarbon exploration. A lack of marine organic matter (alginite, the Type I and II kerogen of many source rock classifications, e.g., Tissot and Welte, 1984) reduces the likelihood of significant oil generation even if optimal thermal maturity was reached and migration into the abundant fluvial sandstone reservoirs occurred. Thus hydrocarbons in the Tertiary Georgia Basin are likely to be gas derived from terrestrial organic matter (mostly Type III kerogen). The one possible exception to this statement is provided in a recent summary of the petroleum geochemistry of Washington State by Lingley and von der Dick (1991). They suggested that some coals and shales in the 1988 AHEL Birch Bay well (well 11, Fig. 3) include Type IIB kerogen prone to light hydrocarbon production.

The generally low thermal maturity values from both previous studies (Bustin, 1990; Mustard and Rouse, 1991) and from this study also suggest the hydrocarbon potential of Tertiary Georgia Basin is restricted to gas. For this study measurements have been made on the TAI (Thermal Alteration Index) using the Chevron colour scale (0-4), which predicts the likelihood of hydrocarbon generation. On this scale, values of about 2.0-2.5 indicate paleotemperatures conducive to gas generation, mainly diagenetic dry gases (methane, CO_2, N_2), values from 2.5 to about 3.2 denote the "oil window", and values about 3.0-3.3 represent the zone for optimum generation of wet gases (condensate).

The TAI from both surface and subsurface samples indicate only marginal maturation levels. This includes samples from the Albian sediments at Blue Mountain (described in Mustard and Rouse, 1991) and at depths of over 3300 m in both the Sunnyside and Point Roberts exploration wells (wells 7 and 8 in Fig. 3). Hence, the prognosis for substantial hydrocarbon generation is not very favourable. This is corroborated by assessment of vitrinite reflectance values by R.M. Bustin (U.B.C., pers. comm., 1990). However, Walsh and Phillips (1983) showed that the rank of Eocene coals in the Chuckanut Formation increases significantly and systematically towards post-Eocene igneous intrusives to the east. The small intrusive bodies and sills present in the Tertiary Georgia Basin could similarly provide localized areas of higher thermal maturity than suggested from the regional data.

Evaluation of subsurface structure and trap potential for Tertiary strata has been hampered by the poor quality of seismic data available. Gordy (1988) evaluated publicly available and proprietary seismic data. Based on this data, well information, and surface geology, he interpreted the important Tertiary structures as high angle reverse and normal faults, trending both easterly and northwesterly. Mapping of structures in the Chuckanut Formation by Miller and Misch (1963) and Johnson (1982) demonstrate that northwest-trending and-plunging folds are common. These are cut by a younger set of northeast- or east-trending faults including some Johnson suggested were active during Chuckanut Formation deposition.

Seismic and gravity data more recent than that examined by Gordy tends to confirm his overall interpretations of subsurface structures. Canadian Hunter carried out an extensive seismic survey prior to drilling a 2900 m hole in 1988 at Birch Bay (well site 10 on Fig. 2). Conoco Ltd., in partnership with Dynamic Oil Co., also obtained recent seismic and gravity data, from which they have identified three structures, one drilled during 1990 (Conoco-Dynamic Mud Bay, well site 5 on Fig. 2) and two in 1993 (well sites 22 and 23 on Fig. 3). Examination of some of the seismic and gravity data by the senior author confirms the two major trends of structures. An older trend of northwestern compressional structures (folds and minor faults) is probably a continuation of the surface features mapped by Johnson (1982) in the Chuckanut Formation to the southeast and regionally part of the compressional event of probable late Eocene age documented by England and Calon (1991) which deformed the Nanaimo Group into a fold and thrust belt. This trend is cut by a younger set of structures, mostly high angle normal and reverse faults which generally have apparent dip-slip offsets of only a few hundred metres and appear to be confined to the middle Tertiary and older strata (pre-Pliocene?). The best example of these faults occurs in the Sumas Prairie area east of Abbotsford (Fig. 3) where high angle normal faults occur on the eastern side of Sumas Mountain (Sumas Mountain Fault) and western side of Vedder Mountain (Vedder Fault), causing a graben structure now partly infilled by Quaternary and recent sediments of the Sumas Prairie (shown schematically in Fig. 15). The Sumas Mountain fault is well-imaged on recent seismic data with an apparent dip-slip downdrop to the southeast of several hundred metres. The trend of these faults parallels linear features Monger (1990) mapped in the Coast Mountains immediately northeast of the Fraser Lowlands. He postulated a mid-Tertiary age for these inferred faults.

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APPENDIX

The following are formal descriptions of one new genus (*Varisulcosporites*) and eleven new species of fungal palynomorphs recognized in this study. The names, citations, and descriptions follow the catalogue "Genera File of Fossil Spores and Pollen" by Jansonius and Hills (1976, 1979), which includes fungal spores. The photographs cited in the descriptions refer to those contained in the plates in the text. In most cases, the magnification of photographs is x1000. Where different, the magnification is noted in the figure legends.

It is apparent from this study that fungal palynomorphs are a highly useful tool in dating and correlation, particularly in Paleogene successions throughout western North America. Because of the warm temperate to subtropical conditions that prevailed throughout much of the Paleogene, it is also apparent that modern counterparts of Paleogene fungal spores, fruiting bodies, and mycelial remnants should be looked for in extant myco-assemblages from the Gulf Coast - Caribbean region, especially in deltaic, shoreline, and low floodplain depositional sites.

Genus *Intratriporopollenites* Thomson and Pflug, 1953

Intratriporopollenites precrassipites sp. nov.

Plate 3, fig. 10, 11

Holotype: Pl. 3, fig. 10, 3rd Beach, Sect. 4, S-12, slide 3, Leitz 21.8; 114.8. Huntingdon Formation, Late Paleocene.

Diagnosis: Amb essentially circular, brevitricolporate, with apertures equatorial, colpi short and wide, with round-angular ends; vestibulum present but weakly developed; annuli expressed only in the outer flanks of the colpus, forming subtriangular pouches in optical section. Sculpture reticulate, with muri clavate in section, lumina <1 µm, essentially uniform in size from apocolpia to margin.

Dimensions: Range in diameter 33-42 µm; holotype 40 µm.

Age range: Late Paleocene

Remarks: The specific epithet "*precrassipites*" alludes to the similarity to *Tilia crassipites* Wodehouse (see Plate 7, fig. 10) that first occurs in the Early Eocene and persists into the Miocene. The chief difference is the weakly developed annulus in *I. precrassipites* relative to the heavy annular collar in *Tilia crassipites*.

This species differs from previously described *Intratriporopollenites* mainly in the weakly developed annulus, and from *I. prevescipites* sp. nov. by its larger size.

Intratriporopollenites precrassipites sp. nov.

Plate 3, fig. 12, 13

Holotype: Pl. 3, fig. 12, 3rd Beach, Sect. 4, S-12, slide 3, Leitz 21.8; 114.8. Huntingdon Formation, Late Paleocene.

Diagnosis: Amb essentially spherical; brevitricolporate, with apertures equatorial; colpi short with rounded ends; vestibulum present but weakly developed; annuli weak, and localized to outer flanks of colpi; sculpture reticulate, with muri reduced-clavate in section; lumina <1 mm, essentially same diameter from apocolpia to perimeter.

Dimensions: Range in diameter 24-29 mm; holotype 25 mm.

Age range: Late Paleocene.

Remarks: I. prevescipites sp. nov. alludes to the similarity to Tilia vescipites Wodehouse (Pl. 7, fig. 11) that ranges from the Early Eocene-Miocene. It differs mainly by having very local development of annulus flanking the outer (mostly) colpal borders. Together with I. precrassipites sp. nov. this species is a good index palynomorph for the Late Paleocene in western and northern North America, where it occurs with Pistillipollenites mcgregorii (Pl. 3, fig. 15, 16), Subtriporopollenites - A (Pl. 3, fig. 21) and Tricolpites - A (Pl. 3, fig. 22). See also Rouse (1977).

Genus *Involutisporonites* Clarke, 1965

Involutisporonites minutus sp. nov.

Plate 4, fig. 6

Holotype: Third Beach, Stanley Park, Zone S-12, Susan Taite, Leitz 37.4; 121.1 G.G. 25, fr. 14.

Diagnosis: Planispiral fungal spores, with few (usually 4) cells encircling a fusiform open area in the centre; outer walls about 0.25 mm thick and leviagate; inner walls of cells about 1.0-1.5 mm; septa about 0.75 mm thick, with a single faint pore.

Dimensions: Range 16-21 mm.

Age: Late Paleocene.

Remarks: This spore is smaller than the type species *I. foraminis*, with fewer and nonlobate cells. It is usually found on fragments of leaf, and hence is likely the spore of an epiphytic fungus.

Genus *Fusiformisporites* Rouse emend Elsik, 1968

Type Species: *Fusiformisporites crabbii* Rouse, 1962

Fusiformisporites paucistriatus sp. nov.

Plate 4, fig. 10, 11

Holotype: Pl. 4, fig. 11, Burrard G.C.-3, Third Beach, Stanley Park, Leitz 3-50.1; 124.0 Neg. G.G.-12, fr. 35.

Diagnosis: Fungal spores dicellate, inaperturate, with striate wall; wall levigate, 0.25-0.5 µm, thickening to 1.5-2.0 µm at apices. Striae mostly very thin and irregular in length and width; occasionally flaring, concentrated in the mid-sections of the wall in both cells. Striae variable in number from 3-10 in each hemisphere.

Dimensions: range in length 39-42 µm; in width 15-22 µm.

Age: Late Paleocene.

Remarks: This species is distinguishable from others by the generally low number of weak, thin, and often short striae concentrated in the mid-section of each hemisphere.

Fusiformisporites lineatus sp. nov.

Plate 13, fig. 1-3

Holotype: Plate 13, fig. 1. Kitsilano, 6th Avenue, No. 1, slide 3, Leitz 3-21.1; 117.2; Neg. G.G. -26, fr. 17.

Diagnosis: Fusiform fungal spores, dicellate, inaperturate, with striate levigate wall 0.75-1.0 µm, uniform in thickness, deep melanin brown; striations varying from prominent and running from one pole to the other, to smaller, shorter, and less distinct extending between the middle septum and either pole; full-length and prominent striae in some specimens extend through polar wall to exterior; polar cap broken away from pole in a few cases (Pl. 13, fig. 1). There are occasional branches in some striae, both prominent (Pl. 13, fig. 1, 2) and weak (Pl. 13, fig. 1).

Dimensions: range of length 58-62 µm; of width 23-29 µm.

Age: Late Eocene - Early Oligocene, Kitsilano Member of Huntingdon Formation

Remarks: This is the largest species of *Fusiformisporites* yet detected in the Tertiary and is a good index fossil for the Late Eocene – Early Oligocene of the western coastal deposits of that age in North America.

Genus *Multicellaesporites* Elsik, 1968, emend

Sheffy and Dilcher, 1971

Multicellaesporites acuminatus sp. nov.

Plate 4, fig. 14, 15

Holotype:- Plate 4, fig. 15. Third Beach, *Pistilli* beds, SKS-5, Leitz coordinate: 37.2; 114.8; Neg. 25, fr. 23.

Diagnosis: Fusiform fungal spores, consisting of 5-6 thin septa in each half; each septum with a small central pore; septa supporting an inner membranous body that is closely appressed to the outer wall in central regions, but contracted away from the outer wall towards the 2 pointed extremities; wrinkles occur sporadically on the thin inner wall that appear as elongate irregular plicae.

Dimensions: range of length: 62-68 µm; of diameter: 17-25 µm.

Age Range: Late Paleocene.

Remarks: Differs from *M. compactilis* Ke et Shi ex Sung et al., in fusiform shape, thinner septa, and inner body with elongate plicae.

Multicellaesporites bilobus sp. nov.

Holotype: Plate 4, fig. 16 Kanaka Creek (1) GER composite (1) Slide ii Left, 42.9; 125.2, neg. G.G.-1 fr. 17.

Diagnosis: Amb elongo-lanceolate; 4-chambered formed by 3 septa, each with a well-defined central pore, wall invaginated inwards around the middle septum forming 2 lobes; shallow trenches encircling the outer wall at each septum; apices moderately acute and thickened to form rounded points; sculpture levigate.

Dimensions: range of length 69-75 µm; of width 18-25 µm; holotype 74 x 28 µm.

Age: Late Paleocene.

Remarks: This species resembles superficially *M. granulatus* Ke et Shi ex Sung et al., 1978 (Pl. 2, fig. 20) in number of chambers (4) and overall length, but differs in being levigate, consisting of 2 distinct lobes of 2 chambers each, and in having thickened pointed apices.

Genus *Diporicellaesporites* Elsik, 1968

Diporicellaesporites extendus sp. nov.

Plate 5, fig. 7

Holotype: Third Beach, Stanley Park, zone 5-12, slide KMP-4: Leitz 4-18.9; 117.1. Neg. G.G. 25, fr. 1.

Diagnosis: Fungal spores elongate fusiform; diporate, with circular pore at each apex; 10 septa, each with a central pore; wall dark brown and levigate, about 0.5 µm thick; septa also about 0.5 µm thick. Apical cells small and triangular, with circular pore and walls extending to form a point.

Dimensions: range of length 64-68 µm; of width 28-30 µm.

Age: Late Paleocene.

Remarks: This species differs from *D. laevigaeformis* Ke et Shi ex Sung et al., 1978 in a more pronounced and elongate fusiform shape, 10 septa vs. 3, and triangular protruding pores.

Diporicellaesporites acutus

Plate 8, fig. 24, 25

Holotype: Plate 8, fig. 25. Second Beach Stanley Park, zone S-16, sample 2B-6, slide 1, Leitz 47.2; 119.5. Neg. G.G. 18, fr. 32.

Diagnosis: Diporate, dicellate-tricellate fungal spores; fusiform; outer wall levigate, about 0.5 µm; pores apical and sharp-ended, each pore forming an inward-directed V notch of about 2-4 µm, wall surrounding notch thickened, forming a diffuse margo; each septum about 1.5 µm thick, with a distinct central pore.

Dimensions: range of length 31-42 µm; of width 12-16 µm.

Age: Late mid-Eocene - Early late Eocene; *Engelhardtia-Castanea* zone.

Diporicellaesporites quadratus sp. nov.

Plate 9, fig. 5, 6

Holotype: Plate 9, fig. 6. SFU, Kitsilano, slide 1, Leitz 43.5; 121.7. Neg. G.G.-19, fr. 22.

Diagnosis: Rectangular fungal spores, diporate, triseptate. Pores apical, subtended by septa with central pores; large central septum about 1 µm thick, with central pore.

Dimensions: range of length 20-22 µm; of width 11-12 µm.

Age: Late mid-Eocene-Early late Eocene, *Engelhardtia - Castanea* zone.

Remarks: Characteristic features are the rectangular amb, small size, and subapical septa.

Diporticellaesporites segmentus sp. nov.

Plate 12, fig. 2-4

Holotype: Plate 12, fig. 2. Kitsilano, 6th Avenue. No. 1, slide KMP-3, Leitz 35.8; 127.9. Neg. G.G. 27, fr. 16.

Diagnosis: Elongate oblong fungal spores; diporate; polyseplate, with (usually) 10 septa, occasionally 9 or 11, each septum dentate with a central pore; septal teeth in each half of the spore generally pointing inwards toward the central septum that is noticeably less dentate than the others. Outer spore wall thin, about 0.25 µm; septal walls 1.0-1.25 µm thick; 2 apical apertures consisting of a combined silt and atriate pore.

Dimensions: range of length 30-40 µm; of width 14-17 µm.

Age: Late Eocene - Early Oligocene, Kitsilano Member of the Huntingdon Formation.

Remarks: *D. segmentus* is a good index fossil for the Late Eocene - Early Oligocene of coastal western Canada.

Genus Varisulcosporites gen. nov.

Type species: *Varisulcosporites eminens* sp. nov.

Generic diagnosis: Unicellate, elliptical to fusiform fungal spores, isopolar, with thickened polar caps including polar walls (Pl. 11, fig. 5, 6). Apertures single, in form of an elliptical oblique sulcus, or a sulcus plus 2 elongate grooves extending between sulcus and poles.

Varisulcosporites eminens sp. nov.

Plate 11, fig. 5, 6

Holotype: Plate 11, fig. 6. Kitsilano, 6th Avenue, No. 1, slide KMP -4, Leitz 22.2; 123.0. Neg. G.G. 21, fr. 16.

Diagnosis and description: Unicellate, elliptical fungal spores. Wall at apices thickened to form apical caps. Spore wall levigate, and dark melanin brown, about 0.5 µm in equatorial regions, thickened to about 1.25 µm at the poles and extended as a thin groove towards both poles, plus second thin groove on the opposite side, also extending to the poles, with the 2 grooves and possibly connected at the poles.

Dimensions: range of length 22-46 µm; of width 18-20 µm.

Remarks: In a few specimens (Pl. 11, fig. 5) the aperture consists of only an oblique sulcus with no extending grooves.

Genus *Pluricellaesporites* van der Hammen, 1954,

emend Clarke, 1965, Elsik, 1968, Sheffy and Dilcher, 1971

Pluricellaesporites magnus sp. nov.

Plate 11, fig. 11

Holotype: Plate 11, fig. 11. Kitsilano, 6th Avenue. No. 1. slide KMP (3), Leitz 3-22.4; 123.0, Neg. 26, frs. 21 & 22 (spliced).

Diagnosis: Monoporate, multiseptate (7-9) with septa dentate and a central pore; spore tapering from wide base to a pointed apex; wall about 1 µm thick, thinning to about 0.5 µm at apex; septal walls about 1.0-1.5 µm thick; spore colour.

Dimensions: range of length 138-185 µm; of width 40-50 µm.

Age: Found in low numbers in Kitsilano Member, Late Eocene-Early Oligocene of the Huntingdon Formation. Apparently limited to sediments of this age, it could occur in younger time zones.

Plate 1

Characteristic palynomorphs from rocks of the Nanaimo Group on Burrard Inlet. Magnification x1000.

Figure 1. *Appendicisporites cf. unicus* (Markova) Singh
Figure 2. *Cicatricosisporites striosporites* Ross
Figure 3. *Ornamentifera baculata* Singh
Figure 4. *Gleicheniidites senonicus* Rouse
Figure 5. *Klukisporites areolatus* Singh
Figure 6. *Cicatricosisporites mohrioides* Delcourt and Sprumont
Figure 7. *Vitreisporites pallidus* (Reissinger) Nilsson
Figure 8. *Vitreisporites* sp.
Figure 9. *Cyathidites minor* Couper
Figure 10. *Deltoidospora microforma* Rouse
Figure 11. *Proteacidites thalmanni* Anderson
Figure 12. *Proteacidites thalmanni* Anderson
Figure 13. *Proteacidites marginus* Rouse
Figure 14. *Tricolpites divergens* Rouse
Figure 15. *Tricolpites divergens* Rouse
Figure 16. *Tricolpites divergens* Rouse
Figure 17. *Quercoidites microhenrica* (Potonié)
Figure 18. *Nudopollis* sp.
Figure 19. *Fraxinoipollenites linguapollenites* Rouse
Figure 20. *Fraxinoipollenites variabilis* Stanley
Figure 21. *Duplopollis carlquistii* Drugg
Figure 22. *Tricolpites compactus* (Norton) Farabee and Canright

Plate 1

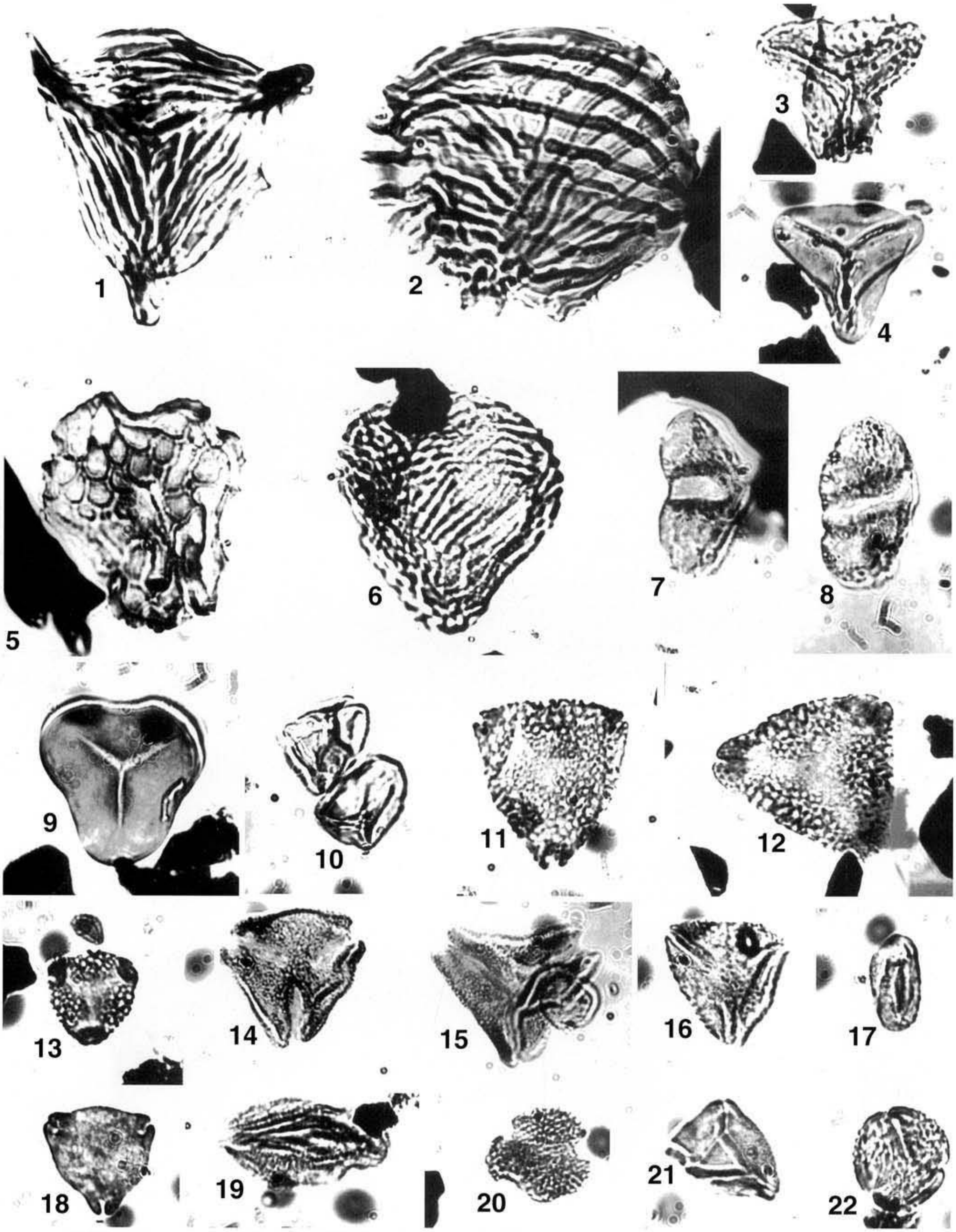


Plate 2

Late Paleocene palynomorphs from the lower Huntingdon Formation at Third Beach, Stanley Park; Kanaka Creek; and Canadian Sumas Mountain region. All figures x1000 except Figure 4 at x500.

Fern Spores:

Figure 1. *Laevigatosporites ovatus* Wilson and Webster

Figure 2. Polypodiaceae - forma 3. Martin and Rouse

Figure 3. *Cicatricosisporites dorogensis* Potonié and Gelletich

Conifer pollen:

Figure 4. *Picea grandivescipites* Wodehouse

Figure 5. *Pinus* sp. - white pine group

Figure 6. *Pinus* sp. - yellow pine group

Figure 7. *Cupressacites hiatipites* Krutzsch

Angiosperm pollen:

Figure 8. *Ericipites* sp. *cf. E. ericius*

Algal Cyst:

Figure 9. *cf. Paralecaniella indentata*

Plate 2

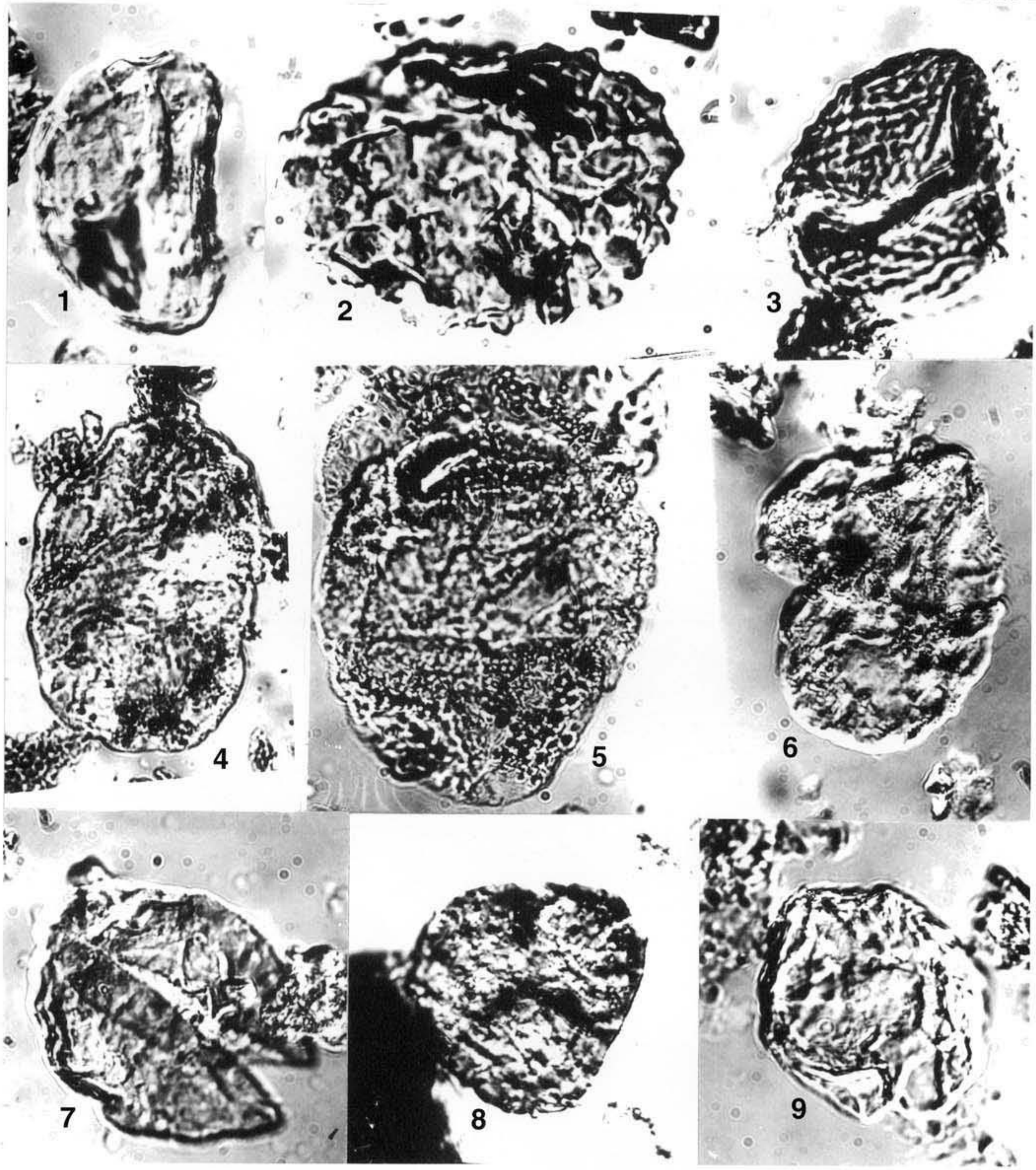


Plate 3

Late Paleocene palynomorphs from the lower Huntingdon Formation at the Vancouver, Kanaka Creek and B.C. Sumas Mountain regions. Magnificaton x1000.

Angiosperm pollen: - (all figures)

Figures 1 & 2. *Triporopollenites mullensis* (Simpson) Rouse and Srivastava

Figure 3. *Myricipites dubius* Wodehouse

Figure 4. *Fraxinoipollenites linguapollenites*

Figure 5. *Paraalnipollenites alterniporus* (Simpson) Srivastava

Figures 6 & 7. *Fraxinoipollenites variabilis* Stanley

Figure 8. *Quercoidites* - A. Rouse

Figure 9. *Quercoidites microhenrica* (Potonié) Potonié

Figures 10 & 11. *Intratriporopollenites precrassipites* (sp. nov.)

Figures 12 & 13. *Intratriporopollenites prevescipites* (sp. nov.)

Figure 14. *Platycaryapollenites* sp.

Figures 15 & 16. *Pistillipollenites mcgregorii* Rouse

Figure 17. *Tricolpites reticulatus*

Figures 18 & 19. *Duplopollis* - B. Drugg

Figure 20. *Kurtzipites* sp.

Figure 21. *Subtriporopollenites* - A. Rouse

Figure 22. *Tricolpites* - A. Rouse

Plate 3

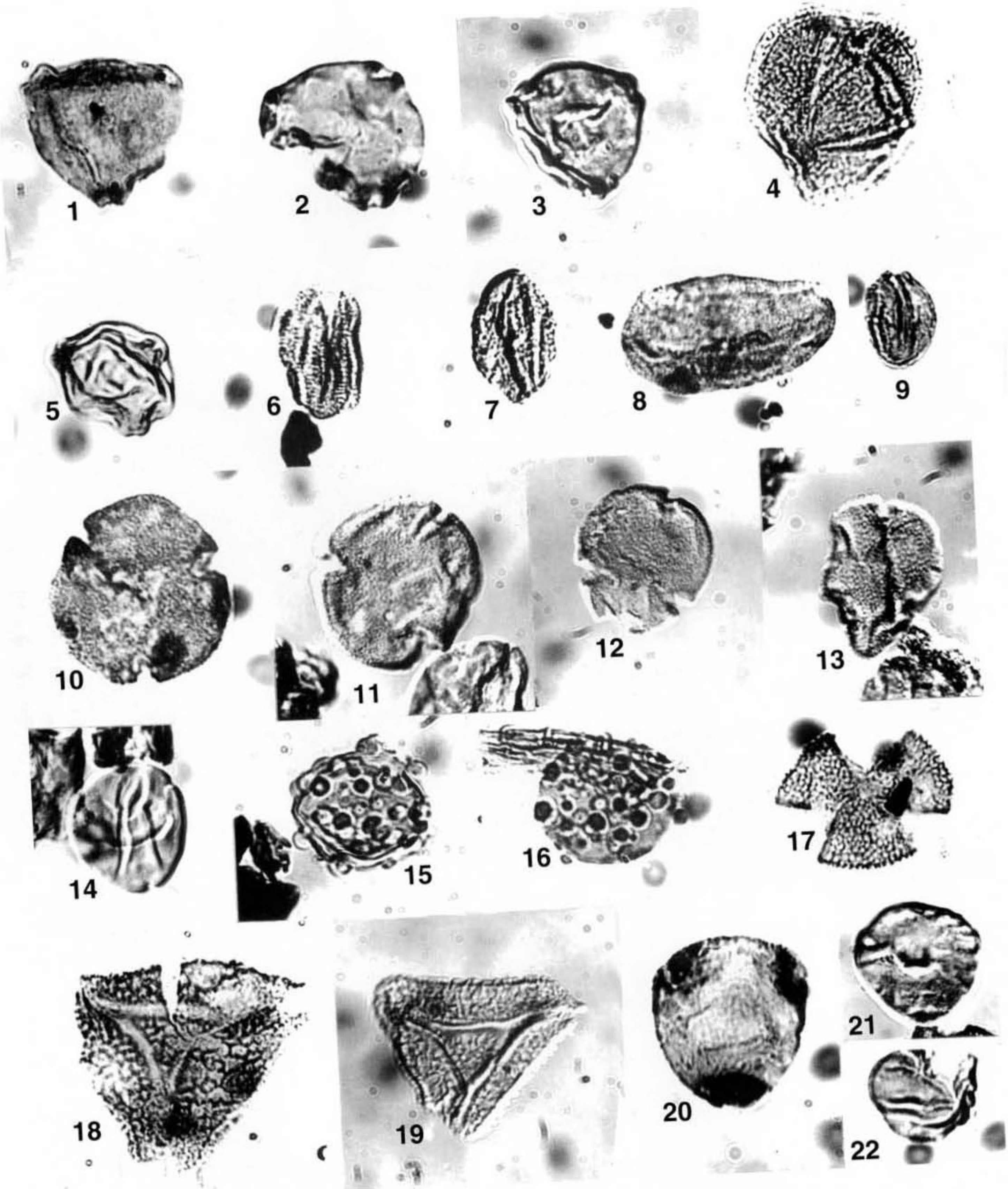


Plate 4

Late Paleocene palynomorphs (fungal spores) from the lower Huntingdon Formation at Third Beach, Stanley Park, Kanaka Creek, and B.C. Sumas Mountain region. Magnification x1000.

Fungal spores: - (all figures)

Figure 1. *Inapertisporites elongatus* Rouse, 1962

Figures 2 & 3. *I. globulosus* Rouse, 1962

Figure 4. *Didymosporisporonites ovatus* Ke et Shi ex Sung et al., 1978

Figure 5. *Pluicellaesporites* cf. *simplicissimus*

Figure 6. *Involutisporonites minutus* sp. nov.

Figure 7. *Dicellaesporites obnixus* Norris, 1986

Figures 8 & 9. *Fusiformisporites striatus* (Ke et Shi ex Sung et al.) comb. nov.

Figures 10 & 11. *F. paucistriatus* sp. nov.

Figure 12. *Pluricellaesporites hillsii* Elsik, 1968

Figure 13. *Multicellaesporites ellipticus* Sheffy and Dilcher, 1971

Figures 14 & 15. *M. acuminatus* sp. nov.

Figure 16. *M. bilobus* sp. nov.

Figures 17 & 18. *M. compactilis* Ke et Shi ex Sung et al., 1978

Plate 4

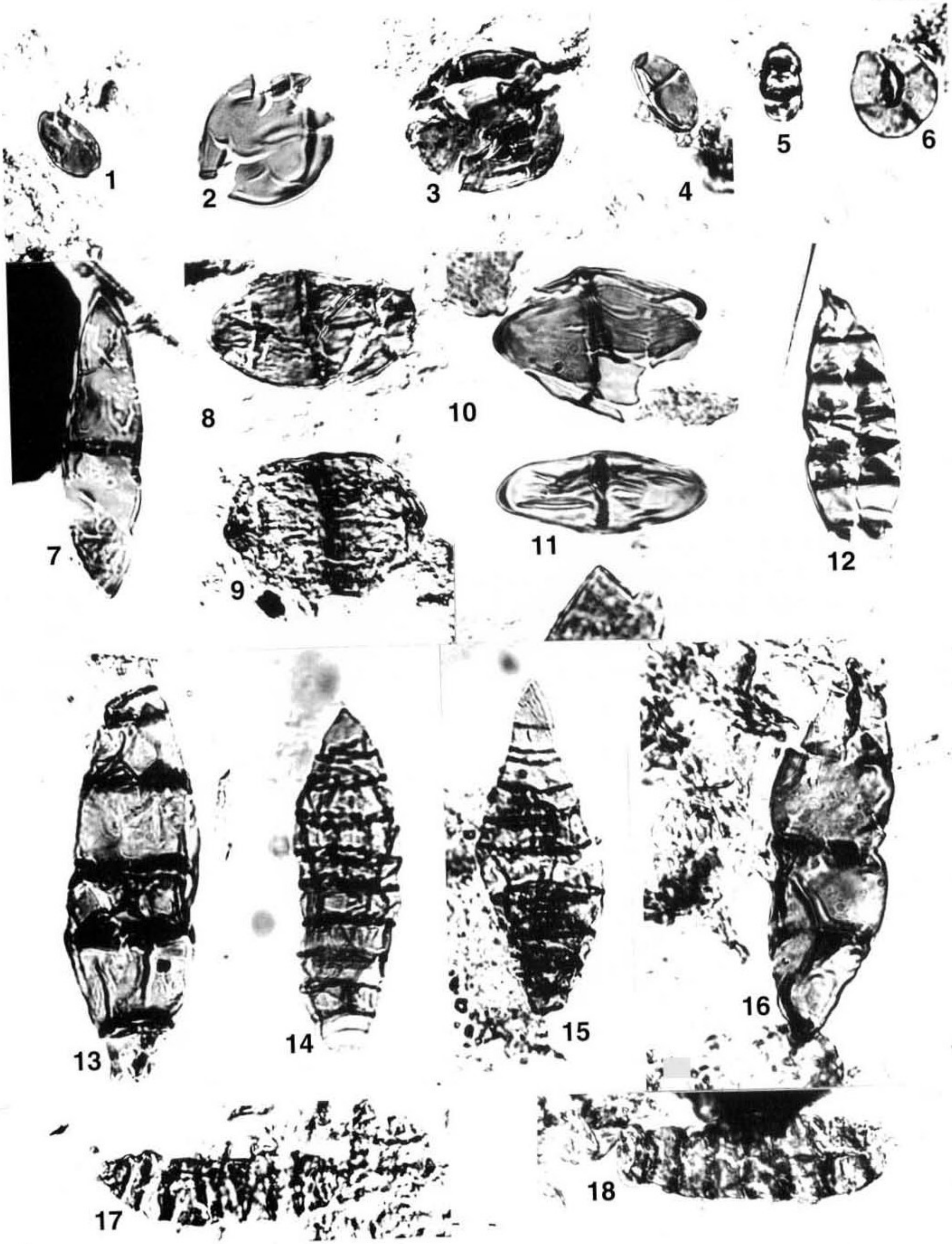


Plate 5

Late Paleocene palynomorphs (fungal) from the lower Huntingdon Formation at Third Beach, Stanley Park, Kanaka Creek, and B.C. Sumas Mountain region. Magnification x1000.

Fungal spores:- (all figures)

Figure 1. *Multicellaesporites conicus* Ke et Shi ex Sung et al., 1978

Figure 2. *Diporisporites oblongus* sp. nov.

Figure 3. *Diporisporites* sp. B.

Figure 4. *Diporicellaesporites oblongatus* Ke et Shi ex Sung et al., 1978

Figure 5. *D. bellulus* Ke et Shi ex Sung et al., 1978

Figure 6. *D. lanceolatus* Ke et Shi ex Sung et al., 1978

Figure 7. *D. extendus* sp. nov.

Figure 8. *Multicellaesporites cf. allomorphus*

Figures 9 & 10. *Brachysporisporites opimus* (Elsik & Jansonius) Norris, 1986

Figures 11 & 12. *B. cotalis*

Figure 13. *B.* cf. *cotalis* (Elsik & Jansonius) Norris, 1986

Figures 14 & 15. *Pesavis paron* Kalgutkno and Sweet, 1988

Figure 16. *Fractisporinites* sp. B. Norris, 1986

Figure 17. *Callimothallus pertusus* Dilcher, 1965

Plate 5

Plate 6

Late Paleocene palynomorphs (fungal) from the lower Huntingdon Formation at Third Beach, Stanley Park, Kanaka Creek, and B.C. Sumas Mountain region. Magnification x1000 except where otherwise indicated.

Fungal spores:- (all figures)

Figures 1 & 2. *Pesavis tagluensis* Elsik and Jansonius, 1974

Figure 3. *Reduviasporonites* sp. B. (x500)

Figures 4-6. *Reduviasporonites* sp. B. (x500)

Figure 7. *Reduviasporonites* sp. C. (x500)

Figures 8-11. *Reduviasporonites* sp. C.

Plate 6

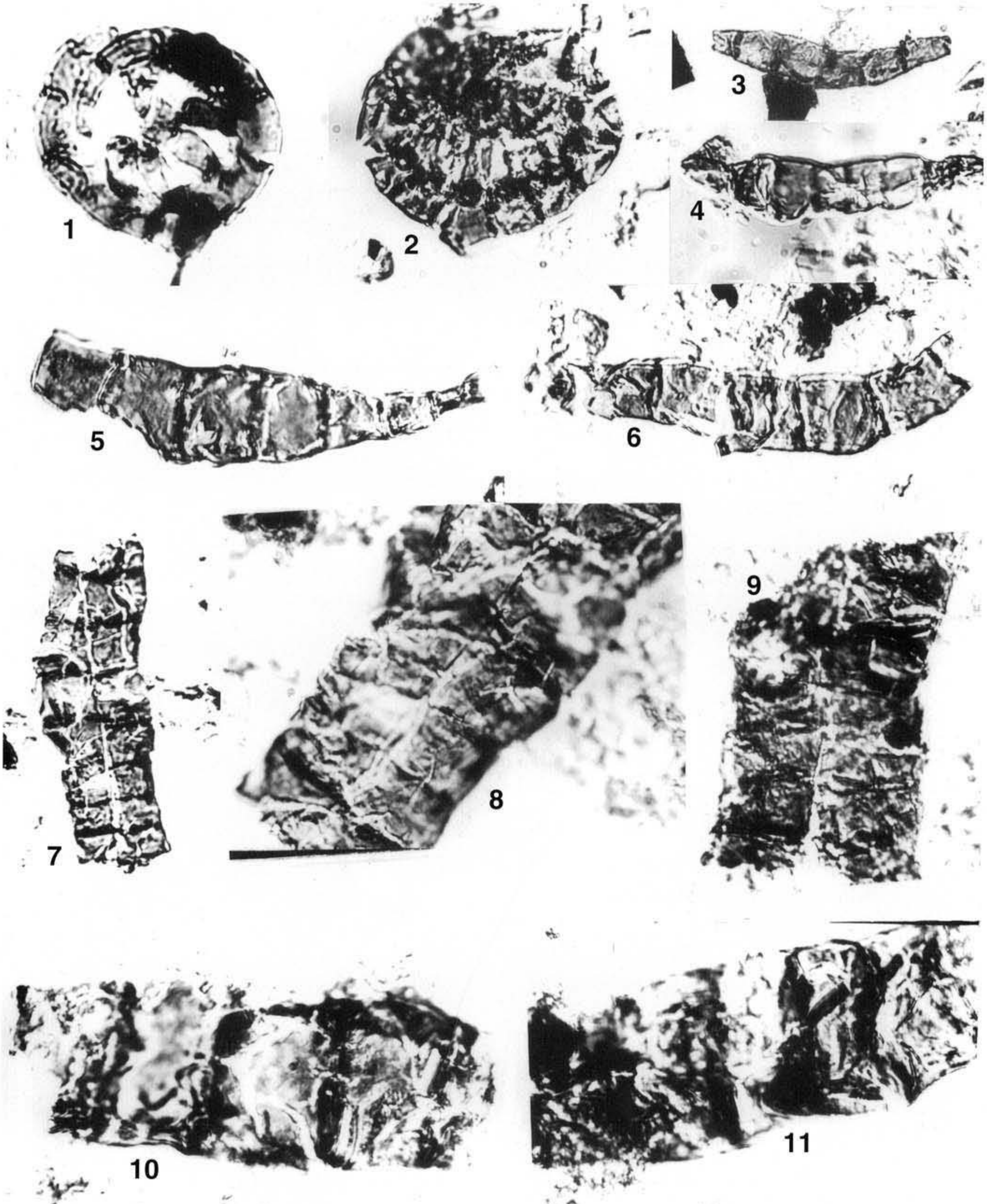


Plate 7

Some characteristic palynomorphs of the Early Eocene, *Platycarya* zone of the the Kitsilano Member, Huntingdon Formation at Ferguson Point, Stanley Park, the Conoco/Dynamic #1 well (Mud Bay, located on Fig. 3 and 15), and Canadian Sumas Mountain sections. Magnification x1000.

Fern spores:

Figure 1. *Cicatricosisporites dorogensis* Potonié and Gelletich, 1933

Figure 2. *Lygodium reticulosporites* Rouse, 1962

Angiosperm pollen:

Figure 3. *Pistillipollenites mcgregorii* Rouse, 1962

Figures 4 & 5. *Platycarya platycaryoides* (Roche) Frederikson and Christopher, 1978

Figures 6 & 7. *Araliaceoipollenites granulosus* (Potonié) Frederiksen, 1980

Figures 8 & 9. *Engelhardtia* sp.

Figure 10. *Tilia crassipites* Wodehouse, 1933

Figure 11. *T. vescipites* Wodehouse, 1933

Figures 12-14. *Castanea/Castanopsis* spp.

Fungal Spores:

Figure 15. *Pluricellaesporites psilatus* Clarke, 1965

Figure 16. *Diporisporites granulatus* Ke et Shi ex Sung et al., 1978

Figure 17. *Didymosporisporonites ovatus* Ke et Shi ex Sung et al., 1978

Plate 7

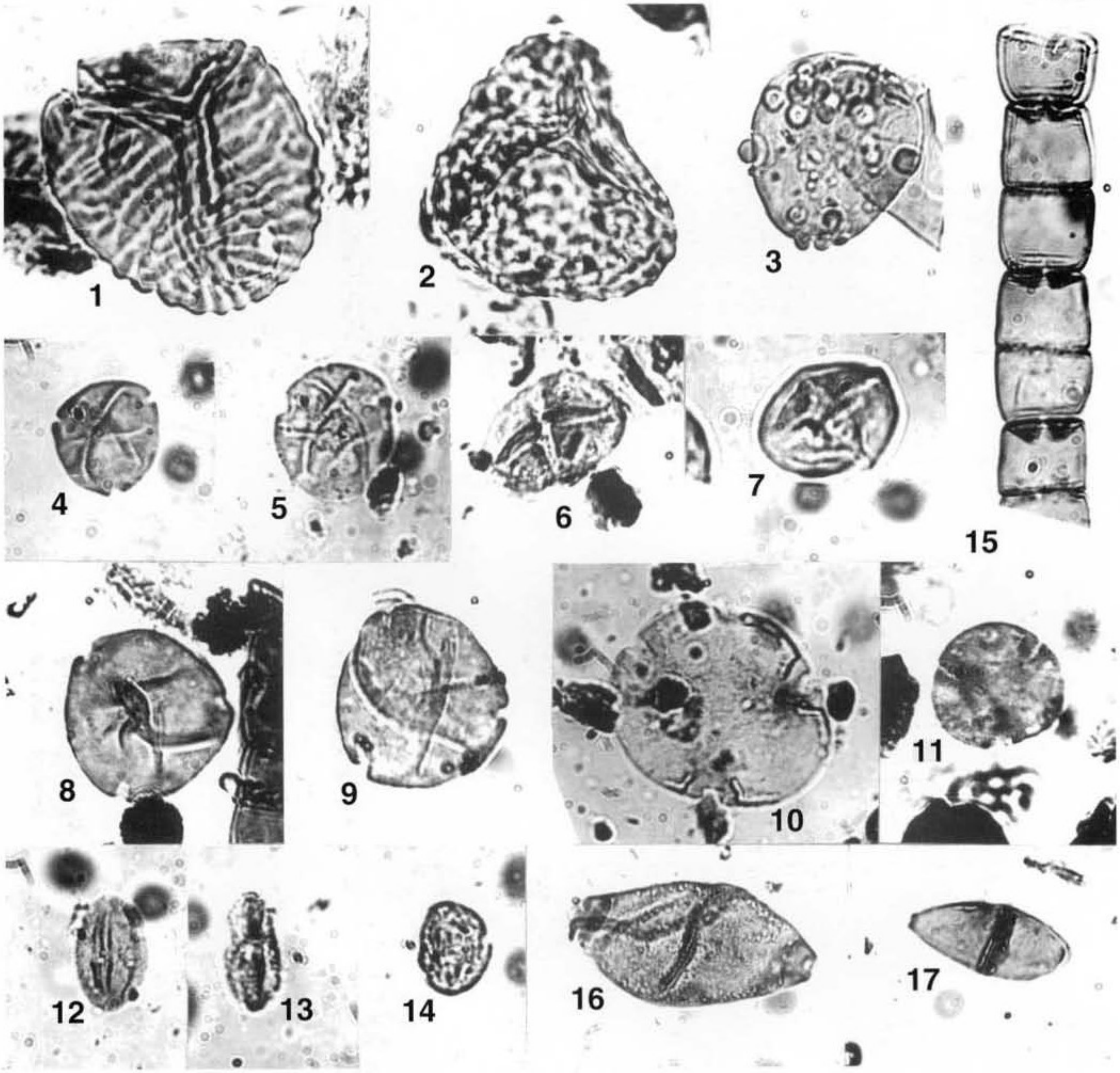


Plate 8

Characteristic palynomorphs from the Late Mid-Eocene - Early Late Eocene interval at Second Beach, Stanley Park and the Conoco-Dynamic Mud Bay #1 well in Boundary Bay (located on Fig. 3 and 15). Magnification x1000.

Angiosperm pollen:

Figures 1 & 2. *Quercoidites microhenrica* (Potonié) Potonié

Figure 3. *Q. granopollenites* (Rouse) Rouse, n. comb.

Figures 4 & 5. *Castanea/Castanopsis* sp.

Figure 6. *Horniella modica* (Mamczar) Frederiksen, 1980.

Figure 7. *Engelhardtia* sp.

Figure 8. *Fraxinoipollenites* sp.

Figure 9. *Momipites* sp.

Figures 10 & 11. *cf. Cercidiphyllum* sp.

Figure 12. *Tricoporopollenites kruschii* (Potonié) Elsik

Figure 13. *Retitricolpites* sp.

Figure 14. *Verrutricolpites* sp.

Figure 15. *Tilia vescipites* Wodehouse, 1933

Figures 16 & 17. *Araliaceoipollenites granulatus* (Potonié) Frederikson, 1980

Fungal Spores:

Figures 18 & 19 *Dicellaesporites popovii* Elsik, 1965

Figure 20. *D. mollis* Ke et Shi ex Sung et al., 1978

Figure 21. *D. aculeolatus* Sheffy and Dilcher, 1971

Figure 22. *Diporisporites* - A

Figure 23. *D. communis* Ke et Shi ex Sung et al., 1978

Figures 24 & 25. *D. acutus* sp. nov.

Plate 8

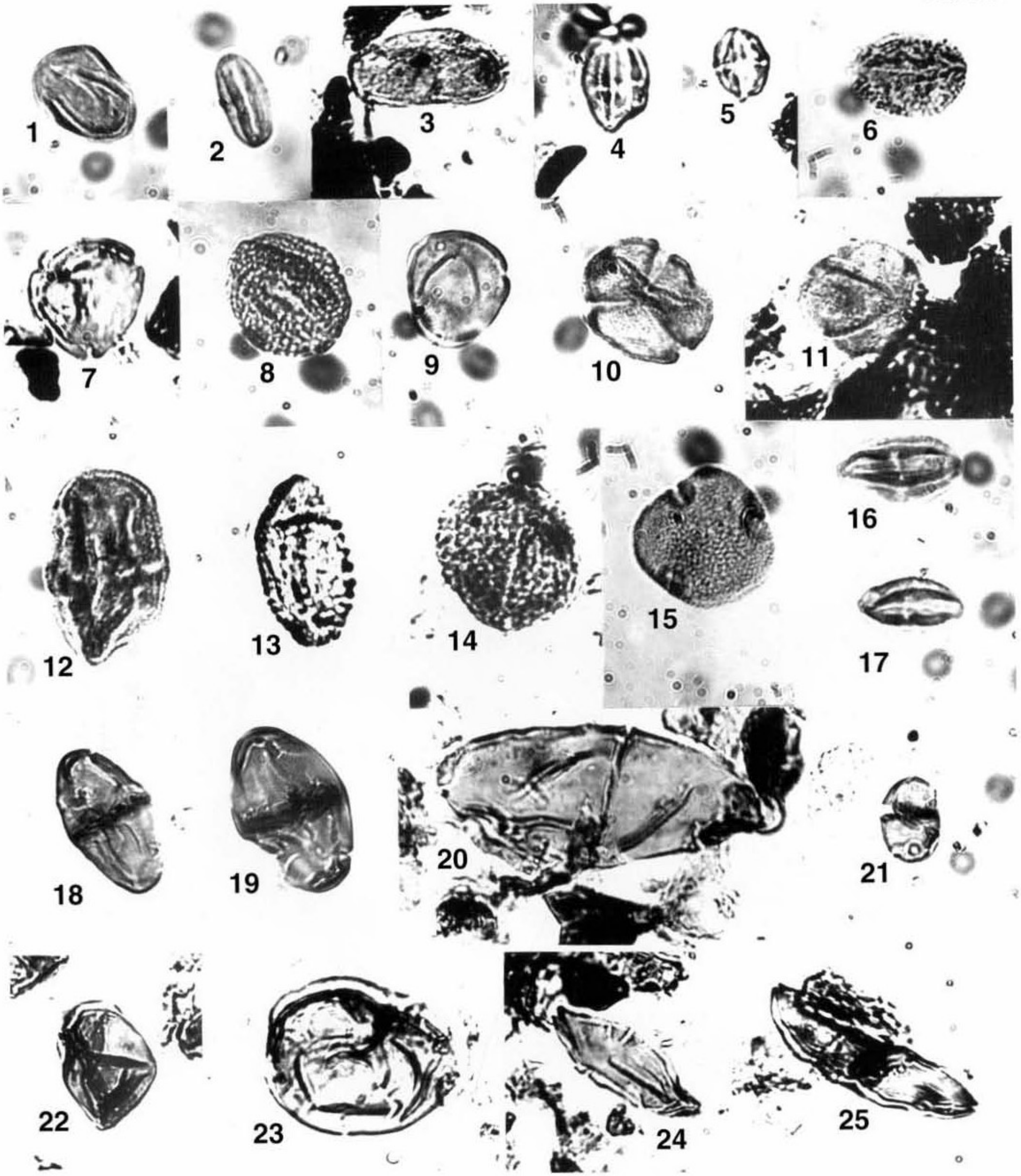


Plate 9

Characteristic palynomorphs from the Late Mid-Eocene - Early Late Eocene interval at Second Beach, Stanley Park and the Conoco-Dynamic Mud Bay #1 well in Boundary Bay (located on Fig. 3 and 15). Magnification x1000.

Fungal spores:- (all figures)

Figure 1. *Diporicellaesporites laevigataeformis* Ke et Shi ex Sung et al. , 1978

Figure 2. *D. giganteus* Ke et Shi ex Sung et al. , 1978

Figure 3. *Fractisporonites* - C

Figure 4. *Diporisporites* - B

Figures 5 & 6. *Diporicellaesporites quadratus* sp. nov.

Figures 7 & 8. *Striadiporites sanctae-barbarae* Elsik and Jansonius, 1974

Figure 9. *S. retistriatus* Ke et Shi ex Sung et al., 1978

Figure 10. *Fusiformisproites microstriatus* Hopkins, 1969

Figure 11. *Multicellaesporites leptaelus* Ke et Shi ex Sung et al., 1978

Figures 12 & 13. *M. desmodes* Ke et Shi ex Sung et al., 1978

Figure 14. *Staphlosporonites conoideus* Sheffy and Dilcher, 1971

Plate 9

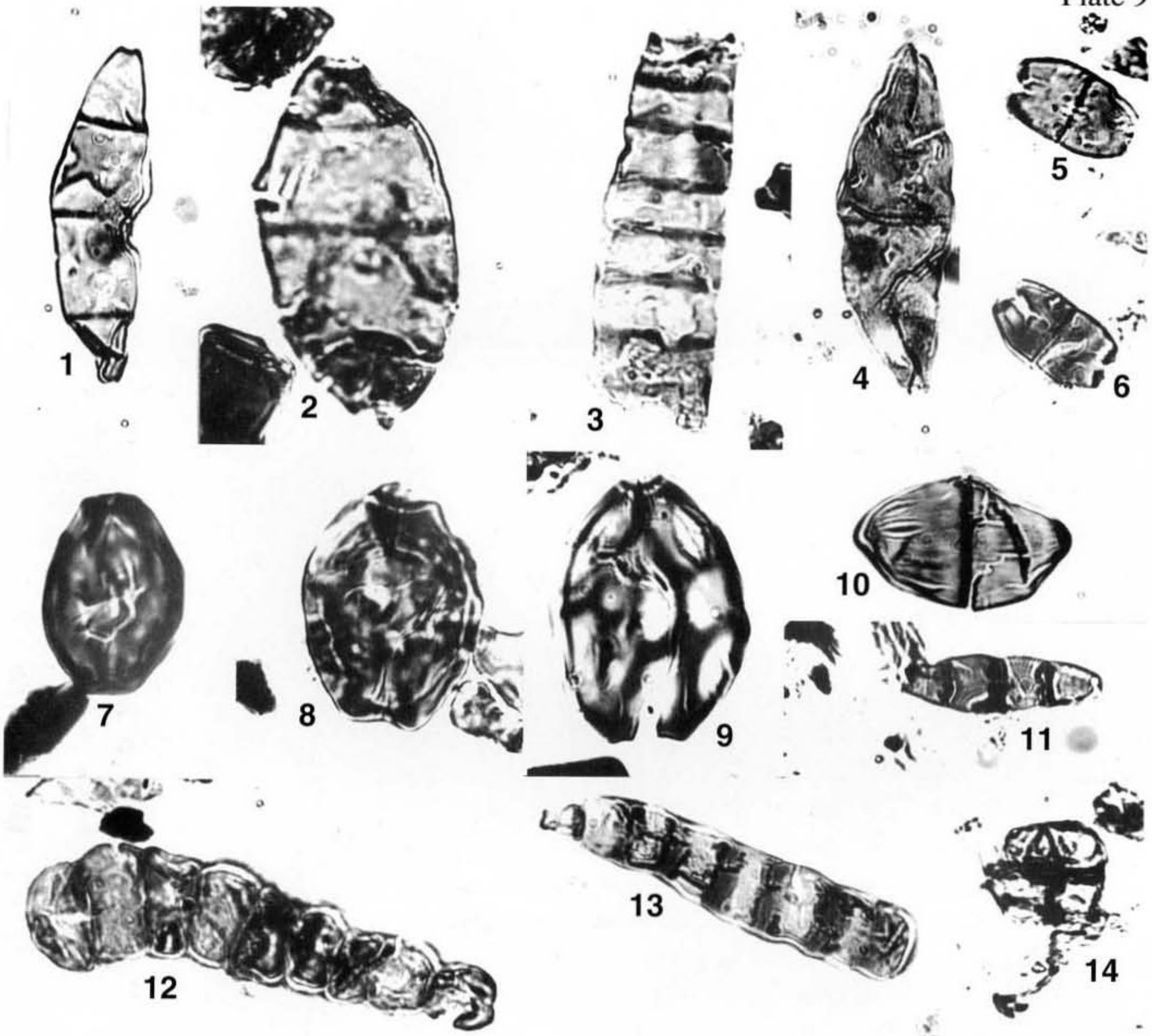


Plate 10

Selected diagnostic palynomorphs from the Late Eocene-Early Oligocene Kitsilano Member of the Huntingdon Formation and equivalent intervals in the Canadian Sumas Mountain region and Boundary Bay wells. Magnification x1000.

Fern Spore:

Figure 1. *Polypodiisporonites favus* Potonié

Angiosperm pollen:

Figures 2 & 3. *Liliacidites tritus* Frederiksen

Figures 4 & 5. *Juglans nigripites* Wodehouse

Figure 6. *Quercoidites inamoenus* (Takahashi) Frederiksen

Figure 7. *Fraxinoipollenites* sp. (cf. Frederiksen, 1989)

Figure 8. *Liquidambar* sp.

Figure 9. *Tilia crassipites* Wodehouse

Figure 10. *Momipites coryloides* Wodehouse

Figure 11. *Nyssa kruschii* (Potonié) Frederiksen

Figures 12 & 13. *Foveotricolporites* sp. (*sensu* Frederiksen, 1980)

Figure 14. *Rhoipites latus* Frederiksen

Figure 15. *Nyssapollenites cf. cruciatus* (*sensu* Frederiksen, 1980)

Figure 16. *Caprifoliipites* - A (*sensu* Rouse, 1977)

Figures 17 & 18. *Tetracolporopollenites lesquereuxianus* (Traverse) Frederiksen

Plate 10

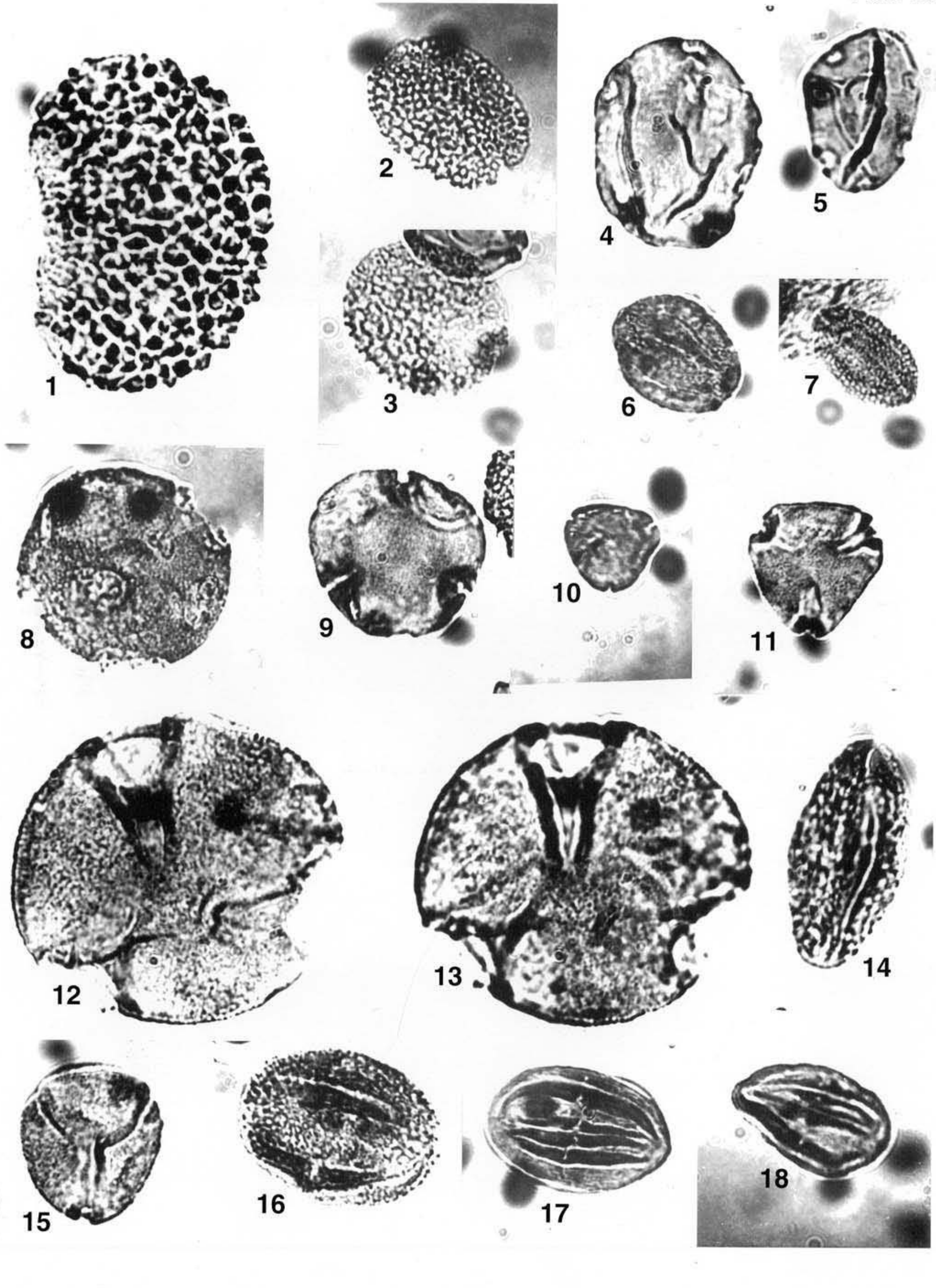


Plate 11

Additional diagnostic palynomorphs from the Late Eocene - Early Oligocene Kitsilano Member of the Huntingdon Formation and equivalent intervals in the Canadian Sumas Mountain region and Boundary Bay wells. Magnification x1000.

<u>Angiosperm pollen</u>:

Figures 1 & 2. *Tricolporopollenites* - A. (sensu Rouse, 1977) Figure 1. equatorial view, Figure 2. polar view

Figure 3. *Platanus occidentaloides* Frederiksen, 1980

Figure 4. *Rhoipites latus* Frederiksen, 1980

Figures 5 & 6. *Varisulcosporites eminens* gen. et. sp. nov.

Figure 7. *Inapertisporites plicatus* sp. nov.

Figure 8. *Schizosporis* sp.

Figures 9 & 10. *Diporisporites* - C.

Figures 11-13. *Diporicellaesporites bellulus*

Figure 14. *Dyadosporites oblongatus* Ke et Shi ex Sung et al., 1978

Figure 15. *Diporicellaesporites laevigataeformis* Ke et Shi ex Sung et al., 1978

Plate 11

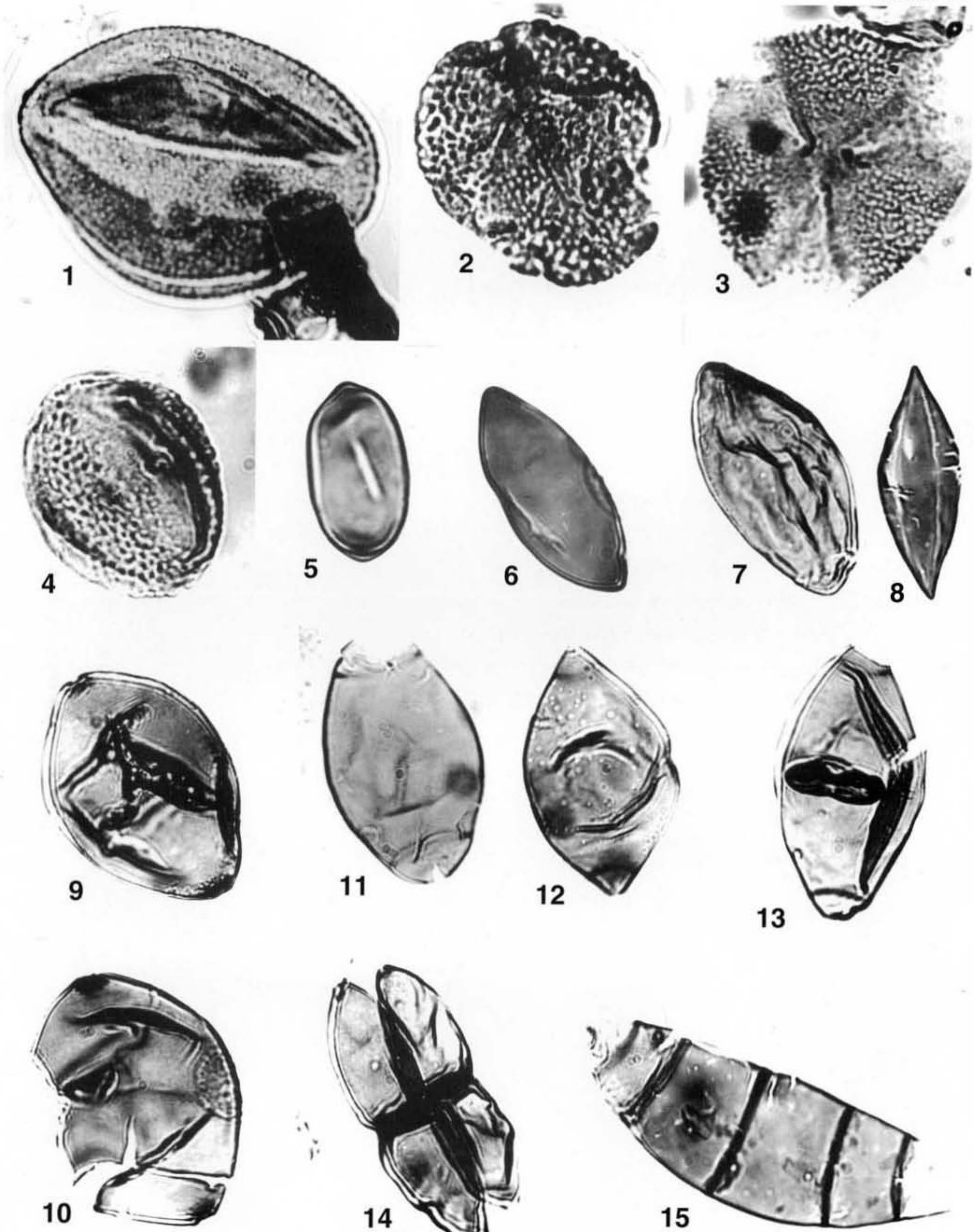


Plate 12

Additional characteristic palynomorphs (fungal spores) from the Late Eocene - Early Oligocene Kitsilano Member in Vancouver and equivalent intervals in the Canadian Sumas Mountain region and Boundary Bay wells. Magnification x1000.

Fungal spores:

Figures 1 & 13. *Diporicellaesporites giganteus* Ke et Shi ex Sung et al., 1978

Figures 2-4. *D. segmentus* sp. nov.

Figures 5 & 6. *Striadiporites bistriatus* (Ke et Shi ex Sung et al.) Norris, 1986

Figure 7. *S. reticulatus* Varma and Rawat, 1963

Figure 8. *Imprimospora* sp.

Figures 9 & 10. *I. tankensis* Norris, 1986

Figure 11. *Multicellaesporites dongyingensis* Ke et Shi ex Sung et al., 1978

Figure 12. *Pesavis tagluensis* Elsik and Jansonius, 1974

Plate 12

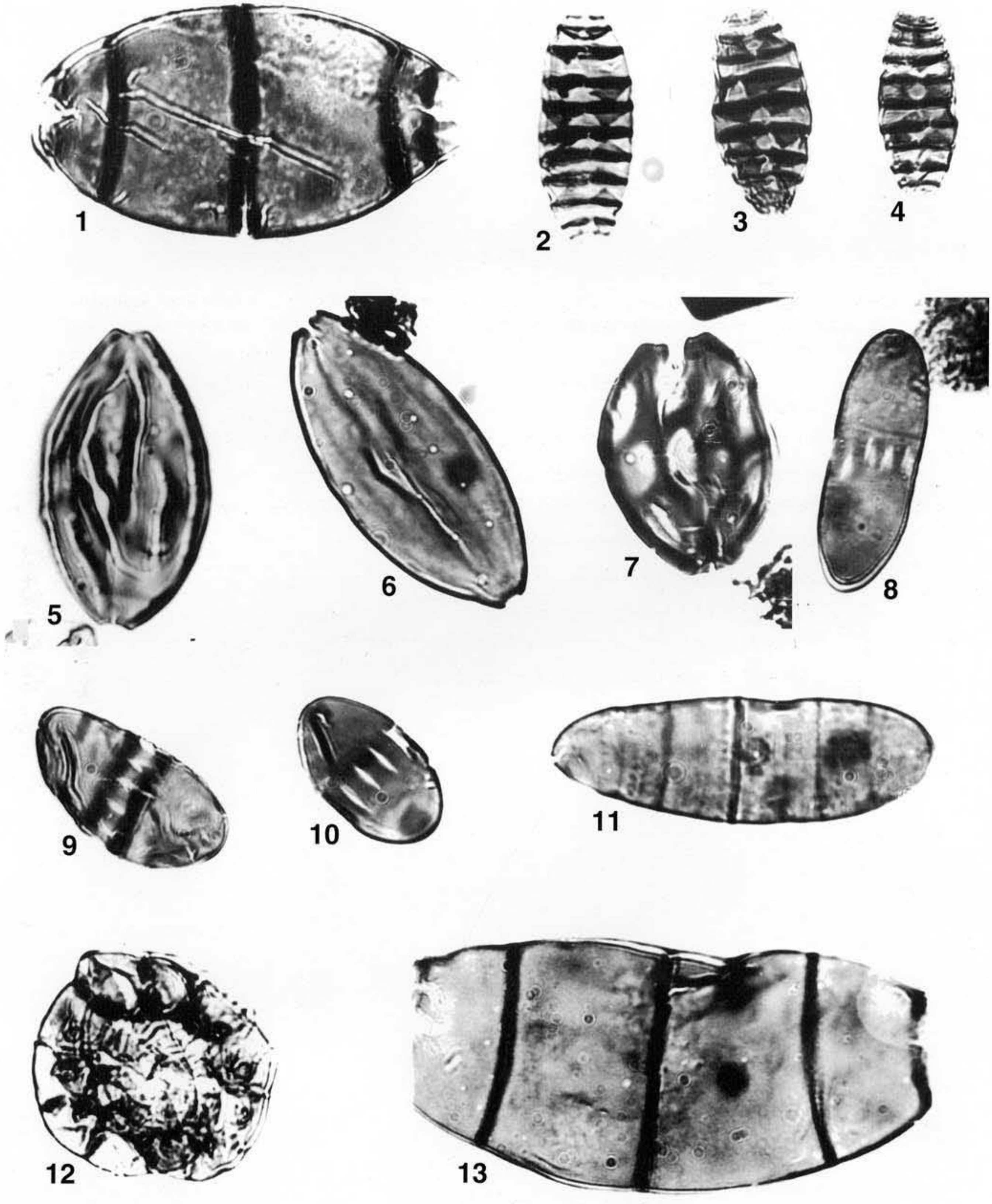


Plate 13

Additional characteristic fungal spores from the Late Eocene - Early Oligocene Kitsilano Member in Vancouver, and equivalent intervals in the Canadian Sumas Mountain region and Boundary Bay wells. Magnification x1000.

Fungal spores:

Figures 1-3. *Fusiformisporites lineatus* sp. nov.

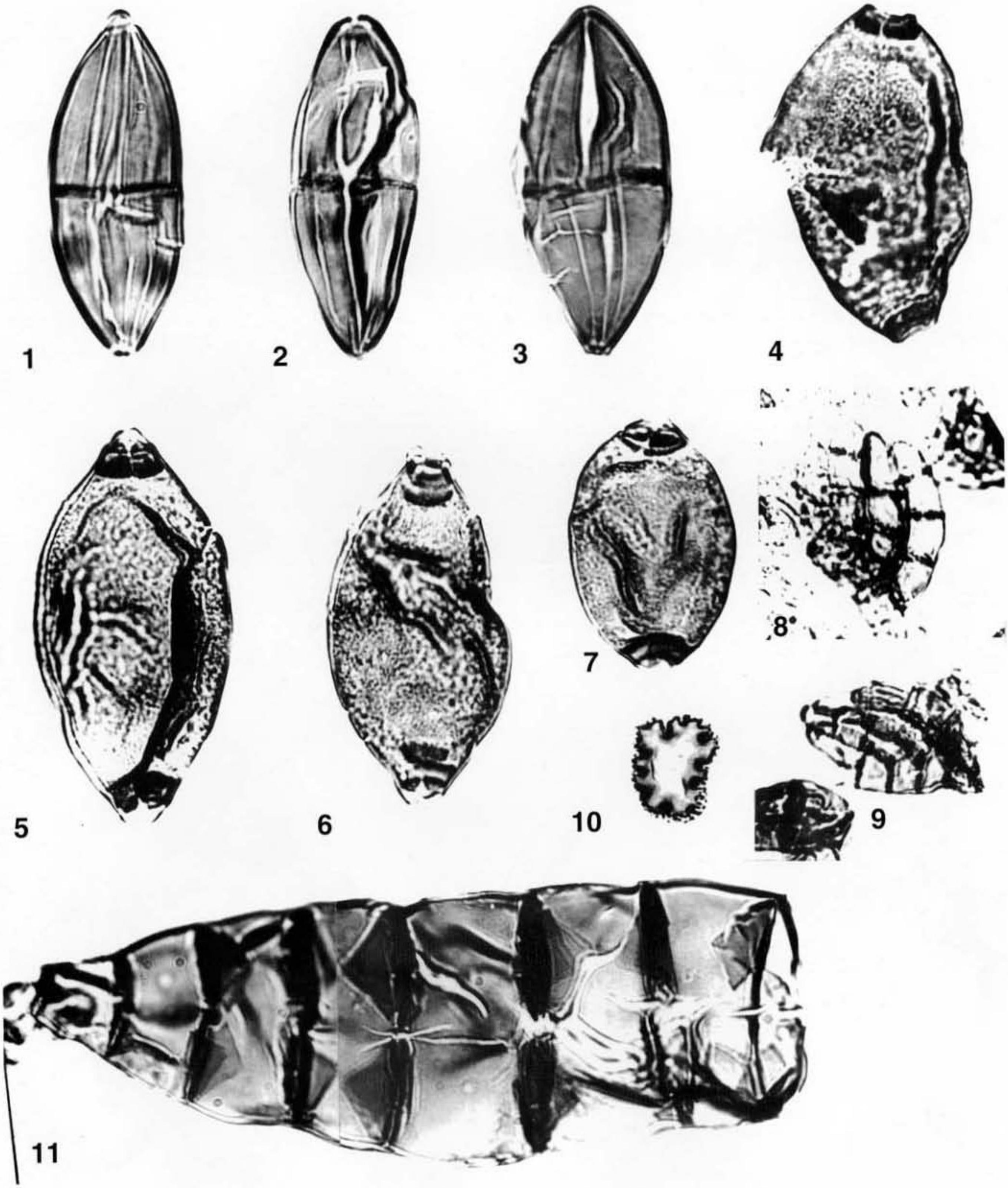
Figures 4-7. *Punctodiporites harrisii* Varma and Rawat, 1963

Figures 8 & 9. *Ctenosporites wolfei* Elsik and Jansonius, 1974

Figure 10. *Desmidiospora* sp.

Figure 11. *Pluricellaesporites magnus* sp. nov.

Plate 13



Middle Tertiary eruptive rocks in the Vancouver area

T.S. Hamilton[1] and J. Dostal[2]

Hamilton, T.S. and Dostal, J., 1994: Middle Tertiary eruptive rocks in the Vancouver area; in Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia, (ed.) J.W.H. Monger; Geological Survey of Canada, Bulletin 481, p. 171-179.

Abstract: Early Oligocene mafic volcanic rocks and related intrusions that cut sediments of the Burrard Formation in Vancouver are a minor component of the Paleogene Georgia-Bellingham basin, which straddles the international boundary between British Columbia and Washington State. In the vicinity of Vancouver they occur as flows, breccias, tuffs, dykes, and sills observed in outcrops generally of limited extent, excavations, and boreholes drilled for hydrocarbon exploration or geotechnical purposes. Petrographically these rocks are aphyric to sparsely porphyritic and may contain phenocrysts of plagioclase, olivine, augite, hornblende, and biotite. Chemically they are calc-alkali basalts and andesites and resemble early tertiary volcanics of the Northcraft and equivalent formations of the early western Cascades arc, and arc assemblages elsewhere. These rocks delineate the western edge of the Cascade arc system in the Early Oligocene.

Résumé : Les roches volcaniques mafiques de l'Oligocène précoce et les intrusions apparentées qui recoupent les sédiments de la Formation de Burrard, à Vancouver, sont une composante mineure du bassin paléogène de Georgia-Bellingham, qui enjambe la frontière internationale entre la Colombie-Britannique et l'État de Washington. À proximité de Vancouver, elles se présentent sous forme de coulées, de brèches, de tufs, de dykes et de filons-couches dans des affleurements d'étendue généralement limitée, dans des excavations et dans des trous forés en vue de la recherche d'hydrocarbures ou de travaux géotechniques. Pétrographiquement, ces roches ont une nature aphyrique à sporadiquement porphyrique et peuvent contenir des phénocristaux de plagioclase, d'olivine, d'augite, de hornblende et de biotite. Chimiquement, ce sont des basaltes calco-alcalins et des andésites, et elles ressemblent aux roches volcaniques du Tertiaire précoce de la Formation de Northcraft et de formations équivalentes du proto-arc occidental des Cascades, et des assemblages d'arc rencontrés ailleurs. Elles délimitent le rebord occidental du réseau de l'arc des Cascades à l'Oligocène précoce.

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INTRODUCTION

Cascades volcanism along the Pacific margin of North America is typical of subduction-related arcs (e.g., McBirney, 1978; McBirney and White, 1982). This arc formed as a result of the oblique subduction of the Farallon/Juan de Fuca plate during the last 40 Ma (Verplanck and Duncan, 1987). While the location of the Cascades arc system is well established in Oregon and Washington (McBirney and White, 1982; Guffanti and Weaver, 1988; Smith, 1989), its northern extent and early history are not as well defined. Souther (1991) interpreted plutons near Pemberton as marking the Miocene arc position, but prior to Miocene there are few data. Occurrences of middle Tertiary igneous rocks in the vicinity of Vancouver are limited and poorly known, but they could provide an important clue as to the former position or extent of the arc. These rocks are important to the understanding of the geological evolution of this region and the thermal maturation of hydrocarbons in parts of the Georgia-Bellingham sedimentary basin. The purpose of this paper is to examine the nature, composition, and origin of these igneous rocks.

GEOLOGICAL SETTING

Middle Tertiary volcanic and related rocks form isolated outcrops in the physiographic uplands along the foot of the Coast Mountains slope, from North and West Vancouver to Mission, British Columbia (Fig. 1). Geologically, this region is the edge of the Georgia-Bellingham basin which straddles the international boundary between British Columbia and Washington State (Mustard, 1994; Mustard and Rouse, 1991, 1992). The rocks were described and mapped by Johnston (1923) prior to extensive urban development, and subsequently by Wootton (1957), Roddick (1965, 1990), Eisbacher (1973), Roddick and Woodsworth (1979), and Roddick et al. (1979). They occur predominantly as dykes and sills, although some tuffs, flows, and pyroclastic breccias are present in the uppermost part of the Paleogene section. Equivalent rocks are extrusives encountered sporadically in excavations and boreholes in Vancouver, near the head of False Creek, and a sill cutting the Huntingdon Formation in an exploration well near Birch Bay, Washington State, U.S.A.[1]. Johnston (1923) reported ash layers containing glass, hornblende, and oligoclase-andesine over a 195 m interval in the Boundary Bay #3 well. While Johnston (1923) conjectured that these ashes were Miocene, biostratigraphy of strata which encompass this interval in the adjacent Mud Bay exploration well[2] yields an Late Eocene-Early Oligocene age. Plagioclase phenocrysts picked from a volcanic ash in the Stateside Campbell[3] well have been submitted for ^{39}Ar-^{40}Ar dating to help resolve the age of these strata. Some small, nearly vertical mafic dykes that cut the basement rocks of the Mesozoic plutonic and metamorphic complex along the edge of the Coast Mountains (Roddick, 1965, 1990) are probably related to these occurrences. In the outcrop areas, the igneous rocks represent only a minor component within Eocene through Lower Oligocene sediments variously referred to as the Burrard, Kitsilano,and Huntingdon formations (Mustard and Rouse, 1994). However, because of their poor exposure and limited exploration, it is difficult to estimate volumetric significance of the igneous rocks in the basin.

The volcanic rocks occur in two associations. They either (1) cut or are intercalated with Paleocene through Lower Oligocene continental sediments (sandstone, shale, conglomerate, coal) which are the fill of the Georgia-Bellingham basin (Hopkins, 1966; Yorath, 1991) or (2) they intrude the plutonic and metamorphic rocks that form the basement to the Georgia-Bellingham basin (Roddick, 1965, 1990). The sites sampled in this study mark a series of small, discreet eruptive centres suggesting that the style of volcanism was dispersed. The vesicularity, quenched glass, curving cooling joints,and complex geometries suggest the intrusions were emplaced at a shallow level with little confining pressure or overlying succession. Where contact relationships are exposed, some of the dykes and sills crosscut sandstones and siltstones of the Middle and Upper Eocene Burrard and overlying Oligocene Kitsilano formations. In Washington State, equivalent strata are called Huntingdon and Chuckanut formations (Johnson, 1984). Occurrences of lavas, breccias,and tuffs are seen to be intercalated with sandstones and siltstones of the Kitsilano Formation. There is some evidence for igneous emplacement penecontemporaneous with normal faulting (Johnston, 1923), and in the Vancouver area, postemplacement tilts are down to the south like the enclosing sedimentary strata. None of the igneous rocks are known to cut the unnamed Miocene strata (Yorath, 1991) higher in the stratigraphic column.

GEOCHRONOLOGY

Whole rock K-Ar ages for selected localities are presented in Table 1. The dates thought to provide the best estimate of the age of this Middle Tertiary mafic volcanism in the Vancouver-Bellingham basin range from 31.5 to 34.6 Ma. This encompasses the localities in West Vancouver, and in the adjacent Coast Mountains. Altered rocks from Grant Hill and Silverdale yielded a single determination whole rock K-Ar age of 17.0 Ma, which is considered to reflect uplift rather than intrusion. The Lower Oligocene dates are consistent with biostratigraphic control on the enclosing sediments from Vancouver to Mission (Rouse, 1962; Mustard and Rouse, 1991, 1992; Mustard, 1994). Mafic volcanism at approximately the same longitude further to the south in Washington State and Oregon, which is related to the early

[1] American Hunter et al., Birch Bay #1, SW sec.32, Twp.N Rge.E, 7040 ft.KB

[2] Conoco Dynamic Mud Bay d-95-D/92-G-2, KB 4.6 m, 822-1150 m, Early Oligocene to Late Eocene shale (J. Britton, Dynamic Oil, pers. comm., 1994).

[3] Conoco Dynamic Stateside Campbell d-3-A/92-G-2, volcanic ash layer 1560-1565 m.

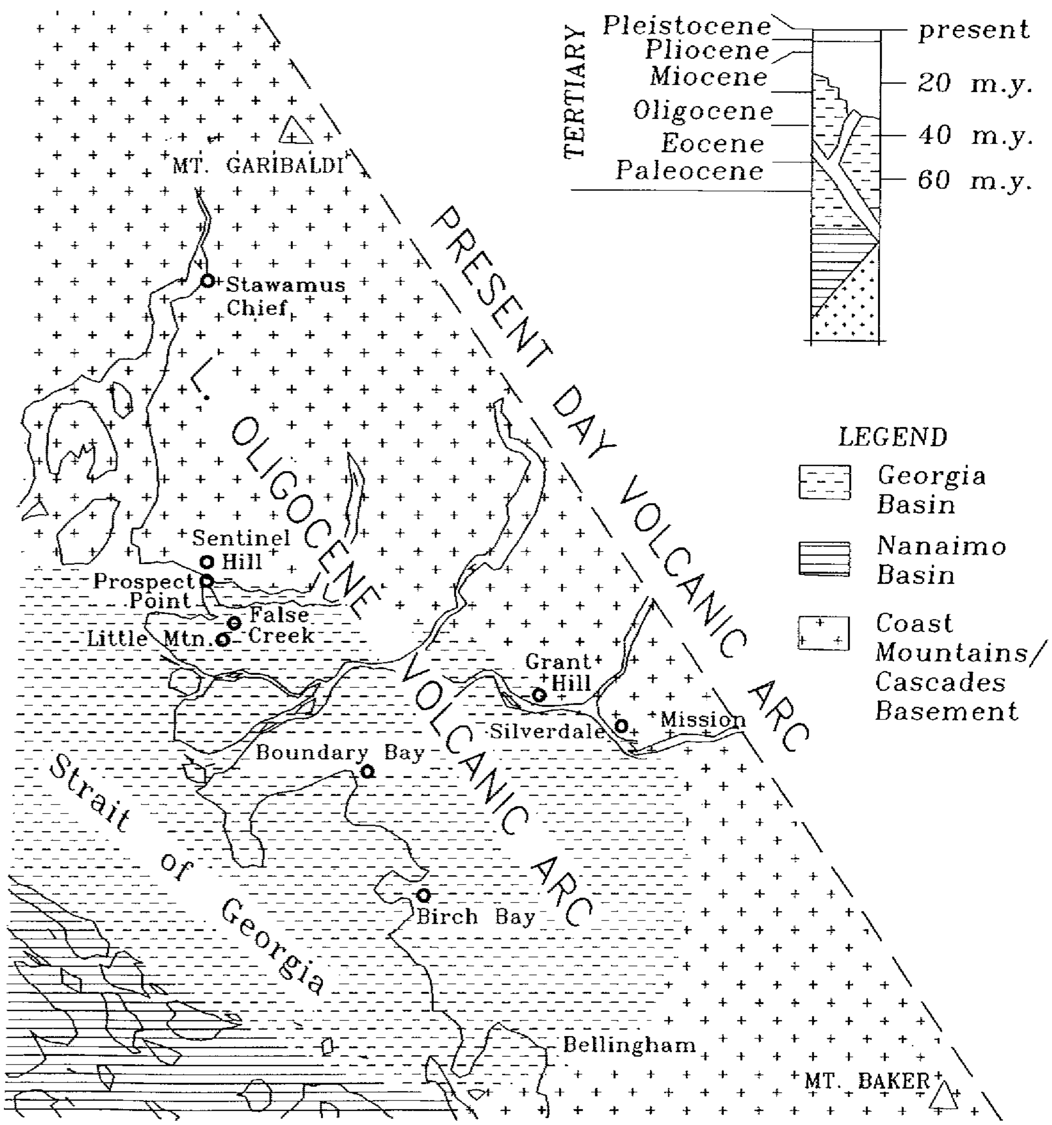

Figure 1. Location map with stratigraphic column. The legend for the basement and basin filling map units is the time-stratigraphic column. Geological epochs and radiometric ages are both shown. Bold triangles and line denote the position of the nearest modern Cascades-Garibaldi volcanoes. Named localities correspond to geochemical samples, stratigraphic control, and radiometric dates referred to in Tables 1 and 2 and in the text. Little Mountain (Locality 1, and LM samples) - marks the sill and quarry in Queen Elizabeth Park. Prospect Point (Locality 2 and PP samples) marks the dyke in Stanley Park. Sentinel Hill (Locality 3, and SH samples) marks the flows and sill in West Vancouver. False Creek (Locality 4, and VV samples) marks volcano-sedimentary outcrops and boreholes at Railroad Flats and False Creek. Grant Hill marking the sill between Kanaka Creek and the Fraser River (GH samples) and Silverdale Mission marking a sill outcropping near those municipalities (SD samples) are collectively described as Locality 5. BIRCH BAY (Locality 6 and BB sample) marks the American Hunter Birch Bay #1 well. Boundary Bay marks the Conoco Dynamic Mud Bay d-95-D/92-G-2 and Boundary Bay #3 exploration wells referred to in text. Stawamus Chief marks a dated dyke in Table 1.

Table 1. Radiometric ages.

LOCALITY	AGE ± ERROR (Ma)	MATERIAL DATED
1) Little Mountain	34.3 ± 1.2	Olivine basalt
2) Prospect Point	31.5 ± 1.2	Dyke
4) False Creek, 6th Avenue	34.6 ± 1.2	Basaltic dyke
7) Stawamus Chief	32.9 ± 2.4	Dyke

These are all whole-rock K-Ar ages. Locations are keyed to the map in Figure 1. The first four were measured by J.E. Harakal at the University of British Columbia Geochronology Laboratory. The date from the dyke cutting the Stawamus Chief was previously reported by Roddick and Woodsworth (1979).

development of the Western Cascades, yields comparable dates (Taylor, 1989; Phillips et al., 1989). Similar K-Ar ages were obtained at the U.B.C. geochronology laboratory on biotites (32.8 to 35.0 ± 2 Ma) from intermediate plutonic rocks of the Silver Creek and Yale stocks about 40 km to the east near Hope, British Columbia. Collectively the calc-alkaline activity between Vancouver and Hope is on trend with the 35 Ma Cascade arc as defined by Guffanti and Weaver (1988), and Wells (1989). The igneous activity at the latitude of Vancouver spans a region comparable to the modern arc width at the latitude of Mount St. Helens and Mount Adams (46.2°N), and defines the width and location of the Canadian portion the Cascades arc in the Early Oligocene. Igneous activity in the forearc region at this time is represented by the trondhjemitic Mount Washington plutonic suite on central Vancouver Island (Massey, 1993).

PETROGRAPHY

The most common rocks are fine grained basalts and andesites that may preserve phenocrysts of olivine but invariably have pale green diopsidic augite. These rocks also contain plagioclase phenocrysts in the bytownite to labradorite range with associated microphenocrysts of Fe-Ti oxides. Crystal clots of olivine-clinopyroxene-plagioclase-oxide at Little Mountain or plagioclase-clinopyroxene-oxide elsewhere probably preserve early phenocryst assemblages from depth. In some andesites, green to brown hornblende occurs as phenocrysts instead of, or as overgrowths on, earlier augite, as in some extrusives near False Creek and andesitic ashes in the subsurface at Boundary Bay (Johnston, 1923). Phenocryst hornblende can in turn be followed by red-brown biotite, as at Grant Hill, Silverdale, and Birch Bay. Hornblende sometimes occurs as the sole or prominent phenocryst in small dykes and as large sprays of blades or needles in pegmatitic apophyses as at Grant Hill. The common paragenetic sequence of mafic minerals from pale green diopsidic augite, to hornblende with a green-brown pleochroic formula to biotite with a red-brown pleochroic formula, suggests related sources and differentiation histories for these mid-Tertiary igneous rocks.

All rocks examined had groundmasses with essential plagioclase laths (labradorite-andesine), equant augite grains, Fe-Ti oxides, glass, occasional hornblende, alkali feldspar, quartz, and accessory apatite. Apatite also occurs as rods included within phenocryst or xenocrystic plagioclase as at Little Mountain and at Grant Hill. Foreign lithic inclusions generally resemble nearby bedrock. Examples include quartzo-feldspathic sandstone and granite at Prospect Point and Sentinel Hill, and pyroxene hornfels at Grant Hill. Variable amounts of secondary minerals are also present including: quartz, chlorite, calcite, zeolites, and clay minerals, either in the groundmass or in veins and amygdales. While many of the finer grained rocks contain a small proportion of glass, Little Mountain rocks contain quenched glassy autoliths with glass enveloping plagioclase microlites, suggesting it was a convecting lava lake or open volcanic vent, rather than a sill. There is petrofabric evidence for postemplacement structural deformation both at Birch Bay and at Grant Hill in the form of kinked or bent biotite and plagioclase grains. This is not surprising in light of the regional geology with evidence of folding, faulting, tilting, and uplift/ subsidence for the basin fill as a whole (Eisbacher, 1973; Roddick, 1990; Yorath, 1991).

GEOCHEMISTRY

The major elements and Rb, Ba, Sr, Y, Zr, Nb, Ga, Ni, Cr, V, Zn, and Cu were determined by X-ray fluorescence at the Geochemical Centre of Saint Mary's University in Halifax. The precision and accuracy of trace element data were reported by Dostal et al. (1986). Alteration that affected some of the analyzed rocks may lead to a redistribution of several major and trace elements. To diminish the effect of alteration, we have petrographically screened them, selecting samples which were relatively fresh and devoid of vesicles, veins, and cleavage surfaces. In confirmation, the observed major and trace element variations are similar to those of modern volcanic rocks suggesting that the elements retain their primary distribution.

Table 2. Geochemistry - middle tertiary eruptive rocks.

	LOW-Nb GROUP											TRANSITIONAL/NORTHCRAFT GROUP							
Sample	BB1-7040	GH-2	GH-3D	GH-3L	GH-4	GH-5	GH-6	SD6-H9	SD8-OR	VV-1	VV-3	LM-2	LM-3	PP-1	PP-3	PP-4	SH1-TV	SH2-TV	Northcraft
Locality	Birch Bay	Grant Hill						Silverdale		6th Ave		Little Mtn.		Prospect Point			Sentinel Hill		
SiO_2	58.07	53.25	54.70	57.34	55.39	53.84	53.42	55.32	51.41	57.24	57.21	54.71	54.42	60.90	60.38	59.12	61.92	61.73	56.59
TiO_2	0.92	0.99	1.01	1.52	1.08	1.03	1.01	1.08	1.05	1.32	1.32	1.47	1.33	1.07	1.09	1.21	0.90	0.95	1.21
Al_2O_3	17.80	17.01	16.87	18.21	17.01	17.01	16.94	17.87	17.39	17.57	17.33	17.51	17.01	16.08	16.26	18.20	16.10	15.97	17.72
Fe_2O_3	1.12	1.28	1.24	1.01	1.18	1.28	1.28	1.19	1.31	1.36	1.38	1.34	1.36	0.96	0.96	1.13	0.91	0.93	1.34
FeO	5.72	6.50	6.34	5.14	6.02	6.53	6.54	6.07	6.70	6.95	7.03	6.83	6.91	4.90	4.89	5.77	4.64	4.75	6.83
MnO	0.12	0.15	0.17	0.12	0.14	0.17	0.15	1.14	0.15	0.13	0.13	0.15	0.15	0.13	0.14	0.17	0.14	0.12	0.14
MgO	4.50	6.70	5.74	2.99	5.14	6.43	6.69	4.47	7.49	5.48	5.36	5.20	5.97	4.63	4.64	2.97	4.55	4.78	3.85
O	6.93	8.75	8.10	6.42	7.82	8.65	8.66	7.78	9.46	6.41	6.80	8.12	8.18	5.30	5.88	4.82	5.21	5.65	8.01
Na_2O	3.31	3.23	3.44	4.22	3.68	3.10	3.17	3.75	3.29	2.47	2.47	3.28	3.32	3.73	3.34	4.08	3.43	3.12	3.38
K_2O	1.26	1.63	1.78	2.40	1.96	1.48	1.62	1.55	1.29	0.89	0.79	1.02	1.04	2.05	2.18	2.23	1.97	1.76	0.68
P_2O_5	0.25	0.52	0.61	0.64	0.59	0.49	0.52	0.47	0.46	0.17	0.17	0.36	0.30	0.25	0.25	0.29	0.23	0.23	0.25
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
LOI	3.60	3.00	2.70	3.30	2.90	4.20	2.90	4.40	5.20	6.90	7.70	1.80	2.60	2.70	2.90	2.70	3.10	2.70	-
Mg#	54.37	60.93	57.84	46.85	56.39	59.87	60.77	54.35	62.86	54.40	53.60	53.58	56.67	58.87	58.97	43.79	59.78	60.40	46.08
Cr	-	184	163	-	124	197	197	105	195	140	121	185	205	181	175	214	187	219	48
Ni	-	133	109	17	86	121	130	69	126	67	61	86	100	61	63	71	87	97	39
V	-	161	178	212	181	211	181	206	162	255	245	156	140	138	145	160	117	122	180
Cu	-	29	42	42	30	15	36	31	39	46	45	35	45	31	33	61	24	-	138
Pb	-	13	10	11	15	32	-	-	-	-	-	-	-	11	10	11	14	-	-
Zn	-	88	112	68	87	149	90	73	65	82	78	90	84	84	85	97	84	40	105
Rb	19	25	26	34	29	22	20	22	21	15	14	21	19	43	38	31	39	31	13
Ba	-	1141	1490	1627	1287	1133	1074	739	696	443	365	391	392	857	750	948	792	786	205
Sr	-	1287	1274	1578	1240	1201	1243	897	1118	1102	971	659	638	381	417	431	391	398	454
Ga	20	20	23	21	21	20	19	20	19	19	21	24	22	20	21	22	20	20	-
Nb	6.5	5.3	-	8.4	6.3	5.3	5.3	5.3	5.4	5.4	5.5	21.9	21.9	20.8	22.0	25.3	18.8	20.8	15.5
Zr	133	158	163	225	180	156	149	149	129	114	107	184	172	105	110	122	94	95	158
Y	20	22	26	38	26	24	22	22	21	21	22	21	20	15	18	16	15	16	19
CIPW-NORM wt%																			
Q	10.0	0.0	1.6	5.0	2.0	1.4	0.0	2.8	0.0	13.4	13.1	5.0	3.5	11.5	11.7	9.2	14.4	15.6	9.2
PL	58.0	54.5	54.5	59.5	55.3	54.5	54.1	59.2	56.7	52.1	54.0	57.9	56.6	51.7	51.2	56.9	51.8	50.9	59.8
DI	2.2	11.1	9.5	4.0	9.4	10.0	10.6	7.1	12.6	0.0	0.0	6.7	8.6	3.1	4.0	0.0	1.6	2.0	5.9
HY	18.4	18.1	19.0	11.8	16.8	20.7	20.7	17.1	7.3	23.3	23.2	19.0	20.4	16.7	16.2	15.4	17.1	17.5	16.4
OL	0.0	2.0	0.0	0.0	0.0	0.0	0.3	0.0	10.9	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
An%	51.8	49.9	46.7	40.0	43.8	51.8	50.5	46.3	51.0	59.9	61.3	52.0	50.4	40.2	44.9	39.4	44.0	48.0	52.2
CI	24.0	34.9	33.7	21.7	30.0	34.6	35.4	28.0	34.8	27.9	27.7	30.4	33.4	23.3	23.7	19.4	21.7	22.8	26.5

The rocks analyzed (Table 2) range in composition from basalts to andesites (Fig. 2A). Like those of the early western Cascades arc, they are subalkaline (Fig. 2A) and display a characteristic calc-alkali trend without any significant Fe-enrichment (Fig. 2B). The basalts and andesites have relatively high alumina contents (typically between 16-18% Al_2O_3), are hypersthene normative, and most of them are medium-K calc-alkali types (Gill, 1981). The major element composition of the mafic rocks is typical of orogenic suites (Fig. 3) and the trace element abundances, particularly Zr, Y, and Ti, confirm calc-alkali affinities (Fig. 4). In contrast to many other continental margin arcs (Leeman, 1983), the volcanism in the Vancouver area is mafic (basalt-andesite). This is apparently typical of some parts of the Cascades arc (e.g., Hughes and Taylor, 1986; Taylor, 1989). All of these rocks show major element variations that on many Harker-type and various bivariate and triangular plots display a single differentiation trend, that includes the mafic compositions from the early western Cascades (Phillips et al., 1989).

Representative MORB-normalized incompatible element abundance patterns for the mafic rocks are shown in Figure 4. The mafic rocks display two distinct patterns which cannot be accounted for chronologically or petrographically. The first type, exemplified by the basaltic samples from the thick sills at Grant Hill and Silverdale and the flows near False Creek, has a high content of large-ion-lithophile elements (LILE; notably alkali and alkali earth-elements – Ba, K, Rb, and Sr), accompanied by relative depletions in high field strength elements (HFSE; especially Nb) and by high LILE/HFSE ratios such as Ba/Nb which are characteristic of many arc magmas (Pearce, 1982; Weaver et al., 1986). The other type, such as that from Little Mountain (sample LM-2 in Fig. 5) or the samples from Prospect Point and Sentinel Hill (Table 2), does not exhibit a distinct Nb depletion but is indistinguishable from those of the early western Cascades arc, including the Northcraft Formation (Fig. 5; Phillips et al., 1989). The compositional differences including Ba/Nb ratios between these two types suggest that these rocks were not derived from the same parental source. Since both Ba and Nb are strongly partitioned into magmas, the Ba/Nb ratio cannot be significantly fractionated during partial melting and fractional crystallization processes (Hofmann, 1988). Furthermore, there is no clear mineralogical distinction between samples having the two trace element patterns as hornblende-oxide (wet) and augite (dry) types are present in both groups. Niobium-depletion is thought to reflect a "source signature" as crustal contamination cannot readily produce this effect in the rocks having rather restricted major element compositions (Hickey et al., 1986).

The close spatial and temporal association of mafic and intermediate lavas with high and low Ba/Nb ratios such as those observed in the Vancouver area are common in the Cascade arc system (Leeman et al., 1989). The compositional differences have been attributed to melting of the heterogeneous mantle beneath the arc (Ellam and Hawkesworth, 1988; Plank and Langmuir, 1988; Leeman et al., 1989). Specifically, the source which is in part characterized by high LILE/HFSE enrichments (similar to those in normal arc magma sources) and in part by MORB- and OIB-like domains (Leeman et al., 1989). Irrespective of the explanation, the occurrence of these two types in the vicinity of Vancouver demonstrates the continuity of Cascades-like sources and/or processes this far north.

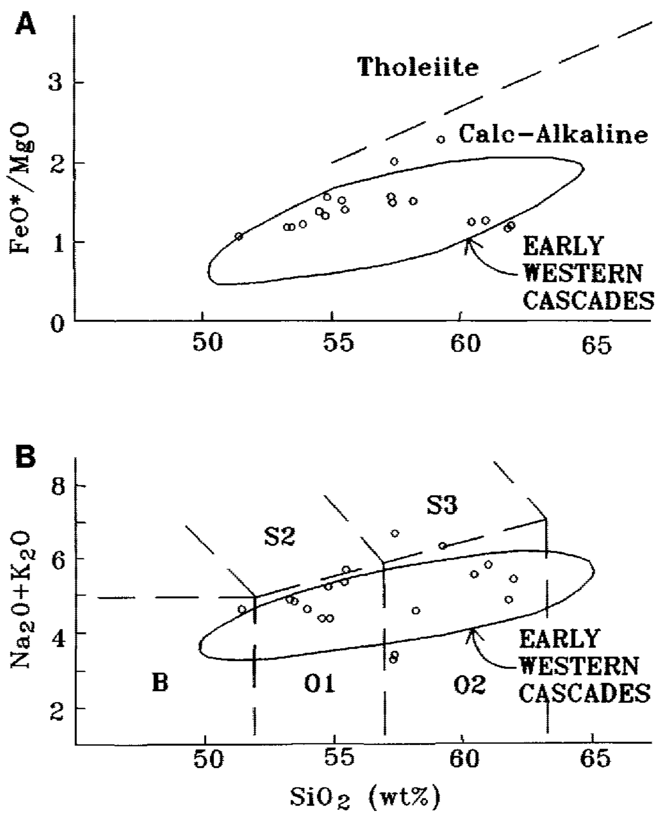

Figure 2. The variations of A) FeO/MgO vs. SiO_2 and B) total alkalies vs. SiO_2 for the Middle Tertiary rocks of the Vancouver area. Dividing line between calc-alkali (CA) and tholeiitic (TH) fields in "A" (FeO*/MgO vs. SiO_2) after Miyashiro (1974). Field boundaries in B after Le Bas et al. (1986). Comparison with the Western Cascades (from Phillips et al., 1989; Taylor, 1989).*

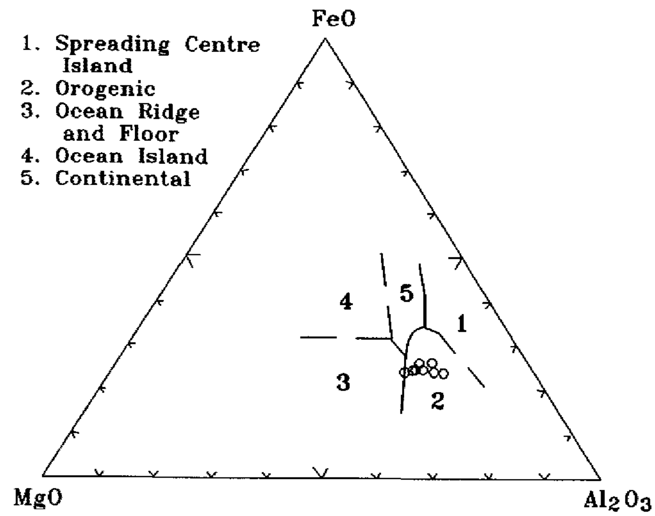

Figure 3. Al_2O_3-FeO-MgO discrimination diagram for the Eocene-Oligocene mafic rocks in the Vancouver area (Pearce et al., 1977).*

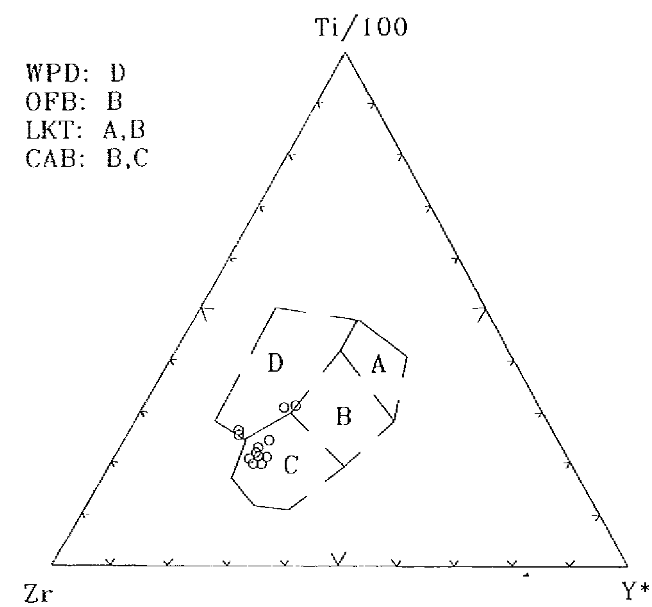

*Figure 4. Ti/100-Y*3-Zr diagram of Pearce and Cann (1973) for the Eocene-Oligocene mafic rocks in the Vancouver area.*

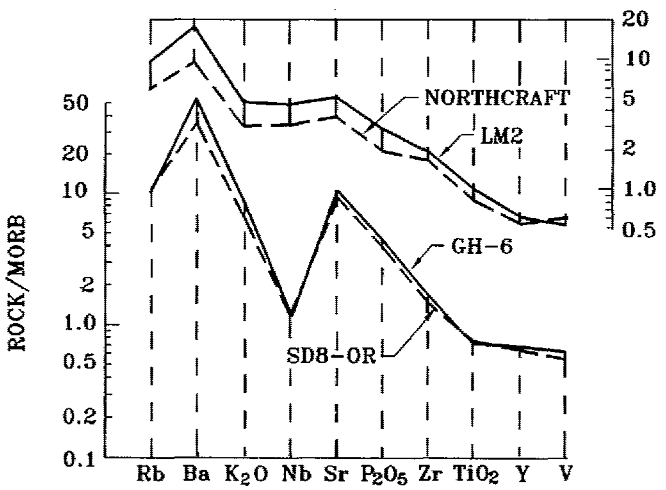

Figure 5. Normalized abundances of incompatible elements in basaltic rocks of the Vancouver area. For ready comparison to published Cascades analyses, normalizing values are after Wood (1979) and Phillips et al. (1989). The average values for the Northcraft Formation are after Phillips et al. (1989).

TECTONIC SIGNIFICANCE

The volcanic and related rocks near Vancouver are the continuation and northern extension of the Cascades arc. The geological development of Early to Middle Tertiary igneous activity in the Vancouver area parallels that of the Cascadia region immediately to the south. The modern subduction regime began with the accretion and uplift of 50 Ma old marginal basin basalts forming a new coast at 43 Ma, reaching from southern Vancouver Island (Massey, 1986) to Oregon (Verplanck and Duncan, 1987). The onset of the silica-saturated mafic volcanism at the longitude of Vancouver and the western Cascades began by Early Oligocene marking the initiation of the Cascades volcanic arc.

SAMPLE LOCATIONS

A brief description of the localities sampled in this study (Fig. 1) is given below. Additional outcrops and more extensive descriptions of local field relations and petrography are presented by Johnston (1923), Wootton (1957), and in field guides (Eisbacher, 1973; Roddick, 1990).

Locality 1: Little Mountain – Queen Elizabeth Park

A thick northwest-trending elliptical body (550 m x 365 m) of massive columnar jointed mafic rocks (LM samples in Table 2) occurs in Queen Elizabeth Park on Little Mountain. Columns are nearly vertical, and up to 70 cm across and the rock ranges from dense to slightly vesicular. The body is probably a sill or vent plug with nearby satellitic dykes. The present landscape is the result of glaciation with local relief associated with the sunken garden and reservoir that are the result of historic quarrying activities in this volcanic unit.

Locality 2: Prospect Point – north end of Stanley Park

Prospect Point at the north end of Stanley Park is a prominent cliff that contains a northeast-trending mafic dyke and an overlying flow or sill which it fed. The dyke (PP samples in Table 2) extends 1220 m across the Stanley Park from Prospect Point to Siwash Rock, where smaller curviplanar and concentric dykes indicate another local intrusive centre. Related dykes also occur along the foreshore further to the south and west at Third Beach, at the north end of English Bay, and at the west end of Kitsilano Beach. The largest of these dykes, at the foot of the Lion's Gate Bridge, is more than 30 m wide and has a prominent set of polygonal cooling joints which trend across the width of the dyke near its base, giving it the appearance of a stack of cordwood. The columnar cooling joints are nearly vertical in the upper part of the dyke suggesting another free cooling surface just above. Sandstones of the Burrard Formation are baked and show slabby exfoliation parallel to the dyke contact; the edge of the dyke is finer grained and contains amygdales filled with chlorite and calcite.

Locality 3: Sentinel Hill – West Vancouver

Sentinel Hill in West Vancouver displays a northeast-trending line of four small outcrops (700 m x 185 m) composed of basalts (SH samples in Table 2) which have polygonal and slabby jointing in various orientations suggesting emplacement in a similar attitude to the present, but with a complex cooling surface. Within the area of basalt outcrop is a block of sandstone, with more sedimentary rocks downhill to the west. The Sentinel Hill body is interpreted to be a sill or laccolith emplaced through diorites of the Coast Plutonic Complex into base of the Tertiary clastic succession. In this general area of West Vancouver there are also thin, vertically oriented mafic dykes that cut the dioritic basement to the west along the foreshore and Tertiary cover formations as at Brothers and Capilano creeks just to the east.

Locality 4: False Creek – Railroad Flats

Within the Vancouver area, numerous small outcrops of basalt, volcanic breccia, and tuff have been noted, chiefly during excavations for buildings (VV samples in Table 2). Approximately 12 m of volcano-sedimentary strata are exposed just to the north of 6th Avenue (Table 1) and to the west of St. Catherine's and the head of False Creek. Related rocks were also penetrated in a geotechnical borehole near the head of False Creek. Tuffs, flows, flow breccias, and cobble conglomerates with volcanic pebbles demonstrate the contemporaneity of volcanism and sedimentation for at least part of the Tertiary succession.

Locality 5: Grant Hill and Silverdale – east of Mission

At Grant Hill (GH samples in Table 2) and Silverdale (SD samples in Table 2) near Mission, British Columbia a pair of thick laccolith or sill-like bodies are exposed in new road cuts along the highway just above the Fraser River and at an elevation of 1000 m up the slope towards the north. Mafic rocks are medium grained with subophitic to granophyric texture and contain small subhorizontal pegmatitic segregations that are more leucocratic and hornblende-rich in their cores. Freshly blasted outcrops reveal large vertical columnar cooling joints and subhorizontal exfoliation joints. Weathered outcrops further uphill frequently show spheroidal or cannonball weathering due to the ophitic texture. No contacts were exposed, but Mesozoic diorite and gabbro outcrop just to the north, while a Tertiary clastic assemblage equivalent to the Huntingdon Formation is exposed in nearby creeks and underlies the Fraser Lowland immediately to the south. The bodies at Grant Hill and Silverdale are thought to be laccoliths emplaced into the base of the Tertiary sedimentary succession. There are large gravity and magnetic anomalies in this vicinity that may indicate more extensive plutonic bodies at depth. These larger Tertiary igneous bodies may have significantly affected the organic maturation of the Tertiary sedimentary strata at this edge of the Georgia-Bellingham Basin.

Locality 6: American Hunter et al., Birch Bay #1 SW sec. 32, Twp. 40N, Rge. 1E

Just to the East of Birch Bay, Washington, in the Birch Bay #1 exploration borehole, a mafic sill was encountered at 2145 m depth (sample BB1-7040 in Table 2). This sill was emplaced into the Lower Oligocene part of the Oligocene Huntingdon Formation, and locally affects the thermal maturation of the formation. Petrographically, it has the same medium grain size and mafic mineral sequence as the bodies across the border to the north at Grant Hill and Silverdale.

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Quaternary stratigraphy and history of south-coastal British Columbia

John J. Clague[1]

Clague, J.J., 1994: Quaternary stratigraphy and history of south-coastal British Columbia; in Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia, (ed.) J.W.H. Monger; Geological Survey of Canada, Bulletin 481, p. 181- 192.

Abstract: Thick Quaternary sediments underlie the lowlands of south-coastal British Columbia. Most of these sediments were deposited in marine, deltaic, and fluvial environments near the margins of Pleistocene glaciers. Heterogeneous units of stratified drift of at least three glaciations are separated by unconformities and, locally, by nonglacial sediments. Surface and near-surface sediments were deposited during the Fraser Glaciation which peaked about 15 ka BP. At that time, lobes of the Cordilleran Ice Sheet covered south-coastal British Columbia to elevations of about 1500 m, and the Strait of Georgia and bordering lowlands attained something close to their present form. During deglaciation, glaciomarine, glaciofluvial, and deltaic sediments were deposited over large areas on the glacially eroded, isostatically depressed landscape. In contrast, sedimentation during postglacial time has been largely restricted to the Strait of Georgia, fiords, and lakes. Sea level along the south coast has fluctuated above and below its present position during Quaternary time owing to glacio-isostatic uplift and subsidence of the land surface, global (eustatic) sea level change, and tectonism. Inferred, minor, sudden changes in sea level in the Victoria and Vancouver areas about 2 and 3.5 ka BP may have resulted from earthquakes that were stronger than any of the historical period. All historical earthquakes in southwestern British Columbia and northwestern Washington State have originated within either the North America or Juan de Fuca plate, but rare, great, plate-boundary earthquakes probably also affect the region.

Résumé : Il existe des sédiments quaternaires épais dans les basses terres de la côte sud de la Colombie-Britannique. La plupart d'entre eux se sont déposés dans des milieux marins, deltaïques et fluviatiles près des marges de glaciers pléistocènes. Des discordances et, localement, des sédiments non glaciaires séparent des unités hétérogènes de drift stratifié issu d'au moins trois glaciations. Des sédiments de surface et de subsurface se sont déposés pendant la Glaciation de Fraser, qui a culminé il y a environ 15 ka. À cette époque, des lobes de l'Inlandsis de la Cordillère couvraient la côte sud de la Colombie-Britannique jusqu'à une altitude d'environ 1 500 m, et le détroit de Georgia et les basses terres avoisinantes avaient presque acquis leur configuration actuelle. Pendant la déglaciation, des sédiments glaciomarins, fluvioglaciaires et deltaïques se sont accumulés sur de vastes étendues du paysage modelé par l'érosion glaciaire et isostatiquement déprimé. En revanche, la sédimentation postglaciaire s'est largement limitée au détroit de Georgia, aux fjords et aux lacs. Au Quaternaire, sur la côte sud, le niveau marin a fluctué au-dessus et au-dessous de sa position actuelle en raison du relèvement glacio-isostatique et de la subsidence de la surface terrestre, des fluctuations eustatiques du niveau marin et du diastrophisme. Des fluctuations présumées, mineures et soudaines, du niveau marin dans les régions de Victoria et de Vancouver, survenues il y a entre 2 ka et 3,5 ka, ont peut-être été causées par des séismes qui ont été plus violents que ceux de la période historique. Tous les séismes de la période historique survenus dans le sud-ouest de la Colombie-Britannique et le nord-ouest de l'État de Washington ont eu leur origine soit dans la plaque nord-américaine, soit dans la plaque Juan de Fuca; toutefois, de rares séismes de grande envergure survenus à la frontière de plaques ont probablement aussi touché la région.

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INTRODUCTION

Quaternary[1] sediments up to 300 m thick underlie lowlands bordering the Strait of Georgia (Fig. 1). In areas where the sediments are thick, for example the Fraser Lowland, there is little or no relationship between the buried bedrock surface and the present land surface; rather, the landscape has been shaped by processes operative during the last glaciation and, to a lesser extent, Holocene time.

The Quaternary succession in south-coastal British Columbia consists of sediments deposited during several glaciations and intervening interglaciations. Thick complex units of till and stratified drift were deposited beneath and at the margins of glaciers that advanced into lowlands from adjacent mountains. These sediments were eroded both by the glaciers that flowed over them and by streams and the sea during subsequent interglaciations. The Quaternary succession, consequently, comprises several packages of drift separated by unconformities and nonglacial sediments.

The repeated waxing and waning of glaciers triggered isostatic movements which, together with eustatic changes, caused sea level to fluctuate up to 200 m relative to the land. Lowlands bordering the Strait of Georgia were repeatedly transgressed and regressed during Quaternary time, and marine and deltaic sediments were deposited there in complex associations with glacial materials.

Little information is available on the Early and Middle Pleistocene stratigraphy and history of the region, but some subsurface deposits date to these times. There is a more complete record of Late Pleistocene and Holocene events, which are emphasized in this paper.

PLEISTOCENE GLACIATION

Cordilleran Ice Sheet

As climate deteriorated early during each Pleistocene glaciation, small mountain ice fields grew and valley glaciers advanced (alpine and intense alpine phases of glaciation described by Kerr, 1934; phases 1 and 2 of Davis and Mathews, 1944). The subsequent coalescence and thickening of glaciers over lowlands and plateaus produced a contiguous system of confluent valley and piedmont glaciers and mountain ice fields known as the Cordilleran Ice Sheet (mountain and continental ice sheet phases of Kerr, 1934; phases 3 and 4 of Davis and Mathews, 1944).

When fully developed, the Cordilleran Ice Sheet enveloped almost all of British Columbia and southern Yukon Territory, as well as parts of Alaska, Alberta, Montana, Idaho, and Washington State (Clague, 1989). The surface of the ice sheet was above 2300 m elevation over much of southern British Columbia, and ice was more than 2000 m thick over major valleys (Ryder et al., 1991). Ice flowing down fiords and valleys in the Coast Mountains covered large areas of the continental shelf (Luternauer and Murray, 1983). Glaciers from the southern Coast Mountains and the Vancouver Island Ranges coalesced over the Strait of Georgia to produce a great piedmont lobe that flowed south into the Puget Lowland of Washington State (Fig. 2; Armstrong et al., 1965; Waitt and Thorson, 1983). Glaciers streaming down valleys in south-central and southeastern British Columbia likewise terminated as large lobes in eastern Washington State, Idaho, and Montana.

The Cordilleran Ice Sheet did not fully develop during all glaciations. At the maxima of lesser glaciations, the major mountain systems supported networks of valley and piedmont glaciers, but much of the interior remained ice free. Much of the Strait of Georgia and Fraser Lowland also may have been ice free at these times, although they were areas of proglacial sedimentation.

The Cordilleran Ice Sheet formed most recently during the last, or Fraser, glaciation (Late Wisconsinan Substage = marine oxygen isotope stage 2). The ice-sheet phase of the Fraser Glaciation (ca. 10-19 ka BP) was preceded by a lengthy period of montane glaciation (ca. 19-30 ka BP), during which glaciers, at times, extended into, but did not completely envelop, the Strait of Georgia and Fraser Lowland (Clague, 1976b, 1977a, 1981). The ice sheet achieved its maximum extent in southwestern British Columbia about 15 ka BP, but disappeared over the next 5 ka (Clague et al., 1980; Clague, 1981).

Geomorphic effects of glaciation

Glaciation has profoundly altered the landscape of south-coastal British Columbia, just as it has other parts of the province (Mathews, 1989). Classic alpine landforms, including cirques, overdeepened valley heads, horns, and comb ridges, abound in the higher parts of the southern Coast Mountains and Vancouver Island Ranges, where some summits stood above névé surfaces. Most mountain valleys have U-shaped cross-profiles, which also is a consequence of glacial erosion. The coastline is indented by fiords that extend as much as 150 km inland and have water depths of as much as 755 m (Peacock, 1935; Picard, 1961). The Strait of Georgia and, to a lesser extent, Juan de Fuca Strait also have been scoured by glaciers.

Some lowlands bordering the Strait of Georgia have been streamlined by southeast-flowing ice. Meltwater channels, raised deltas, kames, kame terraces, outwash plains, and relict shoreline and seafloor deposits also occur in these areas and collectively record the decay of the Cordilleran Ice Sheet at the close of the Fraser Glaciation.

[1] The Quaternary Period spans about the last 1.6 Ma and comprises the Pleistocene (1.6 Ma - 10 ka BP) and Holocene (10 ka BP - present) epochs.

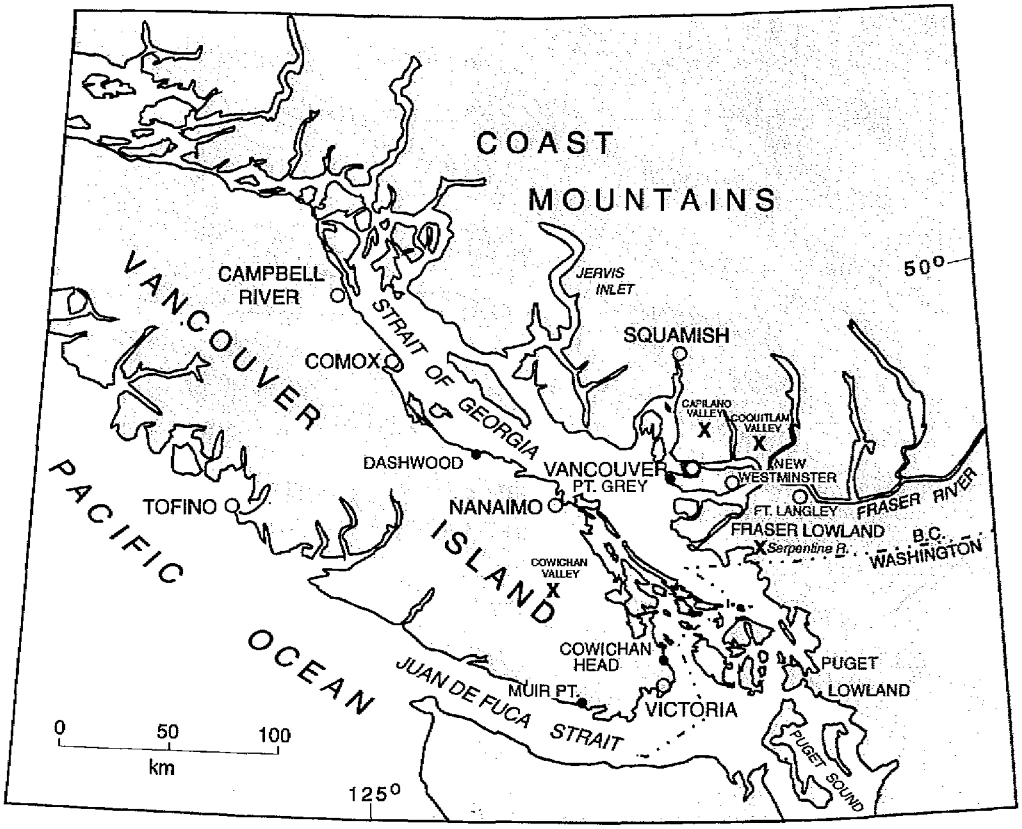

Figure 1. Location map.

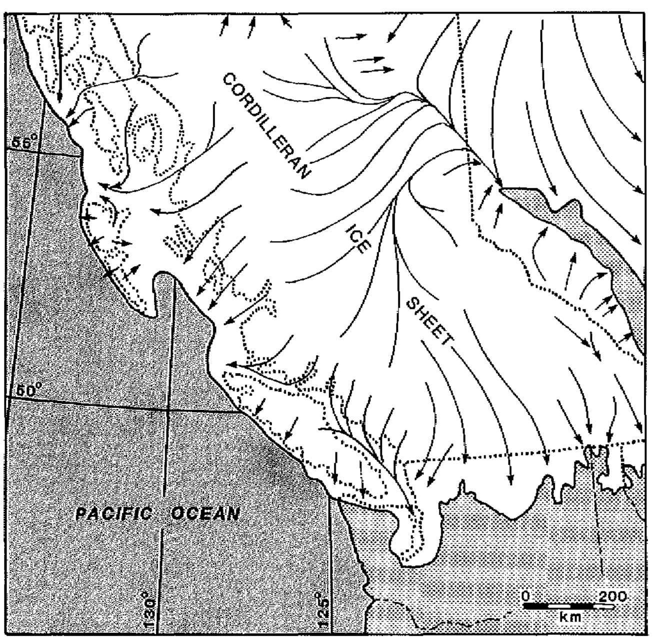

Figure 2. *Cordilleran Ice Sheet at the maximum of the last glaciation; generalized flow pattern is indicated by arrows (adapted from Blaise et al., 1990, Fig. 11).*

QUATERNARY STRATIGRAPHY AND HISTORY

Regional distribution of Quaternary sediments

The south-coastal region contains the most extensive and most intensively studied Quaternary deposits in British Columbia (Fig. 3, 4). Sediments older than the Fraser Glaciation (Cowichan Head Formation and older units; Fig. 4) are widespread on both sides of the Strait of Georgia, but are not well exposed because of the cover of younger materials. They outcrop in sea cliffs, in natural exposures and gravel pits in some mountain and lowland valleys, and on steep slopes in the western Fraser Lowland. In addition, they have been encountered in many drill holes in the Fraser Lowland and underlie banks in the Strait of Georgia (Clague, 1976a).

Thick bodies of sand and gravel deposited early during the Fraser Glaciation (Quadra Sand and correlative units; Fig. 4) are important elements of the Quaternary succession. Quadra Sand is exposed in many coastal bluffs on Vancouver Island and on islands in the Strait of Georgia, and is an important subsurface unit in the Vancouver area (Fig. 5). Its gravelly facies (Saanichton gravel and coarser parts of Coquitlam Drift) is a major, although diminishing source of aggregate on

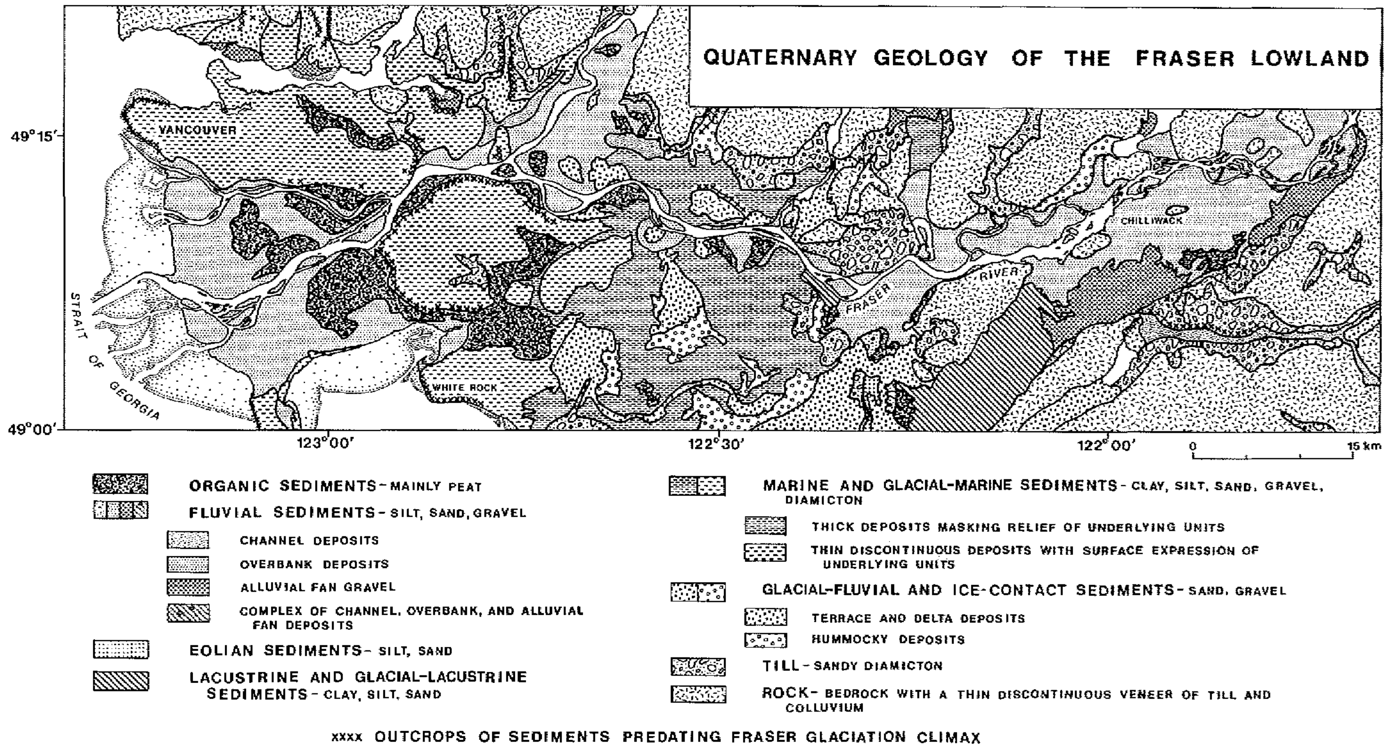

Figure 3. *Generalized surficial geology of the Fraser Lowland (from Clague and Luternauer, 1982, Fig. 22). Surface sediments were deposited during the last glaciation and Holocene time. Older sediments occur widely at depth.*

southern Vancouver Island and in the Fraser Lowland. Correlative, fine grained sediments (clay, silt, and sand) occur in the lower reaches of some mountain valleys on Vancouver Island (e.g., Cowichan valley) and the British Columbia mainland (e.g., Coquitlam and Capilano valleys). These sediments were deposited in lakes impounded at the margin of a piedmont glacier that flowed down the Strait of Georgia and into the Fraser Lowland.

Quadra Sand and related deposits were eroded by the glaciers that covered the south coast during the climactic (Vashon) phase of the Fraser Glaciation. In many places, sandy till overlies Quadra Sand, but in other places, Fraser till is absent and the erosion surface developed on Quadra Sand and older materials is directly overlain by deglacial or postglacial sediments.

Glacial-marine sediments (Capilano Sediments, Fort Langley Formation) were deposited on isostatically depressed lowlands below 200 m elevation at the end of the Fraser Glaciation (Armstrong, 1981). There are large sand and gravel deltas (part of the Capilano Sediments) at the mouths of many mountain valleys, and, in some cases, gravel terraces extend upvalley considerable distances from these deltas. Deposits of silt and clay, commonly containing stones and fossil shells, underlie much of the western and central Fraser Lowland (Fig. 3), as well as parts of eastern Vancouver Island. They are thickest in the central Fraser Lowland where they were deposited on the sea floor adjacent to a fluctuating ice margin. In this area, they are complexly interbedded with diamicton, gravel, and sand, all deposited in close association with glacier ice. Farther east are the slightly younger tills and glacial-fluvial gravels of the Sumas Drift, deposited during a minor readvance after the sea had regressed from much of the Fraser Lowland.

Glacial-marine sediments which correlate with the Capilano Sediments and Fort Langley Formation are present in the Strait of Georgia (Clague, 1977b) and in fiords such as Howe Sound and Bute Inlet. They presumably were deposited from suspension and sediment gravity flows at a time when glaciers were still present in the region. They underlie similar, although more organic-rich, nonglacial sediments that have accumulated in these basins during the last 10 ka (Pharo and Barnes, 1976; Clague, 1977b).

The main areas of sedimentation in south-coastal British Columbia during the Holocene have been floodplains, fans, deltas, lakes, fiords, and the Strait of Georgia. Deltas are present at the mouths of many streams that flow into the Strait of Georgia and also occur at the heads of fiords and large lakes in the southern Coast Mountains and Vancouver Island Ranges. Deltaic deposits commonly extend far inland along river valleys, but are covered by alluvium laid down by prograding streams. Deep basins of the Strait of Georgia and the floors of fiords and large lakes are underlain by thick, fine grained sediments derived mainly from inflowing streams (Gilbert, 1975; Pharo and Barnes, 1976; Clague, 1977a).

Early or Middle Pleistocene deposits

Westlynn Drift (Fig. 4), the oldest named Quaternary unit in southwestern British Columbia, comprises till, glacial-marine diamicton, glacial-fluvial gravel and sand, and rhythmically bedded glacial-lacustrine silt and clay (Armstrong, 1975). This unit outcrops in only a few places near Vancouver and has not been extensively studied. It is thought to predate the last interglaciation (Sangamonian Stage = oxygen isotope stage 5), and thus be older than 128 ka; no radiometric ages, however, are available.

The last interglaciation

Deposits assigned to the Sangamonian Stage have been recognized at only a few localities in south-coastal British Columbia (Clague et al., 1992b). Highbury Sediments (Fig. 4), identified at three sites in the Fraser Lowland, consist of fluvial, deltaic, and marine sediments (mainly silt and sand) underlying drift of the penultimate (oxygen isotope stage 4?) glaciation (Armstrong, 1975; Hicock and Armstrong, 1983). Marine facies are overlain by deltaic and fluvial facies, recording seaward progradation of deltas.

The Muir Point Formation (Fig. 4) on southern Vancouver Island is thought to correlate with Highbury Sediments on the basis of similar sediment types and stratigraphic position. It consists of silt, sand, gravel, and diamicton, deposited as alluvium and colluvium on a coastal floodplain (Hicock and Armstrong, 1983; Alley and Hicock, 1986; Hicock, 1990). At the type section, the Muir Point Formation is more than 30 m thick, is unconformably overlain by the Cowichan Head Formation and Fraser Glaciation till, and is underlain by an older till unit (Fig. 6).

Fossil pollen assemblages from the Muir Point Formation provide a record of vegetation and climate during the last interglaciation (Alley and Hicock, 1986; Hicock, 1990). Douglas-fir (*Pseudotsuga menziesii*) pollen is more abundant in these sediments than in the modern pollen rain in the same area. This, along with abundant pollen of alder (*Alnus*) and western red cedar (*Thuja plicata*), indicates that conditions were warmer and/or drier on the south coast of British Columbia during the last interglaciation than today. The uppermost sediments of the Muir Point Formation record marked declines in Douglas-fir, alder, and western red cedar pollen and corresponding increases in western hemlock (*Tsuga heterophylla*) and spruce (*Picea*) pollen. This signifies a major change in forest structure which perhaps was caused by climatic cooling. It is not known whether this inferred climatic deterioration marks the onset of a glaciation or whether it was a cool, moist phase of the last interglaciation.

Dashwood (Semiahmoo) Glaciation

Mapleguard Sediments, exposed at the base of some sea cliffs on eastern Vancouver Island, may record the transition between the Sangamonian Stage and the penultimate (Dashwood) glaciation (Fig. 4, 7). They consist of bedded silt, sand, and minor gravel of fluvial and perhaps deltaic or marine origin (Fyles, 1963; Hicock, 1980). Hicock and Armstrong (1983) proposed that Mapleguard Sediments are younger than, and separate from, the Muir Point Formation. Mapleguard Sediments may be strictly nonglacial in origin, but more likely are outwash deposits laid down during the early part of the penultimate glaciation (Hicock and Armstrong, 1983).

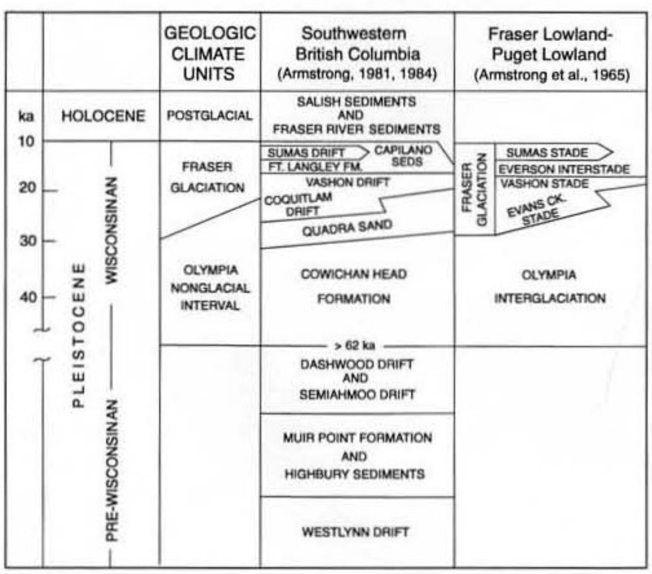

Figure 4. *Subdivisions of Quaternary events and deposits in southwestern British Columbia.*

Figure 5. *Quadra Sand, Point Grey, Vancouver. The lower darker part of the unit contains interbeds of silt and has yielded radiocarbon ages ranging from 26.1 ka BP at the bottom to 24.4 ka BP at the top. GSC 1994-715*

Highbury Sediments and the Muir Point Formation are overlain by a complex of glacial deposits referred to as Dashwood Drift on southern Vancouver Island and Semiahmoo Drift in the Fraser Lowland (Fig. 4, 7; Hicock and Armstrong, 1983). Dashwood Drift consists of till and an overlying unit of glacial-marine silt and silty sand (Fyles, 1963). Hicock and Armstrong (1983) included the previously mentioned Mapleguard Sediments, which they consider to be glacial in origin, in Dashwood Drift. Semiahmoo Drift is more complex, comprising two or more tills interlayered with stratified glacial-marine, glacial-fluvial, and possibly glacial-lacustrine sediments (Armstrong, 1975; Hicock and Armstrong, 1983). Some of the "glacial-fluvial" materials were deposited subaqueously as outwash fans and deltas. The complexity of this unit may be a consequence of it having been deposited near tidewater glaciers that occupied the region during the growth and decay phases of the penultimate glaciation. Dashwood and Semiahmoo drifts resemble, in character and complexity, drift of the Fraser Glaciation, suggesting that the pattern of glaciation on the south coast during the last two glaciations was similar.

All radiocarbon ages from Dashwood and Semiahmoo drifts are beyond the limit of the dating technique (for details on radiocarbon ages see Clague, 1980; Hicock and Armstrong, 1983). It is clear from this and from stratigraphic considerations that the two units are no younger than Early Wisconsinan (oxygen isotope stage 4, ca. 65-80 ka), and that the underlying Muir Point Formation and Highbury Sediments are no younger than Sangamonian. The radiocarbon

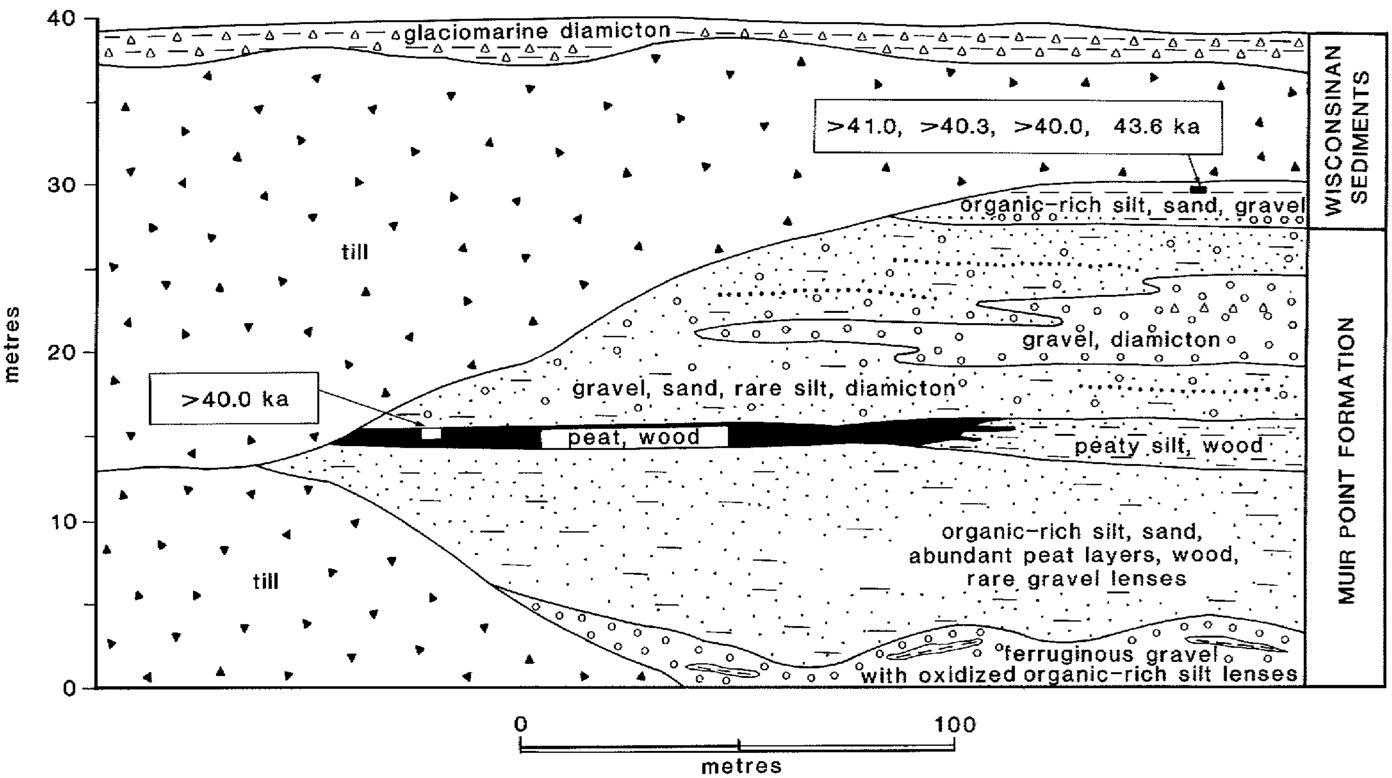

Figure 6. Type section of the Muir Point Formation (adapted from Alley and Hicock, 1986, Fig. 2). Horizontal scale is approximate.

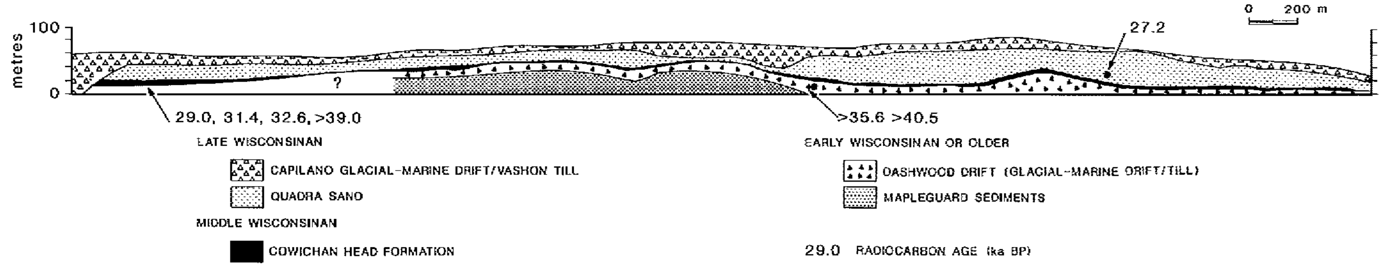

Figure 7. Stratigraphic section, Dashwood sea cliff, Vancouver Island (adapted from Fyles, 1963, Map 1111A).

ages themselves do not preclude the possibility that all of these deposits are older than the Sangamonian Stage. Mean D/L (dextrorotary/levorotary) ratios of aspartic acid in shells of several species recovered from Dashwood and Semiahmoo glacial-marine sediments are greater than ratios from the same taxa in Late Wisconsinan glacial-marine sediments, indicating that the two groups of deposits differ significantly in age (Hicock and Rutter, 1986). Numerical ages, however, have not been determined from these ratios because the temperature history of the fossils is poorly known.

Olympia nonglacial interval

The Cowichan Head Formation overlies Dashwood and Semiahmoo drifts and comprises gravel, sand, silt, and peat deposited in fluvial, estuarine, and marine environments (Armstrong and Clague, 1977). Reliable radiocarbon ages from this unit range from 23.8 ka to 58.8 ka (Clague, 1980), indicating that it was deposited during the Middle Wisconsinan Olympia nonglacial interval, a period when glaciers in British Columbia were confined to mountain ranges (Fulton, 1971; Clague, 1981). In general, the sedimentary deposits, geomorphic framework, and processes of the Olympia interval were similar to those of Holocene time.

Studies of plant microfossils from the Cowichan Head Formation and correlative sediments (Hansen and Easterbrook, 1974; Heusser, 1977; Alley, 1979; Heusser et al., 1980; Armstrong et al., 1985) and of oxygen-isotope fractionation in speleothems dated by uranium-series methods (Gascoyne et al., 1980, 1981) suggest that climate during the Olympia interval varied with time, but generally was cooler and perhaps moister than at present (Clague, 1978; Heusser et al., 1980). Pollen spectra, rich in spruce, western hemlock, mountain hemlock (*Tsuga mertensiana*), and pine (*Pinus*), and with low amounts of Douglas-fir, are markedly different from those of the Muir Point Formation.

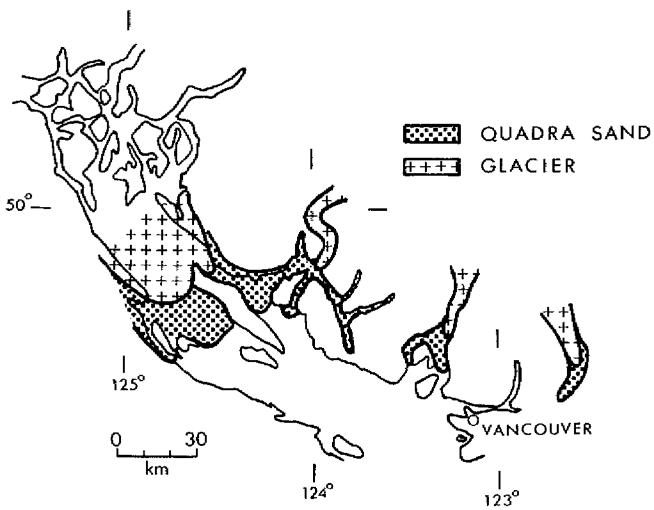

Figure 8. Origin of Quadra Sand (from Clague et al., 1987, Fig. 8). Aprons of sand formed in front of and along the margins of a glacier advancing down the Strait of Georgia during the Fraser Glaciation. The configuration shown in this figure dates to about 25 ka BP.

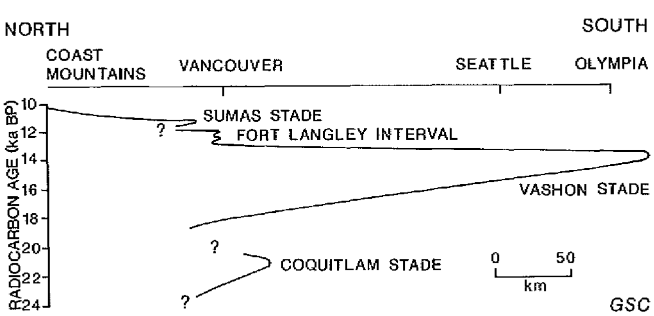

Figure 9. Time-distance diagram showing fluctuations of the margin of the Cordilleran Ice Sheet in southwestern British Columbia and northwestern Washington during the Fraser Glaciation (from Clague, 1991, Fig. 12.7).

Fraser Glaciation

Aprons of thick, well sorted sand (Quadra Sand) were deposited in front of, and perhaps along the margins of, glaciers moving down the Strait of Georgia during the early part of the Fraser Glaciation (Fig. 5, 8; Clague, 1976b, 1977a). The sediment accumulated in channels of braided streams, on adjacent floodplains, and at the fronts of deltas that were prograding into the sea. The unit was progressively overridden and eroded by advancing glaciers. Palynological investigations of organic-rich beds in Quadra Sand and the Cowichan Head Formation indicate a gradual deterioration of climate after about 29 ka BP, marked by the expansion of subalpine plants such as *Bistorta*, *Polemonium*, and *Valeriana sitchensis* into lowland areas (Alley, 1979; Mathewes, 1979). There is no evidence, however, for tundra conditions, even as late as 18 ka BP (Mathewes, 1991).

Quadra Sand was deposited during the long period of glacier growth that preceded the last glacial maximum. This corresponds to the alpine and intense alpine phases of glaciation described by Kerr (1934) and the first and second phases of glaciation of Davis and Mathews (1944), during which glaciers expanded from cirques to form great branching systems of valley glaciers.

Evidence from several mountain valleys north and east of Vancouver indicates that glaciers reached the Fraser Lowland after 25 ka BP. The deposits of this first incursion of Fraser Glaciation ice into the area are termed Coquitlam Drift (Fig. 4; Hicock and Armstrong, 1981). Glaciers retreated from the Fraser Lowland about 19 ka BP when forests, dominated by fir (*Abies*) and spruce, became re-established in the area. These low-elevation forests were adapted to a cool, humid continental climate similar to the present climate of the subalpine zone in the northern British Columbia interior; mean annual temperature was depressed about 8°C and tree line was 1200-1500 m lower than today (Hicock et al., 1982).

Forests in the western Fraser Lowland were overridden by readvancing ice about 17-18 ka BP during the Vashon Stade (Fig. 4, 9). Vashon Drift was deposited at this time, and the

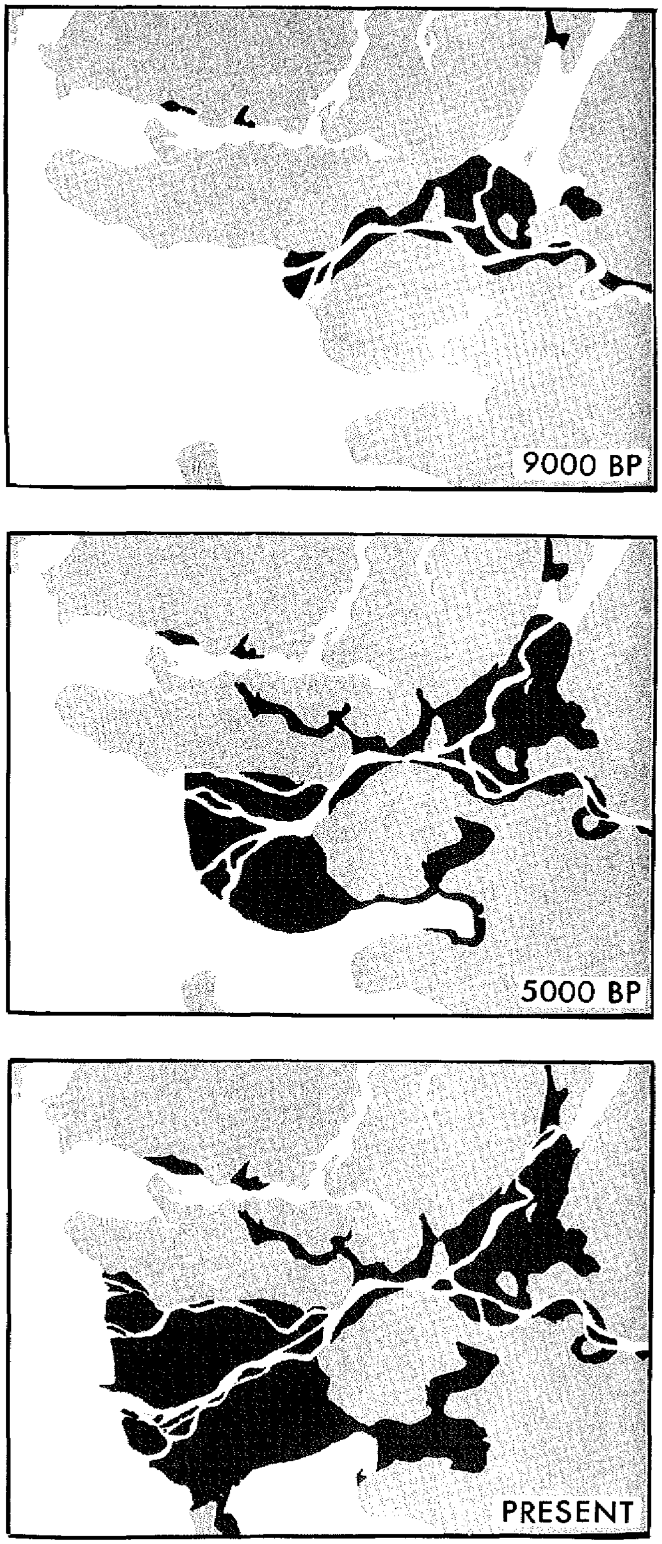

Figure 10. Holocene evolution of the Fraser River delta (from Clague et al., 1991, Fig. 9). Dark shading – Holocene floodplains, fans, and peat bogs; light shading – pre-Holocene landmass. Dates are approximate.

large glacier in the Strait of Georgia advanced to the south end of Puget Lowland (Waitt and Thorson, 1983), achieving its maximum extent about 14.5 ka BP (Mullineaux et al., 1965).

The Puget lobe began to retreat about 14 ka BP, and parts of the Strait of Georgia were ice free by 13 ka BP (Clague, 1980). During retreat, glacial, glacial-marine, and glacial-fluvial sediments accumulated in the Strait of Georgia and on bordering, isostatically depressed lowlands. Glacial-marine sediments deposited beyond the retreating ice margin have been termed Capilano Sediments; interbedded glacial-marine and glacial sediments deposited in eastern Fraser Lowland in an area of fluctuating ice margins are included in the Fort Langley Formation; and till and glacial-fluvial sediments deposited on top of the Fort Langley Formation during a minor readvance near the end of the Fraser Glaciation are termed Sumas Drift (Fig. 4; Armstrong, 1981). Sumas Drift has yielded radiocarbon ages ranging from about 11.5 to 11.1 ka BP (Clague, 1980, 1981; Saunders et al., 1987). The Fraser Lowland probably was completely free of glacier ice by 11 ka BP.

Holocene

The most significant change to the landscape of south-coastal British Columbia during the Holocene has been the formation and growth of the Fraser River floodplain and delta (Fig. 10; Clague et al., 1983, 1991). The lower reaches of the Fraser River occupy a valley that was occupied by the sea and glacier ice at the end of the Pleistocene. The river rapidly prograded its floodplain westward, and began to construct a delta directly into the Strait of Georgia near New Westminster about 8-10 ka BP. The extensive delta plain west and southwest of New Westminster has been built since that time.

Waves and currents have eroded those parts of the coast that are backed by thick Quaternary sediments (Clague and Bornhold, 1980). Some shorelines may have retreated considerable distances during middle and late Holocene time as the sea rose relative to the land (see below).

The climate of south-coastal British Columbia at the end of the Fraser Glaciation was cool and dry, but rapidly ameliorated and probably was as warm as, or warmer than, present from ca. 10 ka BP until at least 7 ka BP (Hansen and Easterbrook, 1974; Mathewes and Rouse, 1975; Heusser, 1977; Heusser et al., 1980; Mathewes and Heusser, 1981; Mathewes, 1985). A wetter cooler climate gradually became established after 7 ka BP, and has persisted to the present.

SEA LEVEL CHANGE AND CRUSTAL DEFORMATION

The Cordilleran Ice Sheet deformed the crust and mantle of western Canada (Clague, 1983). Gradual growth of glaciers during the early part of the Fraser Glaciation induced localized isostatic depression in the southern Coast Mountains. Lateral flow in the asthenosphere away from this area produced an outward-migrating forebulge which, together with eustatic effects, may have caused sea level to fall on the south coast. As glaciers expanded beyond the mountains, however, isostatically depressed areas grew in size, the coast began to subside, and eventually the sea rose above its present level relative to the land.

At the Fraser Glaciation maximum, the entire region was isostatically depressed. The greatest vertical displacements (>250 m) were in areas of maximum ice thicknesses, namely the Coast Mountains, Strait of Georgia, and Fraser Lowland (Clague, 1983). There were lesser displacements on western Vancouver Island and in northern Washington State.

Rapid deglaciation at the end of the Pleistocene was accompanied by isostatic uplift which, in this region, was greater than the coeval eustatic rise. Thus sea level fell as deglaciation progressed (Fig. 11; Mathews et al., 1970; Clague et al., 1982). The elevation of the marine limit declines with increasing distance from the main centres of ice loading. It is highest, about 200 m, on the mainland coast and declines toward the west and southwest to less than 50 m on the west coast of Vancouver Island near the margin of the ice sheet (Mathews et al., 1970; Clague et al., 1982).

Isostatic uplift occurred at different times during deglaciation due to diachronous retreat of the ice sheet (Clague, 1983). In general, areas that were deglaciated first rebounded earlier than those deglaciated later (Fig. 11). Uplift was accompanied by a fall in the level of the sea relative to the land. The pattern of sea level change, however, was complicated by variable rates and directions of eustatic changes, isostatic effects of local glacial stillstands and readvances, and possible displacements along faults.

The sea fell below its present position relative to the land between 12 and 10 ka BP (Mathews et al., 1970; Clague, 1983). Shortly thereafter, it reached its lowest Holocene level, which was at least 10 m below present sea level and perhaps much lower (Fig. 11). A transgression during middle and late Holocene time may be due largely to a eustatic rise in sea level without compensatory isostatic uplift. During the last 2 ka, however, relative sea level has varied no more than 1 m, indicating that isostatic, tectonic, and residual isostatic effects have largely compensated for one another.

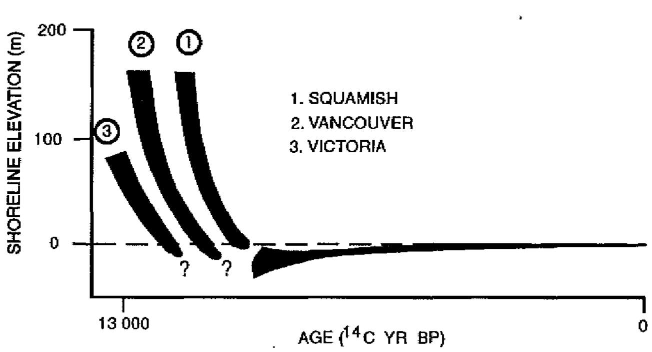

Figure 11. Generalized patterns of sea level change on the south coast of British Columbia since the end of the last glaciation. Deglaciation and isostatic rebound occurred later in the southern Coast Mountains (Squamish) than on Vancouver Island (Victoria). Sea level positions are approximate.

There is evidence for minor (<1 m), rapid changes in sea level on at least two occasions during the last 4 ka, possibly due to large earthquakes. Remnants of late Holocene forests have been found in the intertidal zone at two sites on southern Vancouver Island. The trees were drowned about 2 and 3.5 ka BP when the sea rose suddenly relative to the land. Coincident, rapid, minor (decimetre-scale) uplift and subsidence at several other sites on southern Vancouver Island and mainland British Columbia to the east have been inferred from stratigraphic and plant microfossil data (Mathewes and Clague, 1994). At one site along Serpentine River south of Vancouver, liquefied sediment erupted onto a 2 ka old, downdropped land surface (Mathewes and Clague, 1994). Sand dykes and sand blows are common on the nearby delta of Fraser River and have been attributed to one or more large, late Holocene earthquakes (Clague et al., 1992a)

Earthquakes in southwestern British Columbia and northwestern Washington State may originate: (1) within the North America plate; (2) within the subducting Juan de Fuca plate; and (3) at the boundary between the North America and Juan de Fuca plates (Shedlock and Weaver, 1991; Rogers, 1992). All historical earthquakes in the region are of the first two types. Although the plate boundary is presently aseismic, there is abundant geophysical evidence (e.g., Heaton and Kanamori, 1984; Heaton and Hartzell, 1986; Rogers, 1988; Savage et al., 1991; Hyndman and Wang, 1993) that it is locked, accumulating strain, and thus capable of generating great (M8+) earthquakes. There also is abundant geological evidence that such earthquakes have occurred during late Holocene time, most recently about 300 years ago (Atwater, 1987, 1992; Adams, 1990; Darienzo and Peterson, 1990; Nelson, 1992).

SUMMARY

Events of the Quaternary Period (approximately the last 1.6 Ma) have left a strong imprint on south-coastal British Columbia. Of particular importance has been the repeated waxing and waning of the Cordilleran Ice Sheet. At its maximum, this ice sheet covered the entire south-coastal region and extended southward into Washington State.

Much of the thick succession of Quaternary sediments in the lowlands of southwestern British Columbia was deposited near the margins of the glaciers that periodically advanced into, and retreated out of, the region. Heterogeneous sequences of Pleistocene proglacial marine, fluvial, and deltaic sediments are bounded by unconformities and by nonglacial sediments that are similar to those accumulating in the area today. The unconformities were formed mainly by glacial erosion.

The present landscape has been most strongly shaped by the last, or Fraser, glaciation (ca. 11-30 ka BP). During the climactic advance of the Fraser Glaciation, about 17-18 ka BP, lobes of the Cordilleran Ice Sheet advanced across the

Fraser Lowland and down the Strait of Georgia. Areas below elevations of about 1500 m were covered by ice at the Fraser Glaciation maximum, ca. 15 ka BP. The glaciers eroded Pleistocene sediments and bedrock, and the Strait of Georgia and bordering lowlands attained something close to their present form at this time. During and following deglaciation, glacial-marine, marine, deltaic, and fluvial sediments were deposited in places on the glacially eroded landscape. The locus of much of the postglacial sedimentation has been the Strait of Georgia, fiords, and large lakes.

Growth and decay of the Cordilleran Ice Sheet during Pleistocene time triggered complex vertical movements of the land surface. The expansion of glaciers in the Coast Mountains during the early part of each glaciation was accompanied by localized isostatic depression. As glaciers thickened and advanced into low-lying areas, the area and magnitude of downwarping increased. At times of extensive glaciation, the entire south-coastal region was isostatically depressed, the amount of depression reflecting the thickness of the ice cover. Rapid unloading during periods of deglaciation triggered new isostatic adjustments as material moved laterally in the asthenosphere from extraglacial regions towards the centre of the decaying ice sheet. Peripheral areas, which were deglaciated earliest, rebounded first. As deglaciation progressed, the zone of rapid isostatic uplift migrated in step with receding glacier margins.

Studies of late Quaternary sea level change have shown that most of the south coast experienced rapid isostatic uplift at the end of the last glaciation. Uplift decreased in a nonlinear fashion and, within a few thousand years, shorelines were lower than they are today. The subsequent transgression may record a eustatic rise in sea level that was not compensated by residual isostatic uplift.

Inferred rapid changes in sea level about 2 and 3.5 ka BP may have resulted from subsidence and uplift during large earthquakes. The region almost certainly has experienced other large earthquakes, for which geological evidence has not yet been found. All historical earthquakes in southwestern British Columbia and northwestern Washington State have had hypocentres within either the North American or Juan de Fuca plate, but rare, great (M8+), plate-boundary earthquakes probably also affect the region.

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Contour map of the sub-Quaternary bedrock surface, Strait of Georgia and Fraser Lowland

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Hamilton, T.S. and Ricketts, B.D., 1994: Contour map of the sub-Quaternary bedrock surface, Strait of Georgia and Fraser Lowland; in Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia, (ed.) J.W.H. Monger; Geological Survey of Canada, Bulletin 481, p. 193-196.

Abstract: The sub-Quaternary bedrock surface beneath the Strait of Georgia and Fraser Lowland is depicted on a contour map. For the Strait of Georgia area, map construction is based on marine seismic surveys, whereas for the Fraser Lowland, the map is based on outcrop and limited data from boreholes.

The morphology of the bedrock surface reflects two processes: probable Neogene tectonic subsidence, and Pleistocene glacial erosion. Unconsolidated Pleistocene sediments are 300-500 m thick beneath much of the Fraser delta and the Strait of Georgia, and up to 700 m thick in parts of the Fraser Valley.

Résumé : La surface subquaternaire du substratum rocheux, dans le détroit de Georgia et les basses terres du Fraser, est représentée sur une carte de courbes de niveau. La carte de la région du détroit de Georgia est basée sur des levés sismiques marins et celle des basses terres du Fraser, sur l'étude des affleurements et de données limitées provenant de trous de sondage.

La morphologie de la surface du substratum rocheux reflète deux processus : une subsidence tectonique probable au Néogène, et l'érosion glaciaire au Pléistocène. Les sédiments pléistocènes non consolidés ont de 300 à 500 m d'épaisseur dans une grande partie du delta du Fraser et du détroit de Georgia, et jusqu'à 700 m d'épaisseur dans des portions de la vallée du Fraser.

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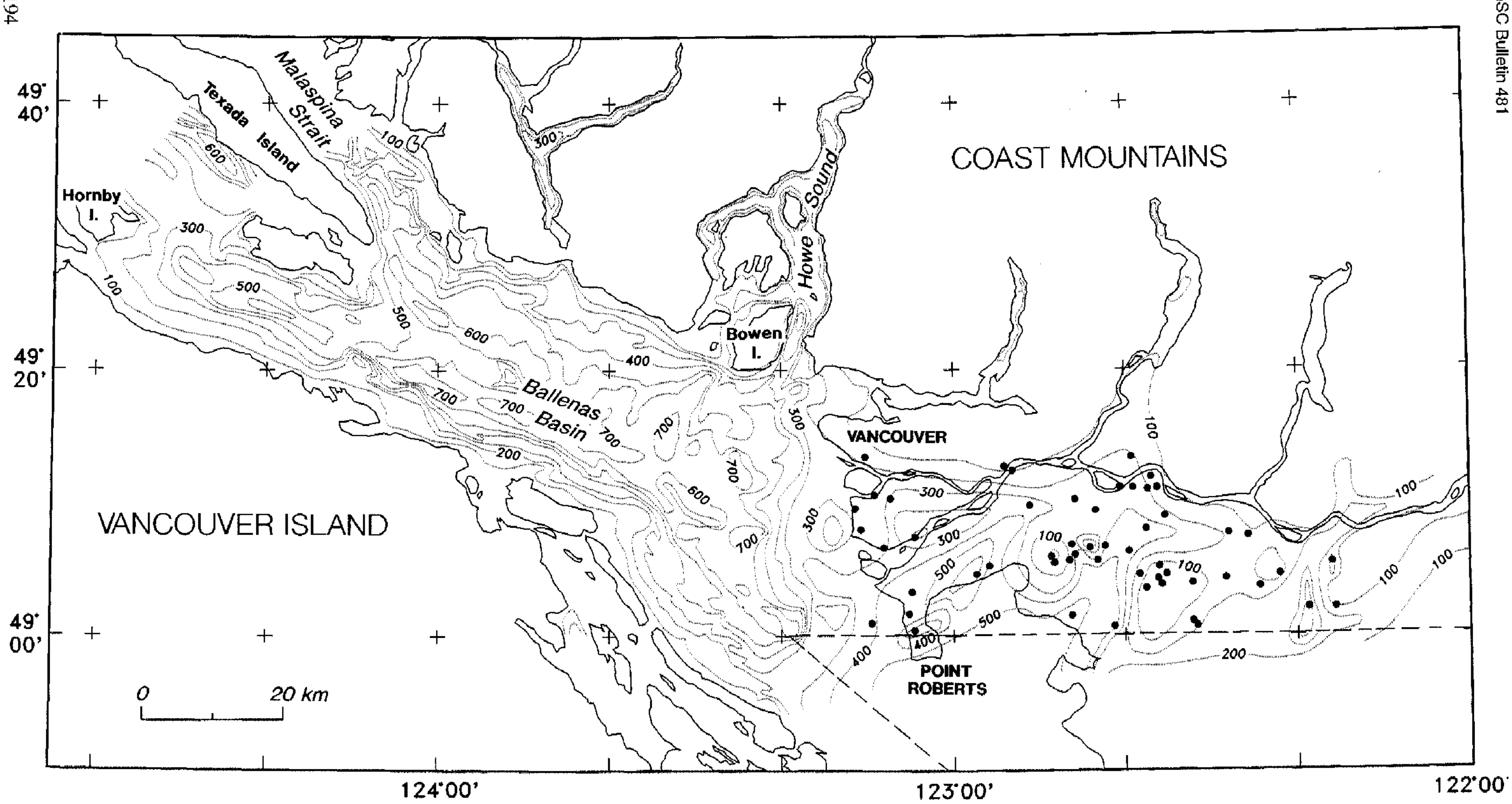

Figure 1. *Subcrop map of the bedrock surface beneath the central and southern Strait of Georgia, Fraser Lowland, and adjacent fiords and fiord lakes. Contour interval is 100 m relative to sea level (e.g., 200 = 200 m below sea level). Note bedrock highs in Surrey and Langley are close to sea level. The black dots indicate wells that control onshore contours. Offshore control is seismic data grid with an average spacing of 2 km (Hamilton et al., 1987). Because a single velocity function was used, thicknesses have an estimated error of ± 25 m. Where thick unconsolidated sediments are mostly low-velocity, Late Holocene delta deposits (lower Fraser River), the depths may be maxima. Where the unconsolidated succession is dominated by high-velocity glacial deposits, the depths are probably minima.*

DATA SOURCES

The subcrop (Fig. 1) map shows the depth to bedrock relative to sea level and contoured at intervals of 100 m. To calculate the total thickness of unconsolidated sediments on land, add the thickness equal to the local elevation above sea level; and offshore subtract local water depth.

Data for the Strait of Georgia section of the map is from an analog seismic survey with a 2 km line spacing (Hamilton et al., 1987); additional digital lines from a multiparameter survey cruise (Hamilton et al., 1990) and lines shown in Hart et al. (1992) and Hart and Hamilton (1993) were also used. A single velocity-depth function (Table 1) was used to convert two-way travel times to depth for the offshore data. There are no offshore wells to confirm the estimated bedrock depths, but seismic velocity measurements in GSC boreholes on land (Luternauer et al., 1986) and in the petroleum exploration borehole at Sunnyside (B.C. Ministry of Energy, Mines and Petroleum Resources, 1981) indicate that the estimates are accurate to within 25 m vertically and 500 m laterally. In nearshore areas and in fiords and lakes, minimum depths to bedrock are based on bathymetry and single seismic profiles along fiord axes.

Onshore borehole data is sparse and derived from hydrocarbon exploration wells (Johnston, 1923; B.C. Ministry of Energy, Mines and Petroleum Resources, 1981), water wells (B.C. Department of Lands, Forests and Water Resources, 1976; Halstead, 1977a, b; Armstrong, 1984), and geotechnical boreholes (Halstead, 1966; Luternauer et al., 1986). Locations of boreholes that provide control on the bedrock surface beneath the Fraser Lowland are shown on the map and the information summarized in Monahan et al. (1993). Much of this information is limiting, in that many (but by no means all) boreholes are completed in Pleistocene sediments above the bedrock.

Seismic data has been collected over parts of the Fraser Lowland by the private sector (B.C. Hydro, 1977; Nevin Sadlier-Brown Goodbrand Ltd., 1985; J. Britton, pers. comm., Dynamic Oil Ltd., 1993) and the public sector (Luternauer et al., 1986; Pullan et al., 1989), but these could not be used rigorously, due to different depths of penetration, processing problems, and proprietary access. Geoelectric sounding data (Nobes et al., 1990) are aerially restricted and their interpretation somewhat uncertain. For these reasons the onshore portion of the map is based solely on well data.

The onshore portion of the map depicts the regionally significant highs and lows on the bedrock surface – the map should only be used as a guide for site-specific studies.

INTERPRETATION

The bedrock surface is a composite erosional surface of Tertiary and Quaternary ages (Peacock, 1935; Mathews, 1972; Hamilton, 1990). Erosional unconformities within the unconsolidated succession (Armstrong, 1984; Hamilton et al., 1987; Hamilton, 1991a) demonstrate that infilling of the antecedent bedrock relief was episodic, and that the landscape was underlain by a combination of bedrock and unconsolidated materials throughout the Quaternary. However, the map provides information that may be usefully applied to a variety of public issues ranging from ground water supply to the response of the ground to seismic shaking.

The most notable feature of the contour map is the northwesterly trend of the contours parallel to the axis of the Strait of Georgia. The bedrock surface beneath the Strait of Georgia slopes 3°-5° from the Sunshine Coast and from Vancouver Island towards the centre of the strait in the southwest. Superimposed on the broad depression underlying the entire Strait of Georgia is a series of smaller elongate depressions.

The bedrock surface is deepest (700 m) beneath bathymetrically deep (350-410 m) portions of the Strait known as Ballenas Basin. This low is not a single trough, but a series of basins extending from Hornby Island to Point Roberts (Hamilton, 1991a, b). Each of these depressions is about 10 km long and 3 km wide. Intervening highs are smaller and more irregular. A smaller trough along Malaspina Strait curves to the south as it enters the Strait of Georgia.

Fiords and fiord lakes occupy deep bedrock depressions. The deepest of these is lower Howe Sound between Horseshoe Bay and Bowen Island.

Most of the area south of the North Arm of the Fraser River and west of New Westminster is underlain by 300 to 500 m of unconsolidated sediments.

Table 1. Two-way time verses depth and average velocities used to reduce offshore seismic data.

Time (ms)	Depth (m)	Velocity ($m \cdot s^{-1}$)
125	100	1600
250	200	1600
365	475	1643
575	500	1684
670	600	1739
760	700	1791
840	800	1842

The gently sloping bedrock basin beneath the Strait of Georgia is probably a forearc basin associated with the Garibaldi Volcanic Belt and the subducting Juan de Fuca Plate (Hamilton and Monger, 1990; Monger and Journeay, 1994). The bedrock surface beneath the fiords and fiord lakes owes its origin to erosion by Pleistocene glaciers flowing out of the Coast Mountains (Peacock, 1935; Mathews, 1972; Hamilton, 1991b). By inference, the northwesterly oriented troughs beneath the Strait of Georgia were carved by large glaciers flowing along the Strait.

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Fraser River delta: geology, geohazards and human impact

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Luternauer, J.L., Barrie, J.V., Christian, H.A., Clague, J.J., Evoy, R.W., Hart, B.S., Hunter, J.A., Killeen, P.G., Kostaschuk, R.A., Mathewes, R.W., Monahan, P.A., Moslow, T.F., Mwenifumbo, C.J., Olynyk, H.W., Patterson, R.T., Pullan, S.E., Roberts, M.C., Robertson, P.K., Tarbotton, M.R., and Woeller, D.J., 1994: Fraser River delta: geology, geohazards and human impact; in Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia, (ed.) J.W.H. Monger; Geological Survey of Canada, Bulletin 481, p. 197-220.

Abstract: The Fraser River delta, south of Vancouver, is the largest delta in western Canada. It is an important agricultural and waterfowl area and a vital link in the Fraser River salmon fishery. It is also an area of rapid urban and industrial growth and lies within the most seismically active region in Canada.

The delta has formed since deglaciation 11 000-13 000 (^{14}C date) years ago. Sand and mud slope deposits constitute the bulk of the deltic sedimentary package.

The delta's distributary channels are estuaries characterized by intrusion of saline water from the Strait of Georgia. Sediment transport is controlled by river and sediment discharges, tidal conditions, and the position of the salinity intrusion. Mud is accumulating on the Sturgeon Bank slope and in the adjacent Strait of Georgia at rates of <1-2 $cm \cdot a^{-1}$. Sand deposition is concentrated off the mouth of Main Channel.

The main geological hazards are failures of the delta slope, burial and scour on the slope, and earthquake-related liquefaction and ground motion amplification. The main areas of concern are the urbanized delta plain, port facilities, a lighthouse, and submarine cables and pipelines on the delta slope.

Human activity has contributed pollutants to the delta and degraded important coastal habitats. Causeways and other large engineered structures, as well as river dredging, have altered sediment and water dispersal patterns and contributed to erosion of parts of the western tidal flats. River training has concentrated sand deposition in small areas, possibly promoting delta slope failures.

Résumé : Le delta du Fraser, au sud de Vancouver, est le plus vaste delta de l'ouest du Canada. C'est une importante région agricole et un important habitat pour la sauvagine, ainsi qu'un chaînon essentiel pour la pêche du saumon dans le fleuve Fraser. C'est aussi une région de développement urbain et industriel rapide qui se trouve dans la région séismique la plus active du Canada.

Le delta se forme depuis la déglaciation il y a entre 11 000 et 13 000 ans (datation par la méthode ^{14}C). Des dépôts sableux et boueux de talus constituent la plus grande partie des sédiments deltaïques.

Les bras du delta constituent des estuaires caractérisés par une marée saline à partir du détroit de Georgia. Le transport solide est contrôlé par les débits fluviatiles et solides, les marées et la position de la marée saline. Les boues s'accumulent sur les talus du banc Sturgeon et dans la région avoisinante du détroit de

Georgia à une vitesse inférieure à 1 ou 2 cm•a-1. L'accumulation de sable se fait principalement au large de l'embouchure du chenal Main.

Les principaux dangers géologiques sont les ruptures du talus du delta, l'enfouissement et l'affouillement survenant sur le talus, et la liquéfaction du sol associée aux séismes ainsi que l'amplification des mouvements du sol. Les principaux secteurs préoccupants sont la plaine deltaïque urbanisée, les installations portuaires, un phare et des câbles et pipelines sous-marins installés sur le talus du delta.

Les activités humaines ont contribué à introduire des polluants dans le delta et ont dégradé d'importants habitats côtiers. Les chaussées et levées et autres grands ouvrages ainsi que le dragage des cours d'eau ont modifié les schémas de dispersion des sédiments et des eaux et contribué à l'érosion de portions des estrans occidentaux. La régularisation des cours d'eau a concentré l'accumulation de sable dans de petits secteurs, et peut-être ainsi favorisé les ruptures du talus du delta.

INTRODUCTION

The Fraser River delta (Fig. 1), just south of Vancouver, British Columbia, is the largest delta in western Canada. It has formed since the disappearance of the Cordilleran ice sheet 11 000-13 000 years ago (Johnston, 1921; Clague et al., 1983). It is an important agricultural and waterfowl area and a vital link in the Fraser River salmon fishery; it is also an area of explosive urban and industrial growth and lies within the most seismically active region in Canada where there is growing concern over the impact of earthquakes (Milne et al., 1978; Byrne and Anderson, 1987; Rogers, 1988; Task Force Report, 1991; Byrne et al., 1992; B.C. Hydro and Klohn Leonoff Ltd., 1992; Rescan Consultants Inc. and Nemetz, 1992; Watts et al., 1992; Luternauer et al., 1993). Urban and industrial activity contributes pollutants to the delta's active sedimentary environments and alters natural sedimentary processes that maintain the delta's ecosystems (Levings, 1975, 1980; Medley and Luternauer, 1976; Birtwell et al., 1983; Standing Committee on the Fraser River Estuary Water Quality Plan, 1990; McLean and Tassone, 1991; Church et al., 1992; Tarbotton et al., 1993). Rising sea level may lead to a loss of coastal wetlands and a rise in the already high water table beneath the urbanized delta plain (Clague, 1989).

The Geological Survey of Canada has addressed geoscience concerns on the Fraser River delta during most of this century, acting either alone or in co-operation with other government agencies, universities, and the private sector. These studies have: a) provided baseline geoscience data against which environmental changes can be measured and which form the underpinning for other natural science surveys, b) identified physical processes that control the delta environment, c) anticipated environmental effects of natural processes and anthropogenic influences, d) developed new technology to more precisely measure and describe the physical character of the delta, e) helped design methods to mitigate unfavourable environmental changes, and f) helped educate people about the geoenvironmental setting in which they live.

This report presents an overview of previous and current earth science studies of the Fraser River delta. A list of referenced literature is included; other pertinent references are provided by Clague (1987) and Missler (1992).

SETTING

The Fraser River delta plain extends 15-23 km west and south from a narrow gap in the Pleistocene uplands at New Westminster and borders the sea along a perimeter of about 40 km (Fig. 1). Twenty-seven kilometres of this perimeter, adjacent to the distributary channels of the Fraser River, face west onto the Strait of Georgia; the remainder face south onto Boundary Bay. These two sections are separated by Point Roberts peninsula, an upland and former island underlain by Pleistocene sediments. Very gently sloping tidal flats and the fringing subtidal part of the delta plain extend up to 9 km from the dyked edge of the delta to the subtidal delta slope. The western delta slope (Fig. 2) is inclined 1-23° (average ~2-3°) towards the marine basin of the Strait of Georgia and terminates at about 300 m water depth, 5-10 km seaward of the edge of the tidal flats. The southern delta slope is ill-defined; it has a gentler gradient than the western delta slope and terminates in much shallower water (about 30 m).

Deposition on the Fraser River delta is controlled by tidal and fluvial processes operating in a high-energy, semi-enclosed marine basin (Luternauer, 1980; Thomson, 1981). Tidal range (4-5 m) and mean tidal height (3 m) in the Strait of Georgia are relatively high, and all distributary channels are tidally influenced. Main Channel, the deepest, is most strongly affected by salt-wedge intrusion, the position of which is controlled by river discharge and tidal height (Kostaschuk and Atwood, 1990; Kostaschuk et al., 1992a).

The Fraser River supplies an average of 17.3 million tonnes of sediment annually to the delta's distributaries, of which 35% is sand (McLean and Tassone, 1991; Church et al., 1992). Main Channel carries 80% of the total load (Milliman, 1980), including about 2.8 million tonnes of fine-to very course-grained sand (0.18-2.0 mm) which constitutes the bed material of the channel (McLean and Tassone, 1991; Church et al., 1992). Sediment and water discharge exhibit strong seasonal variations and are greatest during the late spring-early summer freshet (Thomson, 1981). Prior to dredging, dyking, and jetty construction which began in the late nineteenth century (Public Works Canada, 1949; H.H. Nesbitt-Porter, B.C. Ministry of Environment, pers. comm., 1993), sand was deposited on the floodplain, tidal flats, and delta slope adjacent to distributary channels.

SURFACE SEDIMENTS AND PRESENT-DAY SEDIMENTARY ENVIRONMENTS

Present sedimentary environments of the Fraser River delta include bogs, the floodplain, estuarine river channels, tidal flats, and the delta slope (Clague et al., 1983) (Fig. 1, 2). Large domed peat bogs cover most of the eastern delta plain (Johnston, 1921; Hebda, 1977). They have developed on a poorly drained substrate close to and locally below low tide level. The continuity of the large peat bogs on the eastern Fraser River delta attests to the stability of distributary channels there during late Holocene time. At only one place is there a gap in the peat deposits not now occupied by an active distributary channel (Fig. 1). The stability of channels on this part of the delta contrasts with their instability on the tidal flats, where there have been marked shifts in channel positions historically (Clague et al., 1983; Luternauer and Finn, 1983) (Fig. 3).

Surface sediments of the dyked portion of the delta plain are sandy to clayey silt deposited in overbank and uppermost intertidal environments (Armstrong and Hicock, 1979, 1980; Armstrong, 1984). Prior to dyking, these sediments

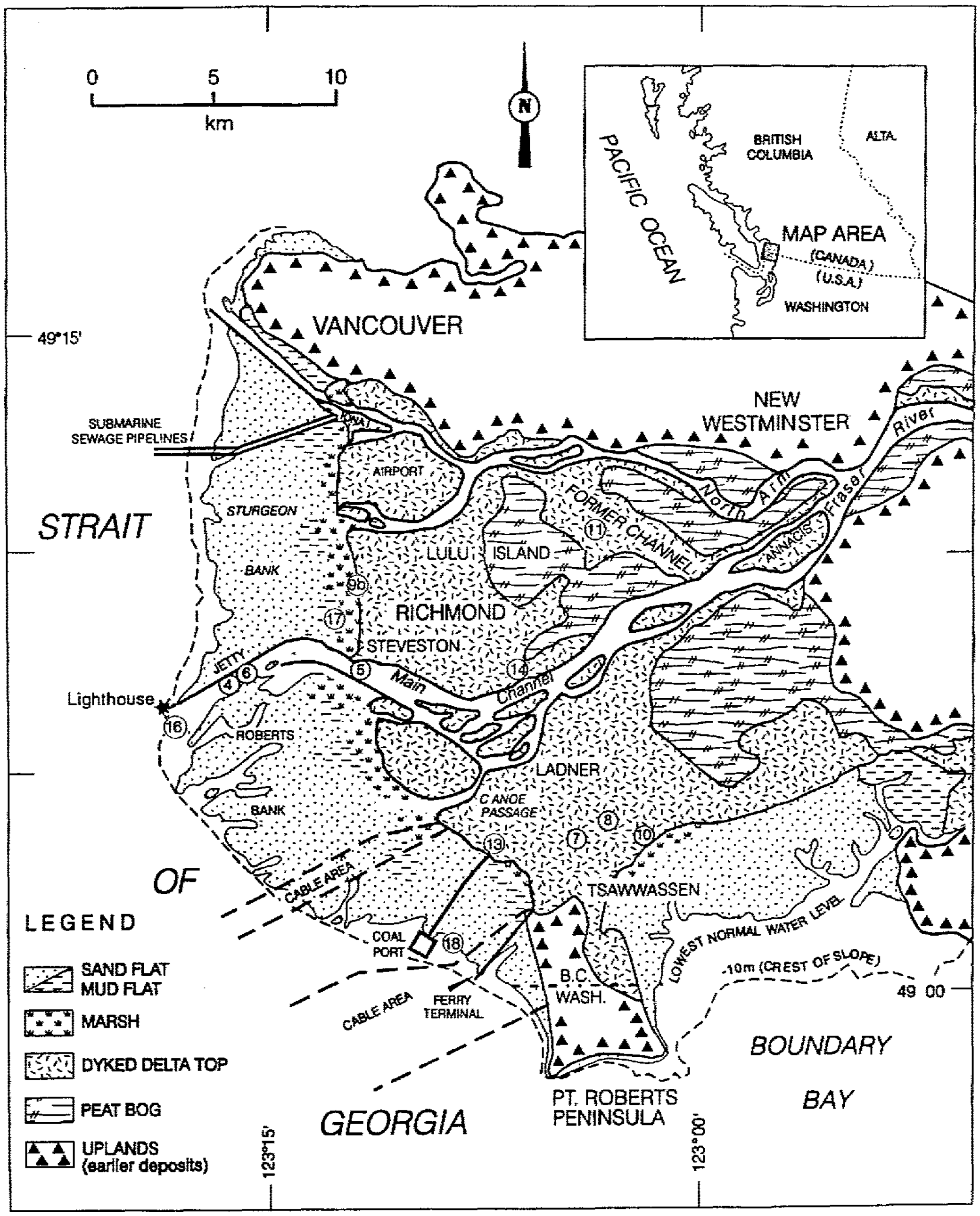


Figure 1. Location and setting of the Fraser River delta. Circled numbers identify general locations of other figures referred to in the text.

accumulated during spring floods and at times of very high tides. The dredged North and Main distributary channels are as deep as 22 m below sea level. They are floored by sand and, locally, gravel, and include sandy islands (Johnston, 1921; Mathews and Shepard, 1962) (Fig. 1).

The estuary of the Fraser River extends 75 km upstream to the limit of tidal rise near Mission, and includes the main distributaries (Main Channel and North Arm) and several smaller channels. Salinity stratification in the estuary varies with tidal range and river discharge (Kostaschuk and Atwood, 1990).

Sand bed-material in Main Channel is moulded by river currents into large dunes which locally are more than 5 m high and 100 m long (Kostaschuk and MacDonald, 1988) (Fig. 4). Replicate surveys of dune trains (Fig. 4) have provided information on bed material transport and thus have helped define

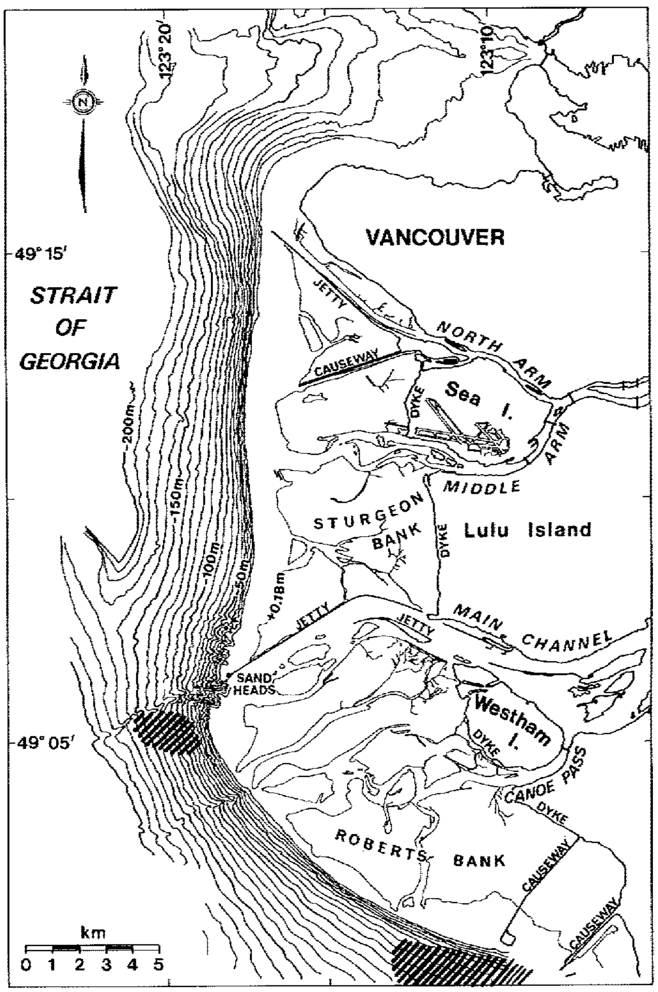


Figure 2. Morphology of the western delta front and prodelta of the Fraser River. Cross hatching delineates two zones of prominent hummocky morphology (Luternauer, 1980). Features within the zone off Sand Heads are interpreted to be shallow rotational slides (Hart et al., 1992a); those on the southern part of the Roberts Bank slope are dunes (Kostaschuk and Luternauer, 1993).

the sediment budget of the delta front. Changes in dune height and length are associated with, but lag behind, seasonal changes in river discharge (Kostaschuk et al., 1989a, 1990) and weekly variations in tidal range (Ilersich, 1992). Sand is transported mainly in suspension, resulting in dunes without lee-side flow separation and with lee-side slope angles far smaller than the friction angle of bed material (Ilersich, 1992). Nevertheless, dunes in Main Channel can cause large turbulent vortices, termed kolks (Kostaschuk et al., 1991; Kostaschuk and Church, 1993). The vortices are generated on the lower upstream sides of dunes (Fig. 5a), appear at the surface tens of metres downstream as boils, and are an important mechanism in entraining and suspending sand (Fig. 5b).

Suspended sediment transport in Main Channel is subject to a time lag, or hysteresis (Milliman, 1980; Kostaschuk et al., 1986, 1989b; Kostaschuk and Luternauer, 1989). Fluvial hysteresis is evident after fine mobile sediment is exhausted early in the spring snowmelt freshet; the peak in daily sediment concentration precedes the peak in river discharge. Tidal hysteresis is apparent during the tidal cycle; the peak in sediment concentration follows the peak in tidal current velocity. This arises from enhanced turbulence as tidal currents decelerate.

Suspended sediment is transported seaward over the salt-wedge in Main Channel (Kostaschuk et al., 1992a) (Fig. 6). The thickness of the fresh water layer above the wedge decreases seaward of the wedge tip, but salinity and velocity increase. Sediment concentration decreases seaward of the wedge tip because of interference in the exchange of sediment between the flow and the bed, reduced turbulence, flocculation of fine sediment, and dilution of the sediment-water mixture. Suspended sediment accumulates near the tip of the salt-wedge, forming a turbidity maximum (Nichols and Biggs, 1985; Kostaschuk et al., 1992a) (Fig. 6). This appears to be an ephemeral phenomenon that is eliminated as the salt wedge is flushed out of the channel at low tide, but it does allow selected pollutants to be further concentrated on host sediments.

Beyond the mouth of the river, Main Channel effluent forms a plume in the Strait of Georgia (Thomson, 1981; Kostaschuk et al., 1989c, 1993; Liedtke et al., in press). When the salt-wedge is pushed out of the estuary channel and river water is in direct contact with the bed, sand bed-material is resuspended and sediment concentrations increase in the plume. When the wedge lies within the channel and flow is stratified at the mouth, concentrations of both fine and coarse sediment are reduced in the plume. Sediment concentration in the plume declines exponentially with distance seaward of the river mouth because the plume is isolated from the bed by a strong pycnocline, there is a seaward decrease in plume velocity, and fine sediment particles flocculate.

Tidal flats are mantled mainly by rippled, fine-to medium-grained sand (Luternauer and Murray, 1973; Swinbanks and Murray, 1981) (Fig. 1). A discontinuous fringe of marsh, underlain by muddy sediments, marks the landward edge of this zone (Luternauer and Murray, 1973; Swinbanks and Murray, 1981; Hutchinson, 1982) (Fig. 1). Sand swells with heights of 0.5 m and wave lengths of 50-100 m are common

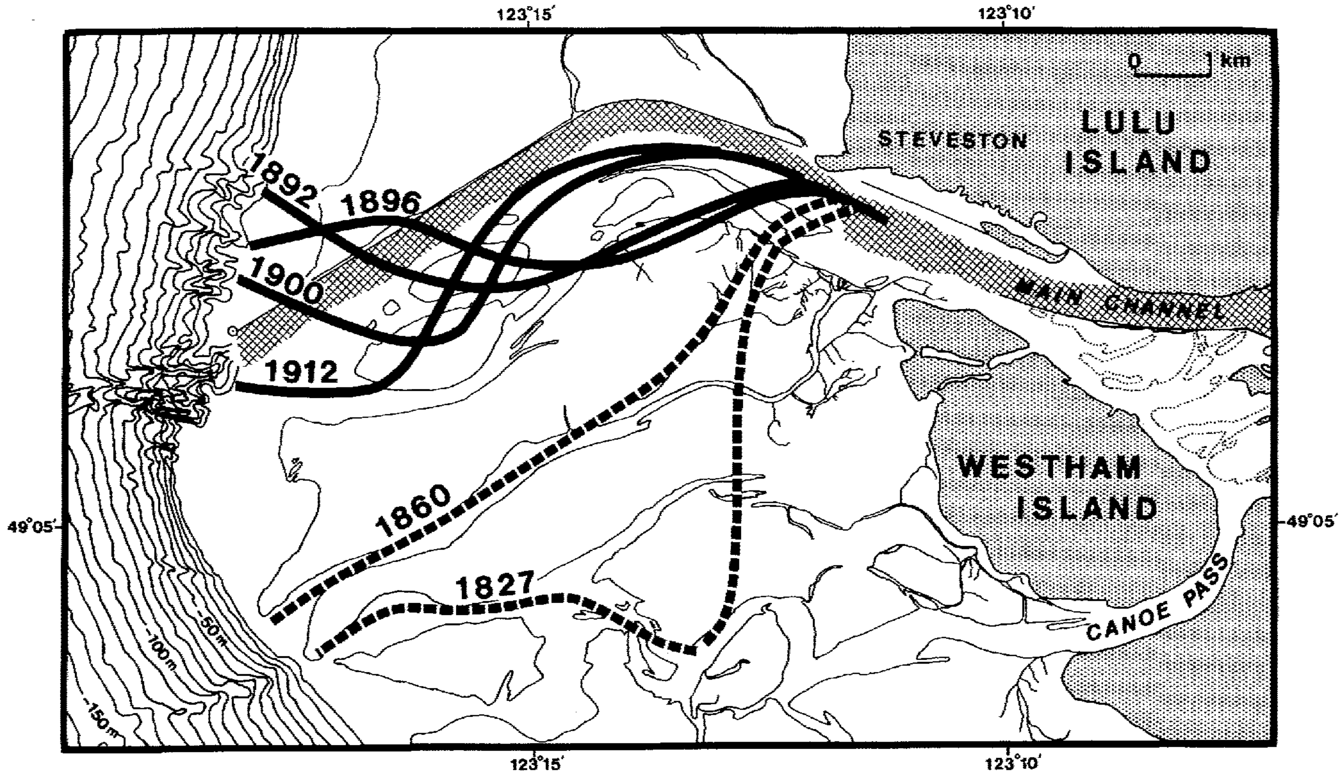


Figure 3. Historical changes of the Fraser River Main Channel on the western tidal flats. (from Luternauer and Finn, 1983).

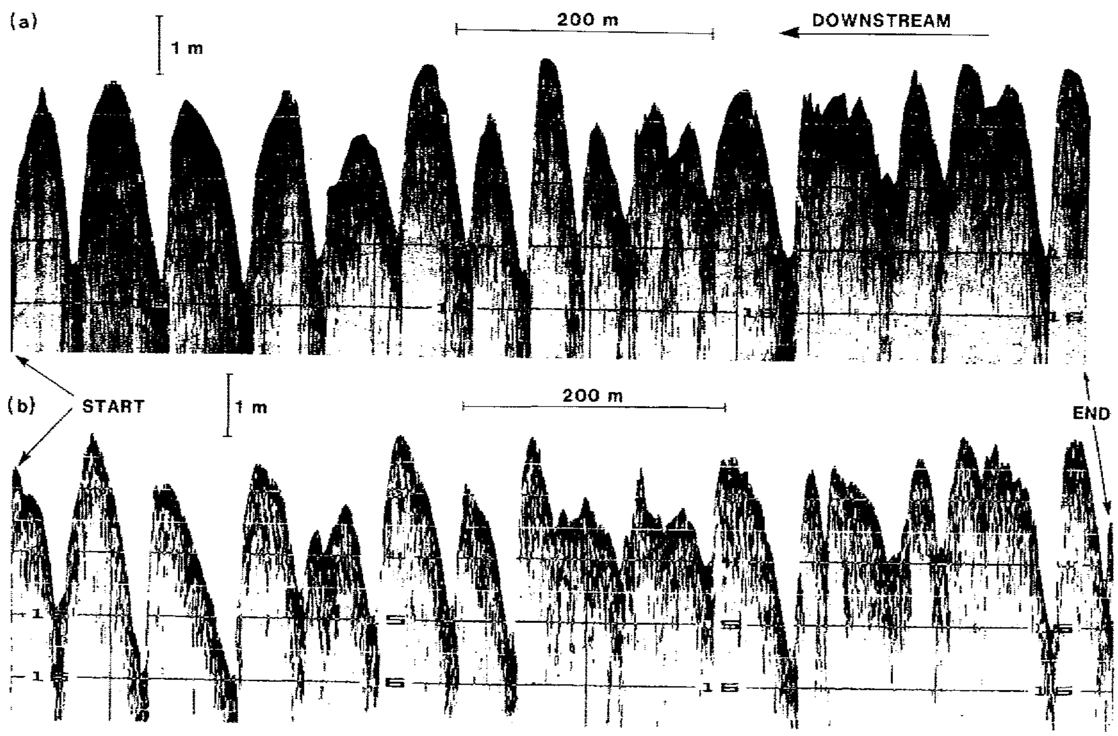


Figure 4. Replicate echo-sounding records of the outer Main Channel, showing dune migration and some evolution during a 24 hour period in June 1986. The average migration over this period was 14.8 m. (from Kostaschuk et al., 1989a).

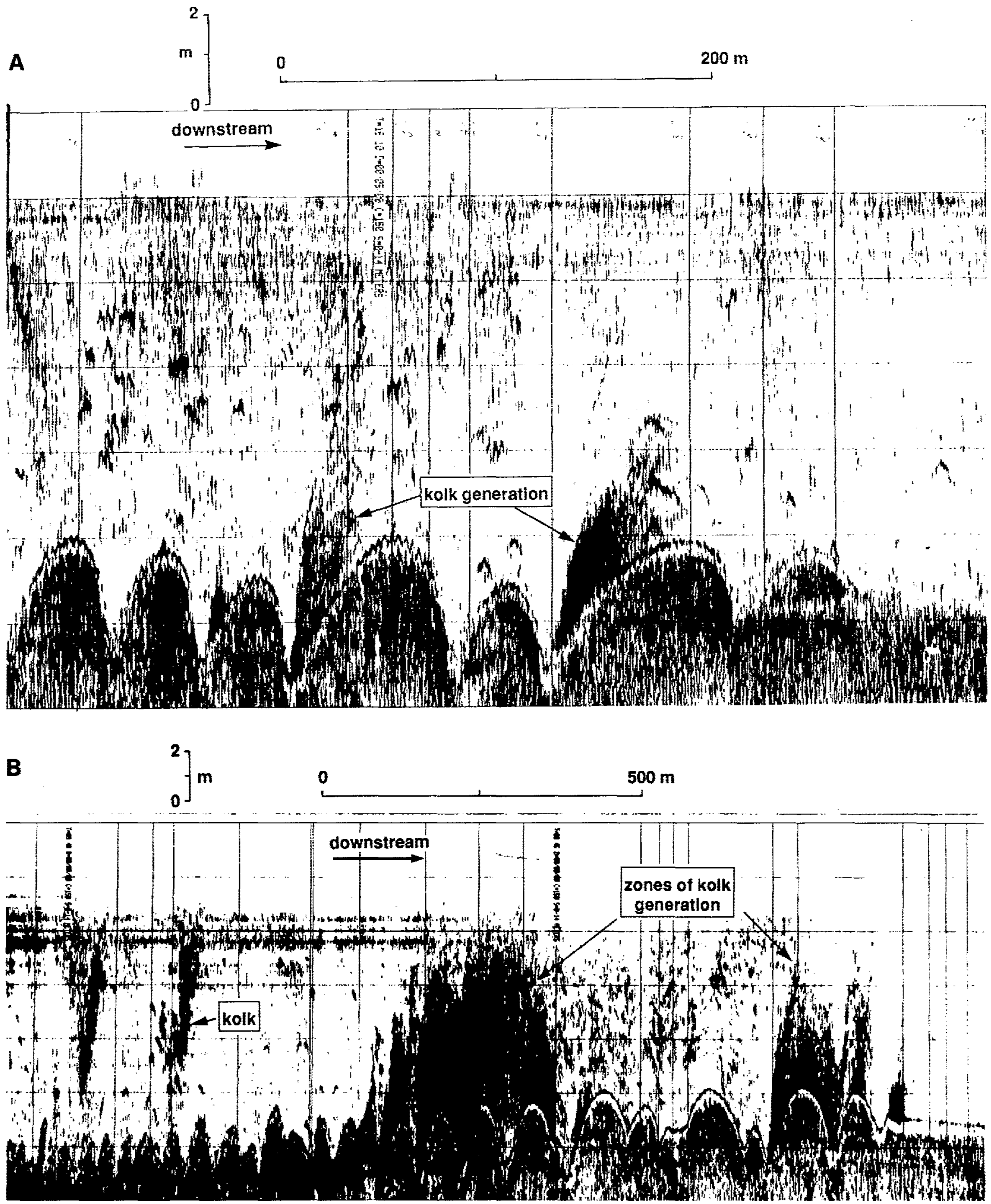


Figure 5 A) Sounding record collected over the dune field in Main Channel near Steveston during a rising tide, showing localized generation of turbulent vortices (kolks) on the lower upstream sides of dunes. The large upstream kolk extends at least 10 m towards the surface (from Kostaschuk et al., 1991), B) Sounding record collected over the dune field in Main Channel near Steveston during a falling tide. Two kolks on the left part of record probably were generated upstream but have sheared from the bed and propagated downstream (from Kostaschuk et al., 1991). The strong acoustic signature of kolks is likely due to high suspended sediment concentrations.

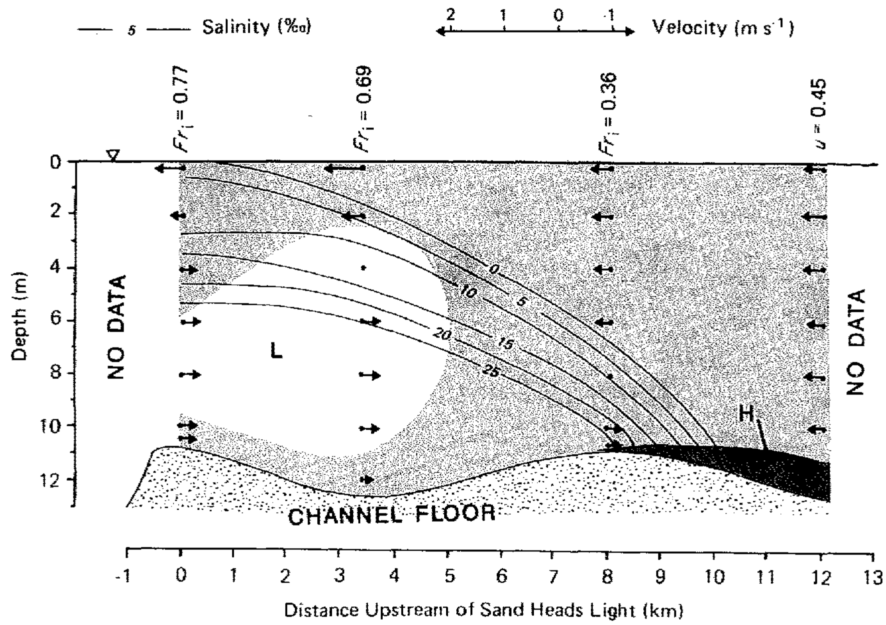


Figure 6. Representative flow characteristics and sediment concentrations in Main Channel from Sand Heads to Steveston during a rising tide; u is mean downstream flow velocity for the site farthest upstream, Fri is interfacial Froude number, and H and L are zones of highest and lowest suspended sediment concentrations, respectively. The river bed is stippled (adapted from Kostaschuk et al., 1992a).

on the tidal flats of the western delta front and result from the movement of sediment by wind-generated waves and currents (Luternauer, 1980).

Sediment dynamics and sediment-pollutant associations on the tidal flats are not well known, but are currently being studied using a variety of instruments and methods, including a field flume called the Sea Carousel (Amos et al., 1992), current meters, sampling of suspended sediment, precise geodetic surveys using global positioning system (GPS) equipment, and the Canadian Coast Guard hovercraft (Aitken and Feeney, 1994; Feeney, 1994). Preliminary results suggest that fine grained sediments on the inner part of Sturgeon Bank off Iona Island (Fig. 1), which contain the highest concentration of pollutants, are being reworked in place, and are not being actively eroded and transported or buried by uncontaminated sediments (Feeney et al., 1994). Contaminated sediments may, therefore, remain available to coastal biota for an indeterminate period in spite of the fact that most effluent is now being discharged offshore.

Delta slope sediments range from fine grained sand to mud (Pharo and Barnes, 1976; Luternauer, 1976, 1980; Clague and Luternauer, 1982; Hart et al., 1992a; Evoy et al., 1993a, 1994; Kostaschuk and Luternauer, 1993). The coarsest sediments occur on the upper slope near the mouths of active distributary channels and over most of the slope south of Main Channel. Sand is widespread in the latter area, not because of abundant supply, but because the oceanographic regime both limits the amount of finer sediment carried to the area and inhibits its accumulation (Luternauer, 1980; Stewart and Tassone, 1989; Kostaschuk and Luternauer, 1993).

West and north of the mouth of Main Channel, delta slope sediments gradually fine from fine grained sand to very fine grained sandy mud or mud with increasing water depth and distance from this primary sediment source. Near the mouth of Main Channel, deposits at the top of the slope are characterized by decimetre- and centimetre-scale interbeds of sand and mud which reflect seasonal and tidal discharge fluctuations (Johnston, 1922; Hart et al., 1992a). Sedimentation rates range from <1 to 2 $cm \cdot a^{-1}$ over much of the Sturgeon Bank slope and in the central Strait of Georgia, but can be an order of magnitude higher at the mouth of Main Channel (Mathews and Shepard, 1962; Hart, 1992; Hart et al., 1992a; Evoy et al., 1993b; Moslow et al., 1993).

Delta growth has been estimated most recently using ^{137}Cs and ^{14}C dating techniques (Clague et al., 1983; Luternauer et al., 1991; Evoy et al., 1993b; Hart et al., 1993; Moslow et al., 1993). In the past, other methods have been employed to estimate sediment age (Williams and Roberts, 1989, 1990; Berger et al., 1990) and to calculate sedimentation rates and identify erosion or deposition of submarine parts of the delta (Johnston, 1921; Mathews and Shepard, 1962; Luternauer and Murray, 1973; Pharo and Barnes, 1976; Luternauer et al., 1983; Stewart and Tassone, 1989).

The slope is cut by submarine channels or sea valleys, formed and maintained by slumps, turbidity currents and, to a lesser degree, debris flows (Mathews and Shepard, 1962; Luternauer, 1980; Kostaschuk et al., 1989c, 1992b; Atkins and Luternauer, 1991; Moslow et al., 1991, 1993; Hart et al., 1992a, b, 1993; McKenna et al., 1992; Evoy et al., 1994) (Fig. 2, 3). Shallow rotational slides and large migrating dunes occur on the northwestern and southwestern slopes of Roberts Bank, respectively, (Luternauer, 1980; Hart et al., 1992a, b; Kostaschuk and Luternauer, 1993) (Fig. 2).

Interstitial gas, generated by bacterial degradation of organic detritus, is widespread in delta slope and prodelta sediments (Hart and Hamilton, 1993) and has been detected beneath the delta plain (Pullan and Hunter, 1987; Pullan et al., 1989).

STRATIGRAPHY

Onshore and offshore drilling and seismic programs conducted by the Geological Survey of Canada and Simon Fraser University, and a large number of engineering test holes, notably cone penetration tests (CPTs), have provided information on the distribution and character of subsurface sediments of the Fraser River delta (Clague et al., 1983,

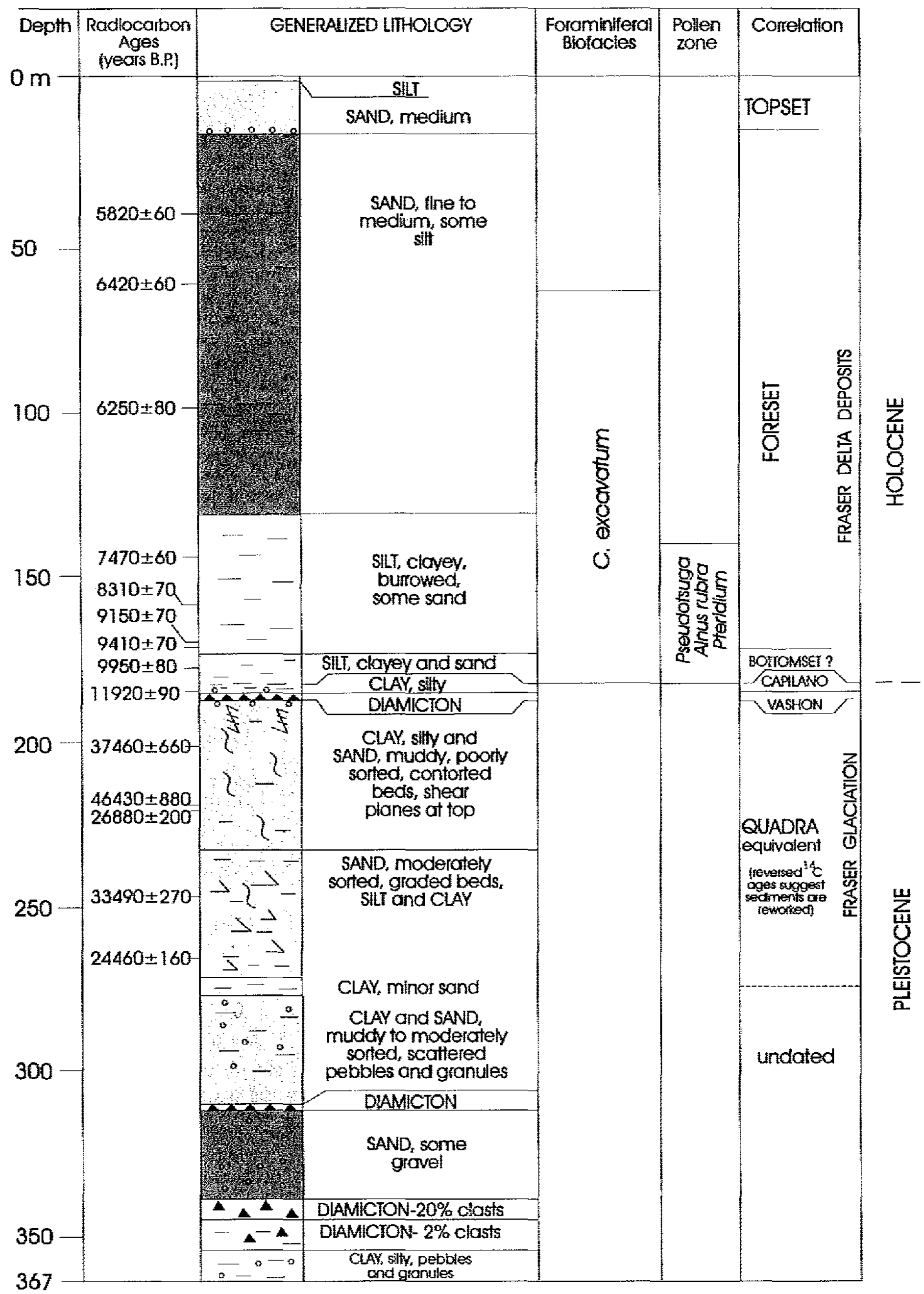


Figure 7. Late Quaternary lithology, chronology, early Holocene biofacies, and proposed stratigraphic correlations for GSC core FD87-1 (Fig. 1) from the southern part of the Fraser River delta.

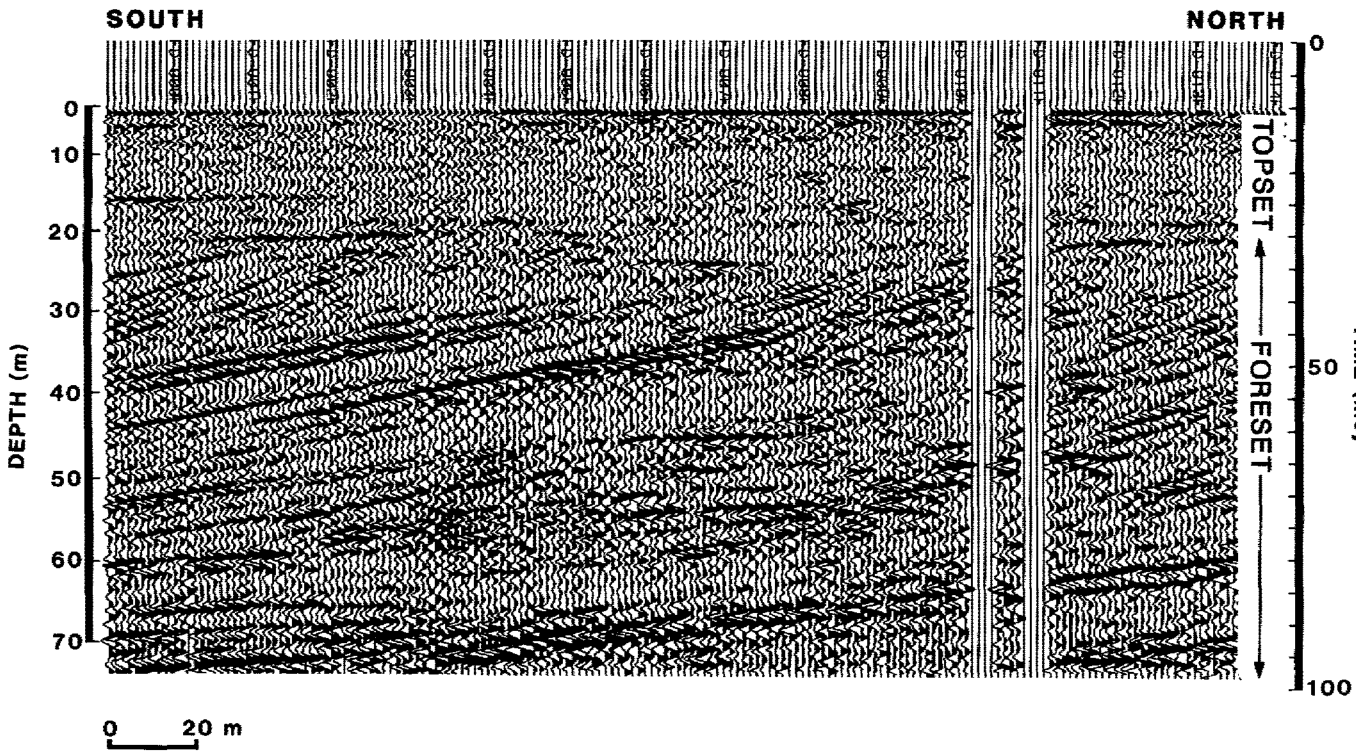


Figure 8. Seismic reflection profile from the southern part of the Fraser River delta showing flat-lying topset beds sharply overlying inclined foreset beds (from Clague et al., 1991b).

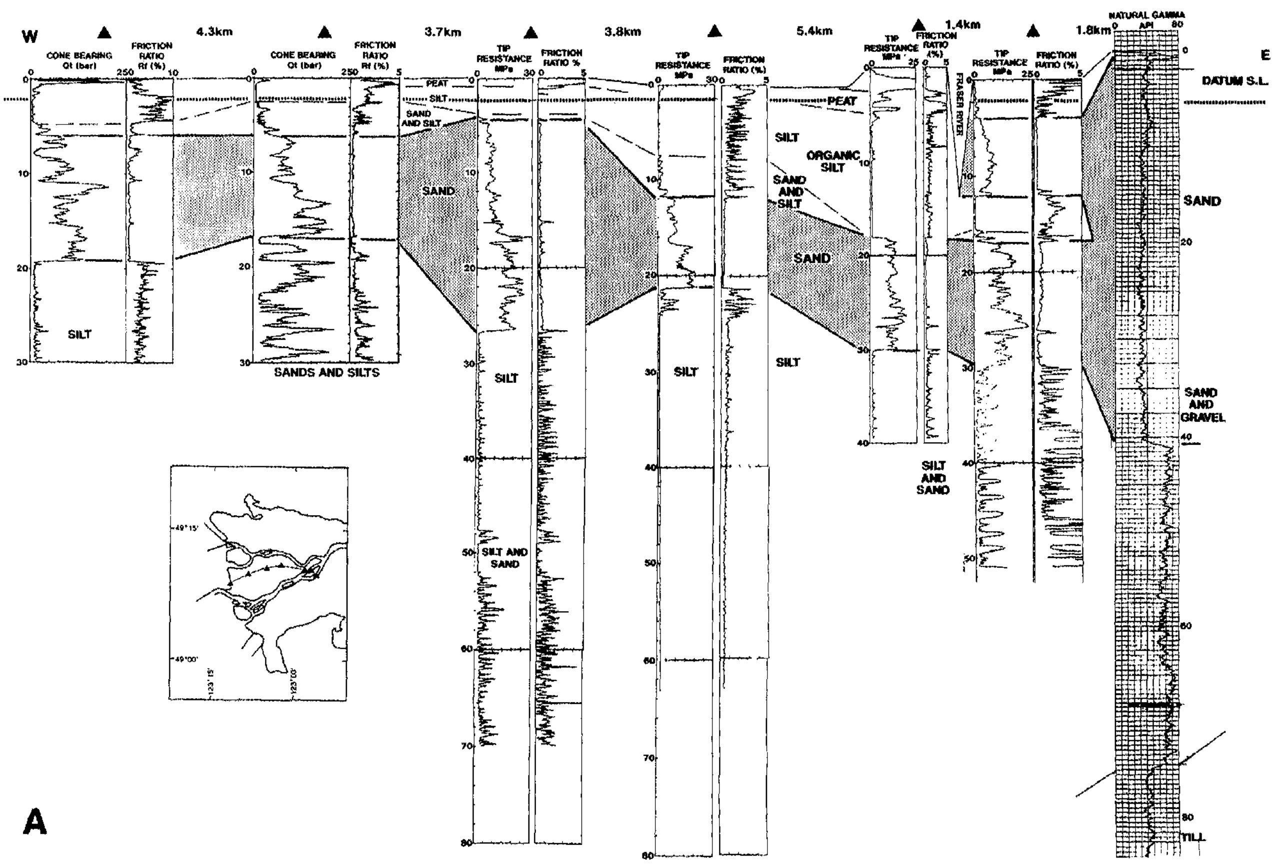
1991b; Roberts et al., 1985; Pullan et al., 1989; Williams and Roberts, 1989; Hamilton, 1991; Jol and Roberts, 1992; Hart and Hamilton, 1993; Monahan et al., 1993a, c; Woeller et al., 1993a, b).

Deltaic sediments have a maximum known thickness of 236 m (unpublished GSC data; see also Johnston, 1921, 1922), and overlie late Pleistocene till and stratified drift (Clague et al., 1983, 1991b; Hamilton, 1991; Hart and Hamilton, 1993; Luternauer et al., 1993) (Fig. 7). Relative sea level rose the time in which the delta prograded into the Strait of Georgia (Clague et al., 1983; Williams and Roberts, 1989); consequently, topset sediments thin to the west from 40 m at the apex of the delta to 20 m or less at the western margin of the dyked delta plain (Clague et al., 1983; Monahan et al., 1993a) (Fig. 8, 9a).

The lowest topset deposits form a sand unit which is 10-20 m thick, commonly has a sharp base with several metres of local relief, fines upward, and is interpreted to be a complex of distributary channel sands (Fig. 8, 9a, b). The sand unit is nearly continuous under the dyked delta plain and the inshore part of the western tidal flats. It is not present, however, beneath the seaward part of the tidal flats, where surficial silty sand has a gradational lower contact with interbedded silt and sand deposited on the delta slope. The distributary channel sand complex was deposited in channels that migrated across and eroded the former tidal flats and, to a lesser extent, the floodplain (Monahan et al., 1993a, b, c). In addition to active channel-fill sand, partially abandoned channel-fill silty sand can be recognized locally. At the apex of the delta, distributary channel sand units are interbedded with silty topset deposits, or are stacked so that the entire topset sequence consists of sand (Fig. 9a). Recognition of this extensive distributary channel sand complex explains the widespread occurrence of sand that is susceptible to liquefaction in the Fraser River delta: channel sand tends to be looser than sand deposited in a shoreface or delta front environment and compacted or densified through wave action (Pryor, 1973; de Mulder and Westerhoff, 1985).

The distributary channel sands commonly are capped by burrowed silty sand and silt containing a marine fauna (Williams, 1988; Williams and Roberts, 1989; Monahan et al., 1993a, c). On the dyked delta plain, organic-rich mud deposited in a floodplain environment generally overlies the tidal flat deposits, but in some areas this mud directly overlies distributary channel deposits. Intertidal and floodplain mud and sand are over 15 m thick near the apex of the delta, where they are as old as 8000 (^{14}C) years. They thin to the west to 5 m at the western margin of the dyked delta plain (Clague et al., 1983; Williams and Roberts, 1989, their Fig. 5) (Fig. 9a). Peat up to 8 m thick covers much of the eastern part of the delta (Fig. 1).

Seismic data (Jol, 1988; Jol and Roberts, 1988, 1992; Pullan et al., 1989; Clague et al., 1991b; Roberts et al., 1992) (Fig. 8, 10) and cone penetrometer resr CCPT correlations (Monahan et al., 1993c) demonstrate that foreset strata dip up to 7° seaward. Foreset sediments on the southernmost part of


W
E
4.3km
3.7km
3.8km
5.4km
1.4km
1.8km
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TIP RESISTANCE MPa
FRICTION RATIO %
FRICTION RATIO (%)
NATURAL GAMMA API
DATUM S.L.
PEAT
SILT
SAND AND SILT
SAND
SANDS AND SILTS
SILT AND SAND
ORGANIC SILT
FRASER RIVER
SAND AND GRAVEL
TILL
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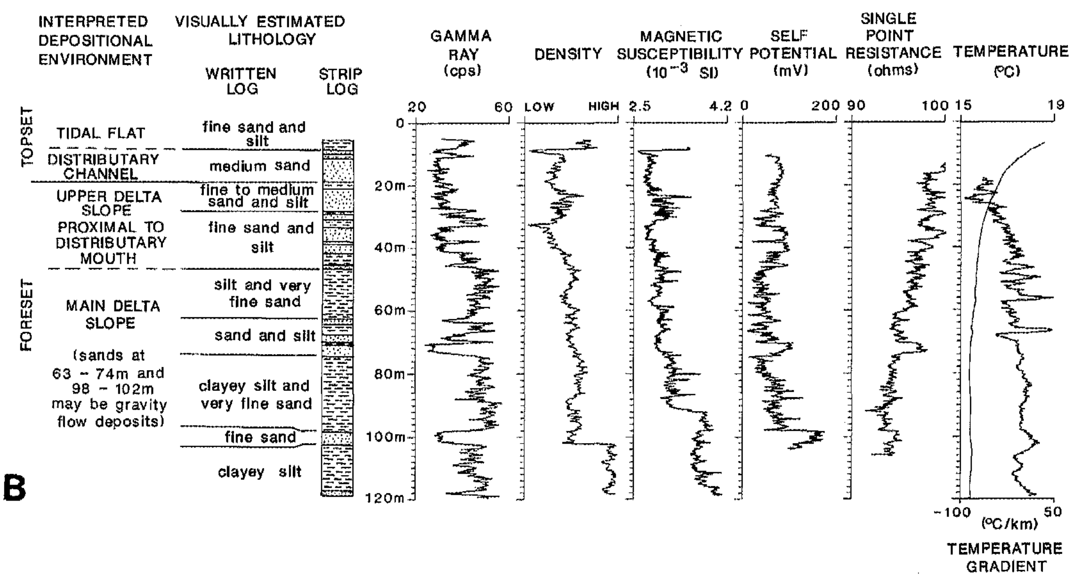

Figure 9 **A)** *East-west stratigraphic section across the subaerial delta plain showing continuity of the topset channel-sand unit (stippled). The data were derived from cone penetration tests and a gamma-ray log, vertical scale in metres (from Monahan et al., 1993a).* **B)** *An example of the display and interpretation of geological and geophysical data, from a borehole in Richmond, complementing those shown in Figure 9a. The gamma-ray log indicates that the highest radioactivity is associated with clayey silt and the lowest with sand. The density log reflects variations in bulk density, porosity, water content, and chemical composition; density information is derived from back-scattered gamma rays in the 180-500 keV energy window. The magnetic susceptibility log provides an indication of the amount of ferromagnetic minerals in the sediment. Self potential and single point resistance identify sand layers which tend to have relatively high resistances. The temperature gradient anomalies between 40 and 70 m probably result from groundwater flow or circulation of drilling fluids.*

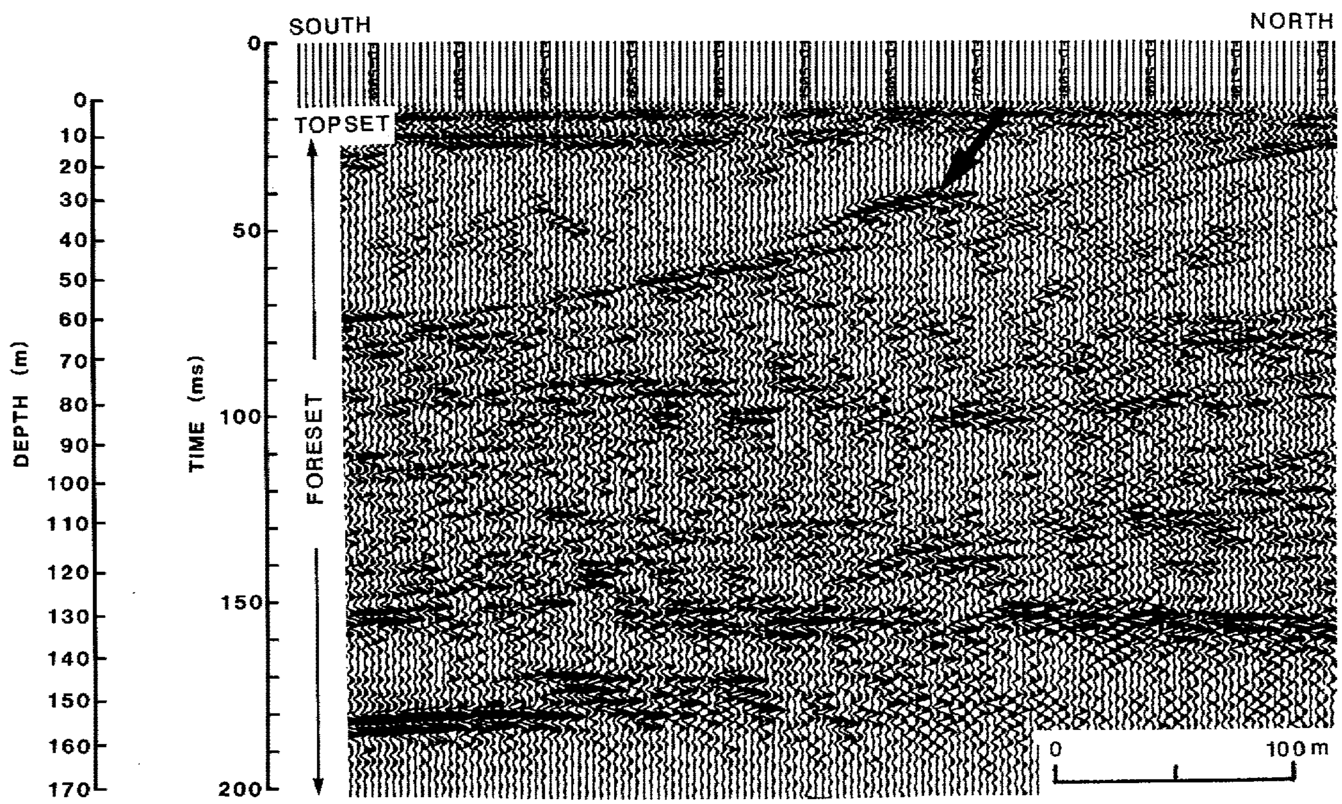

Figure 10. Seismic resolution profile from the southern part of the delta showing topset and foreset beds and a buried failure deposit (arrow) within the foreset sequence (from Roberts et al., 1992).

the delta consist primarily of sand (Fig. 7), but elsewhere they are mainly silt with localized sand-dominated units up to 30 m thick (Clague et al., 1991b; Luternauer et al., 1991, 1993; Monahan, 1993) (Fig. 9a, b). Foreset sand is interstratified with silt, and is interpreted to have been deposited close to an active distributary mouth. Correlations of borehole logs suggest that some sandy units fill channels incised into older foreset deposits. These channels are analogous to recent failures at the modern river mouth reported by McKenna et al. (1992), and to the submarine channel on the delta slope seaward of the river mouth, studied by Kostaschuk et al. (1992b) and Hart et al. (1992b, c). Failure deposits and channels also have been recognized in the foreset sequence from seismic reflection data (Jol, 1988; Jol and Roberts, 1992) (Fig. 10). Deeper in the foreset sequence, upward-fining beds of fine-to medium-grained sand are interlayered with silt (Monahan, 1993) (Fig. 9a). These sand beds are interpreted to be gravity flow deposits that by-passed the upper foreslope, and are analogous to gravity flow deposits reported by Hart et al. (1992a, b, c) and Evoy et al. (1994) near the base of the modern delta slope.

Commonly, foreset deposits unconformably overlie Pleistocene sediments. In low areas on the Pleistocene surface, however, foreset sediments overlie flat to gently dipping clay, silt, and minor sand that accumulated more slowly than the overlying deposits (Fig. 7). These constitute the bottomset of the delta and are analogous to sediments that are presently being deposited on the floor of the Strait of Georgia off the delta front. Bottomset sediments conformably overlie Late Pleistocene glaciomarine deposits (Hamilton, 1991; Hart and Hamilton, 1993) (Fig. 7).

BIOFACIES

Paleoecological studies, based on the examination of pollen, diatoms, foraminifera, and arcellaceans, have helped define environments of deposition and past climatic conditions (Hebda, 1977; Mathewes, 1985; Mathewes and Heusser, 1981; Patterson, 1990; Patterson and Cameron, 1991; Clague et al., 1991a; Jonasson and Patterson, 1992; Patterson and Luternauer, 1993).

Benthic foraminifera occur in distinct biofacies from the high marsh where salinity is very low (<5 ‰) to the deep sea (Lipps, 1982). In contrast, arcellaceans (testate rhizopods) are found in freshwater environments (Medioli et al., 1990). Nine foraminiferal and arcellacean biofacies have been identified on the Fraser River delta using data obtained from surface samples and drill core. Due to processes occurring after death and burial, not all of the modern (surface) biofacies are preserved in the fossil record (down core). Conversely, some of the fossil biofacies have no modern counterparts in the area due to late Quaternary oceanographic changes.

Freshwater biofacies

With the exception of *Centropyxis aculeata*, a species tolerant of very low salinities in splash and storm zones, arcellaceans are restricted to freshwater environments. Clague et al. (1991a) used arcellacea in peat deposits to determine sea level change on the Fraser delta for the period 4000-4500 (^{14}C) years ago. Evoy et al. (1993b) identified intervals in cores collected on the delta slope that consist almost entirely of arcellaceans. The presence of these freshwater organisms in such high concentrations in a marine setting suggests that fluvial sediment may by-pass intermediate environments and be deposited directly in deep water (Evoy et al., 1993b).

Marsh biofacies

Based on foraminiferal biofacies, Fraser River delta marshes can be divided into a high marsh (*Miliammina fusca*) zone, equivalent to Hutchinson's (1982) floral high marsh zone, and a low marsh (*Jadammina macrescens-Trochammina inflata*) zone. The low marsh zone can further be divided into a higher low marsh zone and a lower low marsh zone.

Submarine biofacies

The *Cribroelphidium excavatum* biofacies, dominated by the nominate species, is found in early Holocene sediments of the Fraser River delta (Fig. 7). This species is presently widely distributed in shallow, temperate, and polar seas (Phleger, 1952; Loeblich and Tappan, 1953; Miller et al., 1982), and also is common in Pleistocene marine sediments deposited under near-glacial conditions (Osterman, 1984). It may also occur in warmer waters, with salinities lower than 35‰, such as might be expected off the mouth of a major river (Hald and Vorren, 1987).

Cribroelphidium excavatum is an important component of modern foraminiferal faunas in the Pacific Ocean adjacent to North America, commonly constituting more than 30% of the fauna (Cockbain, 1963; Vilks et al., 1979; Miller et al., 1982). It is even more abundant (up to 85%) in early Holocene samples from the Fraser River delta. Osterman (1984) suggests that there has been a dramatic reduction in the proportion of *Cribroelphidium excavatum* in modern oceans due to the retreat of glaciers and a resultant reduction in habitat for this species. In the Strait of Georgia *Cribroelphidium excavatum* presently comprises no more than 24% of the fauna and occurs in water depths between 50-200 m (Cockbain, 1963). Taking into account the Holocene decline in the species, the *Cribroelphidium excavatum* biofacies found in early Holocene sediments of the Fraser River delta probably represents an environment similar to that of the present-day Strait of Georgia.

Pollen and spores recovered from sediments at the base of the deltaic sequence (Fig. 7) indicate that local climate during early Holocene time (ca. 10 000-7500 ^{14}C years ago) was warmer and drier than today. The forest was very different from that of the present, with no red cedar (*Thuja plicata*) and little hemlock (*Tsuga heterophylla*), but more abundant Douglas-fir (*Pseudotsuga menziesii*) (Mathewes and Heusser, 1981; Mathewes, 1985). Disturbance indicators such as alder (*Alnus*) and bracken fern (*Pteridium*) indicate increased fire frequency.

GEOHAZARDS

Seismic hazards

Southwestern British Columbia has one of the highest earthquake hazard ratings in Canada. The National Building Code of Canada requires that engineered structures in the Vancouver area be able to withstand a Richter magnitude 7.0 local earthquake, with peak horizontal ground accelerations of up to 0.26 g. In addition to the local "design" earthquake, the area could also experience a much larger (M 8+) earthquake centred on the Cascadia subduction zone in the eastern North Pacific Ocean (Rogers, 1988, 1994). In recognition of this possibility, and partly in response to recent damaging earthquakes in California and Mexico, there has been heightened interest in the evaluation of ground response in the region, particularly on the Fraser River delta where thick sequences of loose, water saturated sediments overlie bedrock (Task Force Report, 1991).

It is known that the ground surface response to seismic shaking of thick or soft sediments differs markedly from that of bedrock. The major effects of such sediments include ground motion amplification, shifts in spectral frequency, and cyclic liquefaction. Modelling of ground response requires knowledge of the stratigraphy, depth to bedrock, dynamic moduli (shear wave velocity and shear modulus), and damping characteristics.

Over the past few years, the Geological Survey of Canada has obtained information on the shear wave velocity structure of the Fraser River delta using a variety of methods, including 1) the recording of downhole three-component shear-wave velocities, 2) the sounding of surface shear-wave refraction, 3) seismic cone-penetration testing (SCPT), 4) spectral analysis of surface waves (SASW), and 5) shear-wave refraction profiling (Finn et al., 1989; Hunter et al., 1991, 1992). Data obtained by these techniques are being matched to lithological information derived from boreholes to better calibrate geophysical logs (Hunter et al., 1993).

Ground motion amplification

One-dimensional computer models of ground motion amplification require the shear-wave velocity-depth profile from the ground surface down to, and including, bedrock. Models such as SHAKE (a computer program that analyzes the earthquake response of horizontally layered materials; Schnabel et al., 1972) are routinely used in North America and have been applied to the Fraser River delta (Finn and Nichols, 1988). Local Holocene deltaic sediments generally have much lower shear-wave velocities than underlying Pleistocene deposits; hence, the depth and shape of the Pleistocene surface is an important parameter in modelling ground motion amplification. In

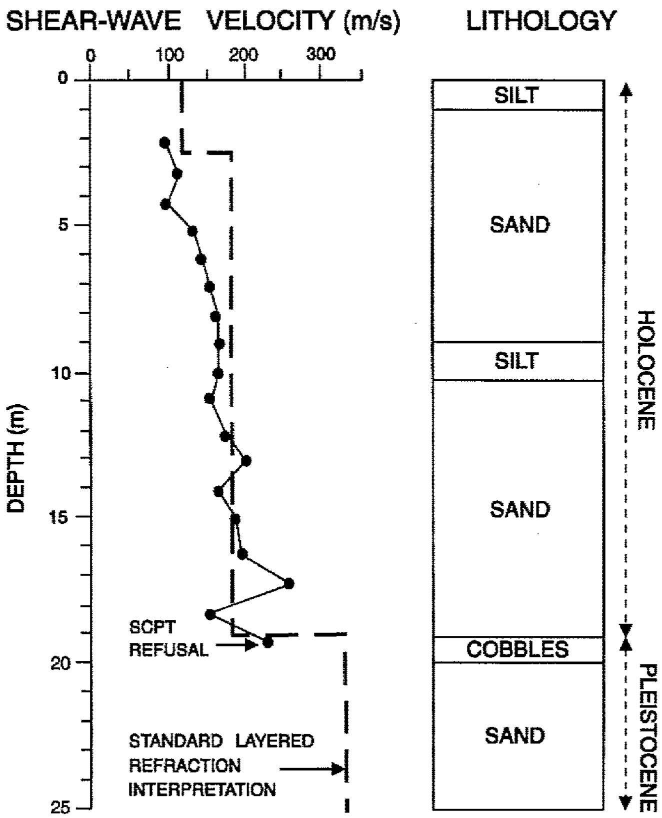

Figure 11. Seismic cone penetrometer and surface refraction results, and borehole stratigraphy for a site on Lulu Island. The data indicate that firm soil (Pleistocene sediments) is present at shallow depth (from Hunter et al., 1993).

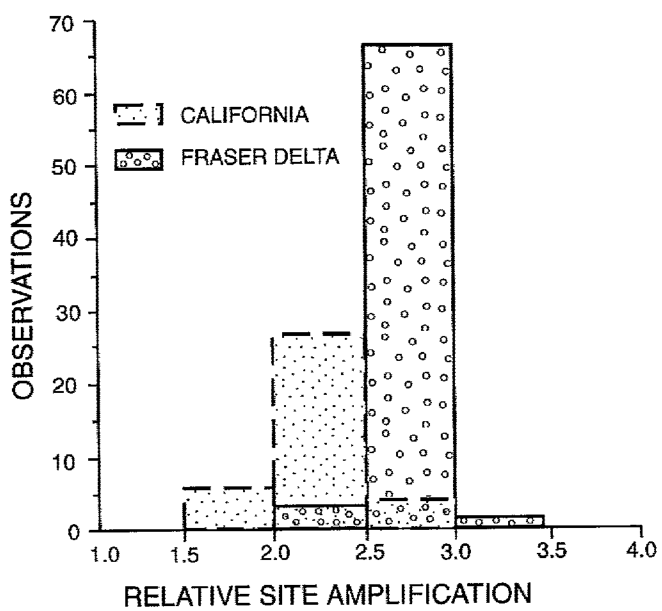

Figure 12. Comparison of relative ground motion amplification on the Fraser River delta with ground motion amplification in areas with similar Holocene deposits in California (San Francisco and Los Angeles) (from Hunter et al., 1993).

a few areas of the delta, where shallow gas is absent, conventional high-resolution P-wave reflection surveys have helped delineate this surface at depths ranging from 8 to 200 m (Clague et al., 1991b). Shear-wave refraction techniques have proven more reliable in areas where gas is present and have successfully detected shallow, high-velocity Pleistocene deposits at a site in central Richmond where thick Holocene sediments were expected (Byrne and Anderson, 1987; Lo et al., 1991; Sy et al., 1991) (Fig. 1, 11). The data indicate that the buried Pleistocene surface has substantial relief and that the Holocene cover varies considerably in thickness. The local presence of Pleistocene deposits at shallow depth significantly lowers the surface ground motion expected during an earthquake.

A simple first-order technique for estimating ground motion amplification in unconsolidated materials using shear-wave velocities has been applied effectively by the United States Geological Survey (Fumal and Tinsley, 1985; Joyner and Fumal, 1985). Assuming that the dominant energy of moderate earthquakes is in the 1 Hz range, one can compare the thickness-weighted average shear-wave velocity of the sediments to a depth of one quarter wavelength, or ~40 m, (V_{AVG}) to the shear-wave velocity of bedrock (V_{ROCK}) to obtain a value of ground motion amplification relative to a bedrock site:

$$Amplification = \sqrt{\frac{V_{ROCK}}{V_{AVG}}}$$

This method has been applied to sites on the Fraser River delta using data from surface shear-wave refraction soundings and assuming a shear-wave velocity in bedrock of 1500 $m{\cdot}s^{-1}$. The thick sequence of low shear-wave velocity sediments has a marked effect on the thickness-weighted average velocity, giving rise to significantly higher amplification factors than have been determined for similar types of deposits in San Francisco and Los Angeles that have been analyzed in the same way (Fig. 12).

Seismic liquefaction

Sediment response to cyclic loading during earthquakes depends primarily on two factors: the intensity and duration of cyclic loading and the behaviour of the sediment during loading. Shear-wave velocity data can be used to identify potentially liquefiable zones. The method may be superior to the SPT (Standard Penetration Test) and the CPT techniques in silty sediments (Seed et al., 1983; Robertson et al., 1992) in that shear-wave velocity is relatively insensitive to the amount of fines in the sediment, whereas the SPT and CPT approaches require large correction factors. Recent research has suggested that a boundary between contractive (potential flow-liquefaction) and dilative behaviour can be defined on the basis of shear-wave velocity (Robertson, 1990; Robertson et al., 1992). Hence, delineation of the shear-wave velocity-depth structure in the upper part of the deltaic pile may identify potentially hazardous areas where further detailed geotechnical investigations should be made (Hunter et al., 1993).

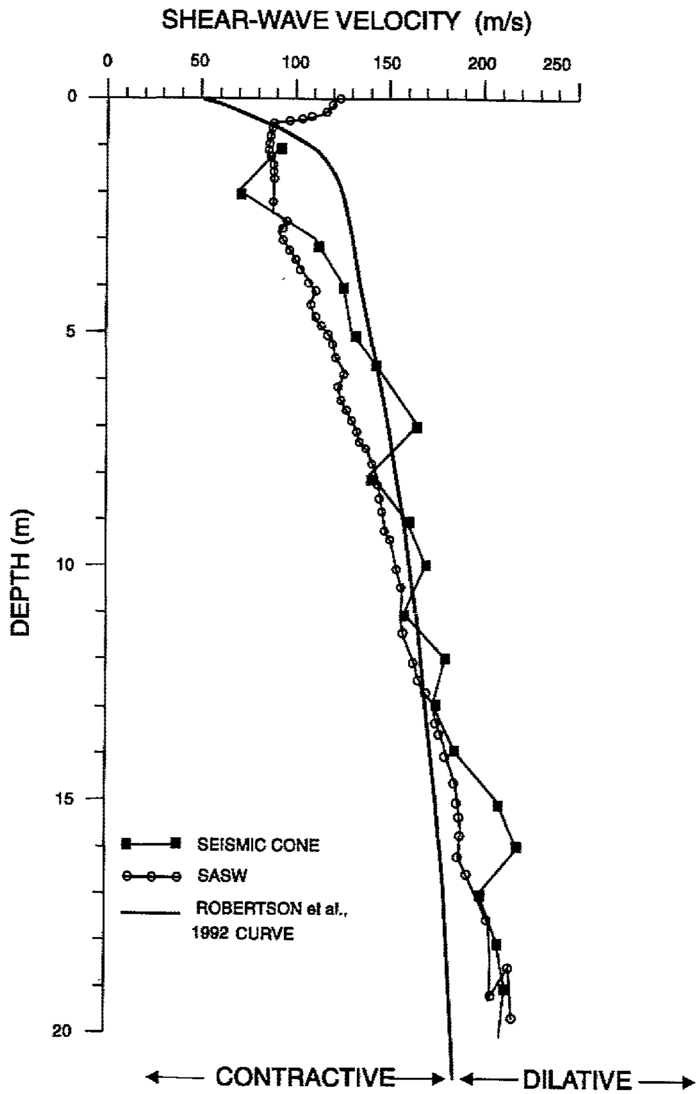

Figure 13. Shear-wave velocity versus depth for a site in Richmond. The data were obtained by seismic cone penetrometer testing (SCPT) and spectral analysis of surface waves (SASW). Also shown is the liquefaction threshold curve of Robertson et al. (1992) (from Hunter et al., 1993).

Figure 13 shows a comparison of shear-wave velocity-depth results obtained by two methods at a site in Richmond (Fig. 1). Spectral analysis surface wave data were obtained using a vertical hammer energy source and geophone spacings ranging from 0.25 to 12 m; multilayered forward modelling was carried out to determine the velocity-depth structure (Addo and Robertson, 1992). Seismic cone penetration data were collected at 1 m intervals using a horizontal hammer energy source at a 0.7 m offset from the cone rods at the surface (Woeller et al., 1993a, b). Also shown in Figure 13 is the "Robertson" steady-state contractive-dilative boundary curve (Robertson, 1990; Robertson et al., 1992); sediments with velocities significantly lower than this line are susceptible to flow-liquefaction during design-earthquake loading. The data in Figure 13 show that water-saturated sand to a depth of at least 6 m and, possibly, to 13 m could liquefy during the design earthquake. Velocities on or near the curve indicate a limited potential for liquefaction, or cyclic mobility, and the possibility of spreading-type failures. The extent of liquefaction and resulting movements are a function of the intensity and duration of shaking, as well as local ground conditions.

Sand dykes and sills, which have been observed in near-surface silty and clayey sediments over much of the Fraser River delta, provide evidence for earthquake-induced liquefaction in the past (Clague et al., 1992; Naesgaard et al., 1992). At a few sites, fluidized sand was expelled onto the former subaerial or intertidal surface of the delta to form sand blows (Fig. 14). All observed sand dykes, sills, and blows are younger than 3500 (^{14}C) years old, and at least some are younger than 2400 (^{14}C) years old.

Submarine slope instability

Delta slope failures could damage or destroy sewage outfall pipelines, as well as submarine transmission cables that supply electricity and telephone service to Vancouver Island. Such failures also could damage structures at the edge of the tidal flats, including a major ferry terminal and Canada's largest coal export facility (Fig. 1, 2, 15). A large rapid submarine failure might generate a tsunami in the Strait of Georgia (Hamilton and Wigen, 1987) that could damage the delta's western dykes (Byrne, 1990).

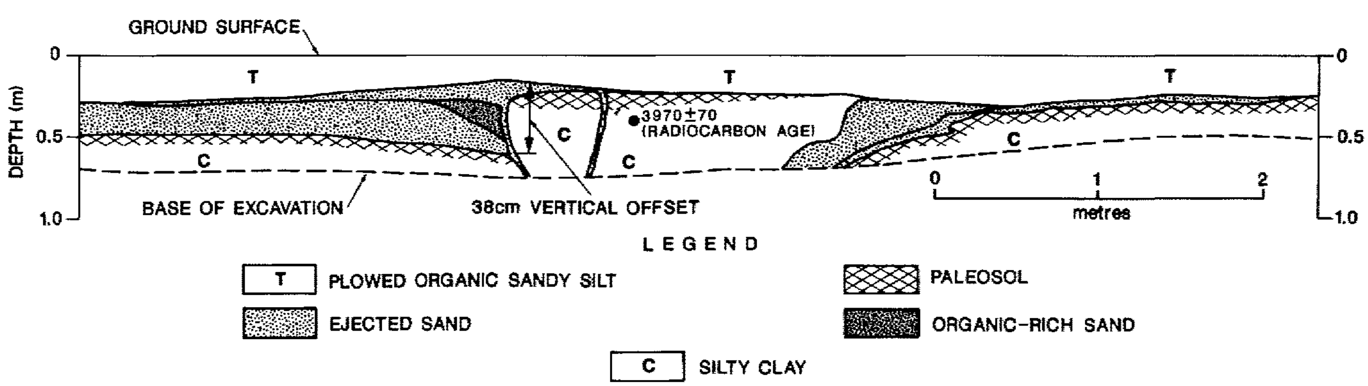

Figure 14. Profile of sand dykes and vented sand exposed in the shallow wall of an excavation on the central Fraser River delta. The vented sand directly overlies a brown paleosol and displays stratification indicative of emplacement during either multiple events or multiple pulses within a single event. Note the vertical displacement of the paleosol and the subjacent mud along the two main feeder dykes (from Clague et al., 1992).

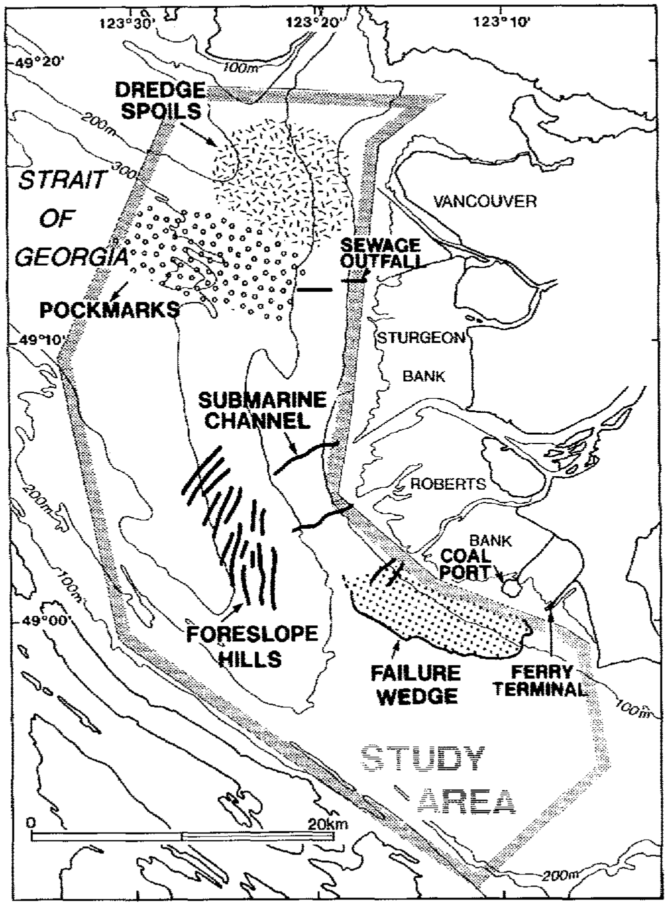

Figure 15. Surficial features of the Fraser River delta slope and adjacent floor of the Strait of Georgia (from Hart et al., 1992b, 1993).

Submarine failures are evident in seismic profiles of the delta (Fig. 10) and also have occurred historically. Deposition of sediment at the river mouth contributes to retrogressive failures at the head of submarine channels; one event in 1985 involved approximately 3 x 10^6 m^3 of material and created a scarp within 100 m of a lighthouse (Atkins and Luternauer, 1991; McKenna et al., 1992) (Fig. 16). On the upper and middle delta slope south of the 1985 failure zone (Fig. 2), the upper 2-3 m of sediment have broken into a series of large blocks tens of metres across that now are slowly moving downslope. Downslope movement may be promoted by interstitial gas and high sedimentation rates which, together, generate high pore pressures in these fine grained sediments (Hart et al., 1992a, 1993).

Two other suspected submarine failure complexes have been identified, one on the delta slope and another at its toe. The Foreslope Hills (Fig. 15), an area of curvilinear ridges and troughs up to 20 m high, covers an area of at least 60 km^2 at the base of the delta slope off northern Roberts Bank (Tiffin et al., 1971). The ridges consist of discrete blocks of prodelta and foreslope sediments bounded by seismic lineations that appear to dip seaward (Hart, 1993b). The deformation may be the result of slumping of delta slope sediments above soft clayey prodelta deposits at the base of the Holocene section (Hart et al., 1992b; Hart, 1993a, b), but probably did not involve massive and rapid downslope movements capable of triggering a tsunami (Hart, 1993b). A second failure complex underlies at least 40 km^2 of the lower part of the southern Roberts Bank slope (Fig. 15). Airgun and high resolution (Huntec, Seistec) seismic records reveal a downslope thinning wedge of chaotic and wavy reflectors, which is locally over 75 m thick (Hart et al., 1992b). The causes of these large failures are unknown, although earthquakes, rapid sedimentation, and interstitial gas may all have contributed.

Dunes occur at depths of 20-120 m on the upper Roberts Bank slope (Fig. 2) and locally exceed 3 m in height and 100 m in wavelength (Luternauer, 1980; Kostaschuk and Luternauer, 1993). The dunes migrate slowly to the northwest in the direction of the flood tidal current. This has resulted in localized burial of electrical transmission cables laid in the mid-1950s (Hart et al., 1993). As there is no obvious source of sand replenishment to the southeast, it appears that the seafloor is being eroded. This may ultimately lead to retreat and oversteepening of the upper Roberts Bank slope adjacent to local port facilities. Comparison of 1968 and 1985 bathymetric surveys within the sand wave field (Stewart and Tassone, 1989) indicates that, over this 17 year period, the slope retreated 40 m at depths of 5-20 m, advanced 10 m at depths of 20-60 m, and retreated 55 m at depths of 60-100 m.

Marine geotechnical research

Marine geotechnical research activities include geophysical mapping of the entire delta front, and geotechnical drilling and in situ testing of sediments on the upper delta slope to evaluate their static and cyclic shearing resistance and collapse potential.

Contemporary instability of the delta slope at Sand Heads affords an exceptional opportunity to study the liquefaction failure process in real-time, which may provide insights into previous river-mouth failures elsewhere along the delta front. One of the major problems in evaluating the potential for slope failure on the Fraser River delta is the nature of the triggering mechanisms. Instability features range from small-scale wave-induced failures at the top of the slope to larger slides that apparently have been caused by some other process. Tidal drawdown has been suggested as a possible trigger for repeated non seismic liquefaction slides adjacent to the Sand Heads lighthouse (Luternauer and Finn, 1983; McKenna et al., 1992; Christian et al., 1994). This hypothesis is now being explored through the development of an in situ monitoring array to continuously measure pore pressures and ground motions.

Collapsive-type or flow-liquefaction gravity flows at Sand Heads have incised a deep channel that extends downslope for more than 7 km, ending in an extensive apron of chaotic debris at the base of the slope (Kostaschuk et al., 1992b; Hart et al., 1992a, b, 1993). A small liquefaction failure at Sand Heads, fortuitously detected by in-ground sensors in November 1992, coincided with the passage of the largest sediment gravity flow of the year, recorded by

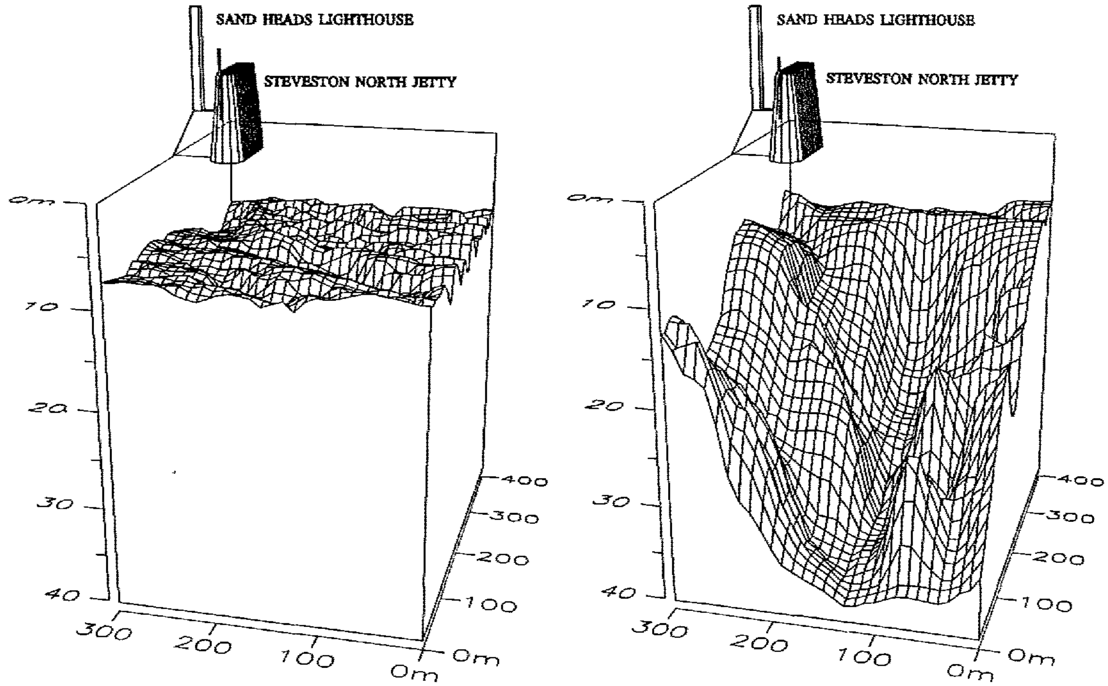

Figure 16. Perspective computer-generated images of the mouth of Main Channel before (June 27, 1985) and after (July 11, 1985) a large failure. Vertical exaggeration is 10; O m = mean low low water (from Atkins and Luternauer, 1991).

turbidity event detectors. Other submarine channels scoured by sediment gravity flows, off Westham Channel and north of Sand Heads on Sturgeon Bank, appear now to be inactive due to a lack of sediment supply. Nevertheless, fine grained turbidity current deposits partially infill depressions between the ridges of the Foreslope Hills (Hart et al., 1992a, b).

HUMAN IMPACT

Urban and industrial development associated with the growth of Vancouver and its satellite communities has adversely affected the Fraser River delta. Pollutants discharged into the Fraser River are derived from a variety of sources and include antisapstain chemicals used in the lumber industry, poly-cyclic aromatic hydrocarbons, and dioxins/furans (Servizi, 1989; Standing Committee on the Fraser River Estuary Water Quality Plan, 1990; Church et al., 1992). Several contaminants are adsorbed onto fine grained sediment particles that are transported in continuous suspension by the river and may be concentrated at a turbidity maximum identified at the upriver end of the salt-wedge (Grieve and Fletcher, 1976b, Kostaschuk et al., 1992a). From there, toxin-laden, fine grained sediments may go on to accumulate in marshes and mud pools of the upper intertidal area of the western foreshore (Grieve and Fletcher, 1976a, b; Luternauer, 1980) or within the adjacent Strait of Georgia (Macdonald et al., 1991).

From 1962 to 1988 Vancouver sewage was pumped from the Iona Island primary treatment plant onto the adjacent tidal flats which led to local concentration of toxic metals and organic detritus (Birtwell et al., 1983). In 1988 pipe was installed to discharge sewage at 100 m depth on the Sturgeon Bank slope (Fig. 1). During severe rainstorms the deep outfall is incapable of handling the volume of flow, and consequently excess waste is still discharged onto the tidal flats. Here, heavy metals such as mercury, cadmium, and vanadium within the effluent (Hall et al., 1974) and adsorbed onto sediments may enter the food chain and accumulate in organisms (Parsons et al., 1973).

The volume of sand dredged for river maintenance and construction material exceeds that supplied to the estuary by the river (McLean and Tassone, 1991). Net erosion of intertidal habitats may occur because the river can no longer adequately replenish sand that is eroded by waves and currents and carried into deep water. In addition, causeways, jetties and other large engineered structures have altered sediment and water dispersal patterns (Levings, 1980) and contributed to the degradation of some coastal habitats (Harrison, 1984; Duggan and Luternauer, 1985; Tarbotton et al., 1993) (Fig. 17, 18). Dredging associated with the excavation of a ship basin and construction of a causeway at the Roberts Bank Coal Port (Westshore Terminals) in 1969-1970 deepened water over parts of the tidal flat. This triggered erosion of dendritic drainage channels within the eelgrass habitat at the head of the basin as ebb tidal flows were focused towards it (Fig. 18).

In spite of local erosion of eelgrass habitat, the areal extent of eelgrass beds between the Coal Port and Ferry Terminal causeways has about doubled since the former was constructed. The 900 m landward migration of the shoreward limit of the beds resulted, in part, from the introduction of a new species of eelgrass, but is mainly attributable to deflection of turbid Fraser River water away from the area by the Coal Port causeway. The resultant increase in light penetration at high tide promotes dense growth of plants which, in turn, permits more storage of water at the landward side of the beds on a low tide and creates new favourable habitat for further colonization (Harrison, 1984).

Expanding eelgrass beds also create conditions that hasten the growth of dendritic channels after they have been initiated. Water stored within beds may fall as much as 30 cm over a few metres into an existing channel at high velocity and continue the process of headward erosion.

The low (~0-1 m high) crest protection structure bordering the Coal Port ship basin (Fig. 18) was designed to mitigate the effects of the excavation on the eelgrass beds. Within the lower energy environment created shoreward of the structure, channel erosion would be inhibited and sedimentation and eelgrass colonization promoted. Eelgrass has recolonized most of the western creek system where erosion has ceased or slowed considerably (Fig. 18). In contrast, to the east, not only has erosion continued apace, albeit within fewer dominant streams, but part of the area also has been altered by sediment accretion. A lobe of sand as much as 50 cm thick, 600 m long, and 100 m wide accumulated at the head of the larger eastern creek between 1989 and 1991 (Fig. 18), smothering underlying eelgrass beds. The sand probably was carried upstream by rising tides from sources along the lower reaches of the larger east channel.

Figure 17. *Erosion of the tidal flat near the edge of the marsh west of Steveston. Sedimentary changes associated with the construction of Steveston North Jetty from 1911 to 1932 (Public Works Canada, 1949) probably inhibited local marsh colonization and promoted creek formation. Geological studies (Medley, 1978; Luternauer, 1980) have indicated, however, that further extension of creeks may help alleviate conditions that limit marsh growth. GSC 1994-714*

Channel migration and storage of sand in channel deposits on the delta plain are inhibited by dykes and jetties. As a result, most of the sand transported by the river is carried directly to the mouth, where its accumulation leads to failures (Monahan et al., 1993b).

Dredge spoils and excavation waste are dumped at a disposal site off the mouth of North Arm (Fig. 15). Over 3×10^6 m^3 of material were dumped here between 1968 and 1987. Side-scan sonar imagery has documented unauthorized dumping outside the limits of the disposal site where it interferes with bottom fishing on the Sturgeon bank delta slope (Hart, 1992). There are other smaller dump sites near the mouth of Main Channel and near the Roberts Bank port facilities.

SUMMARY

The Fraser River delta is the largest delta in western Canada. It is an important coastal ecosystem and an area of explosive urban and industrial growth situated within the most seismically active region in Canada.

The delta started to form about 10 000 (^{14}C) years ago, after the area was deglaciated, and has a maximum reported thickness of 213 m. The basal stratigraphic unit consists of mud deposited at the toe of the delta under oceanographic conditions similar to those in the present Strait of Georgia. South of Main Channel, the dominant distributary of the Fraser River, this fine sediment is conformably overlain by a succession of mainly sandy beds deposited on the former slopes of the delta. North of Main Channel, the paleoslope deposits consist primarily of silt-rich mud with localized bodies of sand up to 30 m thick. Slope deposits constitute the bulk of the deltaic sedimentary package. The uppermost stratigraphic unit is a complex of distributary channel sands capped by mud and peat of intertidal, floodplain, and bog origin.

The delta's distributary channels are estuaries characterized by intrusion of saline water from the Strait of Georgia. Sediment transport in this estuarine system is controlled by (a) river and sediment discharge, (b) tidal conditions, and (c) the position of the salinity intrusion. Sand bed-material in Main Channel is moulded into dunes that can exceed 5 m in height and 100 m in length. The dunes generate large turbulent vortices capable of scouring the sandy channel floor.

Fine grained sediments and associated pollutants transported by the Fraser River are dispersed widely in the southern Strait of Georgia in a well defined plume. Mud accumulates on the Sturgeon Bank slope and in the adjacent Strait of Georgia at rates of $<1\text{-}2\ cm \cdot a^{-1}$. Sand is deposited mainly off the mouth of Main Channel at highly variable rates.

Major geological hazards are slope failures, burial and scour of the delta slope, and earthquake-induced liquefaction. Rapid rates of deposition at the river mouth contribute to slope failures. One such failure, in 1985, involved approximately 3×10^6 m^3 of material and formed a headscarp within 100 m

of a lighthouse. Prehistoric failures, involving sediments up to 75 m thick, underlie at least 40 km^2 of the southwestern delta slope. A field of subtidal dunes, which are locally greater than 3 m high and 100 m long and occur at depths of 20-120 m on the southwestern delta slope, developed in response to sediment transport by tidal currents. The dunes locally bury electrical transmission cables, and movement of sediment in this area may destabilize the slope crest adjacent to major port facilities. In at least part of the dune field, the slope retreated 2-3 m $\cdot a^{-1}$ between 1968 and 1985. Liquefaction of a shallow sand unit, which is present beneath much of the delta plain, probably was caused by one or more large prehistoric earthquakes sometime within the last 2400 (^{14}C) years. The high susceptibility of this unit to liquefaction stems from its origin as distributary channel deposits. Lateral and vertical variability of delta deposits indicates that the pattern of ground motion amplification and liquefaction during a major earthquake will be complex.

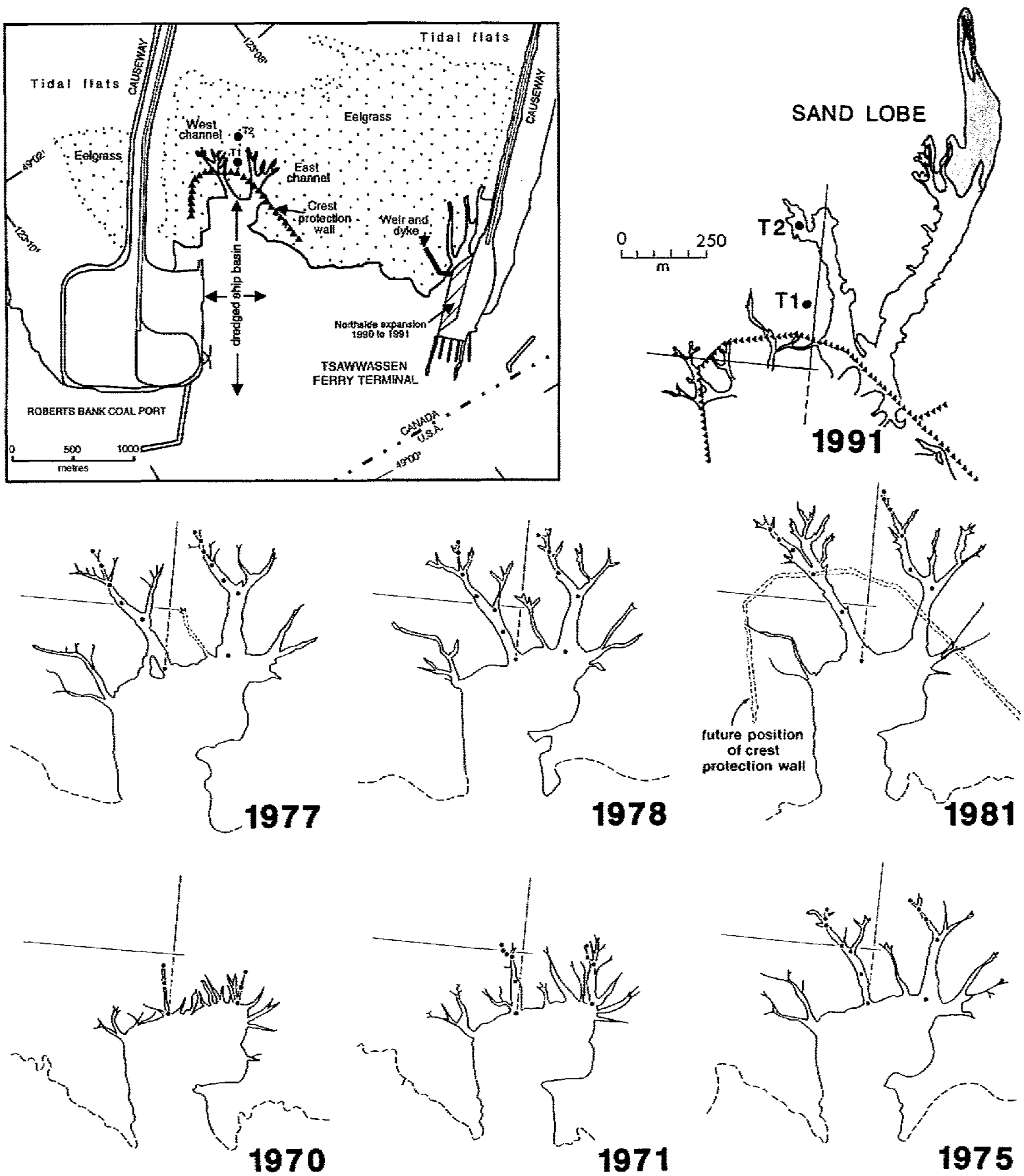

Figure 18. Dendritic creeks at the head of the ship basin at Roberts Bank Coal Port. Severe erosion of eelgrass beds, caused by the excavation of the basin in 1970 (Duggan and Luternauer, 1985), prompted the erection of a crest protection wall which has helped to re-establish eelgrass in the western channel area (Tarbotton et al., 1993). T1 and T2 represent navigation towers.

Human activity has contributed pollutants to the delta and has altered natural sedimentary processes. The discharge of treated sewage onto the tidal flats has led to the concentration of toxic metals and organic detritus within sediments. Causeways and other large engineered structures have altered sediment and water dispersal patterns and contributed to the degradation of some coastal habitats. Training walls along the lower portions of principal distributary channels have eliminated meandering, concentrated sand deposition in small areas, and possibly contributed to local failures at the mouth of Main Channel (the geological record, however, indicates that human interference is not a prerequisite for delta slope failure). River training and channel dredging to maintain navigable depths and provide construction material may lead to erosion of part of the seaward edge of the western tidal flats.

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Earthquakes in the Vancouver area

Garry C. Rogers[1]

Rogers, G.C., 1994: Earthquakes in the Vancouver area; in Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia, (ed.) J.W.H. Monger; Geological Survey of Canada, Bulletin 481, p. 221-229.

Abstract: Vancouver and the densely populated Lower Mainland region of southwest British Columbia are situated over an active subduction zone. The dynamic geological setting makes this region subject to frequent seismic activity and contributes to a higher risk of large damaging earthquakes than in other parts of Canada. Earthquakes that may present a hazard to the area are located in three distinct source regions: earthquakes within the continental crust, deeper earthquakes within the subducted oceanic plate, and earthquakes on the subduction boundary between the lithospheric plates. While Vancouver (incorporated 1886) has not experienced a damaging earthquake in its short history, large earthquakes nearby have been strongly felt and there is paleoseismic evidence for very strong shaking in prehistoric time. Most of the region is placed in Seismic Zone 4 in the 1990 edition of the National Building Code of Canada. Ongoing microearthquake activity and felt earthquakes occurring in most years are reminders of the seismic hazard.

Résumé : Vancouver et la région densément peuplée des basses terres du Fraser du sud-ouest de la Colombie-Britannique sont situées sur une zone de subduction active. Le contexte géologique dynamique fait que cette région est sujette à une activité séismique fréquente; il crée un plus grand risque de forts séismes destructeurs que dans les autres parties du Canada. Les séismes qui peuvent présenter un danger pour la région se manifestent dans trois régions hypocentriques distinctes : les séismes survenant dans la croûte continentale, les séismes plus profonds survenant dans la plaque océanique subduite, et ceux survenant à la limite de subduction entre les plaques lithosphériques. Tandis que Vancouver (constituée en 1886) n'a pas connu de séisme destructeur pendant sa courte histoire, de vastes séismes ont été fortement ressentis à proximité; on a découvert des indices paléoséismiques de très fortes secousses survenues dans les temps préhistoriques. La majeure partie de la région a été placée dans la Zone séismique 4, selon l'édition de 1990 du Code national du bâtiment. L'activité microséismique actuelle et les séismes ressentis presque chaque année sont un avertissement du risque séismique existant.

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INTRODUCTION

The southwest coast of British Columbia, including Vancouver and the densely populated Lower Mainland, is located in a dynamic geological setting (Fig. 1) that makes it one of the most seismically active regions of Canada. While Vancouver (incorporated 1886) has not yet experienced a damaging earthquake, a number of large earthquakes have occurred close enough to have been strongly felt in the city (Table 1) and paleoseismic evidence confirms that large earthquakes have occurred in prehistoric time (Clague et al., 1992). Occurrences of felt earthquakes in most years (Fig. 2) and frequent microearthquake activity (Fig. 3) are a reminder that the area is seismically active.

Southwest British Columbia is situated over an active subduction zone, and thus is in an earthquake environment similar to the east coast of Japan, the south coast of Alaska and most of the west coast of Central and South America. The oceanic Juan de Fuca and Explorer plates are being subducted in a northeast direction beneath the continental North American plate (Fig. 1). Earthquakes that may present a hazard to the area occur in three distinct source regions in this geological environment: continental crust earthquakes, deeper earthquakes within the subducted oceanic plate, and very large earthquakes on the subduction boundary between the lithospheric plates. The continental crust earthquakes, which are the most numerous, are driven by a compressive stress parallel to the continental margin, oriented north-northwest in the Vancouver region, and include the damaging 1946 central Vancouver Island earthquake (M = 7.3) and the large prehistoric Seattle earthquake (M = 7+). The larger subcrustal earthquakes are caused by a tensional stress regime within the subducted plate in a depth range of 45 to 65 km and have included damaging earthquakes in southern Puget Sound in 1949 (M = 7.0) and 1965 (M = 6.5). The subduction boundary, off the west coast of Vancouver Island, has not produced any earthquakes in historic time, but paleoseismic evidence shows they occur at intervals of centuries (Atwater, 1987; Adams, 1990) and analysis of contemporary crustal deformation reveals that strain is accumulating for a future event (Dragert et al., 1994). Subduction earthquakes are among the world's largest earthquakes. Those along the west coast of Vancouver Island are likely to exceed M = 8 (Rogers, 1988a, b), and to have effects comparable to the great Alaska earthquake (M = 9.2) in 1964 (Steinbrugge et al., 1967).

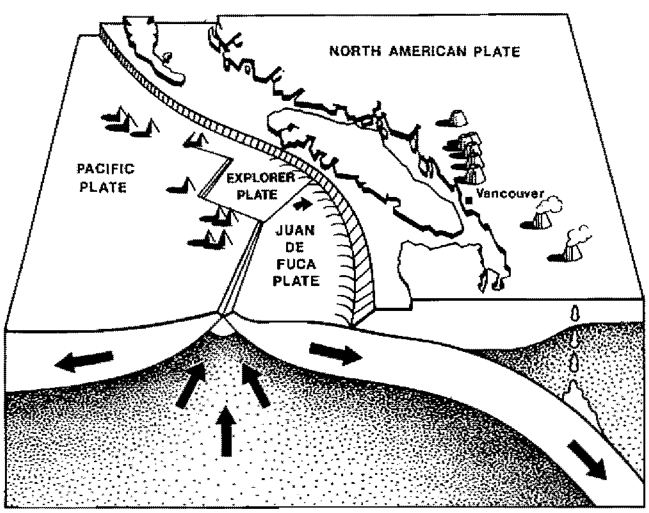

Figure 1. Cartoon of the tectonic setting of southwest British Columbia showing the oceanic Juan de Fuca and Explorer plates subducting beneath the continental North American Plate.

CRUSTAL EARTHQUAKES

About ninety per cent of the small earthquakes in the Vancouver region occur in the continental crust of the North American Plate. Forty years of earthquake monitoring in southwest British Columbia has revealed persistent crustal microearthquake activity (Milne et al., 1978; Rogers and Horner, 1991). Since the present seismograph network for monitoring the Lower Mainland and southern Vancouver Island regions was completed (early in 1983 for the Vancouver region) it has been possible to determine very accurately, to within a few kilometres horizontally and vertically, the locations of most small earthquakes (Fig. 3). Vancouver lies at the north end of an increased level of earthquake activity that extends to the south end of Puget Sound in Washington State. Concentrations of small events occur, but there are no distinct alignments of epicentres to convincingly mark the locations of active faults. However, the earthquakes die out to the east rather abruptly along a linear north-northwest trend, and this may be fault controlled. The vertical distribution of crustal earthquakes (Fig. 4) reveals that most occur at a considerable depth within the crust, on the order of 20 km, and it is perhaps not surprising that there appears to be little correlation with mapped surface faults. This depth range places most of the seismicity in this region deeper than earthquakes in California, where most earthquakes occur in the top 10 km of the crust and events deeper than 15 km are rare. The deeper source of most of the crustal events in southwest British Columbia means that there are fewer aftershocks than typical for California earthquakes (e.g., Page, 1968).

The extra thickness of the brittle portion of the crust, where earthquakes can occur, is a direct result of the subduction environment. The subducting plate is colder than the asthenosphere into which it is descending and absorbs much of the heat flowing from the interior of the earth. The result is that there is very low heat flow through the upper crust and thus a thicker region in the temperature range that is cool enough to support brittle fracture in the form of earthquakes. Just to the west of the Quaternary volcanoes, which are part of the Cascade magmatic arc, the heat flow increases markedly (Lewis, 1991) and a reduction in the seismicity rate appears to correlate with this change to a thinner section of brittle crust.

Table 1. Some significant earthquakes felt in Vancouver.

Date	Location	Magnitude	Comment
~900	47.6°N 122.5°W	7+	Surface rupture and tsunami deposits in Puget Sound.
~1700	West of Vancouver Island	8+	Subduction earthquake. Native houses on Vancouver Island may have been damaged.
1864/10/29	48¾°N 123¼°W	~5.5	Probably deep, no reported aftershocks. Felt strongly in the Lower Mainland.
1872/12/15	47.7°N 120.2°W	~7.4	Location uncertain, based on felt aftershocks. Felt strongly in the Lower Mainland.
1909/01/11	48.7°N 122.8°W	~6.0	Probably deep, no reported aftershocks. Felt strongly in the Lower Mainland.
1918/12/06	49.4°N 126.2°W	~7.0	Felt by most in the Lower Mainland. Damage on west coast of Vancouver Island.
1920/01/24	48.6°N 123.0°W	~5.5	Probably deep, no reported aftershocks. Felt strongly in the Lower Mainland.
1946/06/23	49.8°N 125.3°W	7.3	Much damage in central Vancouver Island. Felt strongly in the Lower Mainland.
1949/04/13	47.1°N 123.0°W	~7.0	Depth 54 km. Much damage in Seattle and Tacoma. Felt by most in Lower Mainland.
1965/04/29	47.4°N 122.3°W	6.5	Depth 63 km. Much damage in Seattle. Felt by most in the Lower Mainland.
1975/11/30	49.2°N 123.6°W	4.9	Shallow, many aftershocks. Felt by many in the Lower Mainland.
1976/05/16	48.8°N 123.4°W	5.4	Depth 60 km. Felt by most in the Lower Mainland.
1990/04/14	48.8°N 122.2°W	4.9	Shallow, many aftershocks. Felt by many in the Lower Mainland.

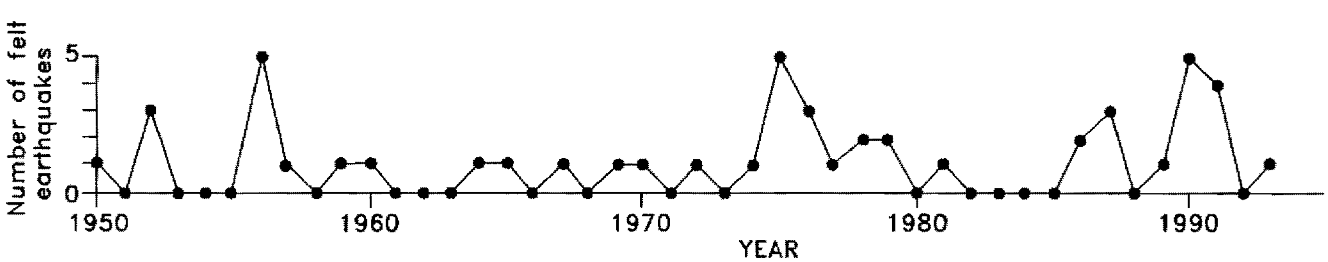

Figure 2. Number of earthquakes felt in greater Vancouver and the Lower Mainland from 1950 to 1993.

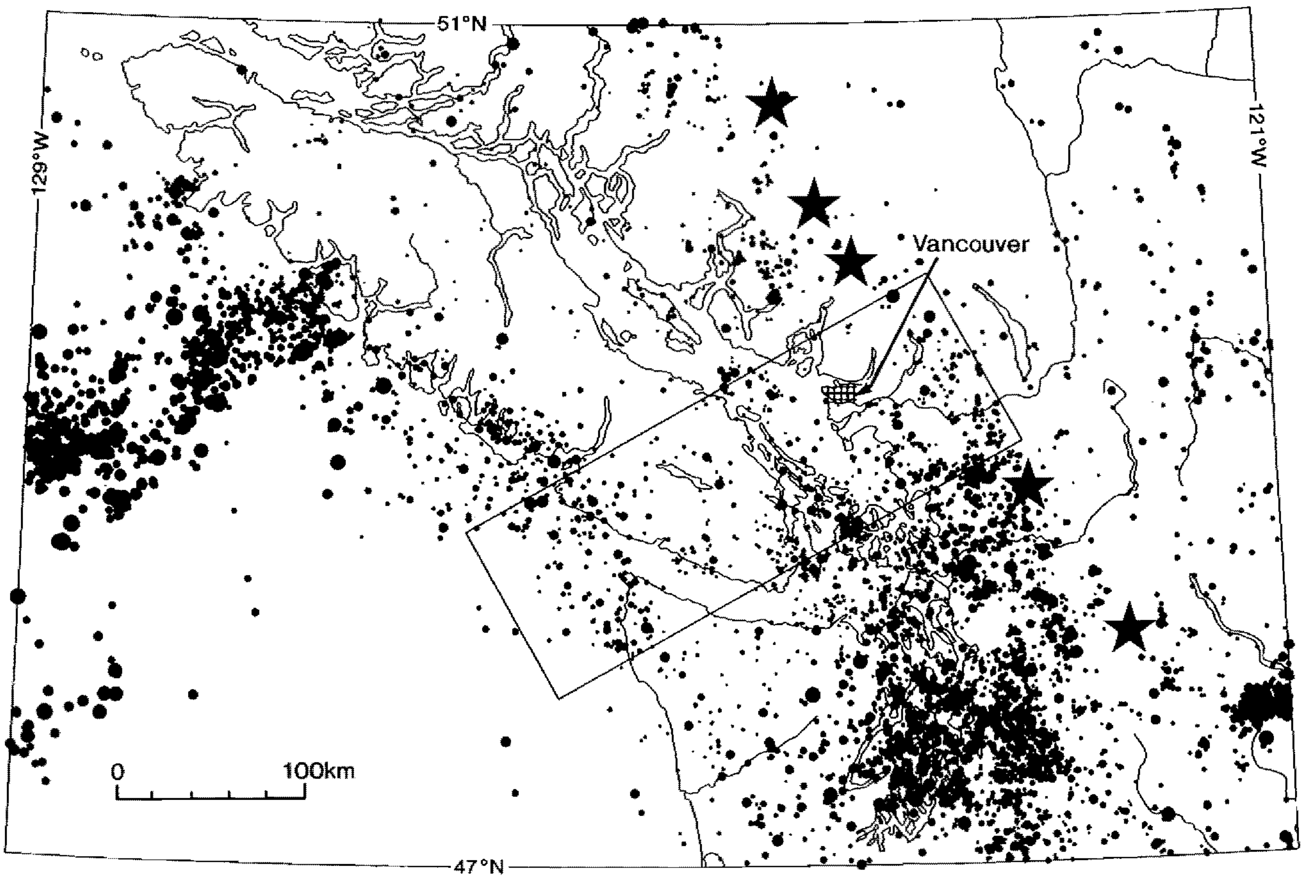

__Figure 3.__ Seismicity in the Vancouver region from 1980 to 1991 inclusive. Dot size is proportional to magnitude. The smallest earthquakes plotted are magnitude 1 and the largest in this sample is an offshore event of magnitude 5.2. Earthquake data from the Geological Survey of Canada and the University of Washington have been combined to make this plot. The rectangle outlines the area used in the cross-section in Figure 6. Stars are Quaternary volcanoes of the Cascade magmatic arc.

There have been three major crustal earthquakes affecting the region in historic time, in 1918 (M=7) and 1946 (M=7.3) on Vancouver Island (Cassidy et al., 1988; Rogers and Hasegawa, 1978), and in 1872 (M = 7.4) in northern Washington State (Malone and Bor, 1979) probably near Lake Chelan (Table 1, Fig. 5). Although there is a concentration of seismicity in the southern Lake Chelan region (southeast corner of Fig. 3), the occurrence of these three large earthquakes is not obviously reflected in the pattern of earthquakes we see today. It should also be noted that all three of these larger earthquakes occurred away from the southern Strait of Georgia-Puget Sound lowland, which is the most intense region of seismicity reflected by the pattern of small earthquakes (Fig. 3).

There is a subset of the ongoing small earthquakes in the upper 10 km of the crust. Some of the larger of these very shallow events have had long aftershock sequences, typical of California earthquakes, and may have occurred on faults that ruptured the surface. The November 30, 1975 earthquake (M = 4.9) in central Strait of Georgia (Rogers, 1979) and the April 13, 1990 Deming, Washington earthquake (M = 4.8), just south of Abbotsford (Qamar and Zollweg, 1990), are two such events (Table 1, Fig. 5). There is also ample geological evidence that a large shallow prehistoric earthquake ruptured the surface in central Puget Sound, near Seattle, about 1100 years ago (Bucknam et al., 1992; Atwater and Moore, 1992). Such extremely shallow earthquakes are rare, but represent the greatest source of uncertainty in assigning seismic hazard in the region because their distribution and maximum magnitude are difficult to assess. Understanding the hazard they pose is important, however, because such near surface sources can have very high accelerations in close proximity to them.

Small crustal earthquakes in this region are a mixture of strike-slip and thrust events with a dominant north-northwest orientation of the principal stress axes (Mulder and Rogers, 1993) suggesting north-northwest compression. This is a different orientation from the apparent north-south compression observed just to the south in Washington State (Ma et al., 1991). It appears as if the compressive crustal stress is

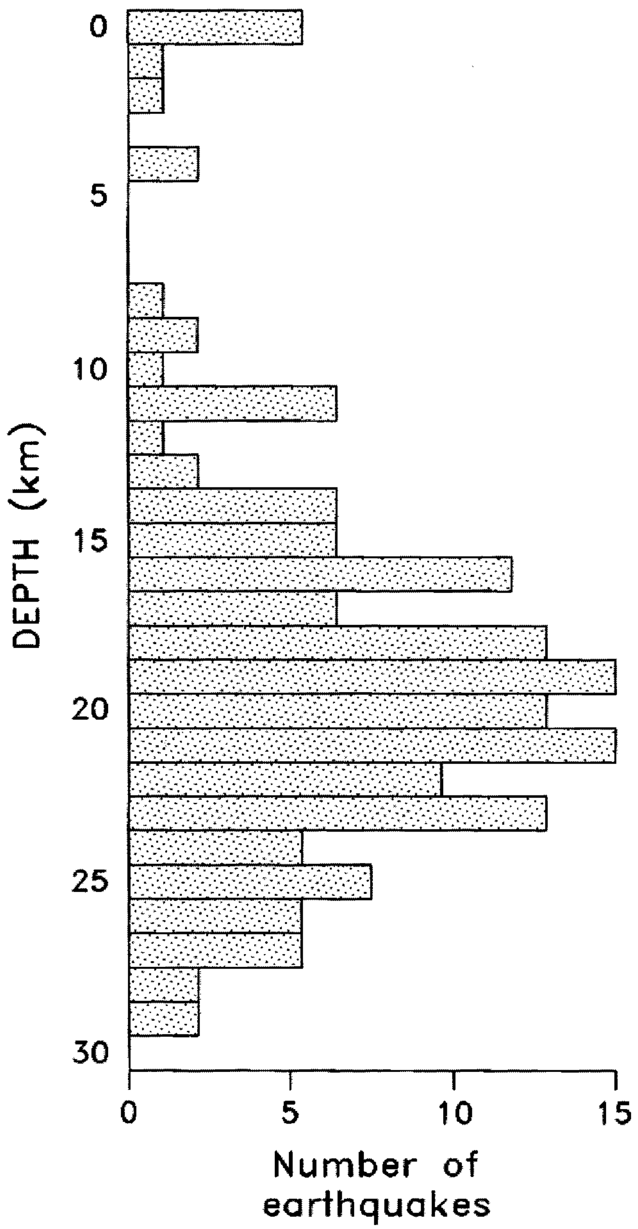

Figure 4. Depth distribution of a three year sample of crustal earthquakes in southern Strait of Georgia in the region bounded by 48°N, 49°N and 122°W, 124°W.

oriented parallel to the continental margin. This does not conflict with the observed crustal shortening (Savage et al., 1991; Dragert et al., 1994) perpendicular to the margin in the direction of the subducting plate, as the crustal deformation represents a change in stress, whereas the focal mechanisms of the earthquakes reflect the regional ambient stress. The origin of this stress regime has not yet been satisfactorily explained, but it may be a result of oblique subduction along most of the Cascadia margin (Rogers, 1979), compression from south of the subduction zone (Sbar, 1982), or rotation of a crustal block (Walcott, 1993).

SUBCRUSTAL EARTHQUAKES

Subcrustal earthquakes in this context refer to earthquakes within the subducting Juan de Fuca Plate. This is the best quantified earthquake source region. Ongoing microearthquake activity positions precisely the subducted plate (Fig. 6). The maximum size of earthquake is constrained to about the 7 magnitude range because the brittle thickness of the young subducting plate is very thin, less than 10 km, and thus places an upper limit to rupture area for typical rupture lengths of less than 100 km. Because of their depth, these earthquakes rarely have aftershocks.

The earthquakes within the descending Juan de Fuca Plate are concentrated in two regions. The first is in the vicinity of the west coast of Vancouver Island where a bend occurs as the plate goes from horizontal below the ocean to a shallow dip of ten to twenty degrees beneath Vancouver Island. The band of seismicity straddling the coast in Figure 3 represents this seismicity. The next concentration is below the Strait of Georgia and Puget Sound, where the plate bends further to a steeper dip of about thirty degrees (Fig. 7). This is probably the region where the buoyancy of the subducting plate changes from positive to negative, because of phase changes in the rocks of the oceanic crust (Pennington, 1983; Rogers, 1983b). It is marked by a band of seismicity, mainly in the 45 km to 65 km depth range, beneath the Strait of Georgia and Puget Sound (Fig. 6). This seismicity is also concentrated in a north-south sense between 47°N and 49.5°N but continues at a much lower rate both to the north and to the south. The arching of the subducting plate in this region to accommodate the bend in the coast line (Rogers, 1983a; Crosson and Owens, 1987) is the likely cause of this north-south concentration.

The extent of the band of small subcrustal earthquakes defines a region where several significant earthquakes occurred before the installation of a modern seismograph network (Table 1, Fig. 5). Earthquakes in 1949 and 1965 at the south end of Puget Sound caused considerable damage. Earthquakes near the damage threshold (in the magnitude 5 to 6 range) occurred in the Gulf Islands/San Juan Islands region in 1864, 1909, 1920, and 1976. The larger subcrustal earthquakes are consistent with a downdip tensional regime in the subducting plate (Rogers, 1983a), but smaller earthquakes show a more complex stress regime (Ma et al., 1991).

SUBDUCTION EARTHQUAKES

The potential for large earthquakes on the subduction interface in the Cascadia subduction zone has been discussed in the scientific literature for less than a decade (Heaton and Kanamori, 1984; Heaton and Hartzell, 1987; Rogers, 1988a) but the wide-ranging evidence for their occurrence has convinced most of the geoscience community that the hazard is real (Rogers, 1988b; Heaton, 1990). A number of different lines of evidence suggest that great subduction earthquakes

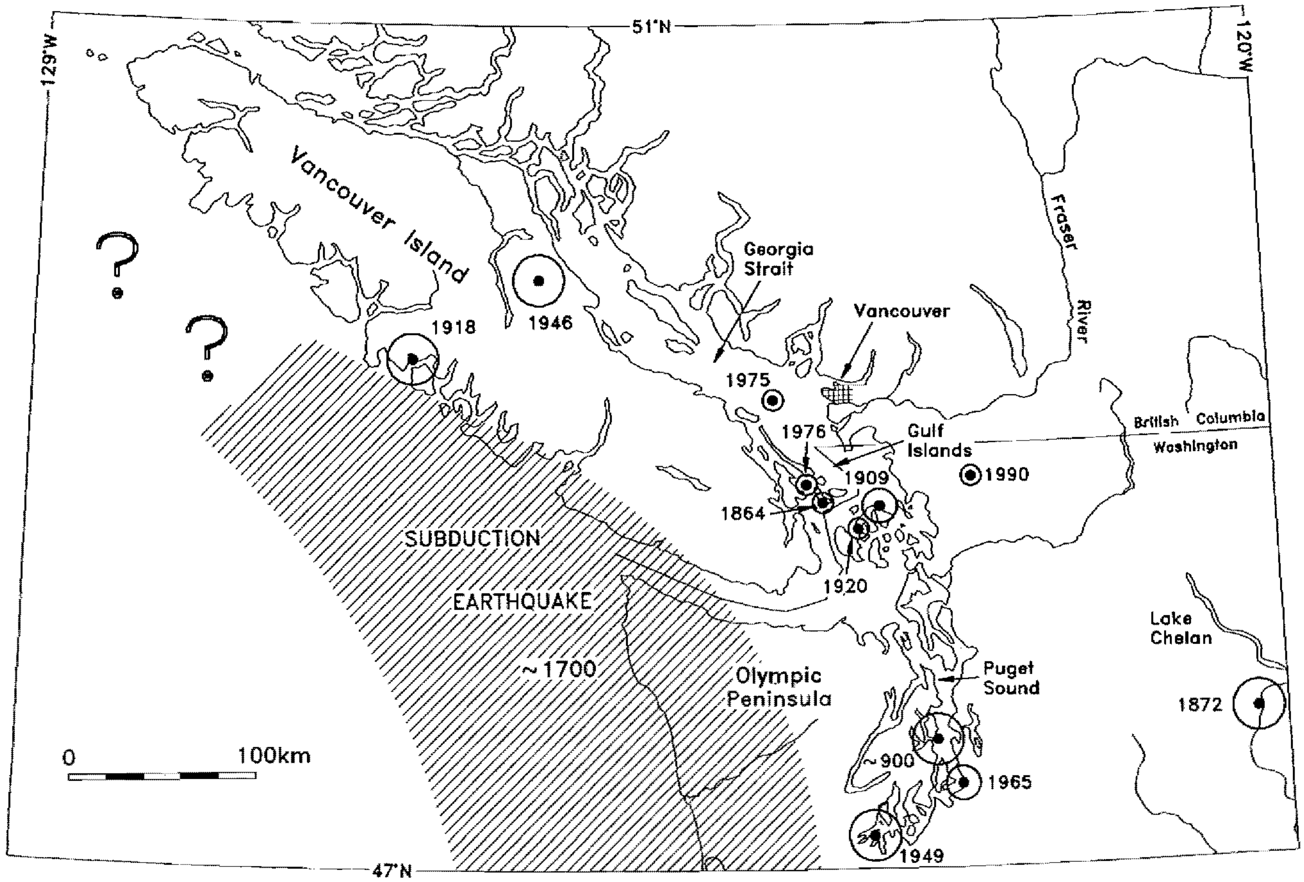

Figure 5. Significant earthquakes felt in Vancouver and the Lower Mainland listed in Table 1. Increasing size of circles indicates magnitudes of 5 and greater, 6 and greater, and 7 and greater. Shaded area along the west coast of Vancouver Island is a maximum potential rupture surface for a subduction earthquake (adapted from Dragert et al., 1994) and it may not extend as far east as shown.

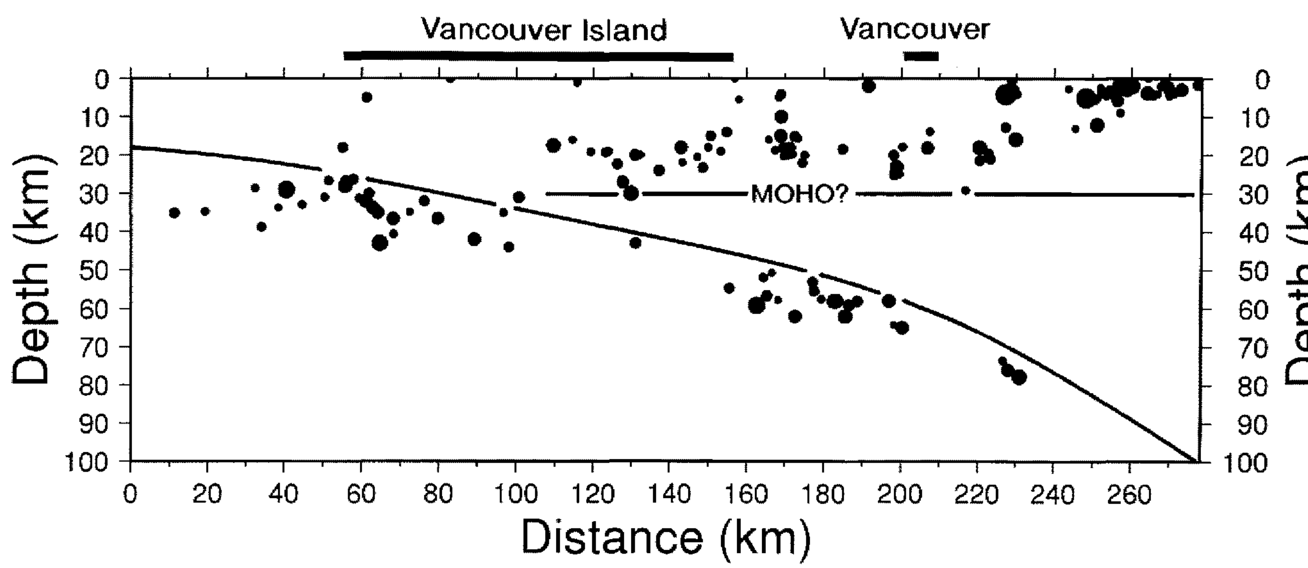

Figure 6. A 100 km wide corridor shown in Figure 3 projected on to a cross-section through Vancouver. Only earthquakes having uncertainties of less than 3 km in depth are shown.

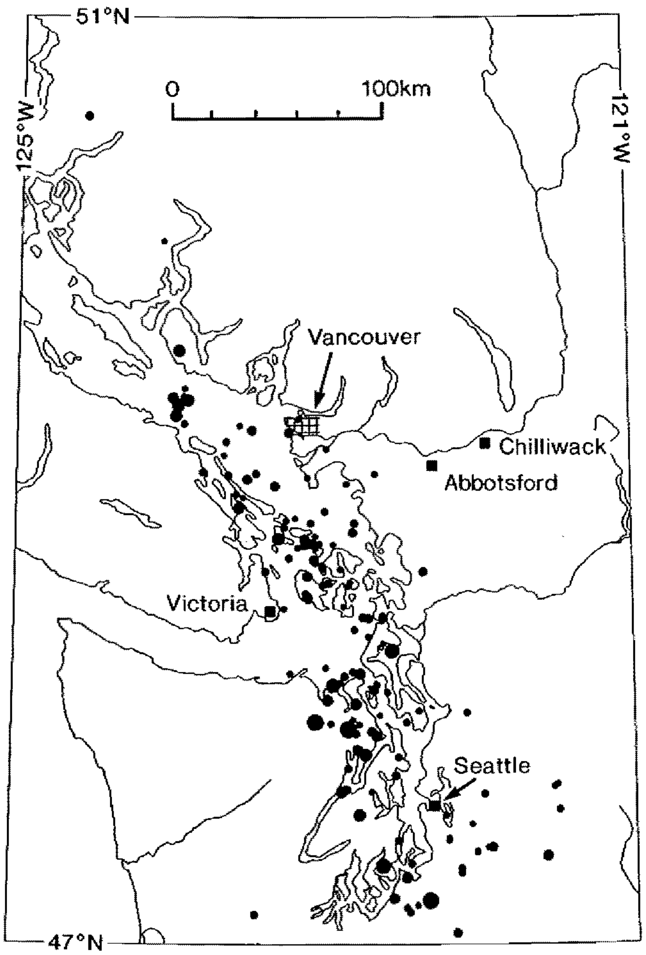

Figure 7. A subset of Figure 3 showing only subcrustal earthquakes with depths of 50 km and greater. The smallest earthquake plotted is magnitude 1 and the largest in the sample is magnitude 3.9.

have occurred in the past. In tidal marshes and estuaries on the west coast of Vancouver Island and along the outer coast of the United States, repeating sediment sequences of peat overlain by mud, often with a sand layer at the interface, are interpreted to result from abrupt subsidence events and tsunamis accompanying great earthquakes (Atwater, 1987; Clague and Bobrowsky, 1994). Layers of coarse sediments in deep sea mud deposits are ascribed to turbidity currents originating from periodic strong shaking of the continental margin (Adams, 1990). Dating, using growth rings of trees killed suddenly by salt water influx due to coastal subsidence (Atwater et al., 1991), places the last great earthquake about 1700 A.D. This date would be consistent with a legend in the oral history of the Cowichan people of southern Vancouver Island, which tells of a great earthquake that threw down houses and caused landslides in the days before the white man (Maud, 1978).

Elastic strain accumulation in the region has also been measured. Contemporary surface tilting and crustal shortening in a direction parallel to the movement of the subducting plate has been observed on Vancouver Island and on the Olympic Peninsula. The rates are too great to be sustained over a long interval. A locked subduction zone that periodically releases the strain in huge earthquakes can explain the observed pattern (Savage et al. 1991; Dragert et al., 1994).

Subduction earthquakes are rare but can be quantified. Themagnitudes are large, M = 8 or greater, but constraints on potential ground motion can be made because the position of the plate interface is known from the microearthquake seismicity (Rogers et al., 1990), seismic reflection (Hyndman et al., 1990) and refraction studies (Drew and Clowes, 1990), and information from other subduction zones can be used to obtain a realistic range of estimates on rupture behaviour, and thus surface ground motion. The most important parameter for hazard estimates is the downdip extent of the seismogenic zone. The maximum down-dip dimension of this potential rupture surface is defined by the length of the contact between the subducting plate and the brittle portion of the overlying continental crust, which is revealed by the depth of crustal earthquake activity (see Fig. 4 and 6). Temperature is the prime factor controlling the depth of brittle behaviour (Hyndman and Wang, 1993). Recent analysis of crustal deformation on Vancouver Island (Dragert et al., 1994) places the currently locked portion of this zone mainly beneath the continental shelf, west of Vancouver Island (Fig. 7). The eastern limit of the down-dip end of the potential rupture coincides at the surface roughly with the west coast of Vancouver Island, at least 150 km west of Vancouver. The Explorer plate subduction regime (Fig. 1), about 250 km from Vancouver, is very poorly understood at present.

SEISMIC HAZARD

In the National Building Code of Canada (NBCC) seismic hazard is defined as the level of horizontal ground shaking (acceleration or velocity) which has a 10% probability of being exceeded over a 50 year period (a probability of 0.0021 per annum). Canada is divided into 7 seismic hazard zones (0 and 1-6) based on the ground motion at this probability level (e.g., Heidebrecht et al., 1983). For each region there are two zoning values Z_a (acceleration – for small structures sensitive to damaging ground motions near 5 Hz) and Z_v (velocity for large and high rise structures sensitive to damaging ground motions near 1 Hz). For most of greater Vancouver and the Lower Mainland the 1990 NBCC zone values are $Z_a = 4$ and $Z_v = 4$, while points further east than Chilliwack in the Fraser valley drop to zones $Z_a = 3$ and $Z_v = 3$. There is a small region immediately adjacent to the U.S. border that technically meets the criteria of $Z_a = 5$ (Weichert, in press). The NBCC is a guide only; it is up to each community to adopt a seismic design value on the basis of municipal boundaries, relative economic considerations, and an understanding of the statistical nature of seismic hazard calculations.

The NBCC ground motion levels are calculated for firm soil. They may be higher on soft soils, which can amplify motions in selected frequency ranges that are defined by thickness and seismic velocity of the soil layer. Soil amplification is dealt with in the NBCC by using a multiplier to be applied to soft soil deposits. The most notable effect in the greater Vancouver region is the likely amplification in the frequency range of high rise structures on the deep soil deposits laid down by the Fraser River (e.g., Sy et al., 1991). Besides shaking, soil failure by landslides or liquefaction also contributes to the seismic hazard. Many examples of soil failure occurred on Vancouver Island and around the Strait of Georgia during the 1946 M = 7.3 earthquake in central Vancouver Island (Mathews, 1979; Rogers, 1980), but Vancouver and the Lower Mainland were just outside the affected region. Some steep slopes in the greater Vancouver region are prone to landslides (e.g., Eisbacher and Clague, 1981), and earthquake shaking may trigger landslides in these areas. Liquefaction deposits have been found in parts of the Fraser River delta (Clague et al., 1992) indicating the liquefaction susceptibility of some of the water saturated sand deposits, and providing direct evidence that strong shaking from large earthquakes occurred in the past.

The1990 NBCC does not consider a subduction earthquake. However, because we know the seismogenic fault is at least 150 km to the west of Vancouver, ground motions can be estimated. Whether attenuation relations based on peak values (e.g., Crouse, 1991) or more complex waveform modelling are used (e.g., Cohee et al., 1991), ground motions at that distance are the same order as those specified by the present NBCC zone classifications for greater Vancouver. The main difference in hazard from great subduction earthquakes is the long duration of strong shaking associated with large rupture surfaces, and the large area of shaking which affects many communities. The long duration can adversely affect certain types of structures, and the liquefaction potential of saturated sands.

Good examples of the effects that a subduction earthquake produces are seen from the damage caused in the city of Anchorage by the great 1964 Alaska earthquake (M = 9.2) (Steinbrugge et al., 1967). Anchorage is about the same distance from the down-dip end of the Alaska seismogenic zone as Vancouver is from the Cascadia seismogenic zone. The estimated three minutes of strong shaking during the Alaska earthquake damaged numerous large structures. However, single family wood frame dwellings and other similar small wood frame structures performed excellently as a class of construction, when not located in areas subject to soil failure.

SUMMARY

The dynamic geological setting of Vancouver and the Lower Mainland makes this region subject to frequent seismic activity and contributes to a higher risk of large damaging earthquakes than in other parts of Canada. Earthquakes that may present a hazard to the area are located in three distinct source regions: earthquakes within the continental crust, deeper earthquakes within the subducted Juan de Fuca Plate, and huge earthquakes on the subduction boundary between the two plates, about 150 km west of Vancouver. While Vancouver has not experienced a damaging earthquake in its short history, large earthquakes nearby have been strongly felt and there is paleoseismic evidence for very strong shaking in prehistoric time. Most of the region is placed in Seismic Zone 4 in the 1990 edition of the National Building Code of Canada. Ongoing microearthquake activity and felt earthquakes occurring in most years are reminders of the seismic hazard.

ACKNOWLEDGMENTS

Chris Spindler assisted with data files and computer plotting. Richard Franklin prepared the diagrams. Bob Horner has had the responsibility for earthquake location in this region since 1986. Precise locations near the international border and plots of data south of the border are made possible by regular data exchange with the University of Washington. Reviews by Dieter Weichert and John Cassidy improved the paper.

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Character of volcanism, volcanic hazards, and risk, northern end of the Cascade magmatic arc, British Columbia and Washington State

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Hickson, C.J., 1994: Character of volcanism, volcanic hazards, and risk, northern end of the Cascade magmatic arc, British Columbia and Washington State; in *Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia, (ed.) J.W.H. Monger; Geological Survey of Canada, Bulletin 481, p. 231-250.*

Abstract: A band of Oligocene (30 Ma) to Recent volcanoes and associated intrusive rocks extend from latitude 44°N in northern California to latitude 52°N in southwestern British Columbia. These centres make up the Cascade magmatic arc. The arc formed in response to subduction of the Farallon Plate (the present day Juan de Fuca Plate).

Volcanism from 5 Ma to the present is restricted to a linear belt of volcanoes broken into six segments. The northernmost segment extends from Glacier Peak in Washington State to the Bridge River cones in British Columbia. The most recent significant volcanic event in this segment of the arc is a Plinian to Peléan eruption from Mount Meager, (2400 BP). This event blocked the Lillooet River and spread ash across southern British Columbia into Alberta. Activity at Mount Baker occurred as recently as 1975.

The geological record of lava flows and volcanic deposits suggests that both basaltic and andesitic eruptions, and infrequent violent dacitic explosive eruptions, may occur. Of the numerous hazards associated with volcanic eruptions, the most likely to affect large numbers of people are airfall tephras and debris flows. A continuing hazard is posed by the extreme relief of many vent areas and the unstable nature of volcanic deposits. At least 44 large postglacial debris flows are known at Mount Baker and landslides from Mount Cayley and the "Barrier" (near Mount Garibaldi) have blocked the Squamish and Cheakamus Rivers. The loose unconsolidated nature of some volcanic deposits leads to increased sediment loads in surrounding drainages and leaching of soluble elements into the groundwater.

Résumé : Une bande de volcans d'âge oligocène (30 Ma) à récent et des roches intrusive associées s'étendent de 44° N en Californie septentrionale à 52° N dans le sud-ouest de la Colombie-Britannique. Ces centres constituent l'arc magmatique de la chaîne des Cascades, qui s'est formé à la suite de la subduction de la plaque Farallon (la plaque Juan de Fuca actuelle).

Le volcanisme survenu entre il y a 5 Ma et la période actuelle a eu lieu le long d'une ceinture linéaire de volcans subdivisée en six segments. Le segment le plus septentrional se prolonge du pic Glacier dans l'État de Washington jusqu'aux cônes volcaniques de la rivière Bridge en Colombie-Britannique. Le plus récent événement volcanique d'importance survenu dans ce segment de l'arc est l'éruption de type plinien à péléen du mont Meager, il y a 2 400 ans. Cet événement a obstrué la rivière Lillooet et répandu des cendres partout dans le sud de la Colombie-Britannique et jusqu'en Alberta. Le mont Baker a manifesté des signes d'activité à une date aussi récente que 1975.

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Les coulées de laves et les dépôts volcaniques portent à croire qu'il peut se produire des éruptions basaltiques et andésitiques, ainsi que des éruptions dacitiques explosives violentes mais peu fréquentes. Parmi les nombreux dangers associés aux éruptions volcaniques, ceux qui sont le plus susceptibles de toucher un grand nombre de personnes sont les tephras et les coulées de débris. Le relief extrême de nombreuses zones de cheminées volcaniques et la nature instable des dépôts volcaniques présentent un danger constant. Au moins 44 grandes coulées de débris postglaciaires ont été identifiées au mont Baker, et des glissements de terrain provenant du mont Cayley et de la «Barrière» (près du mont Garibaldi) ont obstrué les rivières Squamish et Cheakamus. Le caractère meuble et non consolidé de certains dépôts volcaniques accroît la charge solide dans les réseaux hydrographiques environnants, ainsi que la lixiviation des éléments solubles dans les eaux souterraines.

INTRODUCTION

Southwestern British Columbia is at the northern end of the Cascade magmatic arc, which is formed from magmas generated as a result of Cascadian subduction (Fig. 1). The volcanoes and associated intrusive rocks that make up the arc extend from latitude 40°N in northern California to latitude 52°N in southwestern British Columbia. The Cascade are has been active since at least Oligocene time, but most prominent physiographically are the post-5 million year stratovolcanoes and volcanic complexes (Fig. 1). This paper reviews magmatism within the arc from Oligocene time to Recent, discusses segmentation of the post-5 Ma arc, and addresses in some detail possible hazards associated with volcanism and the potential risks.

Geology, volcanic-hazards, and risk are covered in separate sections; the geological review summarizes earlier, more detailed treatments (Souther, 1970, 1977; Green et al., 1988; Guffanti and Weaver, 1988; Sherrod and Smith, 1990; Souther and Yorath, 1991). The hazard and risk sections draw upon work by Hoblitt et al. (1987) and Scott (1990) and previous papers by the author (Hickson, 1991, 1992, 1994). Volcano monitoring is not covered in detail in this paper, but a good review can be found in Ewart and Swanson (1992) and references therein.

In Canada, the Quaternary Cascade magmatic arc consists of three major volcanic complexes: Mount Garibaldi, Mount Cayley, and Mount Meager, all part of the Garibaldi volcanic belt (GVB). In southwestern British Columbia, shallow intrusions and one major preserved stratovolcano, Coquihalla Mountain, are Miocene in age and lie to the east of the main trend of the Garibaldi volcanic belt. Just south of the border in Washington State lies Mount Baker, an active stratovolcano, built upon two older volcanoes. Glacier Peak lies to the south of Mount Baker. These volcanoes, volcanic fields, and complexes make up the northernmost segment of the Cascade magmatic arc. Site of the largest population concentration in British Columbia, the Fraser Lowland lies approximately over the middle of the segment.

In this northern segment of the arc, repeated eruptions at a central vent, building large stratovolcanoes, is not the norm. This eruptive style is typical in the central portions of the arc, but father north volcanism occurs at centres that are closely spaced but not coincident. The result is complex volcanic fields rather than large central edifices. Of the entire Cascade magmatic arc, this northern segment has the lowest magma generation rates. These rates are expressed in cubic kilometres of magma per million years per kilometre of arc length (Sherrod and Smith, 1990). In addition it has the fewest recorded eruption periods in the last 15 000 years (Scott, 1990). Therefore, the short term threat of a volcanic eruption occurring is low. As near as can be determined, intervals between eruptions for these northern volcanic fields and complexes are very long; specific, individual centres may never erupt again. However, significant volcanism has occurred in the geologically recent past at Mount Meager, Mount Baker, and Glacier Peak.

In the event of an eruption, ashfall (tephra) and debris flows (lahars) will pose the greatest problems for downwind and downstream communities. Built in mountainous regions of rugged relief and unstable rock, the volcanoes and volcanic areas are more prone to slope failures than the surrounding peaks. These slope failures, in the form of debris flows and landslides, pose the greatest immediate threat to downslope and downstream communities. Many rock avalanches and debris flows have been documented from all of the volcanic areas and these events will continue to pose a threat in the future. At Mount Baker, 44 postglacial debris flows are known, one of which travelled more than 30 km from the mountain. Landslides from Mount Cayley and the "Barrier" (near Mount Garibaldi) blocked the Squamish and Cheakamus rivers during the nineteenth and twentieth centuries. In addition, landslides from Mount Meager volcanic complex have blocked upper Meager Creek, and debris flows have caused significant damage downstream.

CASCADE MAGMATIC ARC

Tectonic setting

Origin

The Cascade subduction zone formed from a much longer subduction zone that existed prior to 30 million years ago (Atwater, 1970). This older zone was the locus of subduction of the Farallon Plate. With the disappearance of the Kula

Plate under the Aleutian arc (prior to 20 Ma), the triple junction between the Farallon, Pacific, and North American plates migrated southward (Atwater, 1970). Concurrently, transcurrent motion along the San Andreas fault system extended north and south, coincident with changes in the orientation of the Farallon subduction zone and relative motion between the Pacific and North America plates. As the exposed Farallon Plate shrank, plate motions in this fragmenting region became more strongly controlled by the larger Pacific and North American plates. Rotation of the remnant Farallon Plate (the Juan de Fuca Plate), resulted in compression of the overlying accretionary wedge extending from the

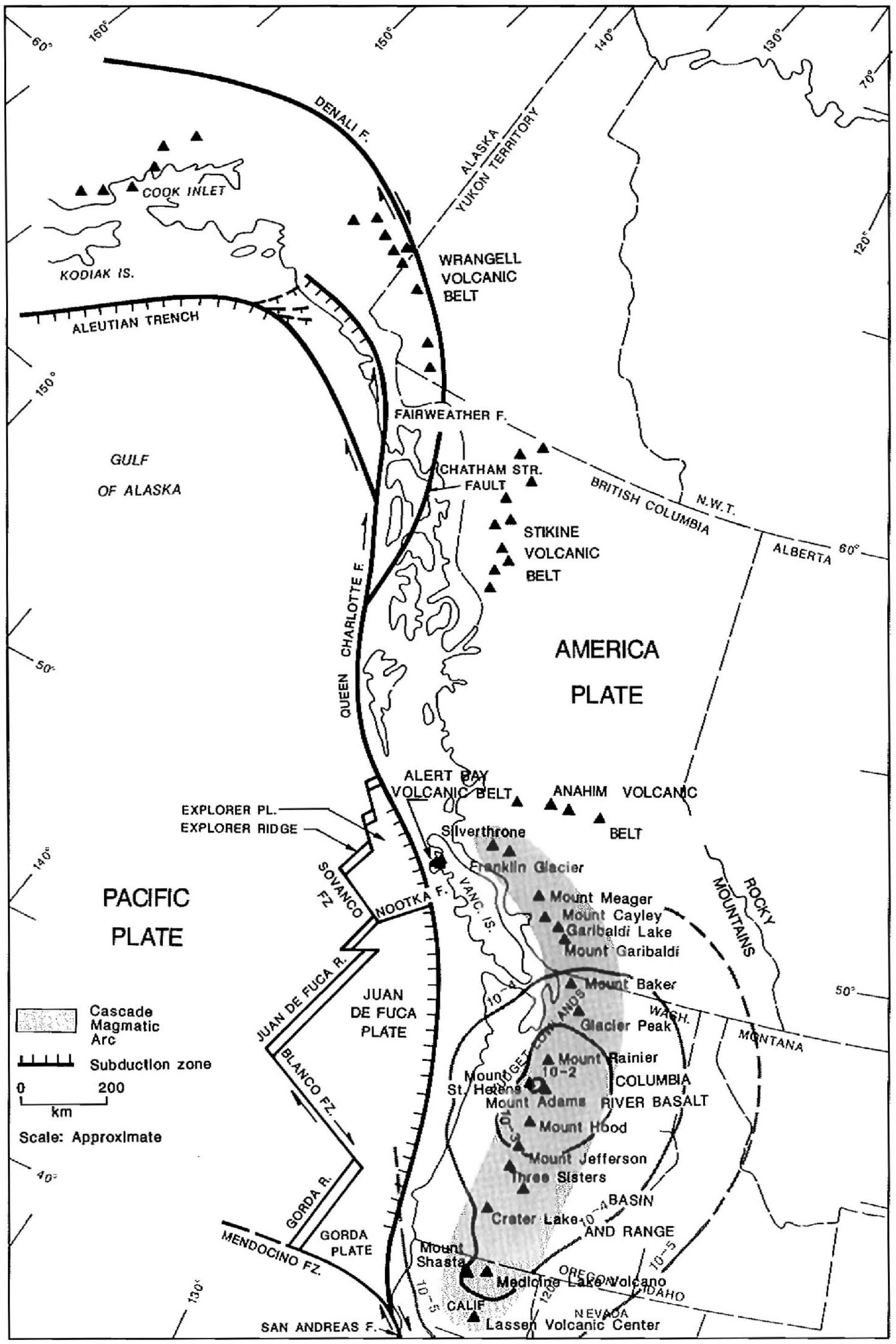

Figure 1. Major tectonic elements, Quaternary volcanic centres of the Canadian Cordillera, and volcanic centres of the Cascade magmatic arc. Numbered lines show the estimated annual probability of tephra fall from volcanoes within the Cascade magmatic arc as calculated by Hoblitt et al. (1987).

Puget Lowlands northward to southern Vancouver Island (Walcott, 1993). Inboard of this region (within the Coast Mountains), the most significant uplift rates in southwestern British Columbia (as much as 4 mm•a^{-1}; Holdahl et al., 1989), are recorded, possibly the result of crustal thickening due to the compression of the overlying accretionary wedge as suggested by Walcott (1993).

Northwestward migration of the southern portions of the subduction zone, due to the relative motions of the Pacific and North America plates, was concurrent with extension across the Basin and Range province (Fig. 1). Extension was accommodated by this motion, as it provided the space into which the Basin and Range extended (Walcott, 1993). Seven million years ago the convergence rate between the Juan de Fuca Plate and the North American Plate dropped by one-half to 3-4 cm•a^{-1}, from 6-7 cm•a^{-1} (Riddihough, 1977, 1984). At 4 Ma the Nootka fault system separated the Explorer Plate from the Juan de Fuca Plate (Hyndman et al., 1979), and Riddihough (1984) suggested that subduction of the Explorer Plate has ceased (Fig. 1).

Timing of volcanism

During this time of plate reorganization, the character of the Cascade magmatic arc was changing. Prior to 20 million years ago magmatism appears to have been widespread. Volcanism stretched from a point a few hundred kilometres inboard of the subduction zone to the limits of deformation in

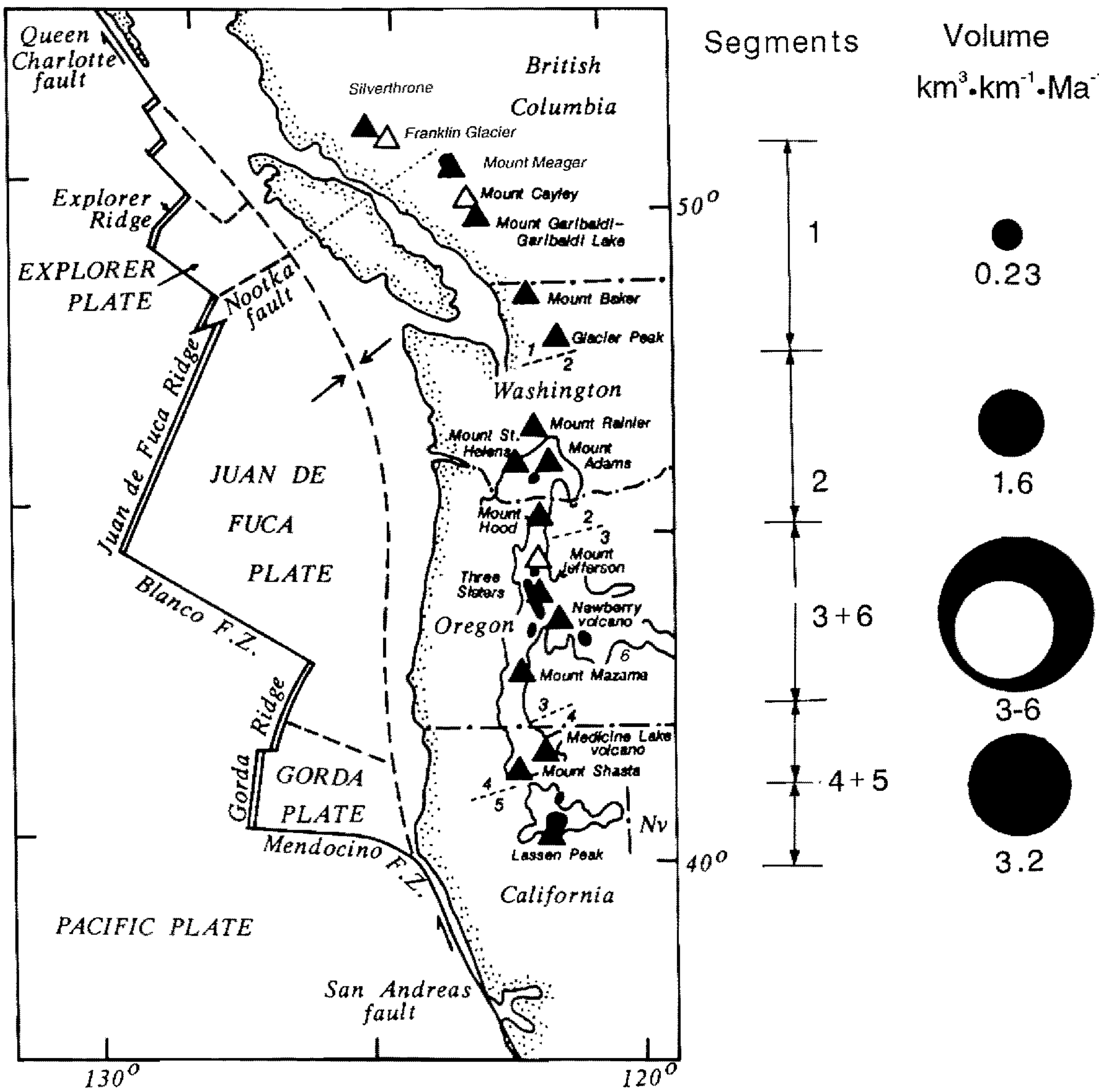

Figure 2. The Cascade magmatic arc, California to central British Columbia. Segments of Guffanti and Weaver (1988) are shown numbered 1-6. Magma generation rates are as calculated by Sherrod and Smith (1990). The volumes are for the age interval 2-0 Ma, note that the volumes for segments 4 and 5 have been combined and the volume of segments 3 and 6 is estimated at between 3 and 6 km^3•Ma^{-1}•km^{-1}.

the Intermontane Belt. In southwestern British Columbia, shallow intrusions and one major preserved stratovolcano, Coquihalla Mountain, lie to the east of the main trend of the Garibaldi volcanic belt. Souther and Yorath (1991) refer to these and similar centres as the Pemberton volcanic belt (PVB) which was active between 30-10 Ma and extends from the Queen Charlotte Islands (Masset Formation) to the U.S. border, crossing the trend of the Garibaldi volcanic belt at the location of Mount Meager. They suggest that these volcanic centres represent the magmatic front associated with subduction of the Farallon Plate. Chemistry of the PVB and the Masset Formation is calc-alkaline (Hickson, 1991) consistent with this hypothesis.

Inboard from the Pemberton volcanic belt, Chilcotin basalt covers much of the Intermontane Belt of British Columbia. These transitional to alkalic basalt flows issued from widely separated centres throughout the Intermontane (Mathews, 1988, 1989), and are the suggested back-arc equivalent of the Pemberton volcanic belt-Garibaldi volcanic belt (Souther and Yorath, 1991). Not unlike the southern portions of the arc, (prior to 20 Ma and extending up to 10 Ma), volcanism was widespread east of the present Quaternary Cascade magmatic arc. Volcanism was prevalent throughout the Basin and Range (Guffanti and Weaver, 1988, their Fig. 1a), and the Columbia Plateau (Columbia River basalt, Fig. 1). Eruptions ceased first in the central portions of the Basin and Range and subsequent volcanism became progressively confined to the western edge where the Basin and Range abuts the Cascade magmatic arc. The Columbia River flood basalt eruptions peaked during the mid-Miocene in a 3 Ma effusion of basalt, then abruptly ceased.

From 10 to 5 Ma, volcanism still had not migrated to the present Cascade magmatic arc, but was becoming more aerially restricted within a broad band to the east and west of the present arc (Guffanti and Weaver, 1988; their Fig. 1b and 2b). In the northern section of the arc, volcanism occurred only at the Franklin Glacier complex and in the Intermontane Belt. It is thought that by this time the triple junction had migrated to a point off the northern tip of Vancouver Island, and has remained there since (Souther and Yorath, 1991). The Alert Bay volcanic belt, a group of late Neogene (8-3.5 Ma) volcanic remnants and related plutons, may have been produced in response to this migration (Armstrong et al., 1985; Souther and Yorath, 1991) (Fig. 1). It was not until the Late Miocene-Pliocene (3-5 Ma) that volcanism became more restricted and formed a linear belt defined by the major Quaternary volcanoes of the Cascade magmatic arc.

Segmentation

The major volcanic edifices that make up the present Cascade magmatic arc are less than 5 Ma old. Unlike the aerially widespread expansive earlier volcanism, this more restricted distribution of volcanoes can be broken into distinct segments. The volcanic front in British Columbia and Washington State lies 300-400 km inboard of the subduction zone. In southern Washington State, Oregon, and California, the front lies 250-300 km inboard. Guffanti and Weaver (1988) analyzed the spatial and temporal distribution of 2821 vents, 5 Ma in age or younger, along the magmatic arc and based on this analysis divided the arc into 6 segments (Fig. 2). Segment 1 includes volcanoes of the Garibaldi volcanic belt, Mount Baker, and Glacier Peak. A gap of 90 km separates Glacier Peak at the end of Segment 1 and Mount Rainier in Segment 2. Segment 2 extends from Mount Rainier to Mount Hood, Segment 3 stretches from south of Mount Hood to the California border, and is noteworthy for the dense distribution of andesitic vents it contains. The fourth segment includes Mount Shasta and Medicine Lake Volcano, and the fifth includes Lassen Peak. Segment 6, referred to as the High Lava Plains, is the belt of volcanoes that trends east southeast from Newberry Volcano near the Cascades into Oregon, and is part of the Basin and Range tectonic province.

The northern and southern segments (1 and 5) differ from central segments, do not have large central volcanoes and are volumetrically smaller. Segment 1 volcanoes are more aerially restricted than those to the south and are volumetrically the smallest (Sherrod and Smith, 1990). Segment 2 volcanoes form a broad band, 150 km wide that contains many basaltic centres. Segments 3 and 4 include large composite volcanoes but these volcanoes are surrounded by more abundant mafic volcanoes and volcanic fields than in segments 1 and 2.

Eruption frequency and rate

Sherrod and Smith (1990) calculated the Quaternary extrusion rate for the arc. They found that central portions of the arc have produced significantly greater quantities of magma, per million years per kilometre of arc length, than segments to the north and south. The extrusion rate for the central portion of the arc (segments 3 and 6) is 3-6 $km^3 \cdot km^{-1} \cdot Ma^{-1}$; the northern end (Segment 1), just 0.23 $km^3 \cdot km^{-1} \cdot Ma^{-1}$ (Fig. 2). Extensive erosion in the northern segment compared with regions to the south may result in a significant error (possibly as much as 100%) in the volume calculation but, even with doubling the figure, the relative rates between the segments remain unchanged. The magma generation rates for the central portions of the arc are comparable with rates for other arcs with similar plate velocities such as the Lesser Antilles (Sherrod and Smith, 1990). Magma generation drops significantly as converging rates (and hence plate velocity) diminish. In Central America convergence is 8.1 $cm \cdot a^{-1}$ and magma generation rates are 31 $km^3 \cdot km^{-1} \cdot Ma^{-1}$ (Sherrod and Smith, 1990). Along the Aleutian arc the average extrusion rate is 0.6 $km^3 \cdot km^{-1} \cdot Ma^{-1}$ (Sherrod and Smith, 1990). Convergence rate diminishes from 5.5 $cm \cdot a^{-1}$ to 0 at the western end of the 4700 km long arc (Sherrod and Smith, 1990). These figures, combined with low magma generation rates, suggest that convergence may diminish from the central region of the arc to the northern boundary. This is consistent with geophysical evidence that suggests that the Explorer Plate has stopped subducting (Riddihough, 1984). In addition, Walcott's (1993) proposed pivoting of the Juan de Fuca Plate suggests more rapid northwestward motion of the southern part of the arc at the expense of the northern section.

The central portion of the Cascade magmatic arc is most active in terms of eruptive periods, regardless of eruption size. Scott (1990, p. 180) defined an eruptive period as "a single eruption or series of eruptions closely spaced in time at a volcano .... that yield a preserved deposit and is differentiated from preceding and subsequent eruptive periods by one or more of the following criteria: (1) separated by an apparent dormant interval of decades to centuries, (2) distinguished by a change in vent location, and (3) marked by a distinct compositional change in eruptive products." These eruptive periods have been compiled by Scott (1990) and are shown in Figure 3. The northern end of the arc is the least active, although Scott (1990) tentatively identified periods during which the entire arc was active during the last 15 ka.

SEGMENT 1

In Canada, the major volcanic edifices included in Segment 1 are mounts Garibaldi, Cayley, and Meager (Fig. 2). Scott (1990) defined the northern end of this segment as a line extending inland from the Nootka fault zone (Fig. 2). This line separates the Bridge River Cones (just north of Mount Meager) from volcanic rocks near Franklin Glacier and Mount Silverthrone. Green et al. (1988) also recognized a change in magmatic style north of this line. Both Silverthrone and Franklin Glacier volcanic complexes are poorly understood in terms of their eruptive histories and tectonic setting. The edifices in Washington State included in Segment 1 are Mount Baker and Glacier Peak (Fig. 2).

Magmas erupted along the length of Segment 1 are diverse in composition. Calc-alkaline and alkalic basalt to calc-alkaline rhyolite are recorded and eruption style has varied between quiet effusion of lavas to cataclysmic caldera forming eruptions. The volcanic areas are characterized by dacitic dome complexes and associated small cinder cones and stratovolcanoes that vary in composition from basalt to andesite to dacite. All of the complexes have had explosive eruptions during their history. Several centres are noteworthy for the eruption of low viscosity dacite that produced flows as long as 18 km. All of the major edifices in Segment 1 are built upon high ridges of basement rock. This gives them anomalous elevation for the actual thickness of the volcanic deposits and volume of the centres. All of the volcanoes in Segment 1 support active glaciers and have been inundated by continental-scale glaciers during the Fraser and previous glaciations.

A number of studies address volcanism in Segment 1, but surprisingly few are aimed at detailed physical volcanological problems. References are given after the discussion of the major centres, but this northern segment of the Cascades remains the least studied, and thus, the most poorly constrained in terms of estimated eruptive volumes and frequency of eruptions. Eruptive frequency drops dramatically from Glacier Peak northward (Scott, 1990; Fig. 3). This apparent drop in frequency might be accounted for, in part, by lack of detailed stratigraphic observations or missing records due to erosion (in particular glaciation), severe weather, and steep slopes. These erosion factors preclude preservation of thin, poorly consolidated volcanic deposits. Through understanding the frequency and volume of eruptions, some measure of the hazards and risks associated with a volcano can be estimated.

Long repose periods, as much as several thousand years, between major explosive events at the volcanoes, typifies Segment 1 (Fig. 4), even at the most active centre, Glacier Peak and the most voluminous one, Mount Baker. Mathews (1958) has suggested there may be a causative link between glacial loading (or unloading) of the continental crust during ice ages and increased rates of volcanism in the Garibaldi volcanic belt during deglaciation. This is difficult to prove because of the incomplete geological record, but there is circumstantial evidence such as the temporal clustering of eruptions synglacially or just postglacial within the Garibaldi volcanic belt, that suggests this may be true.

Recent seismic imaging from Geological Survey of Canada supported Lithoprobe studies in the region of Mount Cayley produced a 'bright spot' which may be attributable to a magma chamber at approximately 15 km depth (R. Clowes, oral comm., 1990; Jones and Dumas, 1993). Hot springs near Mount Cayley, Mount Meager, Mount Baker, and Glacier Peak suggest that magmatic heat is still present. This long history of volcanism along a still active plate margin suggests we have not seen the last eruption in Segment 1.

Major volcanic complexes

The following are summaries of information on the volcanic regions within Segment 1. The reader is encouraged to look at the references at the end of each section which give more detailed stratigraphic, dating, and petrochemical information.

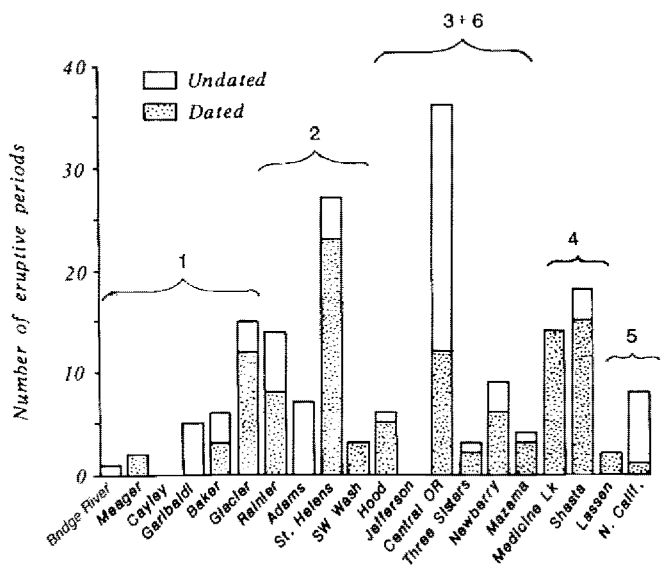

Figure 3. *Histogram showing the number of known eruptive periods for volcanic centres in the Cascade magmatic arc. Brackets enclose centres within the various segments. Compilation is of activity in the last 15 000 years (modified from Scott, 1990).*

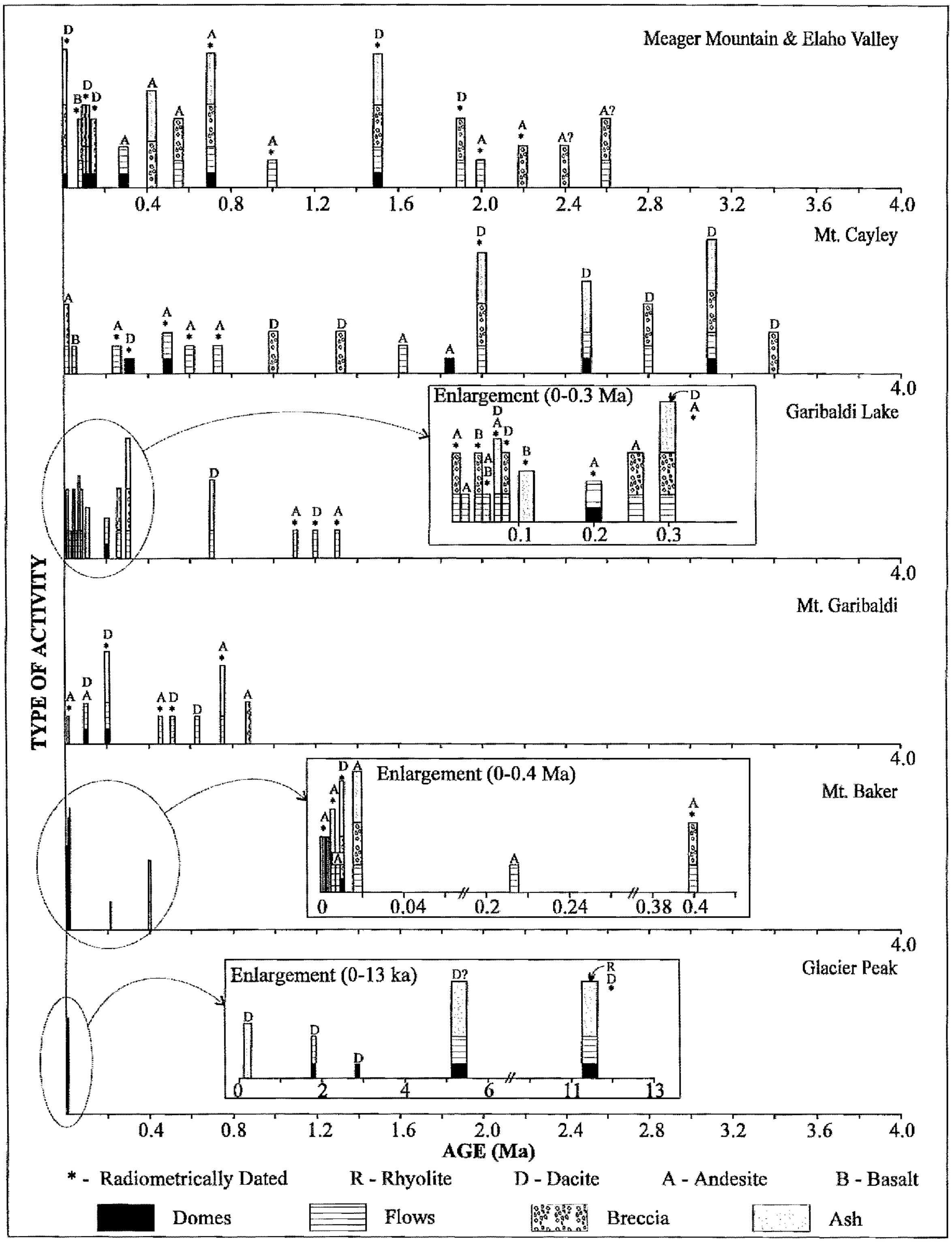

Figure 4. Compilation of eruption information for Segment 1 volcanic complexes. Bars show types of identified eruptive products and composition. Neither bar height nor width have any numerical value. Radiometric data compiled from Hoblitt et al. (1987) and Green et al. (1988); stratigraphic information from the references given within the text.

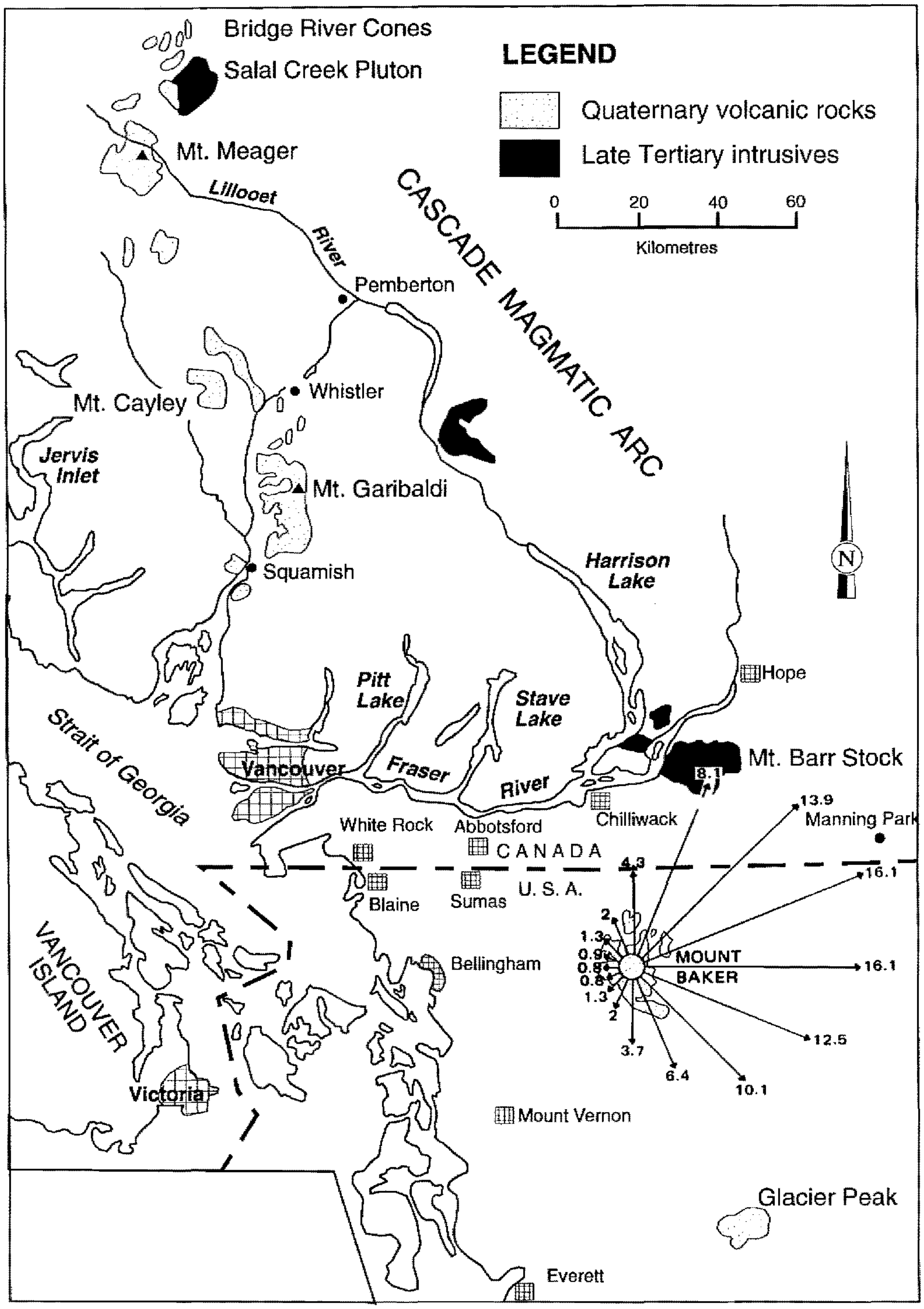

Figure 5. *Segment 1 of the Cascade magmatic arc showing major population centres, drainages and volcanoes in Segment 1, Cascade magmatic arc. Average directional frequency of winds based on 20 year record of winds at altitudes of about 3000-16 000 m averaged from Quillayute, Washington State. Values at arrowtips reflect percentage of frequency of winds blowing in the indicated direction (Hyde and Crandell, 1978).*

Unless noted, volume estimates for the centres were calculated by the author and can be found in Sherrod and Smith (1990).

Glacier Peak

Glacier Peak (Fig. 5) is an active[1] Pleistocene to Holocene stratovolcano built of clustered dacitic domes, very similar both in morphology and size to Mount Meager. Basalt has erupted from satellite vents on the flank of the volcano. The edifice volume is estimated at 29 km^3 (Sherrod and Smith, 1990) and the peak reaches 3000 m above the surrounding terrain. Like other volcanoes in Segment 1, Glacier Peak is perched on a ridge and its volcanic pile is no more than 500 to 1000 m thick. The volcano is relatively young; the cone is Pleistocene and the bulk of the activity Holocene (Fig. 4). Two eruptions, the first 11 000-12 000 years ago (>1 km^3 dense rock equivalent[2] (DRE)) and a second, a few centuries later (0.01-0.1 km^3 DRE), are the most voluminous post-glacial tephras produced in the Cascades, other than the Mount Mazama ash (34 km^3 DRE) (Hoblitt et al., 1987). These tephras have been identified at distances of 800 to 1000 km to the east in Montana (Mehringer et al., 1984) and in southern Alberta (Westgate and Evans, 1978). Pyroclastic flows associated with these tephra-producing eruptions travelled as far as 15 km from the volcano, and lahars reached more than 100 km distance. After a dormant period of 6000 years the volcano began erupting intermittently producing thin tephra falls, pyroclastic flows, and lahars. The most recent eruption was in the mid-1700s. More detailed information can be found in Beget (1982, 1983, 1984).

Mount Baker

Mount Baker is a large active andesitic stratovolcano that rises 2000 m, to an elevation of 3284 m, above the surrounding metamorphic and crystalline basement rocks of the Cascade Range. The estimated volume of the volcano is 72 km^3 making it the largest volcanic complex in Segment 1 and the largest cone of andesitic composition in the arc (Sherrod and Smith, 1990). Mount Baker supports one of the largest thermal fields in the Cascade magmatic arc. Underlying the present cone are remnants of two older volcanoes. The oldest is an early Pleistocene caldera complex 5 x 8 km in size, newly identified by Hildreth (1994). This caldera is comparable in size to Crater Lake and has been named Kulshan caldera. Overlying it is a (~0.4 Ma) dacitic to basaltic volcano, Black Buttes, largely destroyed by erosion. The present cone was formed before the last major glaciation (25 000-10 000 BP) and consists of interlayered andesitic lava flows, and breccia. Holocene activity at Mount Baker includes pyroclastic flows, lava flows and tephra erupted from Sherman Crater, a 450 m wide subsidiary summit crater (Fig. 6). Tephra, as well as at least one lava flow, also issued from a vent near the south base of the mountain at Schriebers Meadow. Increased hydrothermal activity starting in March of 1975 and tapering off in 1977, prompted a number of studies that are reported in Hyde and Crandell (1978). Besides these studies and a detailed petrological study carried out by Swan (1980) little work has focused on the volcanic history of the mountain. At present, detailed work is being carried out by W. Hildreth (geology) and K. Scott (hydrology/debris flows), United States Geological Survey.

Mount Garibaldi volcanic field (MGVF)

Mount Garibaldi and the adjacent peaks, Dalton Dome and Atwell Peak form Garibaldi volcano. Rising to an elevation of 2678 m, Garibaldi volcano is one of the three major composite volcanoes that make up the Canadian portion of Segment 1. The present day volume of the edifice is estimated at 6.5 km^3 (Mathews, 1958) and represents the largest centre in the combined Mount Garibaldi (MGVF) – Garibaldi Lake volcanic field (GLVF). The total present day calculated volume for the combined fields is 26 km^3 (Mathews, 1958). The total volume of magma erupted is probably much larger, perhaps double the volume preserved. Substantial erosion of many pre-Holocene centres has occurred, including collapse of a large segment of Mount Garibaldi itself. The style of volcanism and volume is similar to those for Glacier Peak, though the edifice volume of Mount Garibaldi is lower.

Garibaldi volcano is built upon crystalline basement rock of the Coast Mountains (Monger and Journeay, 1994). Andesitic to dacitic lava flows and pyroclastic rocks (of early to mid-Pleistocene age) filled paleovalleys, but now cap the ridges upon which Garibaldi volcano sits. Garibaldi volcano was built in a series of explosive (Peléan) dacitic phases during the waning stages of the last ice age (12 000-10 000 BP; Mathews, 1952a). These explosions built an apron of blocky, unconsolidated material around the vent area and out onto the surrounding sheet of thick glacial ice. This unconsolidated material was intruded by the growing domes and capped by a lava flow. As the ice receded, the edifice, formed from the coalesced domes and associated breccia, collapsed leaving steep (600 m) southwest-facing cliffs. Erosion of unstable material from this near-vertical face of the mountain contributes enormous amounts of sediment to debris flows originating on the slopes of the volcano. This debris accumulated to form Cheekye Fan that consists of an upper compound kame terrace and a lower alluvial fan. The estimated volume of the upper fan is 150 x 10^6 m^3; the lower fan is presently active and covers an area of 7 km^2.

Pre-Garibaldi volcano volcanism in the Mount Garibaldi volcanic field includes the older drumlin-shaped forms called the Ecnostuck flow and dome-shaped knobs that protrude from the southern flank of Round Mountain. The most recent

[1] "Active" and "dormant" when applied to volcanoes are imprecise terms. Active is generally taken to mean the volcano has erupted within historic time (or in regions with a very short historic record it is sometimes defined as the last 200 years), and is expected to erupt again. While a volcano is actually erupting it can be referred to as "active" or "in eruption". Dormant implies that the volcano has not erupted within historic time, but is expected to erupt again.

[2] Dense rock equivalent is obtained by first estimating the total volume of erupted material. An average porosity is then calculated from tephra samples and this "pore" volume subtracted from the total volume of material to obtain the actual amount of dense (nonporous) rock which was erupted.

eruptive activity is represented by the Ring Creek Lava flow, issued from Opal cone on the southeast flank of Mount Garibaldi. The dacitic lava flow extends 15 km down valley, and recent stratigraphic and radiocarbon dating suggests that the flow is between 10 000 and 9000 years old (Brooks and Friele, 1992).

Further information on activity near Mount Garibaldi can be found in Mathews (1952a, 1958) and Green et al. (1988), and a detailed summary in Green (1990).

Garibaldi Lake volcanic field (GLVF)

Garibaldi Lake volcanic field is immediately north of Mount Garibaldi, and consists of many volcanic vents active during the Quaternary. The two volcanic fields (Mount Garibaldi and Garibaldi Lake) have been separated based on two distinct magma associations. Mount Garibaldi volcanic field lavas are hypersthene-normative and nepheline-normative mugearite with lesser amounts of olivine tholeiites. Garibaldi Lake volcanic field comprises calc-alkaline basaltic andesites through rhyolite (Green, 1990). Black Tusk, an andesitic cone, is a landmark in the area (Fig. 7). It formed in two phases, the most recent phase of which was approximately 0.21 Ma ago (Green et al., 1988). Less than 100 000 years ago, basaltic eruptions produced several cinder cones and small stratovolcanoes, most important of which is the Cinder Cone and the associated 9 km long basaltic andesite flow. During glaciation, andesitic to basaltic eruptions produced the Table, a subglacial volcano, and other subglacial deposits. Activity culminated during glacial retreat when a dome was produced on the flank of Mount Price. Two (>300 m) thick andesite lava flows, referred to as the Barrier and Culliton Creek flows, erupted from Clinker Peak. Additional information can be found in Mathews (1952b), Green et al. (1988), and Green (1990) .

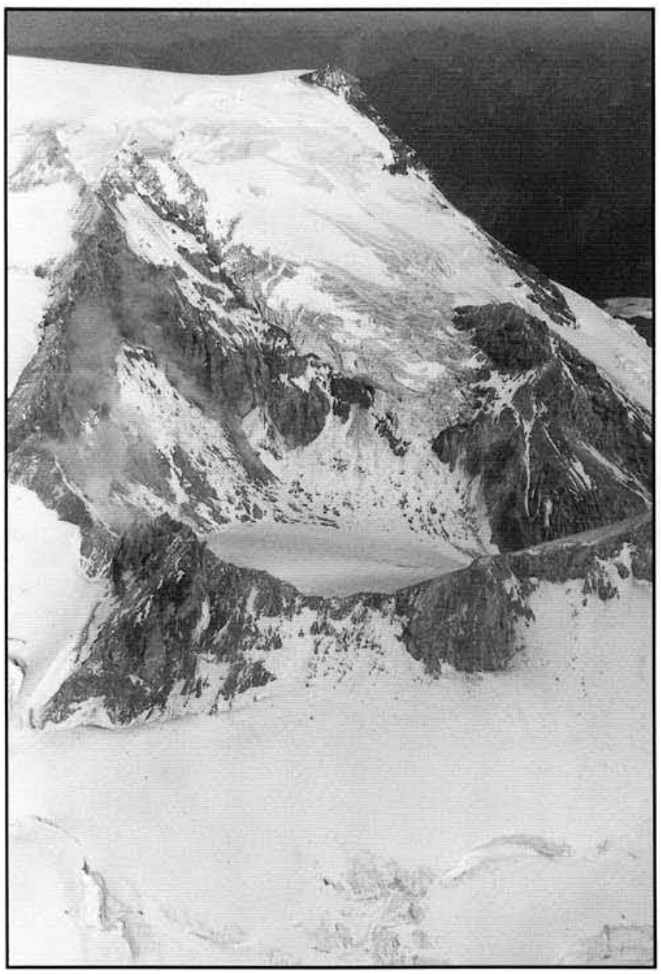

Figure 6. View looking northward at Sherman Crater, 450 m wide, on the southeast flank of Mount Baker. Sherman Crater was the site of the 1975-1977 activity and still has active fumaroles visible on the western rim of the crater in this photograph. Photograph was taken in September, 1992. GSC 1994-713C

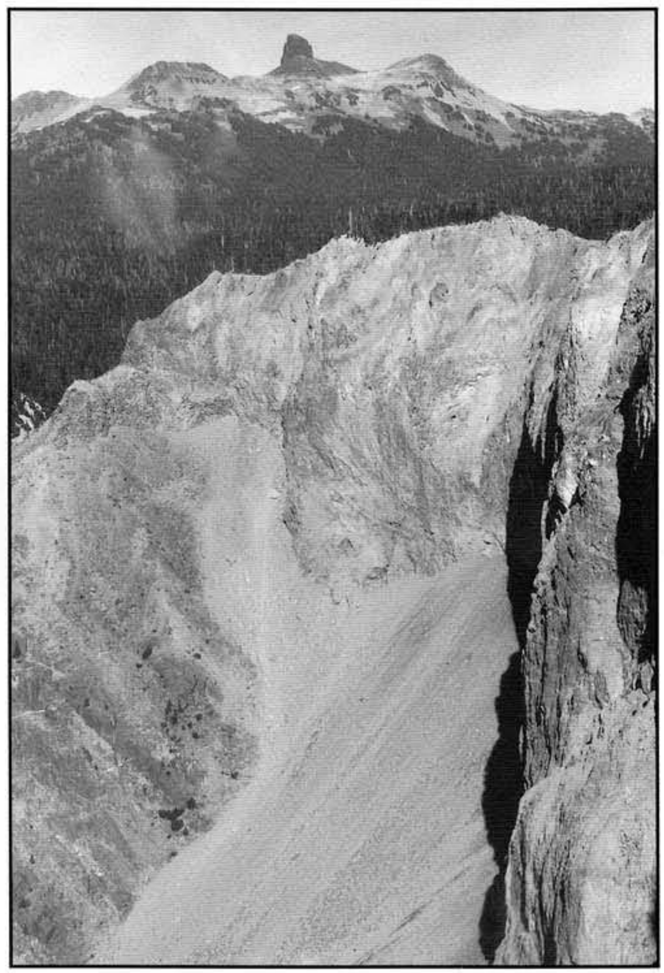

Figure 7. View looking northeastward over the sheer cliff face of the "Barrier" towards Black Tusk. The Barrier is the collapsed front of a lava flow from Clinker Peak. According to Moore and Mathews (1978), the collapse occurred in the winter of 1855/56. GSC 1994-713B

Mount Cayley

Mount Cayley is a substantially eroded edifice (total estimated volume of 13 km^3), that was built in two stages spanning the period from 4 Ma to 0.6 Ma. The first stage produced dacitic breccias and lava flows and culminated in the emplacement of a central dome. The second stage produced welded vent breccia. Effusion of dacite lava flows and satellite domes commenced again between 0.3 and 0.2 Ma extending the complex to the south of the original edifice. There are no known Holocene eruptions, but Mount Cayley is associated with a hydrothermal field. Evans (1990) has suggested that some postglacial landslides and debris flows near Mount Cayley may have been volcanically triggered. For further details including dates on Mount Cayley see Souther (1980, 1990) and Souther and Dellechaie (1984).

Mount Meager

The Mount Meager complex has been the site of volcanism since the Pliocene. It represents the largest stratovolcano in the Garibaldi volcanic belt at a calculated volume of 20 km^3. It is volumetrically and morphologically similar to Glacier Peak. The earliest volcanic deposit is a breccia as much as 300 m thick. It rests on a basement of Mesozoic metamorphic and crystalline rocks that have up to 400 m of relief. The early phases of activity were dacitic to rhyolitic in composition and continued from at least 1.9 Ma to about 1 Ma. This phase was followed by an andesitic, "middle" phase from 1 Ma to 0.5 Ma. Rhyolitic to dacitic volcanism resumed about 100 000 BP and continued into the Holocene. The most recently documented eruption was about 2400 (^{14}C date) BP from a vent on the northeast side of the mountain (Fig. 8). According to Stasiuk et al. (1994) this dacitic eruption started with a vent-clearing Plinian phase that produced tephra that can be traced as far east as western Alberta (Fig. 9). This early phase was followed by block and ash flows into the Lillooet Valley that blocked the Lillooet River. This blockage led to a cataclysmic outburst flood. Final stages of the eruption produced a short lava flow (Fig. 8). Large volume, fine grained, debris flows north of Mount Meager may have been volcanically triggered (P. Jordan and M. Bovis, pers. comm., 1991); if so, then the postglacial eruptive history of the mountain is still poorly known. Hot springs exist in several locations surrounding the mountain. Details on the Mount Meager complex can be found in Read (1977, 1990), Stasiuk and Russell (1989, 1990), and Stasiuk et al. (1994).

Figure 8. View westward into the crater formed during the 2400 BP eruption of Mount Meager. The crater is 1.5 km wide and the mound in the centre of the picture at the bottom of the crater, is the head of the dacitic lava flow that culminated the eruption. GSC 1994-713A

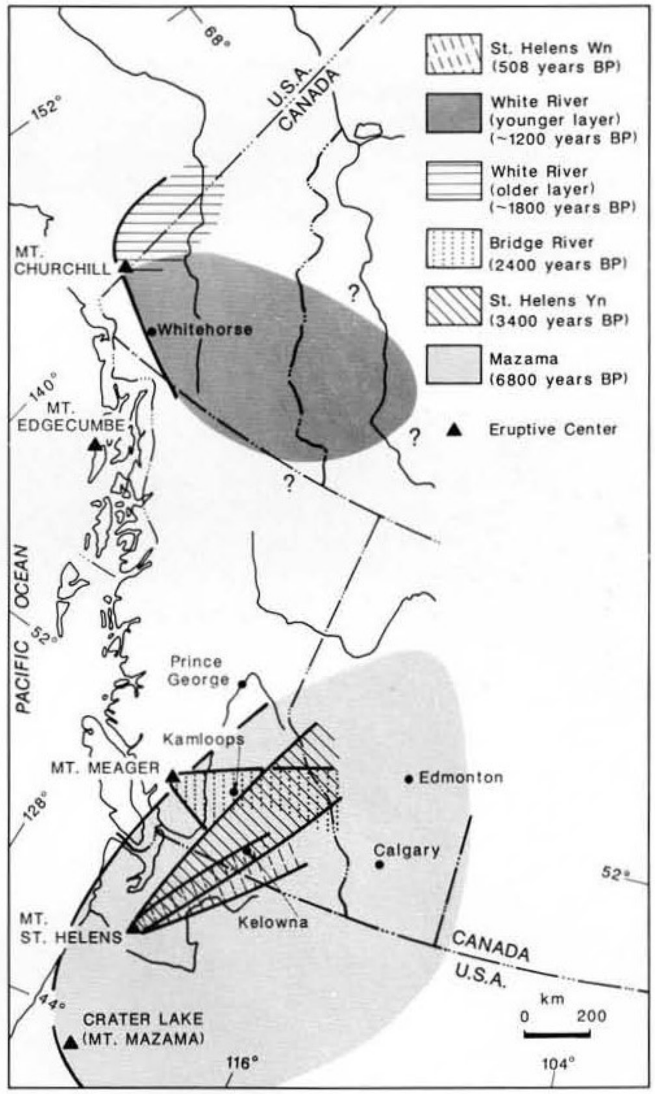

Figure 9. Distribution of major Holocene tephras in the Canadian Cordillera (modified from Clague, 1991). Thin tephras (less than a few millimetres) are only rarely preserved in the geological record and are not indicated on the figure. Ash from the 1980 eruption of Mount St. Helens fell in southeastern British Columbia (<1 mm) and ash from eruptions in Alaska have repeatedly fallen in the Yukon Territory. Most recently, ash from the September 17, 1992 eruption of Mount Spurr fell over central Yukon Territory. In addition to falling ash limiting visibility on the Alaska Highway, international air traffic flying in Canadian airspace was also disrupted

Bridge River cones

The Bridge River cones (sometimes called the Salal Creek cones) are a group of monogenetic cones, tuyas, and small stratovolcanoes. The total estimated volume is 8 km^3. Unlike Mount Meager, just to the south, volcanic rocks from these centres are chemically alkalic in composition, and consist of alkali olivine basalt. Mapping, dating, and geochemistry has been the focus of studies by Lawrence et al. (1984) and Roddick and Souther (1987).

Franklin Glacier and Mount Silverthrone

Small isolated remnants of flows and two larger Pliocene to Quaternary volcanic centres are found northwest of a line projected from the Nootka fault zone (Fig. 1). The larger areas are the Franklin Glacier and Mount Silverthrone complexes, neither of which have been thoroughly studied. Franklin Glacier lies 120 km northwest of Mount Meager at the head of Bute Inlet. Silverthrone lies another 55 km west-northwest between the Kingcome and Machmell rivers. Both complexes appear to be calderas, although most of the rocks preserved at Franklin Glacier are subvolcanic plutons. Dating of the Franklin Glacier Complex suggests that there are two phases of activity, the first between 6-8 Ma and the second between 2-3 Ma (Green et al., 1988). Rocks in the lower parts of the succession are dacitic in composition, and minor andesite flows and breccia are found in the upper portions of the section. Dates and a summary given in Souther and Yorath (1991) are based in part on the work of McKnight (1965) and Ney (1968).

Silverthrone is the larger of the two complexes and the best preserved. A summary of present knowledge of the complex is presented in Souther and Yorath (1991). From limited studies of the area, the complex is composed of rhyolite, dacite, and andesite domes, flows, and breccia. The bulk of the complex appears to have been erupted between 0.5 and 1 million years ago. Postglacial basaltic andesite lava flows fill the valleys of the Kingcome and Machmell rivers. The flow emanating from the north side of the complex (down the Machmell River) seems particularly well preserved. The original flow surface, complete with levees, is preserved along the sides of the valley. A K-Ar date on the flow yielded an age of 0.95 ± 0.2 Ma BP (Souther and Yorath, 1991); however, this date is anomalously old, and a ^{14}C date from barnacles 8.5 km upstream from the mouth of Machmell River and buried by the flow yields an age of 12 200 ± 140 BP (Blake, 1985). This is a limiting maximum age for the flow, which could be much younger. A complete understanding of Silverthrone and Franklin Glacier complexes awaits further study.

VARIATIONS WITHIN SEGMENT 1

Volcanic centres in Segment 1 share many similarities. Unlike many volcanoes to the south, they are all complex centres where volcanism has not been limited to central edifice vents. Mount Baker appears to have had the longest lived central eruptions that built the present day stratocone. Lavas at most centres have spanned the compositional range from rhyolite to basalt. Long time spans appear to elapse between eruptive periods and volcanic activity during these periods is as likely to be explosive as effusive (Fig. 4). The trend of Quaternary volcanic centres crosses the northwestward-trending structural grain of the region (Monger and Journeay, 1994), but is coincident with regions of greatest uplift in the past 4 Ma (Parrish, 1983). Older (Miocene) volcanic rocks are controlled by northeast-trending fault systems (Coish and Journeay, 1992), but the Quaternary structures do not follow these trends. Major structures that run northwest-southeast along Harrison Lake and the Pemberton Valley may control volcanism in the Mount Meager area, or at the least produce structural weaknesses that are exploited by rising magma batches.

Table 1. Volcanic-hazard summary (modified from Thorarinson, 1979 and Blong, 1984).

Volcanic hazard	Frequency of adverse effect/damage/death Distances (kilometres from volcanic vent)					
	< 10 short range	10-30 short range	30-100 mid range	100-500 mid range	500-1000 long range	> 1000 global
Seismic activity and ground deformation	C	C	VR			
Lava flows	F	C	VR			
Ballistic projectiles	C					
Tephra falls	VF	F	F	C	R	
Debris avalanches, edifice collapse, sector collapse	A	F	VR			
Pyroclastic flows, surges	A	F	R	VR		
Lahars, debris flows, and jokulhlaups	F	F	R	VR		
Tsunami	A	F	C	R	VR	
Atmospheric effects	C	C	R	VR	VR	
Acid rains and gases	F	F	R	R	VR	VR

* Hazard level is based on the relative frequency of damage and deaths given that the specific type of activity occurs. A = Always; VF = Very Frequent; F = Frequent; C = Common; R = Rare; VR = Very Rare.

Table 2. General relationships between volcano types, predominant lava, eruption styles, and common eruptive characteristics (from Tiling, 1989; Table 1.1, p. 2).

Volcano type	Predominant magma composition	Relative viscosity of magma	Eruption style	Common eruptive characteristics
Shield[1]	Basaltic (mafic)	Fluid	Generally nonexplosive to weakly explosive	Lava fountains, lava flow (long), lava lakes and pools
	Andesitic	Less fluid	Generally explosive but sometimes nonexplosive	Lava flows (medium), explosive ejecta, tephra falls
Composite[2]	Dacitic to rhyolitic (felsic)	Viscous to very viscous but can be nonexplosive	Typically highly explosive but can be nonexplosive,especially after a large event	Explosive ejecta, tephra falls, pyroclastic flows and surges, and lava domes

[1] Generally located in the interior of tectonic plates ("intraplate") and presumed to overlie "hot spots," but also may occur in other tectonic settings such as arcs. In British Columbia examples of shield volcanoes can be found in the Anahim Volcanic Belt (Fig. 1).
[2] Generally located along or near the boundaries of convergent tectonic plates (subduction zones); also called stratovolcanoes (e.g., Cascade magmatic arc is composed of central composite volcanoes such as Mount Baker).

Rogers (1985) and Walcott (1993) suggest that northern Washington State and British Columbia are in compression compared to regions of the arc farther south. Greater compression in this portion of the arc may explain the lesser volumes of magma produced in Segment 1. Hildreth (1981) suggested that in extensional environments magma can rise quickly as dykes along fractures, providing less chance for differentiation. This appears to be the case within segments 3 and 6. Segment 6 is the Brothers Fault Zone, an extensional environment that impinges on Segment 3. The extensional environment within the fault zone may explain the large volumes of basaltic magma in this central portion of the arc. A low convergence rate in a compressional setting, accommodating large stagnating magma bodies, may provide the answer for the low volume but differentiated magmas seen throughout Segment 1.

Table 3. Volcanic Explosivity Index (VEI) of Mount St. Helens and the deadliest eruptions since A.D. 1500 (modified from Tiling et al., 1990, p. 33).

ERUPTION	YEAR	VEI	CASUALTIES
Pinatubo	1991	6	450
Nevado del Ruiz, Colombia	1985	3	25 000
Mount St. Helens	1980	5	57
Mount Katmai	1912	6	0?
Mont Pelée, Martinique	1902	4	30 000
Krakatau, Indonesia	1883	6	36 000
Tambora, Indonesia	1815	7	92 000
Unzen, Japan	1792	3	15 000
Lakagigar (Laki), Iceland	1783	4	9000
Kelut, Indonesia	1586	4	10 000

VOLCANIC HAZARDS

Volcanoes, when they erupt, produce many hazardous events (Table 1). The style of eruption and type of hazard generally depends to a large degree on the composition of the erupting magma (Table 2). Basaltic eruptions generally pose minimal hazards in comparison to explosive, high silica rhyolitic to dacitic eruptions. Similarly, the hazards associated with an andesitic eruption may have less impact than those associated with a dacitic one. However, many caveats must be applied to these generalizations. Basaltic eruptions occurring during winter months in regions of heavy snow pack, such as the Coast Mountains of southwestern British Columbia, could produce devastating debris flows (lahars) or floods from rapidly melting snow.

Hazards associated with the eruption of intermediate to high silica content magmas can have moderate to extreme impact. The impact depends on the size of the eruption. Eruption size is quantified using a scale called the Volcanic Explosivity Index (VEI) (Newhall and Self, 1982). The VEI considers the volume of eruptive products, height of eruption cloud, duration of the main eruptive phase, and other parameters to assign a number from 0 to 8 on a linear scale. For example, the May 18, 1980 eruption of Mount St. Helens, which destroyed 632 km^2 of land, expelled 1.4 km^3 of magma (DRE) and produced an eruption column that peaked at an altitude of 24 km, had a VEI of 5 (Table 3). Table 3 is a listing of the VEI of some notable volcanic eruptions in relationship to the loss of life. The table shows that there is no direct relationship between size of the eruption and the number of lives lost, even though some of these eruptions have occurred in regions of similar population density. However, a relationship exists between eruption size and economic loss sustained by a region; the larger the eruption the greater and farther reaching the destruction.

Primary volcanic hazards (i.e., those that are directly associated with the volcanic eruption) are listed in Table 1. Besides the primary hazards, there are several secondary hazards associated with volcanoes. Secondary hazards, including mud and debris flows, landslides, and ground and surface water contamination, may persist at a volcano for decades after an eruption. However, whether a volcano is active or long dead, landslides and debris flows are inherent in regions of steep terrain, particularly where hydrothermal alteration has weakened the rocks. Only brief descriptions of the various hazards are given here, more thorough discussions can be found in Blong (1984), Hoblitt et al. (1987), Tilling (1989), and Ewart and Swanson (1992).

Edifice collapse/landslides/sector collapse/debris avalanche

By whatever name, this hazard involves the collapse of a part of the volcano. If the collapse is associated with a volcanic eruption it can result in a larger and potentially more explosive eruption than might otherwise be predicted. This was the case in the May 18, 1980 eruption of Mount St. Helens. Catastrophic depressurization of the volcano, due to the landslide, led to an eruption significantly different in character to that predicted. During inflation before eruption, slopes become oversteepened and unstable. Ground shaking associated with stream venting and magma assent can trigger collapse. Collapse of a large part of the edifice is termed a sector collapse. The landslide can become a debris avalanche as it is mobilized by melting snow and ice, carrying it farther than a similarly sized rock avalanche (Voight et al., 1981; Glicken, 1990). Landslides or rock avalanches also pose a secondary hazard. Those slopes which are steepened by explosions or later erosion, and weakened by hydrothermal alteration (Frank et al., 1975), are susceptible to failure at later times.

Lahars/debris flows

Debris flows, or to use the Indonesian word, lahar, are slurries of water and rock particles. The rock particles can be very fine grained, creating mudflows of milkshake-like consistency that behave like wet concrete. The flows can also be more dilute debris flows carrying many particle sizes. A range of sizes from clay-sized particles up to blocks as large as houses are not uncommon. Lahars are extremely destructive events; they are, however, topographically controlled and usually confined to valley bottoms. Lahars are most common and most voluminous as a result of explosive eruptions on snow-clad volcanoes. Pyroclastic flows (see below) can instantaneously melt ice and snow, creating large volume debris flows.

Debris flows pose a significant secondary volcanic hazard as they can occur days, weeks, or years following an eruption. Explosive eruptions, and to a lesser extent, effusive eruptions, can denude areas around a volcano and disrupt the drainage pattern. This denudation and disruption can lead to long term flooding problems around the volcano. In addition, vast areas around the volcano may be covered by unconsolidated tephra. The tephra is easily mobilized and can be washed away by heavy rainfall forming mudflows and debris flows.

Pyroclastic flows

Pyroclastic flows are dense avalanches of hot gas, fine hot ash, and heated rock particles that cascade down the slopes of the volcano during an eruption. These flows are most commonly associated with explosive eruptions, resulting from the collapse of the rising column of ash above the volcano. They can also form from disintegration of the front of a lava flow or lava dome or the "boiling" over of a crater or caldera. Pyroclastic flows originating from column collapse can have considerable momentum and travel great distances. Speeds of between 50 and 150 $km{\cdot}h^{-1}$ have been measured and distances of 30 km are not unusual (Table 1). Though rare, the largest eruptions have produced pyroclastic flows that extend 100 km from the volcano. All pyroclastic flows are extremely destructive (Table 1), destroying buildings, trees, or any objects that come in their path by impact, burial, or fire. People in the path of a pyroclastic flow have little chance of survival.

Pyroclastic surges

Pyroclastic surges are generated by volcanic explosions, usually in the early phases of an eruption. They are more turbulent and dilute than pyroclastic flows but also contain gas and rock debris. Travelling at greater speeds than pyroclastic flows, they have been clocked at over 360 $km{\cdot}h^{-1}$. Surges are extremely destructive and there is little hope for structures or people in their path.

Lava flows

It is a popular misconception that lava flows always accompany volcanic eruptions. They frequently do, but many eruptions never produce lava flows. Lava flows are among the least hazardous processes associated with a volcanic eruption (Table 1). They normally travel at slow speeds of a few kilometres per hour to a fraction of a kilometer per hour, and people and animals can normally move out of their way. However, structures in the path of a lava flow are usually destroyed.

Tephra/ballistic projectiles

Tephra (or ash) is finely comminuted volcanic rock and accompanies nearly all explosive volcanic eruptions (Table 2). Ballistic projectiles are also produced in explosive eruptions and represent the larger blocks of desegregated and comminuted material that because of their weight are not thrown any great distance from the vent (Table 1). In very energetic, explosive eruptions, fine tephra can be carried upwards into the upper atmosphere and ballistic projectiles thrown several kilometres. Tephra injected into the upper atmosphere can be carried by the jet stream for hundreds and thousands of kilometres. In lower energy eruptions, most of the tephra falls within a few kilometres from the vent and the accompanying ballistic projectiles a correspondingly shorter distance.

Ash, because it can be dispersed widely, may pose significant hazards to health and create economic problems over a wide area. Ash can pollute water supplies, disrupt transportation, and thick accumulation of heavy ash can collapse buildings or other structures. Inhaled ash can aggravate respiratory conditions such as asthma and bronchitis. Coarser particles can lodge in the nose (causing irritation), and the eyes (sometimes resulting in corneal abrasions). Silicosis has been attributed to long term exposure to volcanic ash (Vollmer et al., 1986). Ash, however, rarely causes direct deaths unless the fallout is so severe and rapid that suffocation results.

Ash can damage mechanical and electrical equipment. The ash is abrasive and, at great distances from the volcano, fine enough to work its way into bearings or other moving parts of equipment, potentially causing damage or failure. Computer equipment is particularly sensitive to this type of damage. Sometimes, electrical power can be disrupted because transformers conduct heat poorly and will overheat and explode when covered by only a few millimetres of ash.

Even if ash does not reach the ground it can affect aircraft flying at high elevations. The adverse impact of ash on high performance jet engines is not new. The first well publicized encounter was in 1982 when a Boeing 747 jet on a flight over Indonesia, encountered ash from Galunggung volcano. The jet was severely damaged and lost power in all four engines, but was able to make a successful emergency landing in Jakarta. Since then the number of incidents has been growing and the adverse effects more carefully studied (Casadevall, 1992, 1993, 1994; Anonymous, 1993a, b). Ash, ingested into high performance jet engines, causes melting and abrasion of the engines leading to engine failure. It also abrades the exterior of the aircraft, "frosting" cockpit windows and landing lights. Ingested ash can badly damage the navigation and hydraulic systems of the aircraft.

Thick accumulation of ash can significantly affect forestry and agriculture. Ash can rip the leaves from trees and bury small plants. Ash-covered trees and crops are difficult to work on because harvesting and cultivation equipment and vehicles can be damaged by the ash. Wind and moving equipment also redistribute the ash, prolonging problems. However, the long term impact can actually be beneficial, enriching the soil and occasionally, as with tephra from the 1980 Mount St. Helens eruption, act as mulch.

Acid rains and gases

Volcanoes produce large quantities of gases, mostly H_2O, but can also vent significant amounts of CO_2, CO, SO_2, HF, in addition to Cl and N compounds (Thorarinsson, 1979; Baxter et al., 1982; Baxter, 1990). Loss of life and damage has been attributed to each of these gases. Sulphur dioxide vented from Laki Volcano in Iceland in 1783 damaged crops and killed livestock and people. Those that survived faced starvation and many more died (Thorarinsson, 1979). Sulphur dioxide has produced a number of problems including increased incidents of acute asthma and bronchitis, and significant crop losses during venting of Masaya Volcano in Nicaragua (Baxter et al., 1982). Fluorine killed and disfigured livestock after the 1845 Hekla eruption and again in 1970 (Thorarinsson, 1979). Acid rain formed from the volcanic clouds can damage crops, and in Hawaii, the increased acidity of water collected in cisterns leaches heavy metals into drinking water, the long term health affects of which have not yet been completely assessed (Wright and Pierson, 1992).

Groundwater

A secondary hazard at volcanoes is contamination of ground and surface water. By virtue of the composition and physical attributes of the rocks that compose them, volcanoes are much more susceptible to weathering than most surrounding plutonic and metamorphic rocks. Weathering leads to increased particulate matter in streams draining the edifices and higher concentration of elements easily leached by percolating groundwater. Thermal springs are often associated with the volcanoes. These springs can have acidic waters that will enhance leaching and they will often carry dissolved metals. Because of these attributes, streams draining volcanic areas may have significantly different trace element chemistry than nearby streams draining other regions underlain by more chemically stable rock types (Bortleson et al., 1977).

VOLCANIC HAZARDS ASSESSMENT AND RISK

Erupting volcanoes only become a risk when there is something valued that may be destroyed – either people injured, lives lost, or property or resources damaged. Risk is usually assessed on the basis of the number of human lives that may be lost because of a hazardous event (Morgan, 1991). In fact, the number of fatalities resulting from natural disasters, throughout history, has taken only a small fraction of the number of lives lost in armed conflict. In 1000 years of record keeping, volcanic eruptions have resulted in less than 300 000 deaths (Tilling, 1989). Volcanic eruptions are far more likely to affect the economic base of a region, either by destroying resources or the infrastructure. Even eruptions in remote areas may adversely impact industries such as logging and fisheries.

Volcanic-hazard assessments made at individual volcanoes provide a basis for understanding the type of hazardous activity that might be expected at that volcano in the future, given its previous history. From this and other factors such as population density, an estimate of risk can be made. These assessments have been carried out for the American portion of the Cascade magmatic arc by Hoblitt et al. (1987) and provide a basis for long-range planning in areas likely to be affected by a volcanic event. These assessments are, however, limited by our level of knowledge of an individual volcano. The risk may be significantly underestimated due to incomplete knowledge of the eruptive history of a volcanic area. A summary of the hazards and probability of recurrence is given in Table 4 for Mount Baker and Glacier Peak. The record for the Canadian segment of the arc is too poorly known to calculate recurrence probabilities. In particular the number of documented debris flows is considerably fewer than the number that have occurred.

In general, the risk associated with a volcanic eruption from any volcano in Segment 1 is low, simply because the frequency of eruptions is low. Yokoyama et al. (1984), devised a method for assessing the general level of risk at a volcano. High risk volcanoes "score" 10 or above. Using this scheme and our present knowledge level, no Canadian volcano falls into the high risk category, nor do Mount Baker or Glacier Peak (Table 5).

Growing populations increases the risk posed by volcanoes both in Canada and abroad. For example, neither Pinatubo, Philippines, nor Nevado del Ruiz, Columbia, were

Table 4. Annual minimum probabilities of future eruptive events at major centres within Segment 1, Cascade magmatic arc. No probabilities are given for the Canadian centres in the arc as the compilations are not complete and knowledge is too limited in many areas to give meaningful probability figures (modified from Hoblitt et al., 1987).

Volcano	Time Interval[1]	Ex. Erup. (>0.1 km^3)	Ex. Erup. (<0.1 km^3)	Lahars/ debris flows	Pyroclastic flows	Lava flows
Glacier Peak	1200	2	15	67	120	0
		2E-04	1E-03	6E-03	1E-02	1E-05[2]
Mount Baker	1200	1	3	44	13	3
	1200	8E-05	3E-04	4E-03	1E-03	3E-04
Mount Garibaldi		0	1	—[3]	0	1
Garibaldi Lake	1200	0	2	2[4]	0	2
Mount Cayley	1200	0	0	2[5]	0	0
Mount Meager	1200	1	0	12[6]	3	1

[1] Estimated retreat of glacier ice in major valleys.
[2] A default probability of 1E-05 is assigned to all categories with zero occurrences.
[3] Debris flows with volumes in excess of 100 000 m^3 are expected more than once in 100 years in the Cheekye River. The Cheekye Fan has a total volume in excess of 300 X 10^6 m^3.
[4] Includes collapse of the Barrier in 1855/56 -- 25 X 10^6 m^3 landslide.
[5] 1977 - 300 000 m^3 1984 - 3.2 X 10^6 m^3.
[6] Includes both landslides and debris flows.

Table 5. Proposed criteria for identification of high-risk volcanoes (from Yokoyama et al., 1984). A score of 1 is assigned for each rating criterion that applies; nil if the criterion does not apply.

HAZARD SCORE	Glacier Peak	Mount Baker	Mount Meager	MGVF & GLVF
1) High silica content of eruptive products (andesite/dacite/rhyolite)	1	1	1	1
2) Major explosive activity within last 500 years	-	-	-	-
3) Major explosive activity within last 5000 years	-	-	1	-
4) Pyrsoclastic flows within last 500 years	1	1	-	-
5) Mudflows within last 500 years	1	1	1	1
6) Destructive tsunami within last 500 years	-	-	-	-
7) Area of destruction within last 5000 years is >10 km^2	1	-	1	-
8) Area of destruction within last 5000 years is >100 km^2	1	-	1	-
9) Occurrence of frequent volcano-seismic swarms	-	-	-	-
10) Occurrence of significant ground deformation within last 50 years	-	-	-	-
*RISK RATING				
1) Population at risk >100	-	1	-	1
2) Population at risk >1000	-	1	-	1
3) Population at risk >10 000	-	-	-	1
4) Population at risk >1 million	-	-	-	-
5) Historical fatalities	-	-	-	-
6) Evacuation as a result of historical eruptions(s)	-	-	-	-
TOTAL SCORE	5	5	5	5

* Population living within 50 km of the vent.

considered high risk volcanos. Loss of life from the 1991 eruption of Mount Pinatubo was minimized because excellent monitoring provided enough time to evacuate people from critical areas (Wright and Pierson, 1992). Nevado del Ruiz was monitored but warnings failed to reach the people. When it erupted on November 13, 1985, 25 000 people were killed – the greatest volcanic disaster since the eruption of Mount Pelée at the turn of the century. A poignant point brought out in Voight's (1989) a retrospective review of this event, was the observation that in 1845, a similar event wiped out 1400 people – virtually the entire population of the town at that time. In 1985, 30 000 people lived in the same area and a repeat of the 1845 event resulted in 20 times the loss of life of the earlier event. In a similar vein, the Philippine volcano of Mayon produced pyroclastic flows during its 1814 eruption that killed 1200 people, most of the inhabitants living in the area – 800 000 people now live in the same region (Voight, 1989).

Volcanic monitoring

As noted for Pinatubo, monitoring precursory activity provided warning of the eruption. Monitoring volcanoes has become a more refined art based on the scientific information gained from the eruption of Mount St. Helens, but it is still fraught with difficulties. Geologically well studied volcanoes at which several types of ongoing monitoring have been undertaken, provide the best opportunity for prediction and forecasting of future events. Large, explosive eruptions are usually preceded by precursory activity. If monitoring equipment is in place, it will likely provide some warning of an impending eruption. Basaltic eruptions, especially effusive ones may have little or no precursory activity associated with the rise of magma. Information on monitoring techniques can be found in Tilling (1989) and Ewart and Swanson (1992) and the references therein.

Landslides/sector collapse/debris avalanche

This hazard poses a risk close to the volcanic edifice (Table 1). Considerable threat in the form of large rock failures exists at the volcanic centres throughout Segment 1 (Read, 1981; Clague and Souther, 1982; Evans, 1990). The volcanoes are in extremely rugged regions of high relief and are underlain by unstable, poorly consolidated and/or strongly jointed volcanic rocks. These conditions have already led to many failures (Evans, 1990; Hungr and Skermer, 1992). Examples include the Dusty Creek (Clague and Souther, 1982; Evans, 1990; Hungr and Skermer, 1992) and Avalanche Creek landslides (Cruden and Lu, 1992) at Mount Cayley; collapse of the Barrier Lava flow into the Cheakamus valley (Moore and Mathews, 1978) (Fig. 7); and many landslides at Mount Meager, one of which killed 4 people (Jordan et al., 1986; Evans, 1992, 1987). At Mount Baker, six historic landslides, involving hydrothermally altered material, have been recorded between 1958 and 1975. Sector collapse is also a possibility if renewed activity were to occur near the summit of Mount Baker. Extensive hydrothermal alteration near Sherman Crater predisposes the area to collapse associated with oversteepening produced by inflation that might accompany the emplacement of magma within the edifice.

Lahars/debris flows

This hazard poses a risk to southwestern British Columbia communities. During the Mount Meager, 2400 BP, eruption, ash was carried as far as Alberta (Nasmith et al., 1967), and pyroclastic flows extended down the Lillooet River a distance of 7 km (Stasiuk et al., 1994). Upstream of the eruption, the Lillooet River formed a lake due to damming of the river by pyroclastic flows. This dam failed catastrophically creating a flood that carried 10 m sized blocks of welded pyroclastic debris several kilometres downstream. Very likely this flood reached Lillooet Lake and inundated parts of the valley downstream from the eruption.

Eruptions almost anywhere along the valley corridor between Vancouver and Pemberton may cut the rail and road links, as often happens when nonvolcanically induced landslides and debris flows occur (see Hungr and Skermer, 1992; Evans and Savigny, 1994). Jordan et al. (1986) and Jordan (1990) recorded many large postglacial debris flows unrelated to eruptions at Mount Meager.

An eruption of Mount Baker may affect the region around Sumas Prairie (near Abbotsford) (Fig. 5). High water in the Nooksack River drainage can overflow into the Sumas River drainage basin, flooding parts of the Fraser River valley. This process was demonstrated in the fall of 1990 when the Nooksack overflowed and water drained northward into Canada, near Sumas.

Steep terrain and unconsolidated material continue to pose a debris flow threat in these volcanic regions. Debris flows of comparable volumes generated in volcanic areas have much longer run-out distances than those generated in nonvolcanic areas, in part, because of a greater percentage of fine material in "volcanic" debris flows (Jordan, 1990).

Other flowage phenomena/ballistic projectiles

Pyroclastic flows, surges, and lava flows, besides ballistic projectiles, pose a significant risk only to communities and resources close to the volcanic edifice (Table 1). However, studies suggest that areas within 50 km of an erupting volcano may be severally affected by these and other volcanic hazards (Hoblitt et al., 1987). Beyond this limit the main hazards (and thus risk) are from tephra and debris flows (Table 1), the effects of which can be localized due to prevailing weather patterns and topography.

Lava flows

Lava flows pose little threat or significant risk unless the eruption occurs in the winter or occurs beneath or near glacial ice. If lava flows over significant amounts of snow, it may lead to melting and the production of debris flows that may extend farther than the lava flow. In addition, water entering a basaltic vent can produce a large explosion, much larger than normally associated with a basaltic eruption. Consequently, the presence of water, snow, or ice will increase the risk of significant impact from an eruption. Subglacial volcanism has also created large destructive jokulhlaups (Table 1).

Tephra

Impact from tephra poses the greatest risk to southwestern British Columbia. In the past 12 000 years, volcanoes in the Cascade magmatic arc have erupted more than 200 times (Scott, 1990). Several of these eruptions have deposited significant quantities of ash in southern British Columbia (Fig. 9). However, all major cities of greater than 100 000 population are located west of the Cascade magmatic arc and prevailing high altitude winds blow easterly. As a result these population centres are less likely to have significant accumulations of tephra. Hoblitt et al. (1987) calculated the recurrence interval for tephra fall out for the Cascade arc (Fig. 1). In the Fraser Lowland 10 cm of ash can be expected once in 10 000 years, and 1 cm once in 1000 years. Small amounts of ash can be expected more frequently. The May 18, 1980 eruption of Mount St. Helens deposited a millimetre thick layer of ash from southeastern British Columbia to Manitoba. The most likely source of small amounts of ash is Mount Baker. Figure 5 gives the frequency of wind directions that would influence dispersal of ash from Mount Baker.

CONCLUSIONS

Segment 1 of the Cascade magmatic arc contains two active volcanoes and three dormant volcanic complexes. Eruptions in historic times have been recorded at both Mount Baker and Glacier Peak. Explosive eruptions have occurred, the most recent at Mount Meager. This being said, Segment 1 of the Cascade magmatic arc poses the lowest risk to surrounding populations from the effects of volcanic eruptions. The compressional tectonic regime and low convergence rate of the Juan de Fuca Plate combine to reduce the frequency and volume of eruptions within the northern part of the arc. However, long repose periods and explosive volcanism are also hallmarks of this segment. Though the probability of an eruption is low, potentially it could be of significant size. Peterson and Tilling (1993) have commented that countries, even those that are technologically advanced, often fare poorly when faced with an impending volcanic eruption. The frequency of events is so low that there is often disbelief among the public that an eruption may occur and unwarranted panic when it does occur. Modern monitoring techniques should give sufficient warning of an eruption to allow government agencies, the private sector, and the public to prepare themselves for an eruption. Plans are already in place to deal with notification of a volcanic event (Hickson and Spurgeon 1993; Hickson, 1994).

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Landslides in the Vancouver-Fraser Valley-Whistler region

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Abstract: A great diversity of landslide types occur in the Vancouver region in response to high relief, steep slopes, heavy rainfall, seismicity, and a variety of landslide-prone materials. Rockfalls and small rock avalanches (less than a million cubic metres) are a significant hazard to land use development but their biggest impact has been on the transportation network and the Fraser River fishery. The deposits of larger rock avalanches (greater than a million cubic metres) are common throughout the region and have occurred along major transportation routes in the Fraser Valley and the Squamish-Pemberton corridor in the last 10 000 years. Noncatastrophic mountain slope deformation is also widespread. Such processes form linear topographic features such as cracks and fissures. Volcanic rocks of the Garibaldi Volcanic Belt are particularly prone to massive rapid landslides, some of which have blocked major rivers and formed temporary lakes upstream. Since the late Pleistocene, major collapses have taken place on the western flanks of Mount Garibaldi and Mount Cayley volcanoes. Large landslides continue to occur in the historical period and are a major consideration in land development in the Belt. Channellized debris flows within steep mountain watersheds triggered by heavy rains occur throughout the region. Debris flow defensive structures have been constructed by provincial authorities at numerous locations to protect transportation routes and/or communities. Landslides in Pleistocene sediments are also important. In addition, a number of cases of catastrophic seepage erosion have been documented. Submarine failures (outside the Fraser Delta) occur on delta fronts in both marine and lacustrine environments. The expansion of development in the Vancouver region is increasing the vulnerability of communities, transportation routes, and the resource base to landslides.

Résumé : Une grande variété de glissements de terrain se produisent dans la région de Vancouver en réponse au relief prononcé, aux fortes pentes, à l'abondance des précipitations, à la séismicité et à la présence de divers matériaux susceptibles de glisser. Les écroulements et les petites avalanches de pierres (moins d'un million de mètres cubes) constituent un danger significatif pour la mise en valeur et l'exploitation des terres, mais leur plus forte incidence a eu lieu sur le réseau de transport et les pêches du fleuve Fraser. Les grandes avalanches de pierres (plus d'un million de mètres cubes) sont survenues le long des grandes voies de transport dans la vallée du Fraser et du couloir de Squamish-Pemberton au cours des 10 000 dernières années; leurs dépôts sont communs dans toute la région. Les déformations non catastrophiques de versants montagneux sont également répandues. Ces processus créent des détails topographiques linéaires tels que des fentes et des fissures. Les roches volcaniques de la ceinture volcanique de Garibaldi sont particulièrement susceptibles de glissements de terrain rapides et massifs, dont certains ont obstrué de grands cours d'eau et donné naissance à des lacs temporaires, en amont. Depuis le Pléistocène tardif, de grands effondrements se sont produits sur les flancs occidentaux des monts Garibaldi et Cayley. De grands glissements de terrain ont continué à se produire pendant la période historique et sont un important facteur à considérer en ce qui concerne l'utilisation des terres dans la ceinture volcanique de Garibaldi. Des coulées de débris, qui sont canalisées dans des bassins hydrographiques de régions montagneuses escarpées et déclenchées par des précipitations abondantes, ont lieu dans toute la région. Les autorités provinciales ont

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construit des structures en de nombreux endroits pour protéger les voies de transport ou les collectivités contre les coulées de débris. Les glissements de terrain survenant dans les sédiments pléistocènes sont également importants. En outre, de nombreux cas d'érosion catastrophique causée par des infiltrations ont été documentés. Des effondrements sous-marins (à l'extérieur du delta du Fraser) se produisent sur des fronts deltaïques marins et lacustres. L'intensification du développement de la région de Vancouver accroît la vulnérabilité des collectivités, des voies de transport et de la base de ressources aux glissements de terrain.

INTRODUCTION

The Vancouver region (Fig. 1) is situated in the southwest corner of the Canadian Cordillera where a wide variety of landslides styles and processes have been described (Eisbacher, 1979; Eisbacher and Clague, 1984; Evans, 1982, 1984, 1990a, 1992a; Cruden, 1985; Cruden et al., 1989; Evans and Gardner, 1989; Clague, 1991; Clague and Evans, 1994b; Evans and Clague, unpublished data). Evidence for landslide activity in the prehistoric past is seen in the widespread distribution of landslide deposits throughout the region, and in the historical period (taken to be 1855 to the present), in frequent landslide events as documented by Evans and Clague (unpublished data).

The interaction between high relief, steep slopes, heavy rainfall, seismicity, a complex active tectonic history, and landslide-prone materials in the region, gives rise to a great diversity of landslide types, each with particular geotechnical characteristics, geological causes, and dynamic behaviour. This presents a considerable challenge to geoscientists who face the task of assessing the hazards posed by landslides (e.g., Eisbacher, 1982; Evans and Gardner, 1989; Morgan, 1986, 1992; Morgan et al., 1992; Hungr et al., 1993; Fell, 1994), to civil engineers in their design of engineering works to mitigate landslide hazards (e.g., Hungr et al., 1984, 1987; Hungr, 1993; Martin et al., 1984; Moore et al., 1992), and to landuse planners and legislators who face the task of regulating development in response to these hazards assessments (e.g., Lister, 1980; Cave, 1992a, b; Berger, 1973; Buchanan, 1983). Geological hazards are one of the many challenges presented by the physical environment to living in or near the mountains (Spearing, 1976).

The Vancouver region has inherent strategic importance as the gateway to transcontinental transportation corridors. In the past, landslides have had important impacts on communities, transportation routes, and on valuable forestry and fishery resources. The objective of this paper is to review the diversity of landslides that occur in the region. The review will provide background for the development of hazard mitigative strategies that increasingly have to be formulated within the context of development pressures and decreasing land availability in one of Canada's most rapidly developing regions.

LANDSLIDE TYPES

A lengthy discussion of landslide types and terminology is outside the scope of this paper. A number of landslide classification schemes are in use in the geotechnical/geological literature (e.g., Varnes, 1978; Hutchinson, 1988; Skempton and Hutchinson, 1969) and reflect, to a large extent, the regional experience of the authors. No single classification scheme has found universal application to the large variety of landslides encountered in the mountains of North America; in this paper, for example, we shall encounter especial difficulty in classifying rapid flow type movements which involve a range of materials and vary in size over several orders of magnitude.

For present purposes, landslides in rock are distinguished from those involving surficial materials. In the rock grouping those involving nonvolcanic rocks are differentiated from those involving Quaternary volcanic materials. Landslides in surficial materials include debris flows originating in colluvial and/or glacial materials in steep mountain watersheds, and landslides involving Pleistocene sediments.

An informative brochure is available describing landslide types in British Columbia (British Columbia Ministry of Energy, Mines, and Petroleum Resources, 1993).

ROCK SLOPE MOVEMENTS IN NONVOLCANIC ROCKS

Definitions

Rockfall involves the detachment of rock fragments from rock slopes and their fall and subsequent bouncing rolling, sliding, and stopping (Varnes, 1978; Hutchinson, 1988; Evans and Hungr, 1993). It may also begin by the detachment of a more or less coherent block that then disintegrates during the course of downslope movement. Rockfalls may be transitional to rock avalanches. A rock avalanche involves a very rapid downslope movement of fragments of bedrock that have become shattered and pulverized during travel which typically results from a rockfall or rockslide in mountainous terrain. It is distinguished from a rockslide by the fact that all of the debris leaves the sliding surface and travels down the valley side, generally to the valley floor, and sometimes along the valley. Rock avalanches with volumes above about one million cubic metres generally show excessive mobility and travel long distances; those with volumes below this threshold behave similarly to rockfalls (e.g., Scheidegger, 1973). Rockslides involve the movement of a rockmass on a defined shear surface or shear surfaces; most of the debris remains in contact with the sliding surface but disruption of the moved mass is considerable. Rockslides are transitional to the complex movement mechanisms associated with mountain slope deformation where discrete sliding surfaces may not be formed and the degree of disruption of the rock mass may be minimal.

Rockfall

Rockfall is a common process in the mountainous terrain of the Vancouver region, where its continued occurrence has formed talus slopes, and it is also common on rock slopes adjacent to linear transportation facilities. Although, as described by Evans and Hungr (1993), rockfall hazards have affected land-use development at the base of talus slopes, as in the example of Silverhope (Fig. 1), the main impact of rockfall in the Vancouver region has been on transportation routes (Hungr and Evans, 1988, 1989; Peckover and Kerr, 1977; Piteau and Peckover, 1978; Theodore, 1986; VanDine, 1992) and on the fisheries resource of the Fraser River (Evans, 1986).

Howe Sound

Rockfalls are common along the Squamish Highway (also known as the Sea-to-Sky Highway) and the adjacent B.C. Rail track along the east side of Howe Sound (Nasmith, 1972; Moore, 1983; Hungr and Skermer, 1992; Fig. 1) where they have frequently caused road and rail traffic interruptions for example, a rockfall in October 1990, near Loggers Creek (Fig. 2) involved 10 000 m^3 of debris and blocked the highway for 12 days thus severing the road link between the communities in the Squamish-Pemberton corridor and Vancouver. The cost of the rockfall, involving repairs and the construction of preventative structures, was approximately $7 million (British Columbia Ministry of Energy, Mines, and Petroleum Resources, 1993). In addition, in response to this event, the

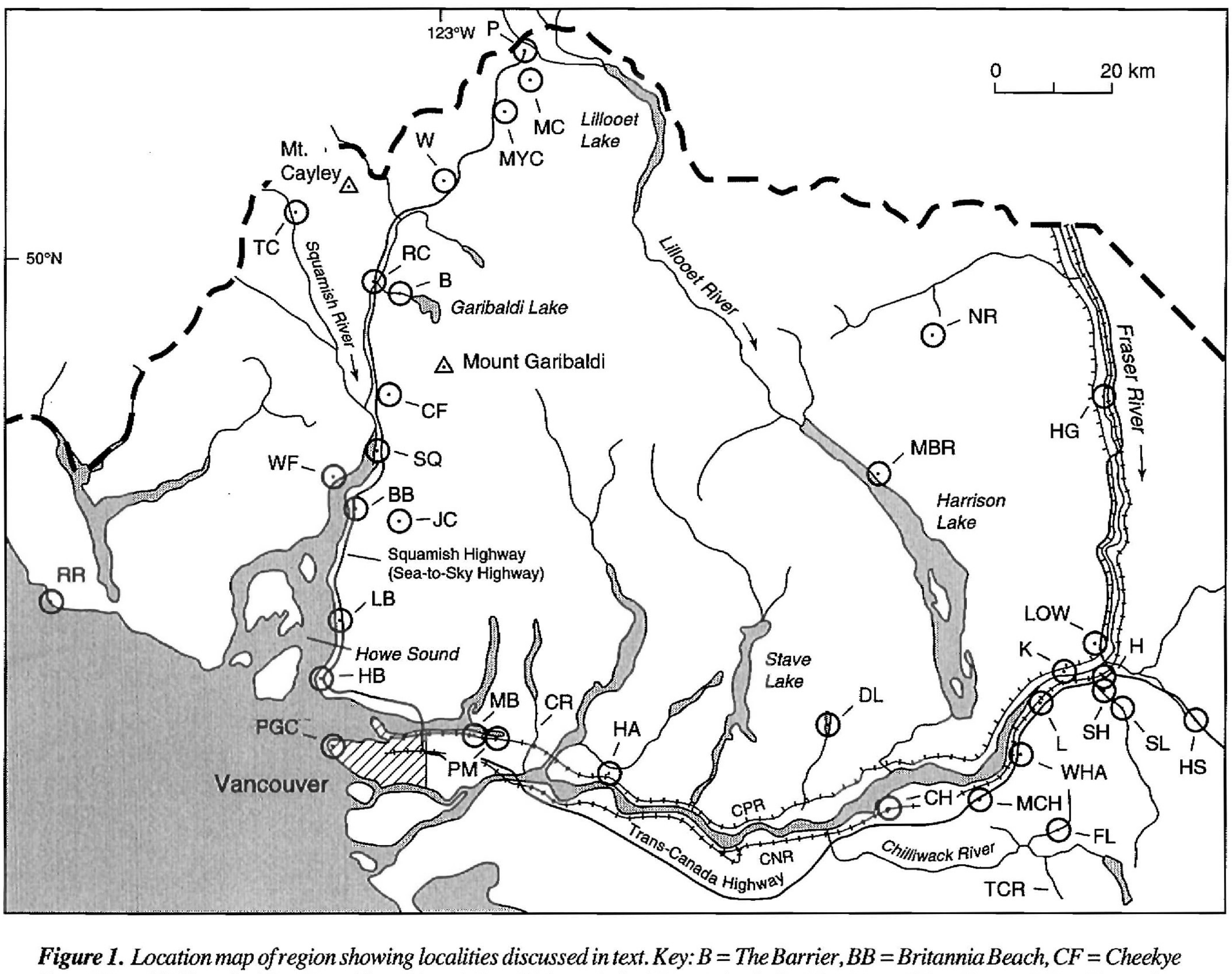

Figure 1. Location map of region showing localities discussed in text. Key: B = The Barrier, BB = Britannia Beach, CF = Cheekye Fan, CH = Chilliwack, CR = Coquitlam River, DL = Dickson Lake, FL = Foley Lake, H = Hope, HA = Haney, HB = Horseshoe Bay, HG = Hell's Gate, HS = Hope Slide, JC = Jane Camp (former site of), K = Katz slide(s), L = Laidlaw, LB = Lions Bay, LOW = Lake of the Woods (also known as Schkam Lake), MB = Mount Burnaby, MBR = Mount Breakenridge, MC = Mount Currie, MCH = Mount Cheam slide(s), MYC = Mystery Creek, NR = Nahatlatch River, P = Pemberton, PGC = Point Grey Cliffs, PM = Port Moody, RC = Rubble Creek, RR = Redroofs escarpment, SH = Silverhope, SL = Silver Lake, SQ = Squamish, TC = Turbid Creek, TCR = Tamihi Creek, W = Whistler, WF = Woodfibre, WHA = Whaleach Power Station; dashed line represents boundary of study area.

provincial government constructed an emergency ferry terminal to be used should the highway be blocked by a similar landslide in the future.

Rockfalls have resulted in the deaths of several motorists on the Squamish Highway since 1969. One such rockfall accident occurred in January 1982 when a single block fell off a rock face above the highway and killed a passenger in a car. The trajectory was analyzed (Fig. 3) by Hungr and Evans (1988) who calculated that the boulder had an impact velocity of 28 $m \cdot s^{-1}$.

Fraser Canyon

Rockfalls are also common in the Fraser Canyon (Peckover and Kerr, 1977; Piteau, 1977) and have impacted upon the Trans-Canada Highway and both intercontinental railway lines, Canadian Pacific Railway (CPR) and Canadian National Railway (CNR) (Fig. 4, 5), causing traffic delays, derailments, and numerous deaths. Since the 1960s, the B.C. Highways Department and both railway companies have spent many millions of dollars to mitigate the rockfall hazard, either by rock slope treatment (e.g., scaling, rock bolts, shotcrete, buttresses, and drainage) or the construction of protective measures such as sheds, catch fences, meshes, ditches, and warning devices (Peckover and Kerr, 1977; Wyllie, 1991).

Data provided in Theodore (1986) shows that the Mountain Region of Canadian National Railways spent almost $2 000 000 per year between 1971 and 1986 on rock slope stabilization work, mostly between Hope and Kamloops. Since the inception of a major rockfall protection program in 1971 the number of derailments has been substantially reduced despite a significant increase in traffic densities.

Figure 2. *Aerial photograph of rockfall which covered the Sea-to-Sky Highway and B.C. Rail tracks, 4 km north of Lions Bay, October 1990 (Photo courtesy of Selkirk Remote Sensing, Richmond, British Columbia; negative number SRS 4466-5).*

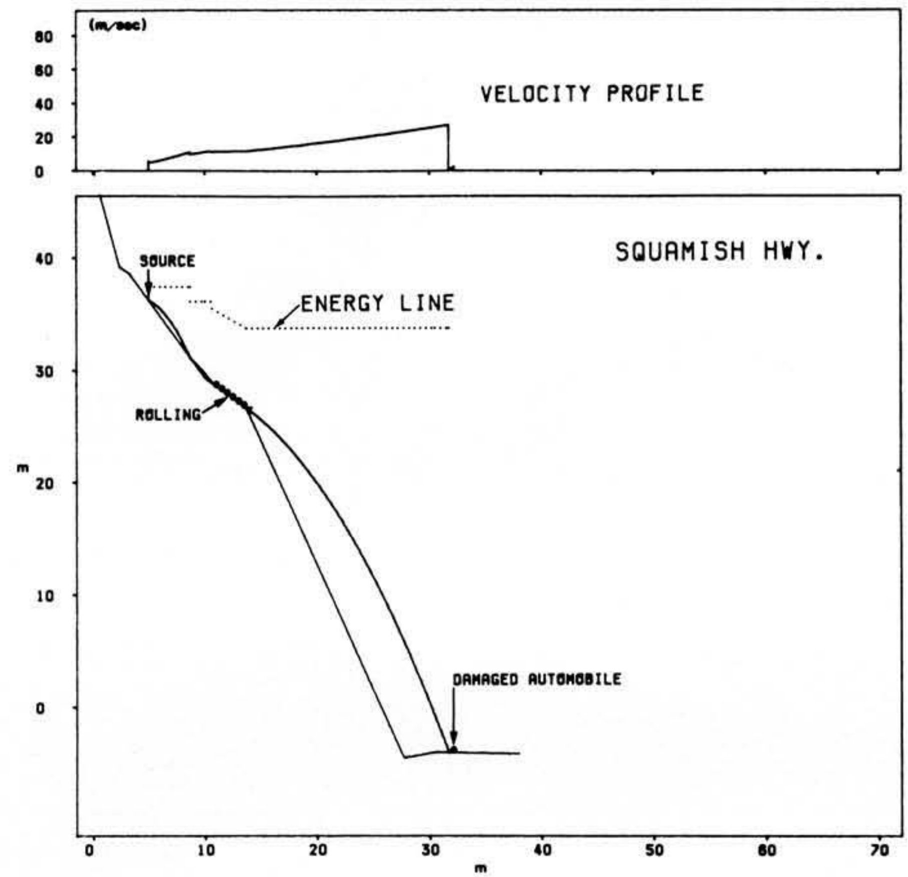

Figure 3. Simulated trajectory of a single boulder which caused the 1982 rockfall accident on Sea-to-Sky Highway, 22 km north of Horseshoe Bay (from Hungr and Evans, 1988).

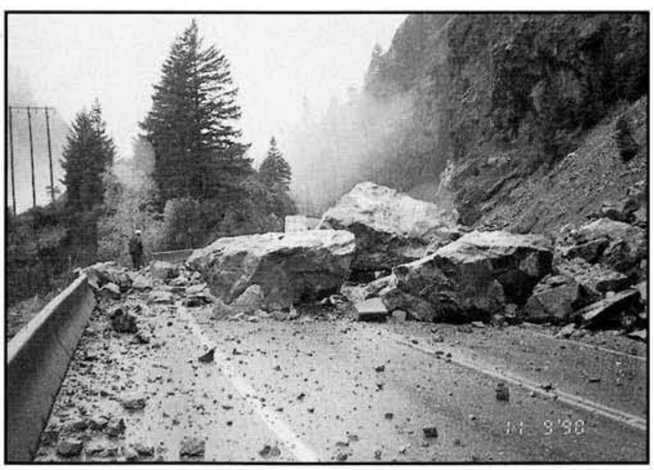

Figure 4. Rockfall on Trans-Canada Highway, Fraser Canyon, near Yale (courtesy of D. Wyllie).

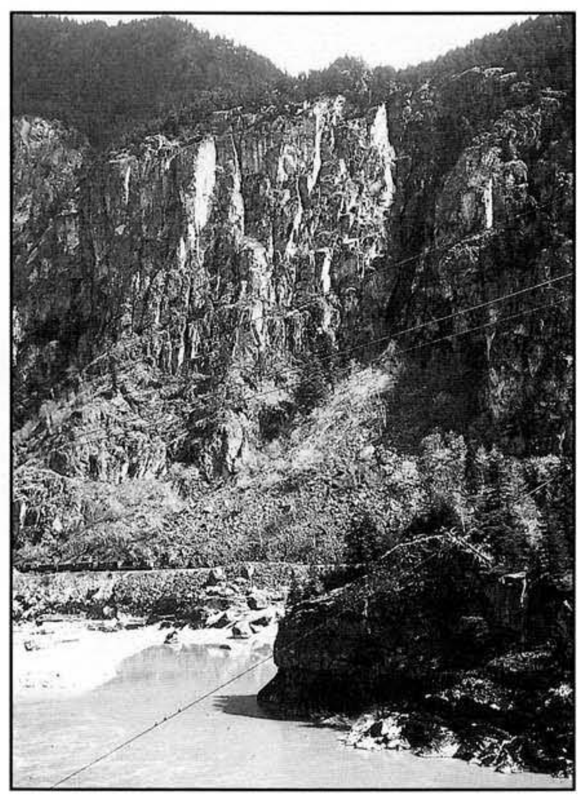

Figure 5. Typical steep jointed rockslope in the Fraser Canyon, which is subject to rockfall. Canadian National Railway tracks run acoss the base of the slope which is located just south of Siwash Creek. GSC 1994-709A

Rockfall frequency shows a marked correlation with the weather (Peckover and Kerr, 1977; Fig. 6). Figure 6 shows data for rockfall frequency on Canadian National Railway tracks in the Fraser Canyon. The highest frequency is found in January, February, and March, the time when freeze-thaw cycles are most frequent, thus reflecting the role of frost wedging in rockfall release. When the mean monthly temperature is above zero (March-November), the number of rockfalls per month varies more or less directly in proportion to mean monthly precipitation (Peckover and Kerr, 1977).

A rockfall (estimated volume 75 000 m^3) at Hell's Gate in the Fraser Canyon (Fig. 1), caused by railway construction activity in 1914, had a major impact on the Fraser River salmon fishery from which it has not yet recovered (International Pacific Salmon Fisheries Commission, 1980; Evans, 1986). The debris greatly increased an obstruction that had been initiated by previous railway construction activity in 1913 and prevented many migrating salmon from returning to their spawning grounds in the vast Fraser watershed, covering about a quarter of the province of British Columbia, above Hell's Gate. The rockfall has had a major impact upon salmon returns in subsequent years; in 1978 dollars and landed values, the loss to sockeye fishery alone, resulting from a diminished return in 1914, amounted to $1.7 billion between 1951 and 1978 (International Pacific Salmon Fisheries Commission, 1980). Fishways were constructed between 1944 and 1966 (Fig. 7), at a cost of $1.36 million, to provide passage for the salmon past the obstruction in an attempt to restore the sockeye and pink salmon runs to their historical abundance.

Rockfall is not confined to steep mountain slopes in the Vancouver region. It may also occur on steep, rocky, coastal cliffs such as along the Stanley Park seawall from Prospect Point to Siwash Rock (Armstrong, 1984).

Small rock avalanches (less than one million cubic metres)

Rock avalanches of less than one million cubic metres are common in the Vancouver region (Fig. 8). In 1915, a rock avalanche involving about 100 000 m^3, killed 56 people at Jane Camp (Fig. 9), a mining community in the Britannia Creek watershed (Evans and Clague, unpublished data). Although this event is Canada's second largest landslide disaster (after the 1903 Frank slide) little is known about the rock avalanche, except that cracks were inspected above Jane Camp several days before the landslide took place but the danger was not realized (Ramsey, 1967).

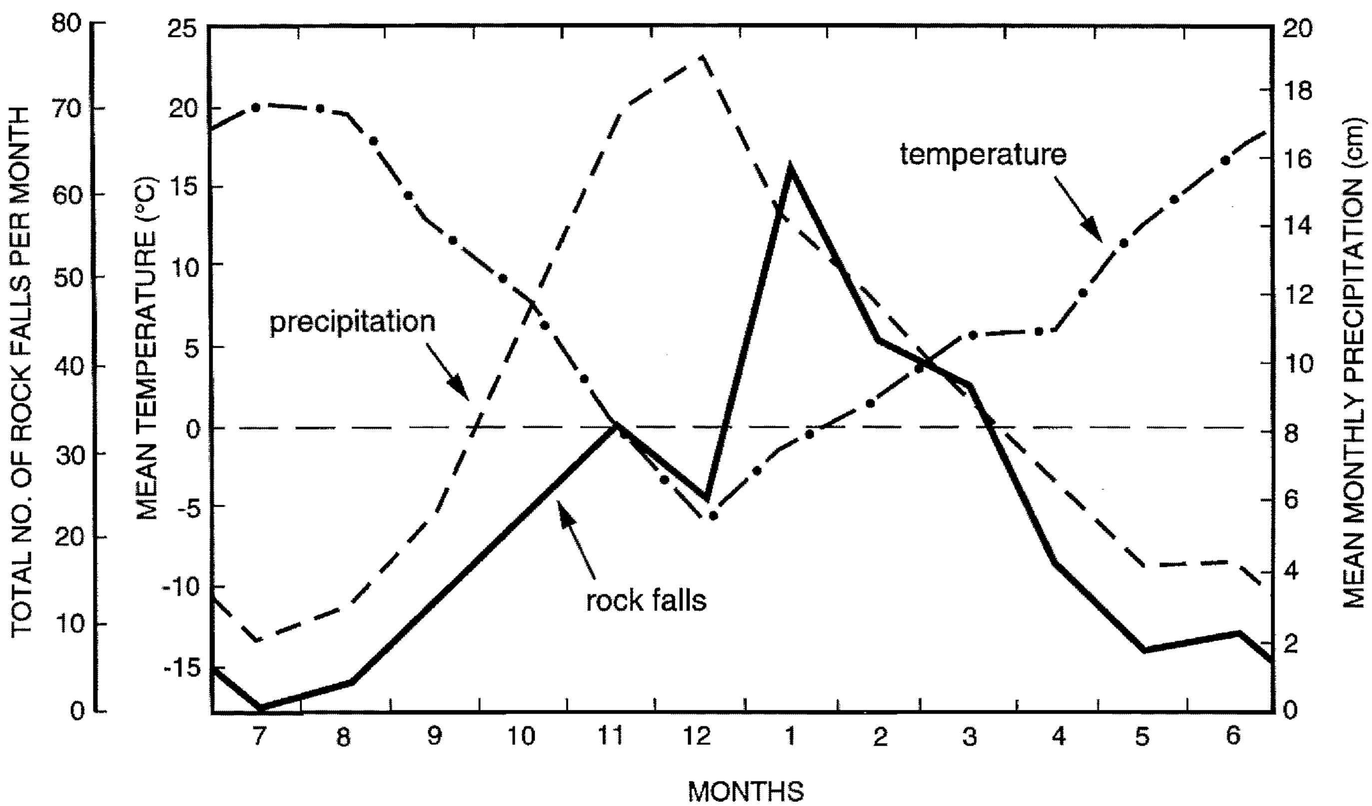

Figure 6. Rockfall frequency and weather over the period 1933-1970, Fraser Canyon, Yale subdivision, Canadian National Railway (modified from Peckover and Kerr, 1977).

Large rock avalanches (greater than one million cubic metres)

Rock avalanches are relatively common in certain geomorphic and geological environments in British Columbia (e.g., Clague and Evans, 1987; Cruden, 1985; Cruden et al., 1989; Eisbacher, 1979; Evans, 1984, 1988, 1989a, b, c, 1990a; Evans and Clague, 1988; Evans and Gardner, 1989; Evans et al., 1989).

A complex set of structural factors, detachment mechanisms, and triggers leads to the occurrence of a rock avalanche. Detachment is favoured on steep rock slopes where planar structural elements, such as joints, bedding planes, and foliation, combine to form a detachment surface that may consist of a single surface (as in rock avalanches involving dipping sedimentary rocks of the eastern part of the Cordillera) or multiple surfaces (which are typical of the rock avalanches found in the plutonic and metamorphic rocks of the Coast and Cascade Mountains) that result in more complex movement mechanisms.

Four rock avalanches, which have occurred in major modern transportation corridors, and that are illustrative of the scale and nature of this landslide type in the Vancouver region, are described below.

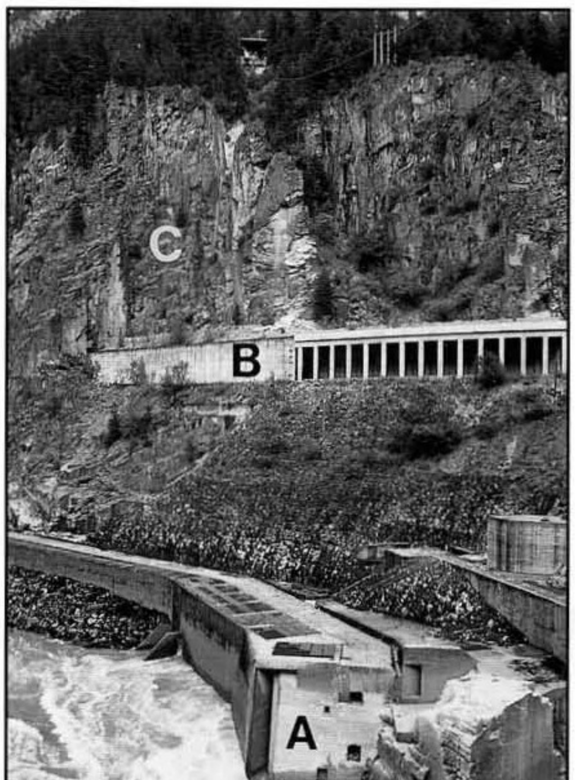

Figure 7. *View of Hell's Gate fishways,* ***A****) looking upstream from the west bank of the Fraser River in 1985. Note Canadian National Railway rockfall protection structure,* ***B****) and steep jointed rock slope of Fraser Canyon wall above* ***C****) source of 1914 rockfall. GSC 1994-709B*

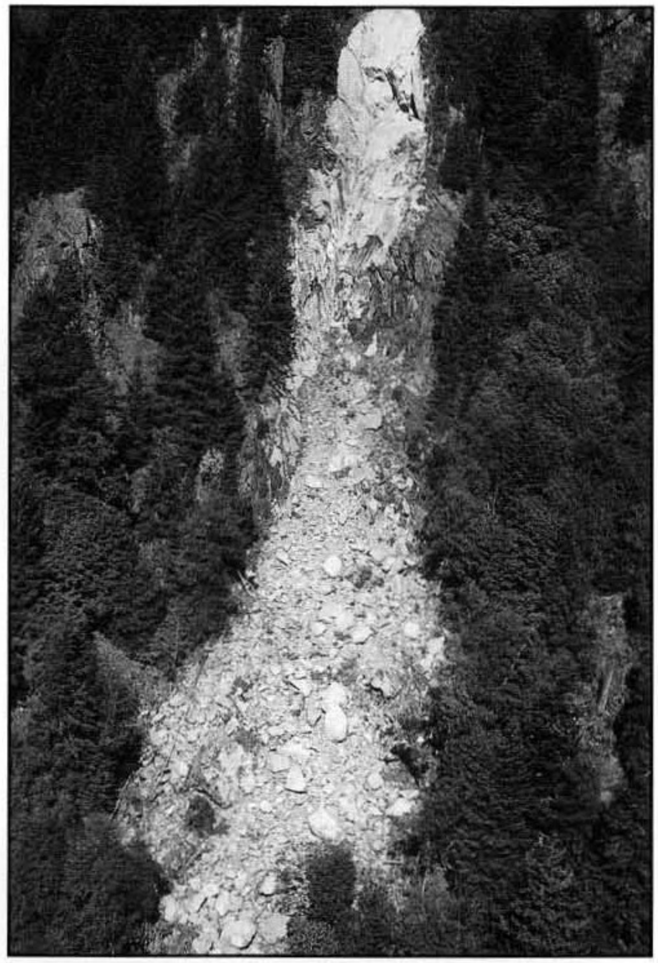

Figure 8. *Typical small rock avalanche (estimated volume 20 000 m^3) in jointed plutonic rock, north side of Fraser Valley, 2 km northwest of Hope. GSC 1994-492G*

Figure 9. *Destroyed buildings and rock debris at Jane Camp, Britannia mine, following the 1915 rock avalanche which resulted in 56 deaths. The Jane Camp tragedy is Canada's second largest landslide disaster, after the 1903 Frank Slide (photo by British Columbia Museum of Mining).*

Hope Slide

The 1965 Hope Slide (Fig. 1, 10) is the largest historical rock avalanche known to have occurred in Canada. It involved 48 000 000 m^3 of rock which descended the southwestern slope of Johnson Peak in two phases separated by about 3 hours. The landslide inundated several kilometres of the Hope-Princeton transportation corridor (Fig. 10), burying three vehicles and claiming four lives.

As noted by G.M. Dawson, the first Geological Survey of Canada geologist in the region (Dawson, 1879), Johnson Peak was the site of a prehistoric rock avalanche. A radiocarbon date obtained from organic material found beneath the debris yielded a radiocarbon age of 9680 $\pm$ 320 BP (GSC-1433) (Mathews and McTaggart, 1978), representing a minimum age for the landslide, which was of comparable size to the 1965 events. The 1965 events deepened the prehistoric slide surface and extended the headscarp almost to the top of the ridge. The rock avalanche developed mainly in greenstone belonging to the Hozameen Complex of Permian to Jurassic age. Felsite intrusions occur as sheets within the greenstone that dip toward the valley; some of these made up part of the sliding surface (Mathews and McTaggart, 1978; Von Sacken, 1991). Contacts between the felsite and the greenstone are sharp and are commonly marked by a weathered clay gouge.

As documented by Von Sacken (1991), the 1965 failure occurred on multiple discontinuity surfaces that dip toward the valley bottom at variable angles. The failure surface in the lower portion of the slope was controlled largely by felsite sheets (Fig. 11A) whereas steeply dipping intersecting joints controlled the upper portion (Fig. 11B).

The 1965 event was proposed by Mathews and McTaggart (1978) to have been triggered by two small earthquakes recorded at approximately the same time as the landslide in the pre-dawn hours of January 9. The possibility of a seismic trigger was further investigated by Wetmiller and Evans (1989) who re-examined the seismic records associated with the landslide. Their attempt to demonstrate an indisputable seismic trigger for the Hope slide was inconclusive. A re-analysis of the 1965 Hope Slide by Weichert et al. (1990, in press), in the light of the seismic signatures generated by the collapse of an open pit mine slope in the interior of British Columbia in 1990, however, suggests that the "earthquakes" associated with the landslide may have been generated by two phases of the landslide itself. Von Sacken (1991) and Von Sacken et al. (1992) found field evidence to support the two slide scenario. It is thought that the debuttressing of the upper slope by the first slide event, as a result of failure along felsite sheets in the lower part of the slope, led to the second event a few hours later. This scenario eliminates an obvious trigger for the Hope slide.

Analysis has shown that the prefailure slope was in a stage of limiting equilibrium before the 1965 slide (Bruce and Cruden, 1977; Wetmiller and Evans, 1989; Von Sacken, 1991) yet it had withstood substantial seismic accelerations in the past including the M = 7.4 North Cascades earthquake of 1872 (Wetmiller and Evans, 1989). It is not clear how the slope that failed in 1965 withstood such forces since, according to slope stability analysis (Wetmiller and Evans, 1989; Von Sacken, 1991), modest seismic forces should have been high enough to result in detachment.

Katz slides

The Katz slide is located on the north side of the Fraser Valley between Hope and Chilliwack (Fig. 1). Linear facilities occupying this section of the Fraser Valley transportation corridor include Canadian National and Canadian Pacific rail lines, the Trans-Canada Highway, trunk gas and oil pipelines owned by Trans Mountain Pipe Line Company Ltd. and Westcoast Energy Inc., respectively, the B.C. Tel fibre optics telecommunications line and several B.C. Hydro power grids; severance of these lines by a modern Katz slide would have a major impact on Vancouver and the Lower Mainland. As indicated on Figure 12, the corridor has been partially inundated by two prehistoric rock avalanche events. These have not been studied in detail. The following description is taken from preliminary accounts by Naumann (1990) and Savigny and Clague (1992).

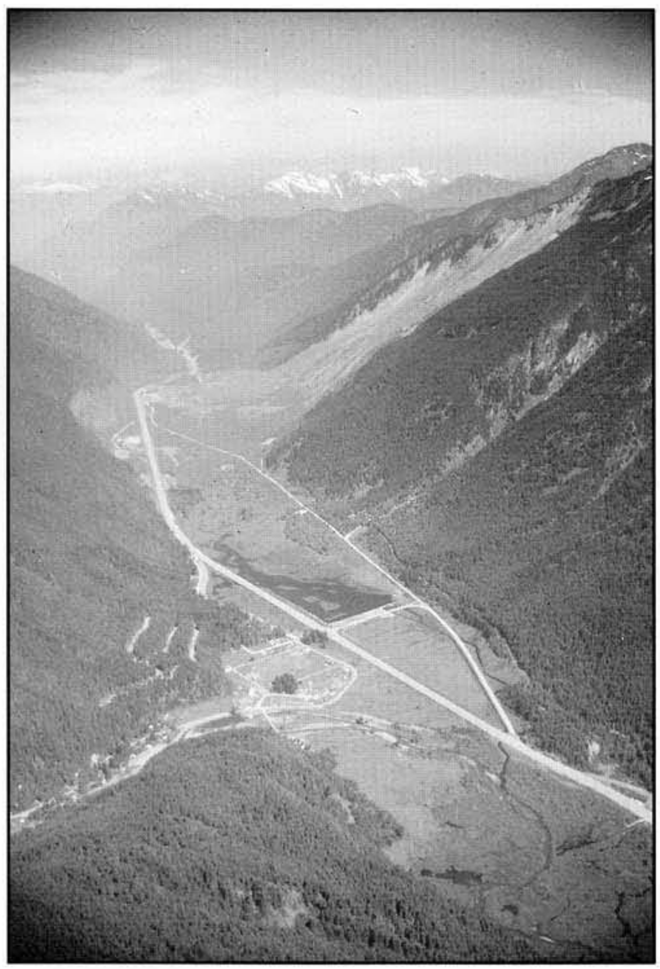

Figure 10. *Oblique aerial photo of 1965 Hope Slide; view to northwest. (Photo by K.W. Savigny).*

The exposed slide debris consist mainly of quartz diorite belonging to the Spuzzum Pluton. A fault, possibly an extension of the northeast-trending Vedder Fault, passes through the headscarp area. A large graben, approximately 120 m wide and at least 35 to 45 m deep, has formed along the fault trend and appears to result from slope distress on the southeast-facing flank of an unnamed mountain overlooking the Fraser River valley. The failure surface appears to be an exfoliation plane.

Katz slide is believed to have occurred as at least two rock avalanches separated by a period of hundreds to thousands of years (Savigny, in press). The first extended across the Fraser Lowland (Fig. 12 and 13) probably blocking the Fraser River, forming a landslide dam and creating a small lake. No volume estimate has been made of the debris because a fan delta quickly prograded through the lake covering all but the largest blocks of the debris. At the time of the second rock avalanche, broad, shallow channels carried flow of the Fraser River along the northwest (right) and southeast (left) sides of the valley (Fig. 12). The second event extended about half way across the valley blocking the northwest channels and diverting all flow to the southeast side (Fig. 13). Debris from the second event covers an area of 1.1 km^2 and has a volume of 15 000 000 m^3. An organic sample from one of the abandoned northwest channel-fill sequences yielded a maximum radiocarbon age for the second rock avalanche of 3260 ± 70 BP (SFU W-02).

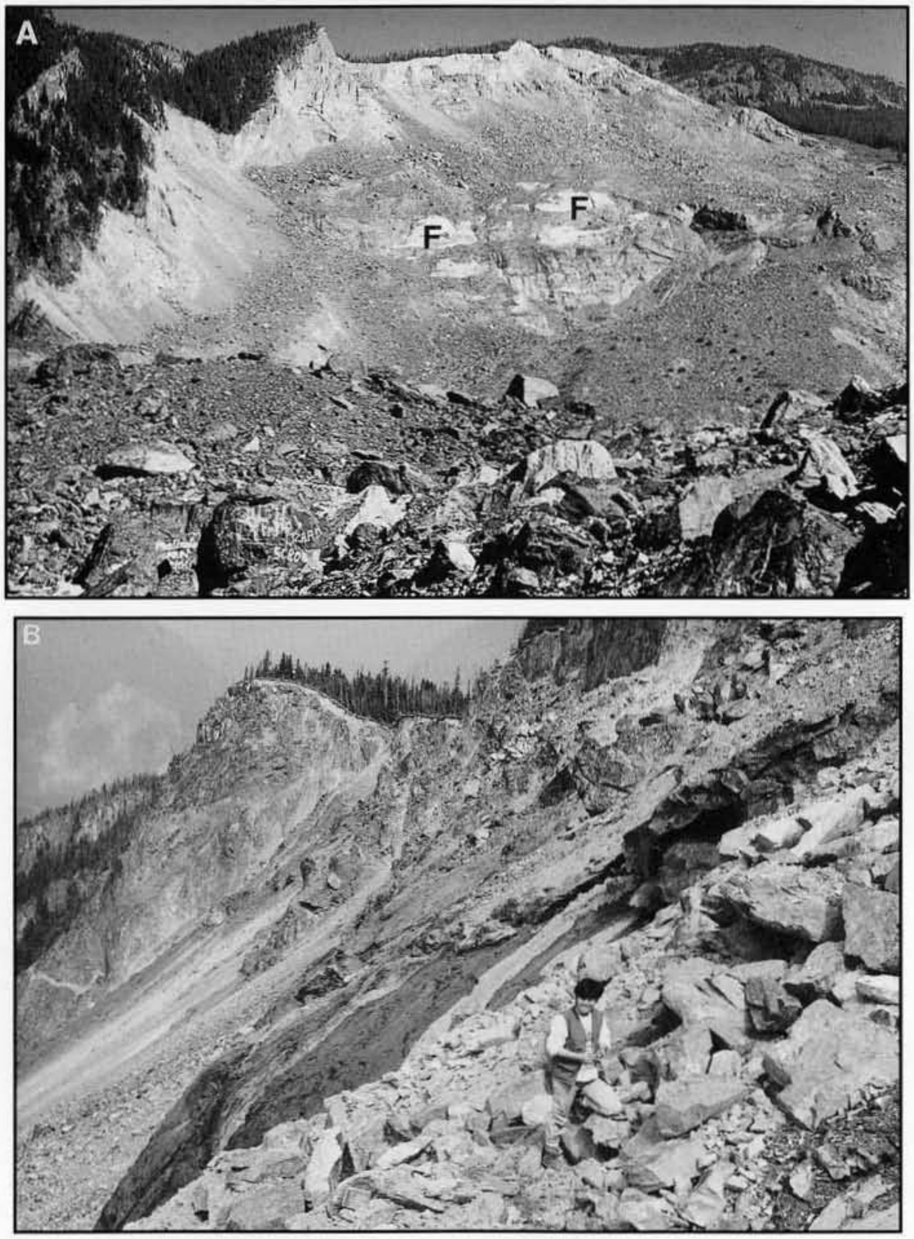

Figure 11. ***A****) 1965 Hope Slide sliding surface and debris; felsite sheets (F) in lower part of the slope. GSC 1994-738* ***B****) Sliding surface in upper part of the slope. Note steep irregular surfaces in greenstone. GSC 1992-114B*

Cheam slides

Cheam slide is a prehistoric rock avalanche complex located on the southeast side of the Fraser River valley 20 km east of Chilliwack (Fig. 1 and 14). The landslide debris forms a hummocky surface which contrasts sharply with the surrounding flat Fraser Lowlands and the adjacent steep slopes of the North Cascade Mountains. The landslide was first examined by Smith (1971); the debris was mapped by Armstrong (1980) as a slope deposit. Naumann (1990) undertook a detailed assessment of both the source and deposition areas and Naumann and Savigny (1992) reported a detailed numerical analysis.

Figure 12. Oblique aerial view of Katz rock avalanche(s) looking southwest in the downstream direction of the Fraser River (F). The source area of the rock avalanche(s) is indicated with a vertical arrow. The debris in the Fraser Valley (D) blocked a channel of the Fraser River (C) and results in a constriction of the present Fraser River channel at point "N". (Photo by K.W. Savigny)

The rock avalanche is believed to have begun as a large asymmetric wedge on an unnamed mountain immediately southwest of Mt. Cheam; a northeast-dipping thrust fault and a steeper, southeast-trending, southwest-dipping joint set may constitute the two slide surfaces in Devonian to Permian Chilliwack Group rocks which consist of volcanic arenites, argillites, and cherty or argillaceous limestones (Naumann, 1990; Naumann and Savigny, 1992). The intersection of these surfaces outcrops in the slope approximately 400 m above the Fraser Lowlands. The source area volume is estimated to be 150 000 000 m^3 (Naumann, 1990). The volume of debris is barely one-third of this, a discrepancy which remains unexplained. The debris shows evidence of multiple events, possibly involving failure onto Late Wisconsinan ice. The debris contains fragments of trees; radiocarbon ages from these wood fragments range between 4350 ± 70 BP (SFU-W-04) and 5010 ± 70 BP (GSC-4004, collected by J.J. Clague) indicating a mid-Holocene age for the events.

Mystery Creek

The Mystery Creek rock avalanche (Fig. 1, 15; estimated volume 40 000 000 m^3) is located 20 km north of Whistler and involved the failure of a portion of the east side of the Green River valley. The debris covers an area of 1.2 km^2 in the bottom of the Green River valley, and is traversed by a B.C. Hydro main transmission line and B.C. Highway 99. The landslide was first reported by Eisbacher (1983), and described briefly by Clague et al. (1987) and Evans (1992a). The rock avalanche involved the failure of a mountain slope consisting of foliated, hard intrusive rock of the Pemberton Dioritic Complex. Slopes adjacent to the scar of the rock avalanche show indications of mountain slope deformation (Fig. 15).

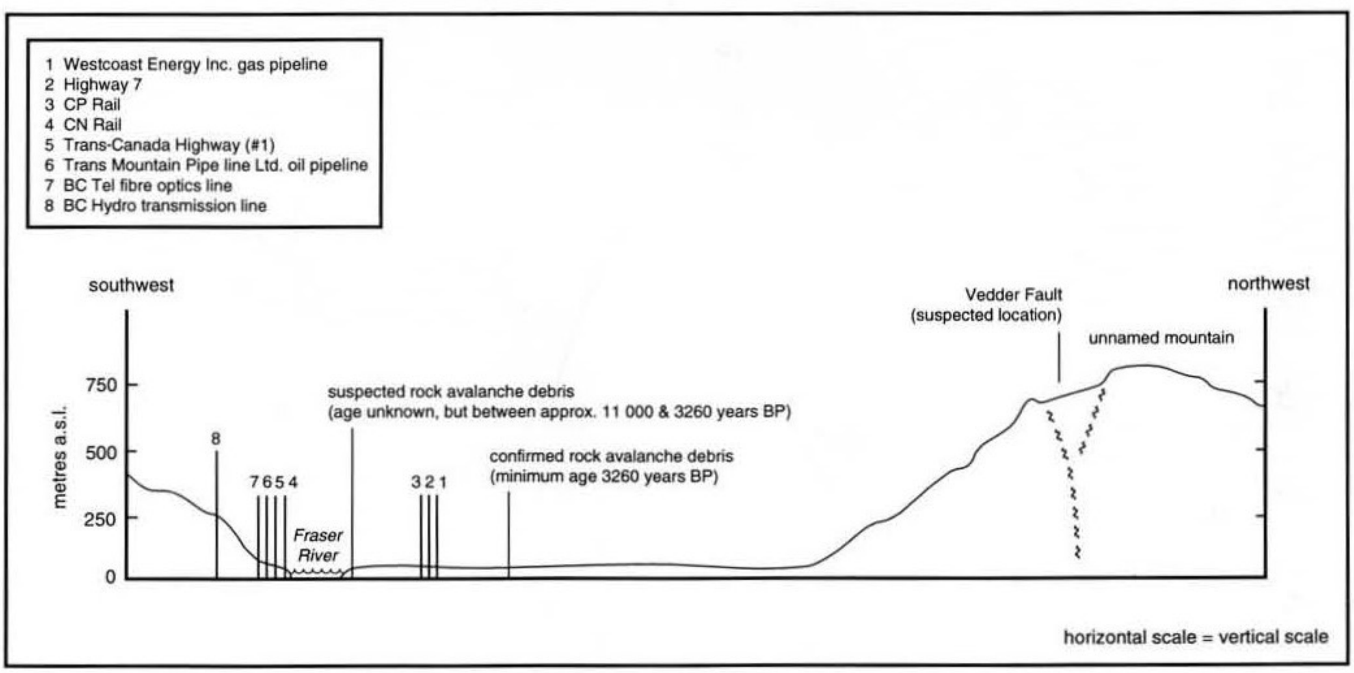

Figure 13. Profile of Katz rock avalanches in relation to linear infrasructural elements and the Fraser River.

A radiocarbon date from charcoal dug out from beneath a large boulder in the debris yielded a radiocarbon age of 880 ± 100 BP (GSC-4237) and is thought to represent a minimum age for the landslide.

Detachment on a low angle (18°) joint surface dipping out of the slope appears to have been preceded by toppling toward the Green River valley involving flexural slip on steep foliation surfaces dipping into the slope (Evans, 1992a). Antislope scarps formed by toppling are present in displaced rock masses along the southern margin of the scar (Fig. 16). The characterization of the process by which this type of slope deformation terminates in catastrophic detachment remains a current research problem in géotechnique.

Other features related to rock avalanches

Some rock avalanches, which have occurred in similar terrain in other parts of the Coast Mountains, exhibit dramatic mobility (e.g., spectacular run-ups, marked changes in direction, and superelevation of debris in curves) and travel long

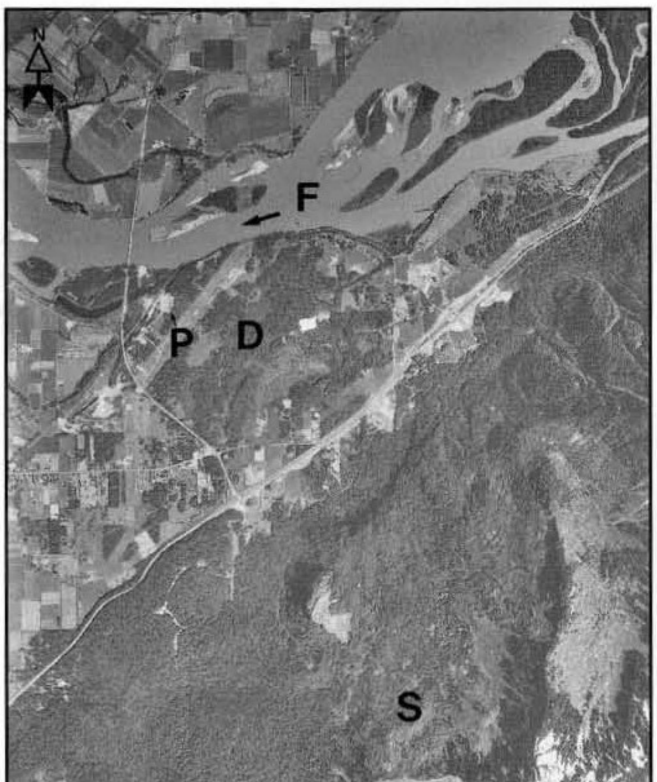

Figure 14. Airphoto of Cheam slide(s), Fraser Valley. Both source area (S) and extent of debris (D) have yet to be defined in detail. The debris is crossed both by the Canadian National Railway and the Trans-Canada Highway. Radiocarbon dates quoted in text were obtained from wood in debris in pit marked P. Fraser River (F) and flow direction are indicated. NAPL A27109-40

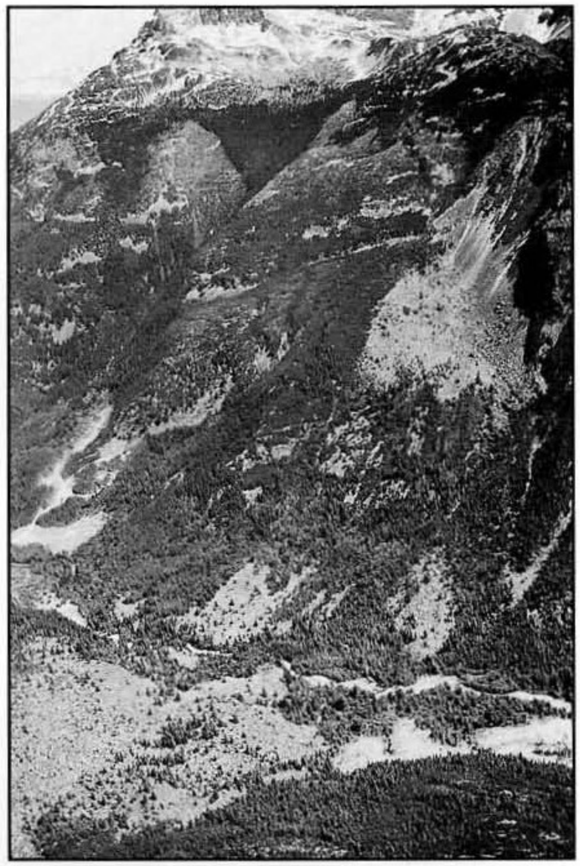

Figure 15. Oblique aerial view to the east of the prehistoric Mystery Creek rock avalanche, 20 km northeast of Whistler. Toppling shown in Figure 16 occurs on right hand (southern) margin of scar. GSC 204107G

Figure 16. Antislope scarp formed by toppling on southern margin of Mystery Creek rock avalanche. Downslope is to the left. GSC 1994-709E

distances from their source, an example being the Pandemonium Creek event (Evans et al., 1989) which occurred in 1959, 360 km northwest of Vancouver, in Tweedsmuir Provincial Park. The debris travelled up to 9 km from its source over a vertical distance of 2 km; an analysis of the event indicated that the velocity of the debris may have reached 100 $m \cdot s^{-1}$ (Evans et al., 1989). Although this type of highly mobile rock avalanche has not been documented so far in the nonvolcanic rocks of the Vancouver region, in view of the similarity in geology and terrain, the potential for such an occurrence is thought to exist.

Rock avalanches may produce important secondary effects, including the damming of rivers and streams to form landslide dammed lakes and landslide-generated waves.

Figure 17. *Landslide dammed lake in the Nahatlatch River watershed. The rock avalanche occurred in plutonic rocks. GSC 1992-0815*

There are several lakes in the Vancouver region which are dammed by rock avalanche debris (Fig. 17; Evans, 1986; Clague and Shilts, 1993; Clague and Evans, 1994b). They include Dickson Lake, Lake of the Woods (also known as Schkam Lake), Silver Lake, and Foley Lake (Fig. 1).

Landslide-generated waves, which extend the zone of potential damage well beyond the limits of the rock avalanche debris, have not been documented in the Vancouver region but have been reported from nearby Vancouver Island (Evans, 1989b).

Rockslides and mountain slope deformation

Numerous rockslides involving noncatastrophic movement of rock masses on defined shear surfaces, and in which the debris largely remains on the sliding surface, have been mapped in the Vancouver region (e.g., Armstrong, 1984; Savigny, in press). On the north slope of Mount Burnaby for example, (Fig. 1, 18) small rockslides have developed in southerly dipping Tertiary sediments (Armstrong, 1984).

The deformation of steep mountain slopes is a common slope movement process in the Vancouver region. It is manifested in topographic features such as cracks, fissures, trenches, antislope scarps at mid- or upper slope locations, collectively known as linears, and, in some cases, bulging at lower slope locations. Frequently, these features occur without well defined headscarps, or lateral scarp or lateral shear zones suggesting that slope movement is occurring without well defined shear surfaces having been formed, unlike rockslides described above. These characteristics often lead to problems in the identification and interpretation of mountain slope deformation as the following examples illustrate.

Figure 18. *Airphoto taken in 1959 of rockslides on north slope of Mount Burnaby. The scarps of the rockslides are indicated with arrows. NAPL A16830; 121*

Wahleach

The best documented example of mountain slope deformation in the Vancouver region is that at B.C. Hydro's Wahleach Power Station in the Fraser Valley (Moore et al., 1992; Savigny and Rinne, 1991). The Wahleach power station generates electricity by water flows from a reservoir at Wahleach Lake through a complex conduit system (Fig. 19) which consists of a 3 m diameter, 3500 m long upper tunnel, a 600 m shaft inclined at 48°, a 300 m lower tunnel, and a 485 m surface penstock to the power house located adjacent to the Trans Canada Highway. A total of 620 m of head is developed. In 1989 the steel lining of the upper tunnel was ruptured by slope movement (Fig. 19) and water was released into the slope.

The slope at Wahleach consists of hard, strong granodiorite cut by minor dykes and has total relief of approximately 920 m and an average slope of 25° (Fig. 19). The rock mass is characterized by closely spaced fractures and shear zones. There is a gradual increase in rock quality with depth (Moore et al., 1992). Throughgoing discontinuities with downslope dips of less than 45° are absent. According to Moore et al. (1992), the Wahleach slope has undergone prehistoric movement down to average depths of 150-200 m as indicated by loosened rock down to this elevation, and the linear troughs and scarps in the upper part of the slope (For location, see D in Fig. 34); the current movement involves rock to depths of 60-120 m (Fig. 19). Between 1990 and 1992 the movements have resulted in surface displacements of 4-40 mm•a^{-1} and, on the basis of their distribution and timing are concluded by Moore et al. (1992) to be consistent with gravitational creep which may also involve elements of sliding and block rotation. Moore et al. (1992, p. 106) concluded that, "lack of throughgoing adversely oriented discontinuities, the long history of diffuse, slow movements and the insensitivity of these movements to groundwater, indicate that the present movements will continue for considerable time and that a large rockslide is not imminent".

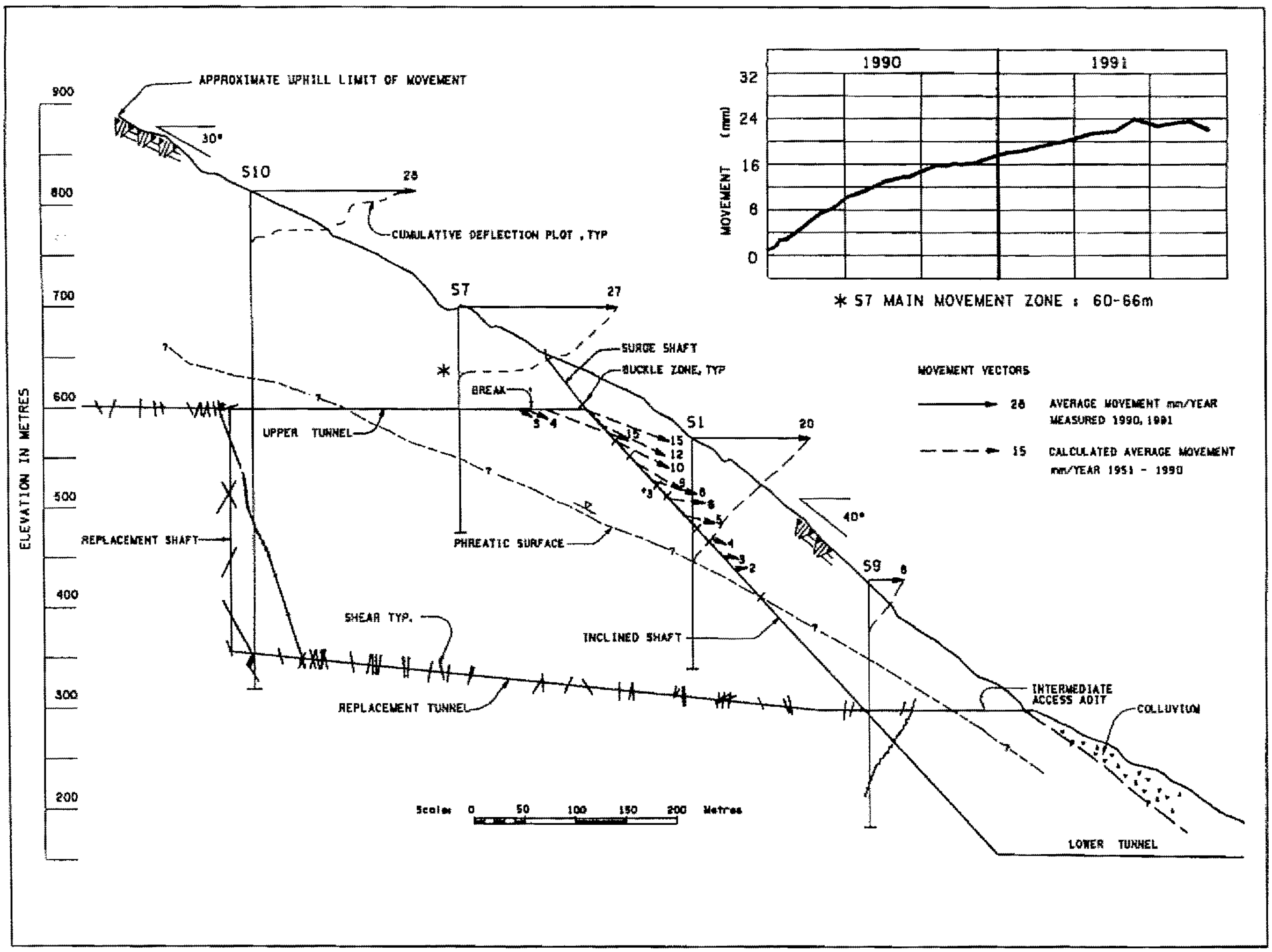

Figure 19. Mountain slope deformation at B.C. Hydro's Wahleach Power station in the Fraser Valley. Section showing movement and geology (after Moore et al., 1992).

Figure 20. *View of mountain slope deformation at Mt. Breakenridge, Harrison Lake. GSC 1994-709F*

Figure 21. *Displaced rock masses resulting from mountain slope deformation, Mt. Breakenridge, Harrison Lake. GSC 1994-709G*

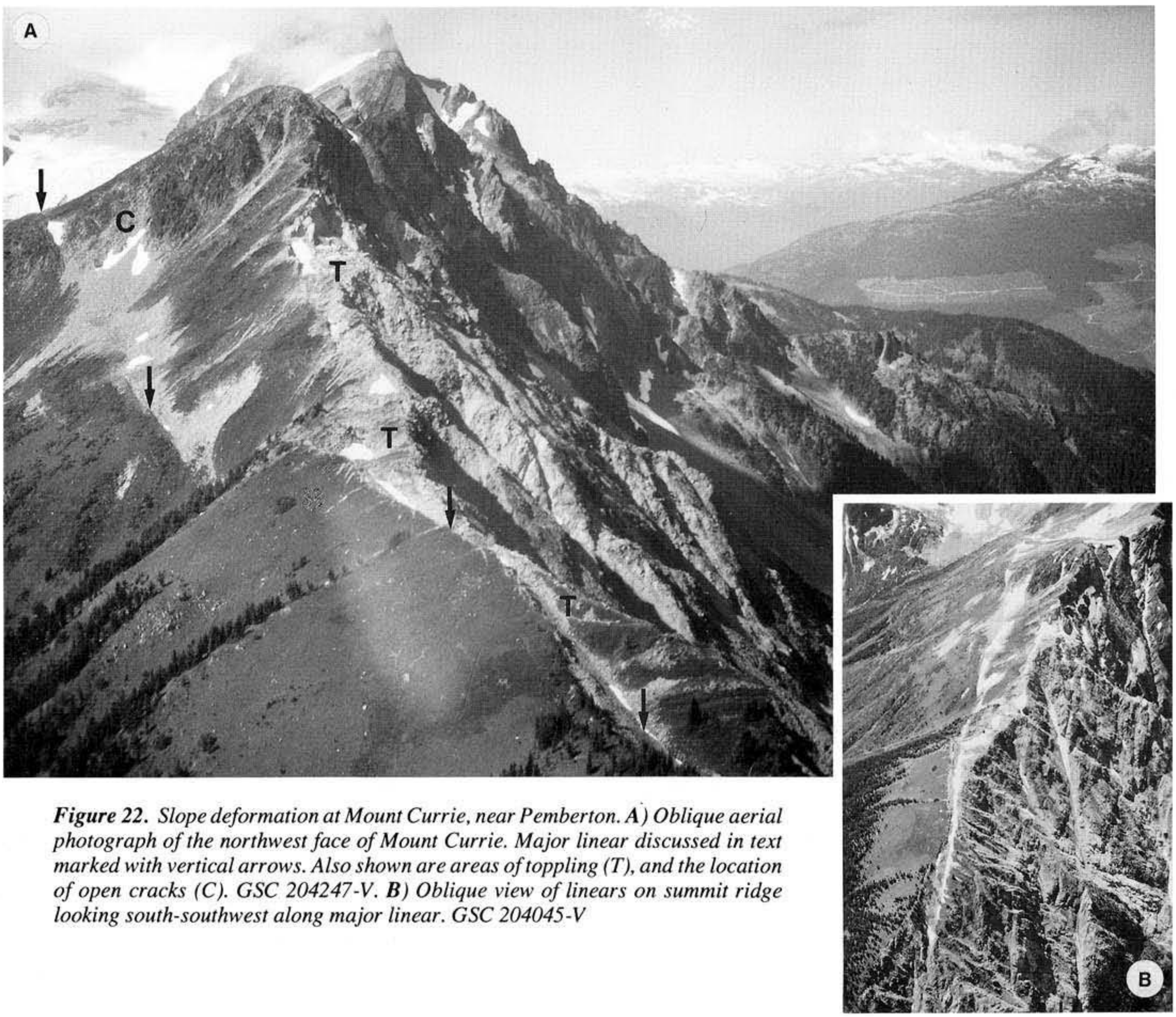

Figure 22. *Slope deformation at Mount Currie, near Pemberton.* ***A****) Oblique aerial photograph of the northwest face of Mount Currie. Major linear discussed in text marked with vertical arrows. Also shown are areas of toppling (T), and the location of open cracks (C). GSC 204247-V.* ***B****) Oblique view of linears on summit ridge looking south-southwest along major linear. GSC 204045-V*

Mount Breakenridge, Harrison Lake

The geotechnical interpretation of mountain slope deformation features, particularly with respect to movement mechanism and catastrophic potential, is sometimes problematical. At Mount Breakenridge, for example, on the eastern shore of Harrison Lake (Fig. 20), a steep mountain slope has undergone considerable deformation (Fig. 21). The slope consists of metamorphic rocks of the Mesozoic Slollicum Schist; it extends from lake level to an elevation of 1445 m and runs beneath Harrison Lake to elevation -200 m, resulting in a total relief of 1645 m. The broken ground on the summit ridge of the slope consists of toppled blocks, scarps, and cracks and extends for 2.5 km along the ridge. The attitude of the foliation in displaced rock masses suggests that slope deformation has resulted from toppling. Concern was raised about the possibility of catastrophic failure at the site which could generate a displacement wave in Harrison Lake (Vancouver Sun, June 30, 1989), which in turn would impact on the tourist resort of Harrison Hot Springs at the southern end of the lake, 48 km from the site, and on the Port Douglas Indian Reserve at the head of the lake, 15 km from the site. Subsequent investigations by provincial authorities indicated that the likelihood of this scenario occurring was low. An acoustic survey of Harrison Lake by Desloges and Gilbert (1991) found no evidence of any disturbances of the lake floor sediments that could have resulted from a previous catastrophic failure of the Mount Breakenridge slope.

Mount Currie

In some parts of the Coast Mountains, mountain linears have previously been interpreted as tectonic features rather than the result of gravitational processes (e.g., Clague and Evans, 1994a). In the Vancouver region, a problematical example of slope deformation is that which exists on the northeast ridge of Mount Currie, overlooking the town of Pemberton (Fig. 1). A 1.75 km long southwest-trending linear, obvious from the air and on aerial photographs (Fig. 22), was first described by Eisbacher (1983) who ascribed a tectonic origin to the feature. A maximum of 20-30 m vertical displacement has taken place along the linear. The northeast ridge is made up of hard gneissic rocks of the Pemberton Dioritic Complex, the structural fabric of which is dominated by a near vertical foliation (Evans, 1987). Rock mass disruption and gaping tension cracks indicate that at least some of the vertical displacement is due to gradual large scale slope movement of the summit ridge. Toppling of the gneissic foliation is also present (Evans, 1987).

Hell's Gate

Slope deformation may also be triggered by large scale construction. During excavation of a 65 m high rock cut at Hell's Gate Bluffs during the construction of the Trans-Canada Highway in the Fraser Canyon, cracks exceeding 1 m wide in places, were discovered in hard granodiorite containing steeply dipping discontinuities. Construction was halted but movement continued. As described by Piteau et al. (1979), the cracks developed along a set of steeply dipping faults and analysis showed that the movements consisted of the outward overturning, or toppling, of fault defined blocks (Kalkani and Piteau, 1976). Movement was found to be a direct function of precipitation. The movement was stabilized by excavation of the head of the slope and the implementation of measures to prevent infiltration.

DEBRIS AVALANCHES IN QUATERNARY VOLCANIC ROCKS, GARIBALDI VOLCANIC BELT

Garibaldi Volcanic Belt

The Garibaldi Volcanic Belt is the northward extension of the Cascade Volcanic Belt. Quaternary volcanic rocks of the Garibaldi Group occur in three major centres, viz. Mt. Garibaldi, Mount Cayley, and Mount Meager (Fig. 23).

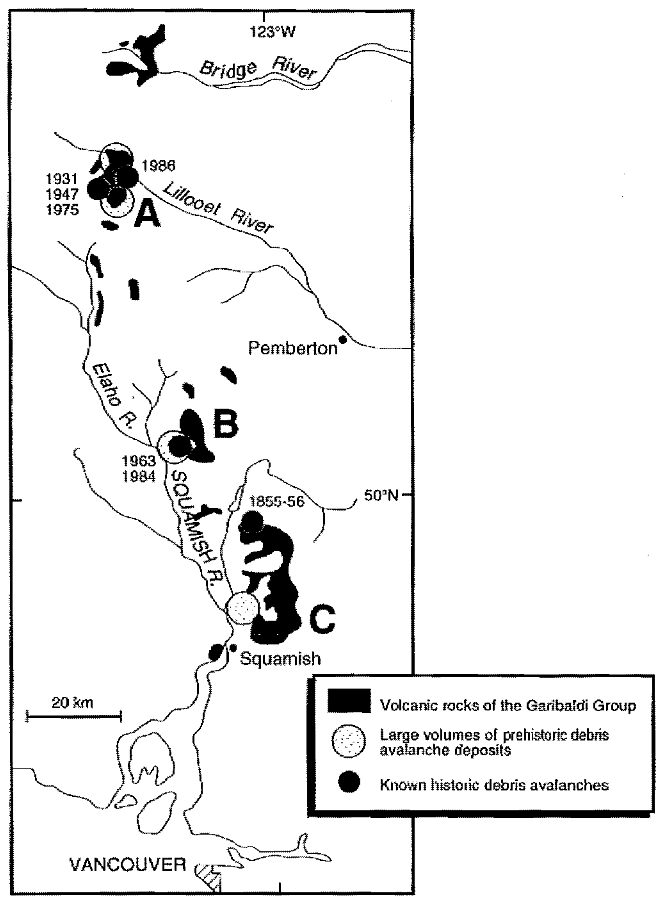

Figure 23. Map of Garibaldi Volcanic Belt, southwestern British Columbia showing main volcanic centres (A = Mount Meager, B = Mount Cayley; C = Mount Garibaldi), location of large volumes of prehistoric debris avalanche deposits, and the location and dates of known historic debris avalanches (modified from Evans and Brooks, 1991).

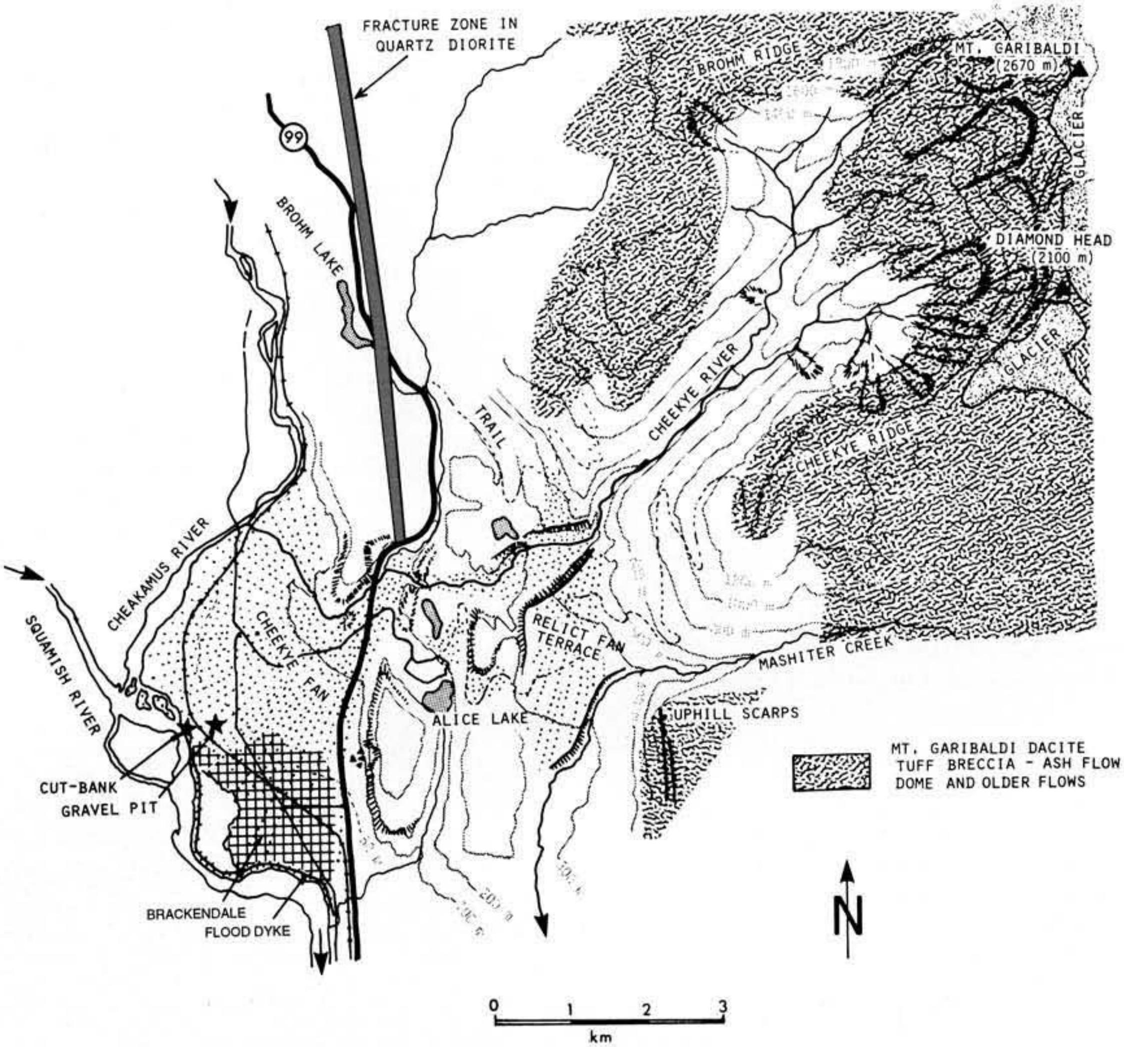

Figure 24. *Sketch map of the western flank of Mount Garibaldi and the Cheekye River basin (after Eisbacher, 1983).*

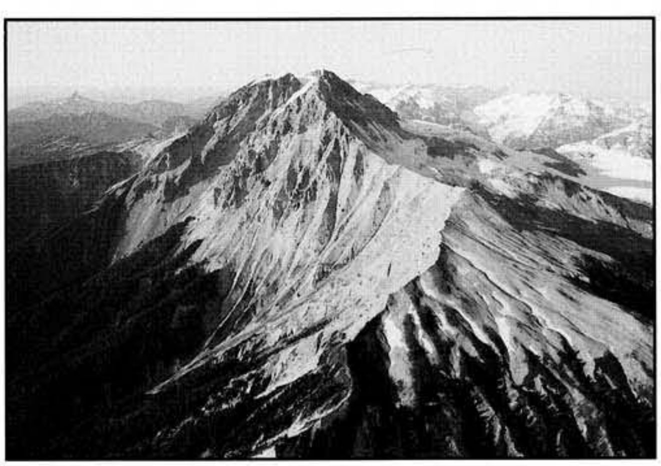

Figure 25.

Oblique aerial view of Mount Garibaldi viewed from the south. The steep western face of the volcano is essentially a scarp formed by the Late Pleistocene flank collapse of the volcano. GSC 1994-709H

The most recent eruption in the Belt was at Plinth Peak, within the Mount Meager complex, at about 2350 BP (Read, 1990; Evans, 1992b) which deposited the so-called Bridge River Ash to the east (Nasmith et al., 1967).

The term "debris avalanche" is used here to describe the transformation of a volcano slope failure into what Schuster and Crandell (1984, p. 567) described as "a sudden and very rapid flowage of an incoherent, unsorted mixture of rock and soil material...Movement of the mass is characterized by flowage regardless of whether it is wet or dry..."

Debris avalanches in the Mount Garibaldi Complex

Large landslides have taken place in two types of settings within the Mount Garibaldi Complex; from the flanks of the volcanoes themselves (e.g., Mount Garibaldi) and from the high precipitous margins of lava flows at some distance from the source vent (e.g., Rubble Creek).

Major debris avalanche deposits have been documented in the Mount Garibaldi-Cheekye River area and Rubble Creek.

Mount Garibaldi-Cheekye River

Debris avalanche deposits were first described in the Mount Garibaldi-Cheekye River area by Mathews (1952a, 1958). They cover a large area of the Squamish valley (Fig. 24) and consist of large dacitic blocks set in a matrix of pulverized tuff/tuff breccia, typical of debris avalanche deposits described elsewhere (e.g., Crandell, 1971; Evans and Brooks, 1991).

Mathews (1952a) has argued that the Mount Garibaldi volcanic cone was partially built over Fraser Glaciation ice, the melting of which during deglaciation removed support from the volcanic edifice resulting in the collapse of its western flank.

The area of the debris (including the Cheekye Fan) is 25 km^2 (Evans, 1990b). Assuming a mean thickness of 100 m this yields a volume of approximately $2.5 \cdot 10^9$ m^3. This is identical to Mathews (1952a) estimate and compares favourably to his estimate of the missing volume from the western flank of Mount Garibaldi ($2.9 \cdot 10^9$ m^3).

The debris avalanche deposits originated in the dacitic lavas and tuff-breccias which make up the western flank of Mount Garibaldi. The amphitheatre-shaped headwater region of the Cheekye River (Fig. 25) is in effect a massive landslide scar created by multiple failure events. Successive failure events may have built up what Mathews (1952a) termed the 'terraced fanglomerates' at the mouth of the Cheekye valley. Unpublished radiocarbon dates obtained by S.G. Evans and Thurber Engineering/Golder Associates (1993) suggest that large landslides continued to occur on the western slopes of Mount Garibaldi and travelled down the Cheekye valley to the Cheekye Fan through prehistoric time. The occurrence of these events has had a major impact on recent land-use decisions regarding the fan (Hungr et al., 1993; Thurber Engineering/Golder Associates, 1993).

Debris flows have continued in historical times. As described by Jones (1959), following heavy rains in August 1958, a debris flow swept down the Cheekye River and formed a 5 m high temporary dam across the Cheakamus River at its mouth. Local residents reported that a similar debris flow occurred in the 1930s (Jones, 1959). Both flows were of the order of 100 000 m^3 (Thurber Engineering/Golder Associates, 1993).

Rubble Creek

The Rubble Creek basin has been the site of at least two large debris avalanches and several debris flows during the Holocene (Mathews, 1952b; Moore and Mathews, 1978; Hardy et al., 1978). The source of the landslides is The Barrier, a precipitous face forming the margin of a dacite lava flow that erupted from Clinker Peak in the late Pleistocene (Fig. 26; Mathews, 1952b). Much of the debris has accumulated in the

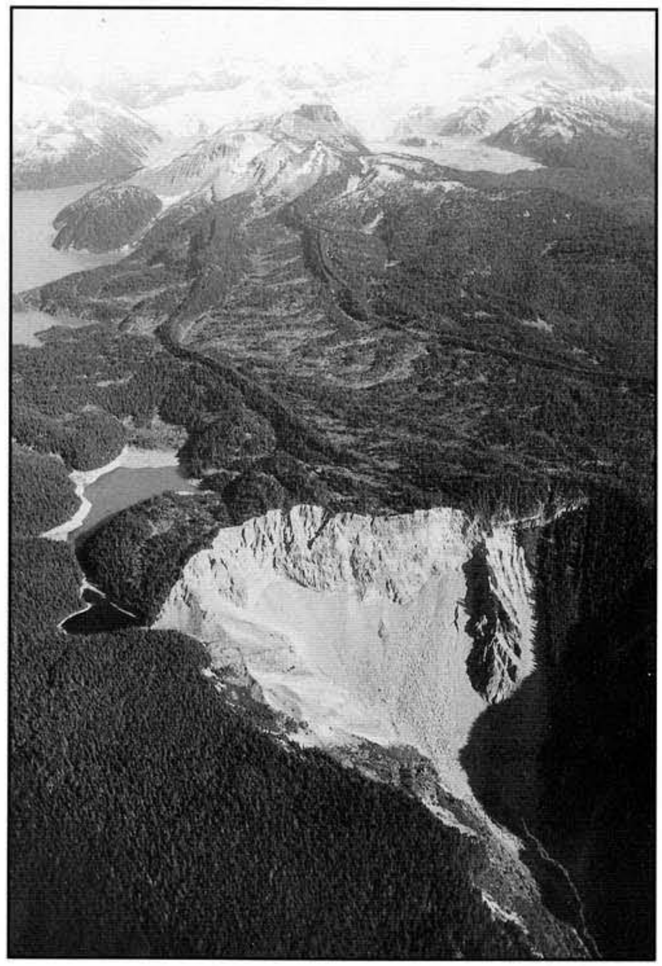

Figure 26. *Aerial view of The Barrier, a steep rock face formed by the successive failure of the margin of the Clinker Peak lava flow, the most recent failure being the 1855-56 rock avalanche. Clinker Peak is visible as the obvious source of the lava flow. Mount Garibaldi is visible in right background. GSC 1991-300*

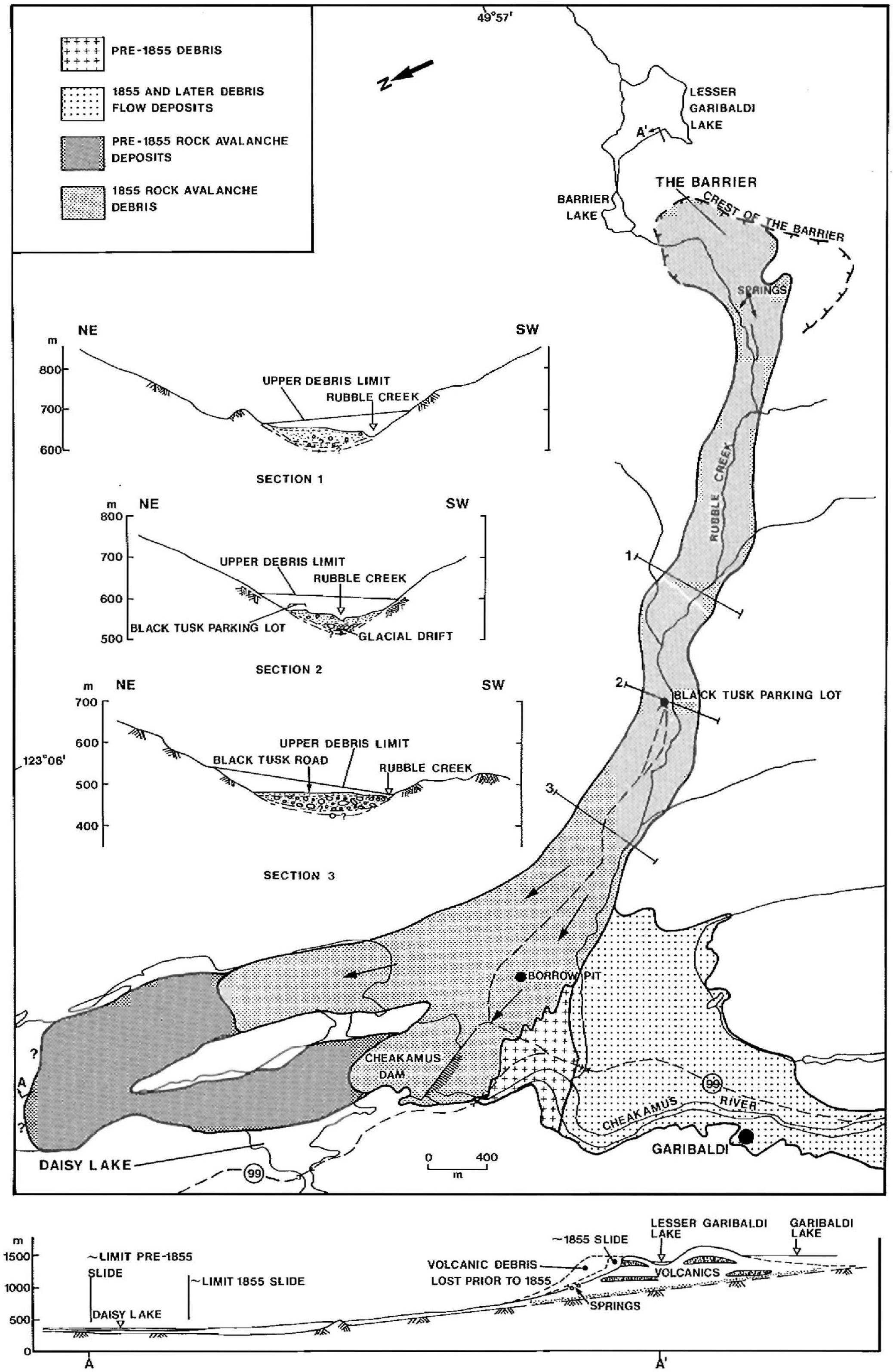

Figure 27. Landslides in Rubble Creek, Mount Garibaldi volcanic complex; map, longitudinal profile, and cross-sections showing upper limit of debris of 1855-56 rock avalanche (reproduced from Clague et al., 1987 which was redrawn from Hardy et al., 1978).

Cheakamus valley in a large fan at the mouth of Rubble Creek. Subsurface investigations indicate that the volume of the fan is between 156-186•10^6 m^3 and contains between 5-10 separate landslide units averaging 5-10 m in thickness (Hardy et al., 1978). A weathered surface exposed near the mouth of Rubble Creek separates historic landslide debris from similar materials which are older than about 600 calendar years (Hardy et al., 1978).

The most recent major failure occurred during the winter of 1855-56, when a major part of The Barrier failed along vertical fractures producing a large debris avalanche (estimated volume 30-36•10^6 m^3; this volume estimate is from Hardy et al. (1978)). Earlier estimates ranged from 15-25•10^6 m^3 (Mathews, 1952b; Moore and Mathews, 1978)). The debris travelled 6 km down Rubble Creek to the Cheakamus valley on an average gradient of 7° (Fig. 27; Moore and Mathews, 1978; Hardy et al., 1978). Based on superelevation data (Fig. 27) the debris reached velocities in excess of 20-25 m•s^{-1} (Moore and Mathews, 1978). A more complex analysis of the movement in Hardy et al. (1978) suggested that velocities may have reached 60 m•s^{-1} in the upper part of the path and that the landslide decelerated down the valley emerging from it onto the fan at about 25-40 m•s^{-1}.

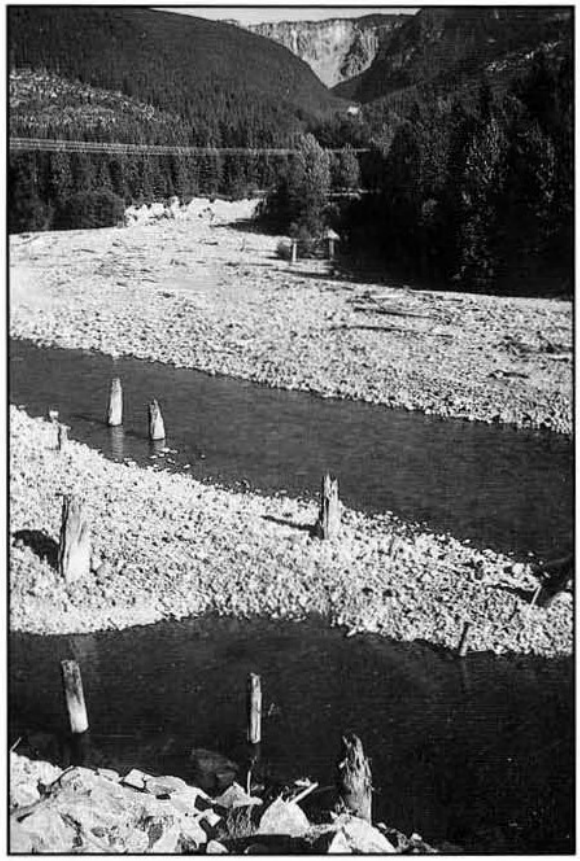

Figure 28. *View across Cheakamus River up Rubble Creek to the Barrier showing trees killed in growing position by the 1855-56 debris avalanche and/or subsequent debris flows. GSC 1994-7091*

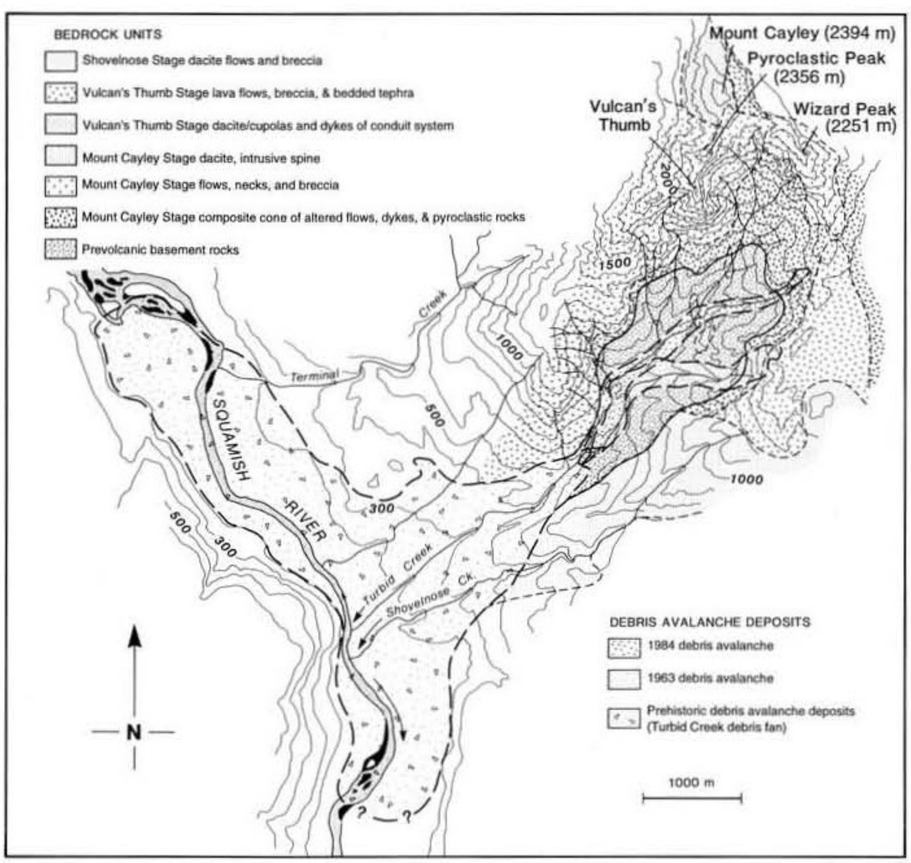

Figure 29.

Map of the Mount Cayley area showing the geology of Turbid Creek basin (after Souther, 1980), limit of prehistoric debris avalanche accumulation and historic debris avalanche paths (after Evans and Brooks, 1991).

The main debris stream spread over the northern half of the Rubble Creek fan and blocked the Cheakamus River (Evans, 1986). Debris flows associated with and following the rock avalanche covered the southern sector of the fan (Hardy et al., 1978). Debris floods, initiated when the Cheakamus River overtopped the landslide dam, buried tracts of forest on the floor of the Cheakamus valley up to 3.5 km below Rubble Creek; numerous rooted stumps of trees killed by these floods are still visible in the channel of the Cheakamus River (Fig. 28).

Between 1955 and 1957, B.C. Hydro constructed an earth and rockfill dam (Cheakamus Dam) across the Cheakamus River less than 1 km north of Rubble Creek (Fig. 27). The southeast abutment is located on the 1855-56 rock avalanche debris. Material obtained from a borrow pit in the 1855-56 debris was incorporated into the core of the dam (Terzaghi, 1960a, b).

A ban on the development of a housing subdivision on the fan was upheld by the B.C. Supreme Court in 1973 (Berger, 1973) because of the risk of another catastrophic landslide from The Barrier and adjacent steep slopes at the margin of the Clinker Peak lava flow. In 1980, Provincial Order in Council 1185 under the Emergency Program Act, designated the Rubble Creek area too hazardous for human habitation. Property owners in the area were bought out, or relocated, at a cost of $17.4 million (Morgan, 1992).

Debris avalanches from Mount Cayley Volcano and the damming of the Squamish River

Investigation of diamicton units exposed in an extensive accumulation of volcanic debris in the Squamish valley, west of Mount Cayley volcano (Fig. 29), has yielded evidence for the occurrence of at least three major debris avalanches, initiated by the collapse of its western flank in the mid-Holocene (Evans and Brooks, 1991, 1992; Brooks, 1992).

Radiocarbon dates obtained from tree fragments (Fig. 30) contained in the deposits indicate that the events took place at approximately 4800, 1100, and 500 BP. All three events dammed the Squamish River and formed temporary lakes upstream of the debris (Brooks, 1992; Brooks and Hickin, 1991) in which fine grained sediments accumulated (Fig. 31).

Figure 30. *Broken trees in debris avalanche deposits in Turbid Creek. Tree above person's head gave radiocarbon date of 950 ± 80 (GSC-5195). GSC 1994-709J*

Figure 31.

Varved lacustrine sediments deposited in Squamish valley in lake formed by the blockage of Squamish River by mid-Holocene collapse of Mount Cayley. GSC 1994-709L

As described by Evans and Brooks (1991), failure of the cone took place after considerable dissection of the original edifice had exposed weak pyroclastic materials at the base of the steep upper slope of the volcano. No evidence of older debris avalanches from Mount Cayley has been discovered.

Smaller scale debris avalanches involving mechanically weak pyroclastic materials continue to occur from Mount Cayley's western flank in historic time. A 1963 event (Fig. 32; estimated volume $5 \cdot 10^6$ m^3) has been described by Souther (1980) and Clague and Souther (1982). The fahrböschung[1] of the landslide was 22° and velocities, calculated from superelevation data, reached 15-20 $m \cdot s^{-1}$.

In 1984 a similar debris avalanche took place from Mount Cayley's western flank (Fig. 32) and resulted in a debris flow which temporarily dammed the Squamish River (Evans, 1986; Jordan, 1987; Cruden and Lu, 1989). Two different interpretations have been made of the event. According to Cruden and Lu (1992) approximately 3.2 million cubic metres of material travelled down into Turbid Creek and blocked it. The breaking of the dam then caused an extremely fast debris flow in Turbid Creek which produced wind gusts up to 34 $m \cdot s^{-1}$ and travelled down to the Squamish River. By contrast, Evans (1993) suggested that the 1984 event, contrary to an initial estimate of (Evans and Gardner, 1989), was an order of magnitude smaller (estimated volume $0.5 \cdot 10^6$ m^3). He did not find evidence to suggest that the debris stopped in Turbid Creek to form a dam. Instead, he proposed that the movement was continuous over a vertical distance of 1.18 km, to 3.46 km from its source. Below this point a debris flow was then initiated which travelled down the lower reaches of Turbid Creek and blocked the Squamish River as described by Cruden and Lu (1992). Evans (1993) concluded that the 1984 event showed hyper-mobile characteristics, i.e. the distance of travel of the debris was typical of a debris avalanche an

[1] The Fahrböschung of a landslide is the angle of the line joining the top of the headscarp and the distal limit of the debris.

Figure 32. *Aerial view of historic debris avalanches in Turbid Creek on the western slopes of Mount Cayley volcano; "A" is the source of the 1963 debris avalanche and "B" is the source of the 1984 event. Both landslides involved initial failure in Pleistocene pyroclastic rocks. This photograph, taken in 1985, shows the swath cut through the tree cover by the 1984 debris avalanche. GSC 204857-F*

order of magnitude greater. The fahrböschung for the 1984 landslide was 19° and based on superelevation measurements, velocities reached at least 31 $m \cdot s^{-1}$.

Debris avalanches from Mount Cayley and the effects of a possible damming of the Squamish River are major geomorphic hazards to public safety and economic development in the Squamish valley.

DEBRIS FLOWS

Channellized debris flows (also known as debris torrents) occurring in steep mountain watersheds and triggered by heavy, intense rains have been responsible for much damage in the Vancouver region (Evans and Clague, unpublished data; Evans, 1982; VanDine, 1985). They are defined by VanDine (1985, p. 44) as "a mass movement that involves water-charged, predominantly coarse grained inorganic and organic material flowing rapidly down a steep, confined, pre-existing channel". They are relatively frequent and are generally less than 100 000 m^3 in magnitude; many are less than 50 000 m^3.

Debris flows are initiated by shallow slides in thin mantles of colluvium or till on slopes above a creek channel bed into which they travel and where they become transformed into debris flows by undrained impulse loading of saturated channel materials (Bovis and Dagg, 1992). Rockfall may also provide a debris flow trigger by undrained impulse loading of channel materials. In other cases, debris flows become mobilized by a large increase in creek discharge (e.g., by the bursting of a debris dam), to a critical stream discharge, which causes the destabilization of channelled bed materials (VanDine, 1985; Bovis et al., 1985; Boris and Dagg, 1988).

Shallow failures occurring on steep, open slopes do not necessarily become channellized, but may still develop into what are termed debris avalanches or open slope debris flows (Evans, 1982). Figure 33 shows an example from the northern side of Britannia Creek which occurred as a result of heavy rains in September 1991. The landslide overrode part of the abandoned townsite of Tunnel Camp.

Figure 33. Debris avalanche (or open slope debris flow) which occurred in August 1991 as a result of heavy rains on north side of Britannia Creek, 4 km upstream of Britannia Beach. The landslide overran part of the abandoned townsite of Tunnel Camp. GSC 1994-492B

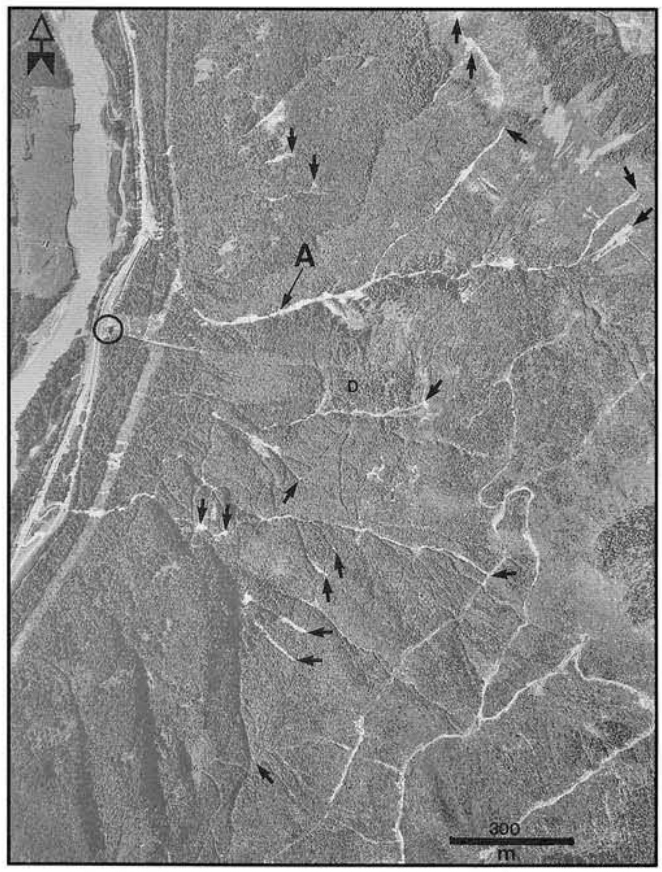

Figure 34. Aerial photograph showing 1983 debris flows in vicinity of Wahleach Power Station (circled) in the Fraser Valley. Multiple starting points of debris flows are arrowed. Note at least three sources for the debris flow in Ted Creek (A) discussed in text. Point "D", located just south of Ted Creek, is an area of topographic linears associated with rock slope deformation illustrated in Figure 19. (BC 83020-104)

Lower Fraser valley/Hope-Chilliwack

Debris flows triggered by intense rains have occurred frequently in the lower Fraser valley in this century (Miles and Kellerhals, 1981; Evans and Clague, unpublished data) resulting in numerous deaths and extensive property damage. In July 1983, for example, a series of debris flows triggered by a severe local rainfall blocked the Trans-Canada Highway and the Canadian National Railway mainline for three days at a number of locations between Chilliwack and Hope (Evans and Lister, 1984) triggered by a locally intense rainfall (Slaymaker et al., 1987). The debris flows originated as shallow failures on steep mountain slopes covered with a thin veneer of colluvial and/or till materials that had been logged in the recent past. Some debris flows had multiple initiating point slides (Fig. 34). At least 14 debris flows reached the Trans-Canada Highway. No loss of life was associated with the debris flow activity although one house was partially engulfed. The cost of repairs to road and railway was estimated to be \$300 000. The debris flow in Ted Creek (Fig. 34) had a volume of about 60 000 m^3, one of the largest debris torrent events to be documented in the Vancouver region (Slaymaker et al., 1987), and the velocity of the debris flow, estimated from superelevation data, reached at least 9.4 $m \cdot s^{-1}$ (Hungr et al., 1984).

Howe Sound-Whistler area

Debris flows, triggered by heavy rains, have occurred in the steep watersheds along the east side of Howe Sound fiord (Fig. 1). These have impacted on communities along the fiord as well as on the British Columbia railway and the Squamish Highway which run along its east side (Thurber Consultants, 1983; Jackson et al., 1985). In the 25 years between the completion of construction of both the railway and the highway in 1958, and 1983, 14 debris torrents occurred on six creeks resulting in 12 deaths, the destruction of 11 bridges, four houses, and other property damage (Lister et al., 1984). Four debris flows were described by Lister et al. (1984) (Fig. 35). They involved volumes in the range of 10 000 to 20 000 m^3; velocities were in the order of 3-6 $m \cdot s^{-1}$ and discharges were in the range 100-350 $m^3 \cdot s^{-1}$.

On October 31, 1981, a debris flow of about 20 000 m^3 in M Creek (Fig. 35, 36) swept down the steep mountain watershed over a distance of 2.2 km and destroyed the highway bridge crossing the creek. Nine people were killed when several vehicles drove into the chasm spanned by the destroyed bridge.

A similar debris flow occurred in Alberta Creek on February 11, 1983 and had a disastrous impact on the community of Lions Bay (Fig. 35, 37). The debris flow consisted of six surges during a period of approximately two and one-half hours in the early hours of the morning (Lister et al., 1984). In addition to substantial damage to transportation facilities, three houses in Lions Bay were destroyed and a house trailer crushed; two people in the trailer were killed.

In an earlier disaster, a debris flood in Britannia Creek in 1921, caused by the collapse of a blocked culverted, fill during heavy rains, devastated Britannia Beach destroying 60 houses and causing 37 deaths (Hungr and Skermer, 1992; Evans and Clague, unpublished data).

Triggers

Although debris flows are usually associated with heavy rainfall, attempts to specify weather conditions (e.g., total rainfall or intensity thresholds) which trigger debris flows have not been successful (e.g., Church and Miles, 1987). Factors that complicate such attempts include the role of snowmelt, antecedent moisture conditions, the availability of debris, and the fact that rain gauges are generally spaced too widely to detect localized high intensity rainfall cells which may trigger debris flows (e.g., the 1983 Wahleach events described above).

A second complex causative relationship is that between debris flow initiation and change in land use in the source watershed either by deforestation or forest fire. Studies by O'Loughlin (1972) and Howes (1987) for example, found that clearcut logging and logging road construction increased the occurrence of initial shallow landslides of the type that could be transformed into debris flows. In contrast, many major historical debris flows have been initiated on slopes that have not been logged (e.g., the M Creek event described above) and slopes with natural vegetation can undergo significant landslide activity (Howes, 1987).

Defensive structures

As described by Hungr et al. (1987), Slaymaker et al. (1987), and Kellerhals and Church (1990) a variety of debris flow defensive structures have been constructed at numerous locations in the Vancouver region in recent years, following the construction of debris flow defensive works to protect the Vancouver Island community of Port Alice (Nasmith and Mercer, 1979). They include debris retention structures that stop and dewater debris in containment basins upstream, channellization works that confine the debris in its passage over a fan surface, and deflection berms that either divert the flow away from potential impacts on infrastructural facilities into a predetermined deposition area, or create open containment areas (Hungr et al., 1987).

The most sophisticated structures in the region have been constructed at several locations along Howe Sound in the vicinity of Lions Bay (Fig. 1) for a total cost of about \$20 million (Evans and Clague, unpublished data). At Charles Creek, (Fig. 35) a large debris retention basin (Fig. 38) was constructed in the mid-1980s at the head of the bay, to protect transportation routes and expensive homes below on the fan. A similar structure was constructed on Harvey Creek at the head of the fan which has a debris retention capacity of 70 000 m^3. At Alberta Creek, scene of the 1983 fatal accident, channellization of the creek (Fig. 39) was carried out to confine flood or debris flow discharges in its passage over the Alberta Creek fan within a gently curving concrete-lined channel (Hungr et al., 1987).

Less sophisticated debris retention structures have been built or have been planned in the region. Hungr (1993), for example, described proposed debris retention structures to protect part of Whistler (Fig. 1) from possible debris flows in Whistler Creek. They consist of two 11 m high debris flow barriers consisting of a series of triangular buttresses of tubular steel, connected by massive steel grating (Fig. 40) which would create a total retention capacity of about 24 000 m^3.

Since the introduction of deflection berms as debris flow defensive structures at Port Alice, Vancouver Island in the 1970s, a number have been built in the Vancouver region. Deflection berms have been constructed at several locations in the vicinity of Wahleach to protect the Trans-Canada Highway from debris flows (Fig. 1, 41). At Ted Creek, for example (see location in Fig. 34) a gravel borrow pit on the Ted Creek fan was shaped to act as a retention basin and a

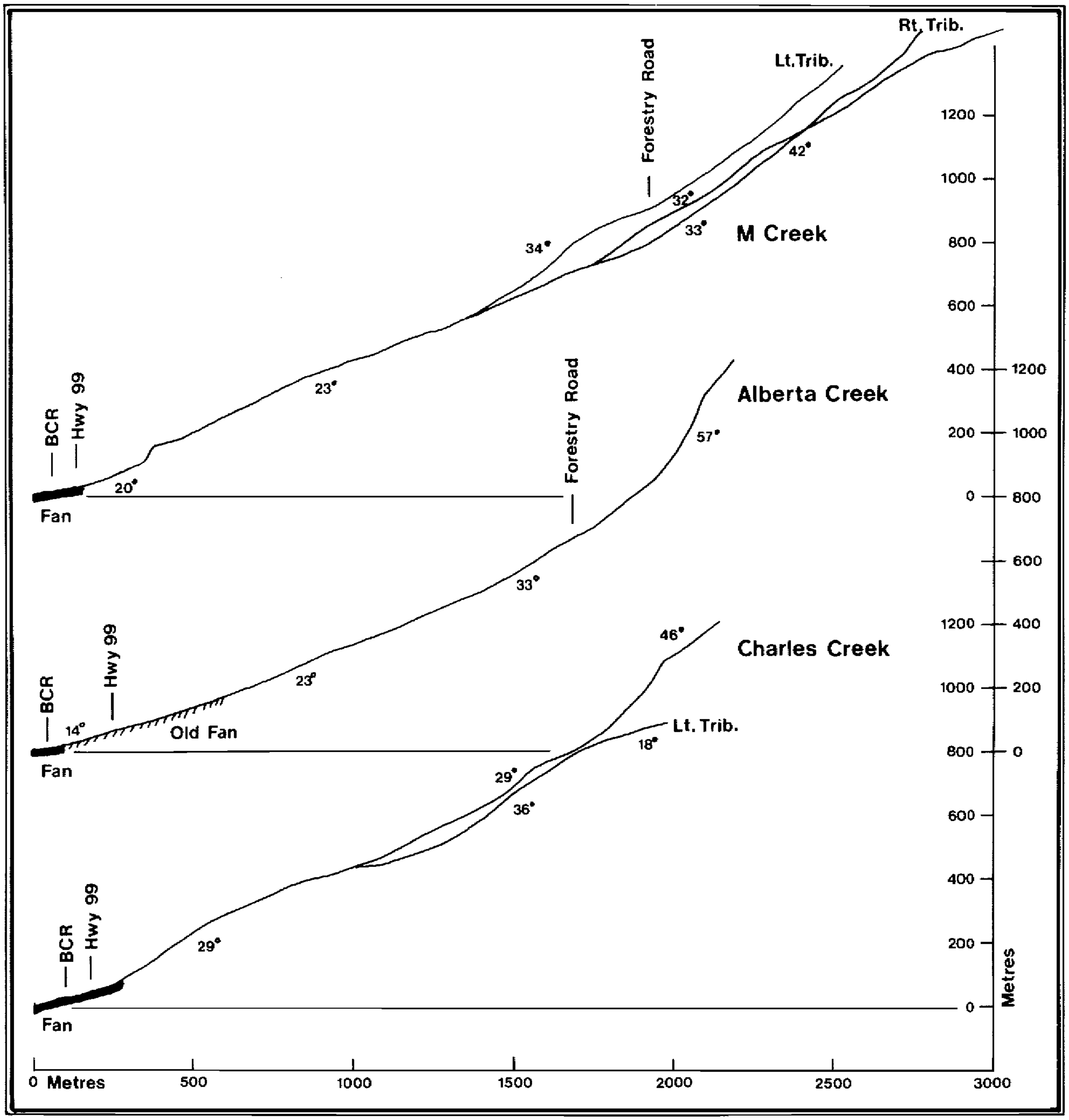

Figure 35. Profiles of creeks on east side of Howe Sound in which debris flows occurred between 1981 and 1983 (after Lister et al., 1984).

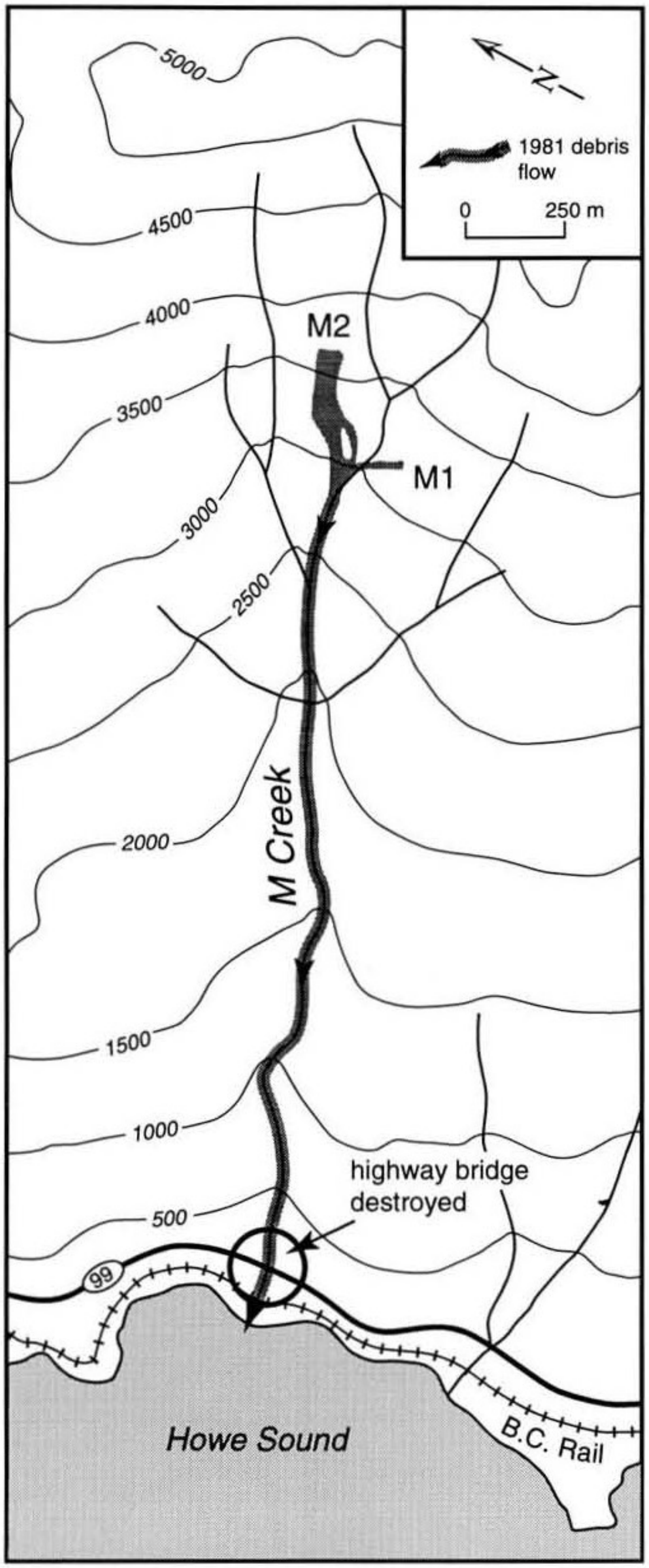

***Figure 36.** Map of M Creek debris flow which occurred on October 28, 1981, 2.5 km north of Lions Bay, east side of Howe Sound; M1 and M2 are probable source landslides for the debris flow (modified after Bovis and Dagg, 1992).*

***Figure 37.** Alberta Creek debris flow, Lions Bay, which occurred on February 11, 1983. Note damaged buildings and bridges (photo courtesy of British Columbia Ministry of Highways and Communications).*

***Figure 38.** Debris flow retention structure on Charles Creek, Howe Sound constructed in the mid-1980s; view downstream. GSC 1991-298*

deflection berm for a retention capacity of 60 000 m^3 (Hungr et al., 1987); deflection berms were also constructed to protect the Agassiz Mountain Correctional Institution from debris flows (Martin et al., 1984) creating a total containment capacity of 12 000 m^3. In addition channellization works and deflection berms have been constructed in several locations along the Coquihalla Highway (Slaymaker et al., 1987) for a total cost of $1.1 million.

LANDSLIDES IN PLEISTOCENE SEDIMENTS

Numerous landslides have taken place in the Pleistocene sediments of the Vancouver region in the historical period, particularly in the Lower Mainland (Eisbacher and Clague, 1981; Evans and Clague, in press). They are most common along the escarpments composed of unconsolidated Wisconsinan sediments that define the edge of terraced

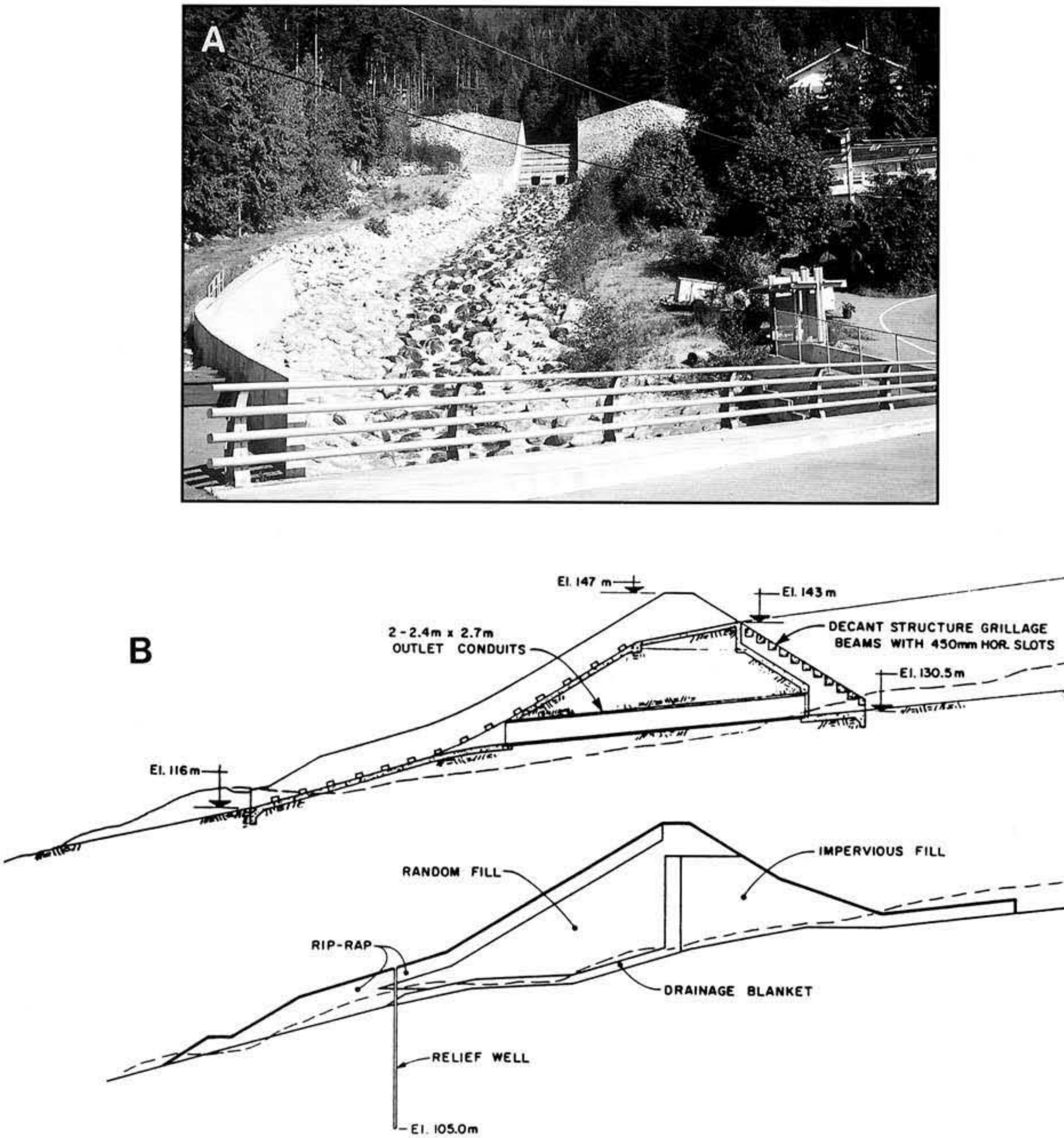

Figure 39. Debris flow retention dam at Harvey Creek, Lions Bay, Howe Sound. ***A)*** *View upstream; GSC 1994-709K* ***B)*** *Structural cross-section (after Hungr and Skermer, (1992) from a sketch drawing by Ker Priestman Associates).*

uplands in the Greater Vancouver area (see Fig. 2 in Eisbacher and Clague, 1981). Materials involved in these events consist of glacial, glaciofluvial, glaciomarine, and glaciolacustrine sediments (Armstrong, 1984) that exhibit considerable geological, hydrogeological, and geotechnical complexity (Eisbacher and Clague, 1981; Evans, 1982; Hungr and Smith, 1985). Problems related to landslides in Pleistocene sediments are not restricted to the Lower Mainland. Recently, Gerath (1993a, b) has pointed out the hazards posed by landslides and related processes on the Redroofs Escarpment on the Sunshine Coast (Fig. 1).

Rapid earthflow in sensitive glaciomarine sediments

Glaciomarine sediments are widespread in the Lower Mainland (Armstrong, 1984) but only one large landslide has occurred in these deposits in historical times. On January 30, 1880, a major landslide (estimated volume 10^6 m^3) occurred at Haney (Fig. 1) in glaciomarine sediments on the eroding north bank of the Fraser River (Fig. 42; Evans, 1982). Eyewitnesses reported that they heard the cracking of the ground and watched as a "great .. moving mass of earth and trees...slid

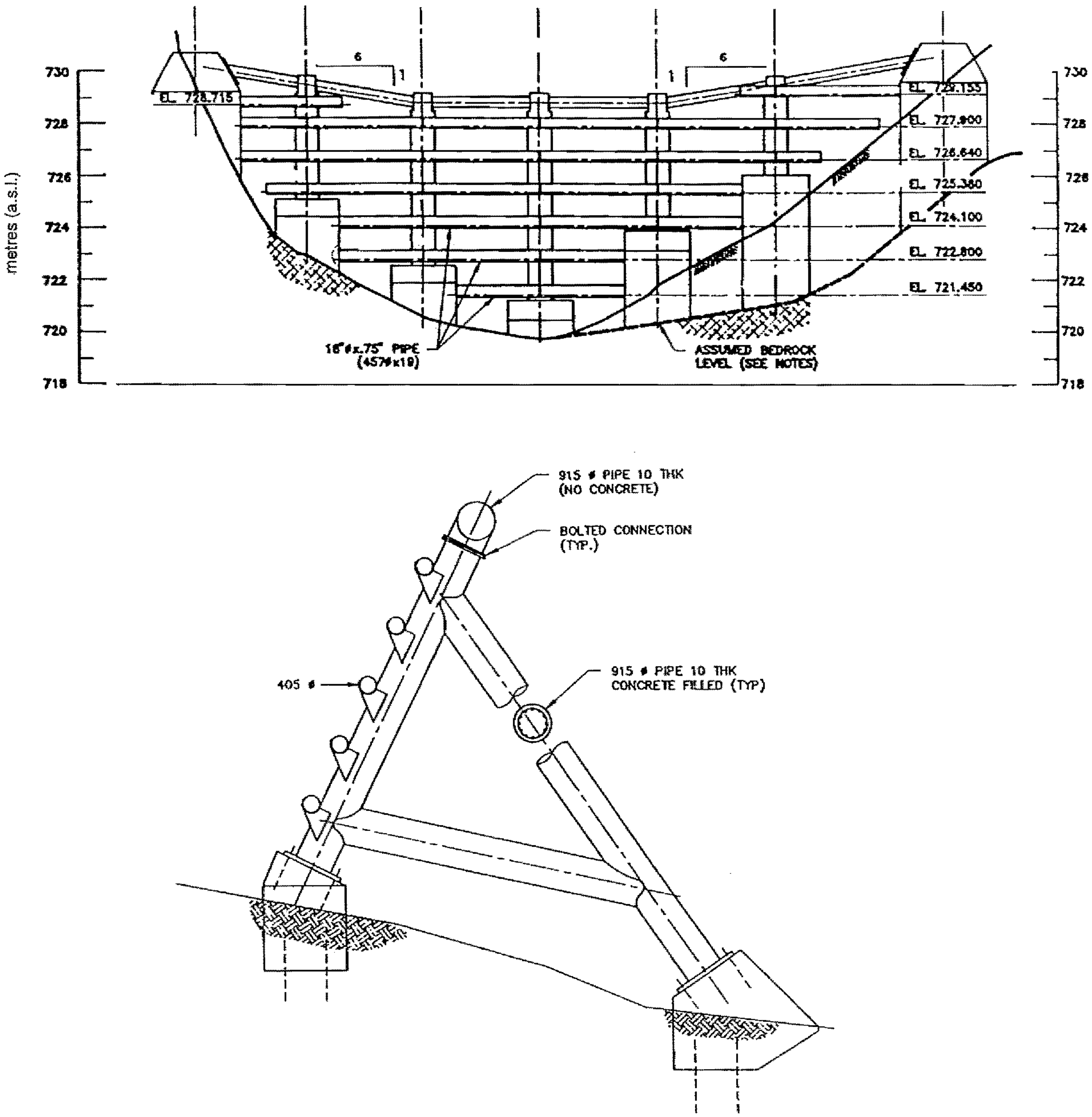

Figure 40. Conceptual design of debris retention barrier for Whistler Creek (from Hungr, 1993, courtesy Ker Priestman Associates).

into the Fraser River" (Victoria Daily Colonist, February 5, 1880). The slide partially blocked the Fraser River and resulted in the death of one person (killed by the 12 m high displacement wave caused by the slide). Substantial damage to docking facilities occurred along the Fraser River (Evans and Clague, unpublished data). Excess pore pressures within sandy interbeds in the sensitive glaciomarine silts and clays and erosion at the toe of the slope by the Fraser River are probable causes of the slide (Evans, 1982).

Debris avalanches and debris flows

The occurrence of debris flows and debris avalanches along escarpments in Pleistocene sediments in the Greater Vancouver area have been described by Eisbacher and Clague (1981), Woods (1984), and Hungr and Smith (1985). They are triggered by heavy autumn and winter rains. The steep ravines and gully walls along the escarpments, which range up to 125 m in height, produce channellized debris flows by two initial mechanisms (a) the failure of a relatively thin (1-2 m) cover of colluvium on slopes steeper than 30° or (b) the slumping of material at the head of a gully. Both these initial failures are transformed into debris flows which run out on areas below the escarpments. The possibility of initial failure may be increased by the placement of loose fill at the heads of the gullies or steep ravines (Hungr and Smith, 1985) which has the effect of loading the slope or impeding drainage.

Open slope debris flows or debris avalanches also occur (Eisbacher and Clague, 1981).

A typical example of this style of landslide is the Port Moody debris flow of December 1979 (Eisbacher and Clague, 1981; Woods, 1984) (Fig. 43). Triggered by torrential rains, a slide involving 4000 m^3 of dumped fill at the crest of a ravine, entered a steep sided gully and surcharged loose material in the floor of the gully. Under undrained loading the debris began moving on a slope of 15°, entrained other materials from the gully bottom, and travelled a distance of 600 m over an average slope of 9°. The debris flow demolished one house and inundated other houses and apartments with debris (Fig. 43). Evidence noted by Armstrong, (1984) suggests that the escarpment had been subject to large debris flows in the prehistoric past.

Figure 41. Debris deflection barrier (or berm) constructed to protect Trans-Canada Highway and Canadian National Railway rail track, 5 km southwest of Laidlaw in the Fraser Valley. GSC 1994-492E

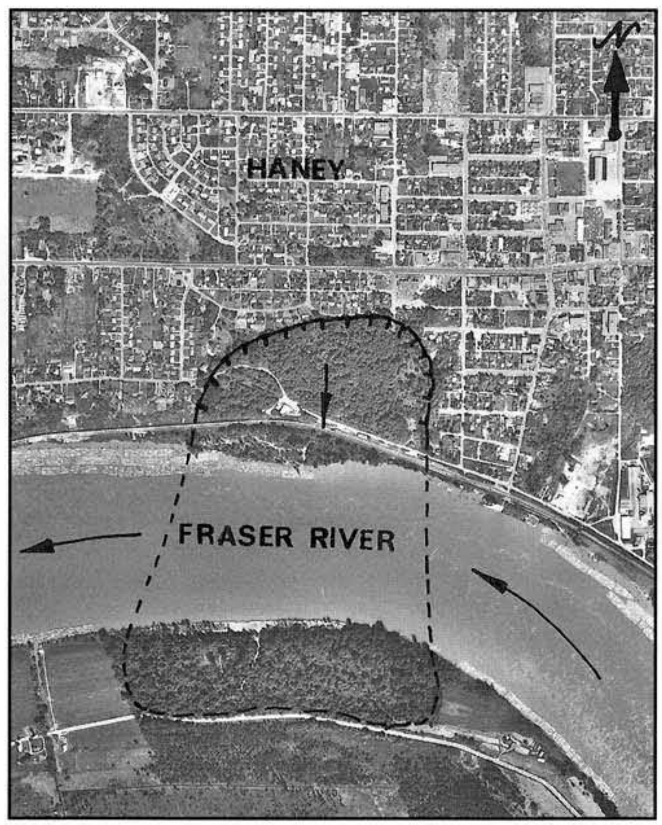

Figure 42. Airphoto of site of 1880 Haney landslide (from BC 7056-116).

Catastrophic seepage erosion

Catastrophic seepage erosion (Hutchinson, 1982), which is also known as caving erosion (Hungr and Smith, 1985), results in what are locally referred to as "washouts"; it occurs both on natural slopes on the escarpments, and in the excavated faces of gravel pits (Allan, 1957; Armstrong, 1984). The process has considerable destructive potential and, according to Armstrong (1984), is the dominant mass movement phenomena occurring in the Fraser Lowland.

Seepage erosion (Fig. 44) occurs in steep slopes where thick, pervious sand, sandy silt, or gravel beds are confined between overlying impervious Late Wisconsin till and/or glaciolacustrine stony clay (Hungr and Smith, 1985) and underlying impervious or less pervious materials. Seepage erosion in the pervious layer undercuts the overlying material resulting in its collapse. Under sustained or increased seepage this type of erosion may increase progressively by the rapid retrogression of a seepage front thus creating significant gullies in a matter of hours. The failed material, which is easily liquified, flows rapidly from the seepage front to the toe of the escarpment where it accumulates in a fan. Seepage erosion may therefore result in inundation of areas downstream in the depositional fan area and also endanger areas upstream by the rapid retrogression of the seepage front.

On natural slopes, the famous UBC Campus washout of 1935 in the Point Grey seacliff, resulted from seepage erosion (Armstrong, 1984). At the cliffs, which are up to 70 m high, a thin veneer of till overlies Quadra Sand and, at the time of the washout, cliff slopes were being steepened by marine erosion. According to Armstrong (1984), the lowest 15 m of Quadra Sand in the section contains impervious clayey silt and organic sediment and much seepage occurs at the upper boundary of this zone. In January 1935, following two days of torrential rain after a week of heavy snowfall, seepage erosion removed about 100 000 m^3 of Quadra Sand, creating an instant canyon about 100 m long in less than two days. During the period of erosion, which drew a crowd of spectators, the steep walls of the canyon repeatedly collapsed sending surges of sand and water down to the sea (Eisbacher and Clague, 1981).

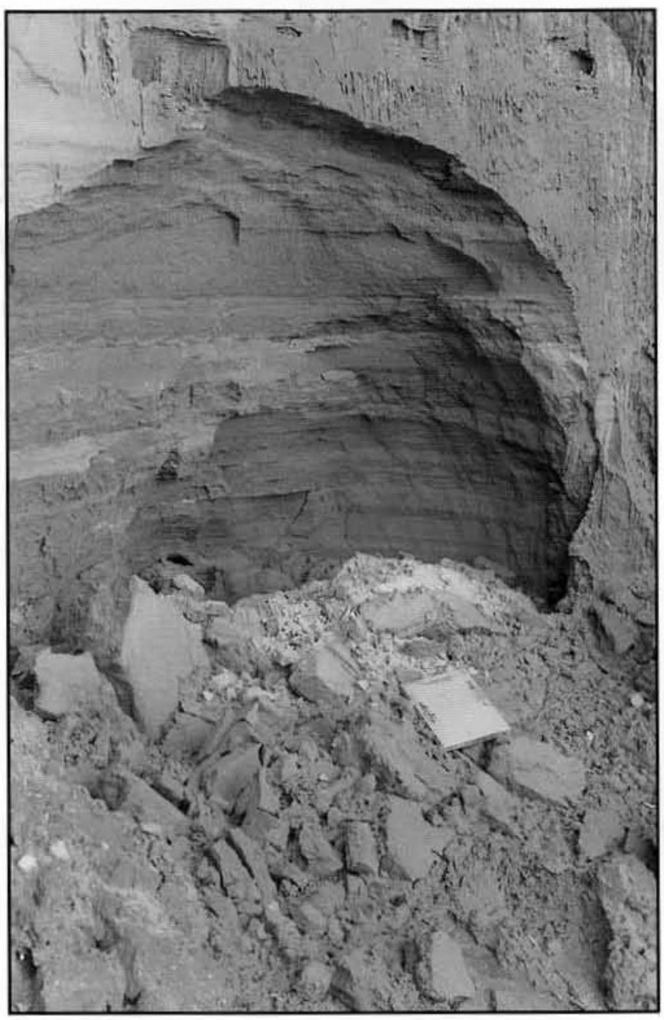

Figure 43. Seepage erosion in glaciofluvial sand at gravel pit in Coquitlam River valley. Note field book for scale (courtesy of O. Hungr, Thurber Engineering).

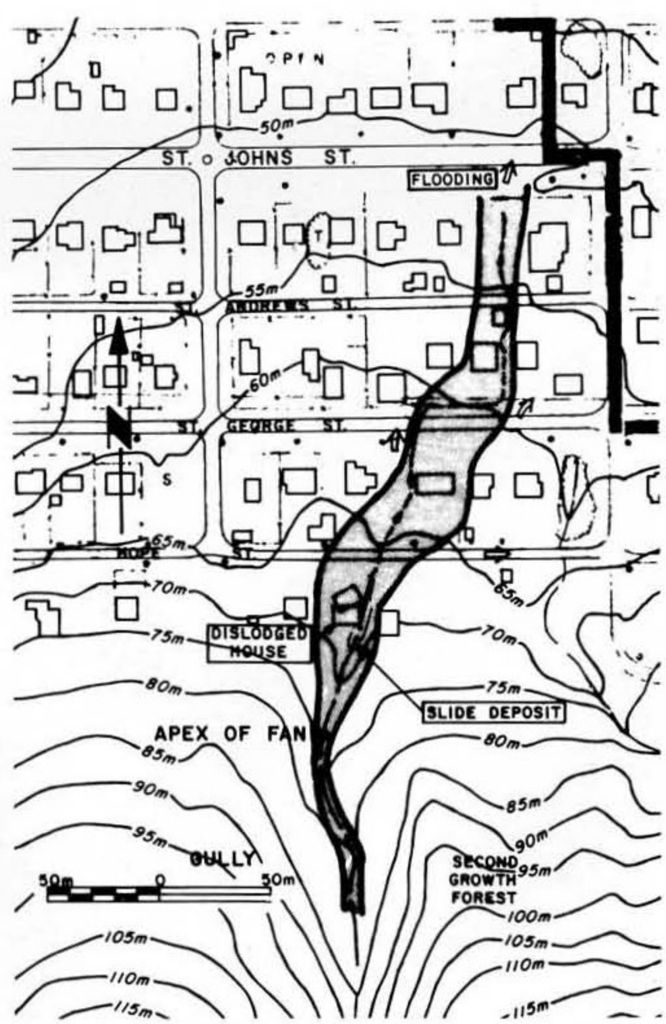

Figure 44. Map of run-out of 1979 Port Moody debris flow (after Woods, 1984).

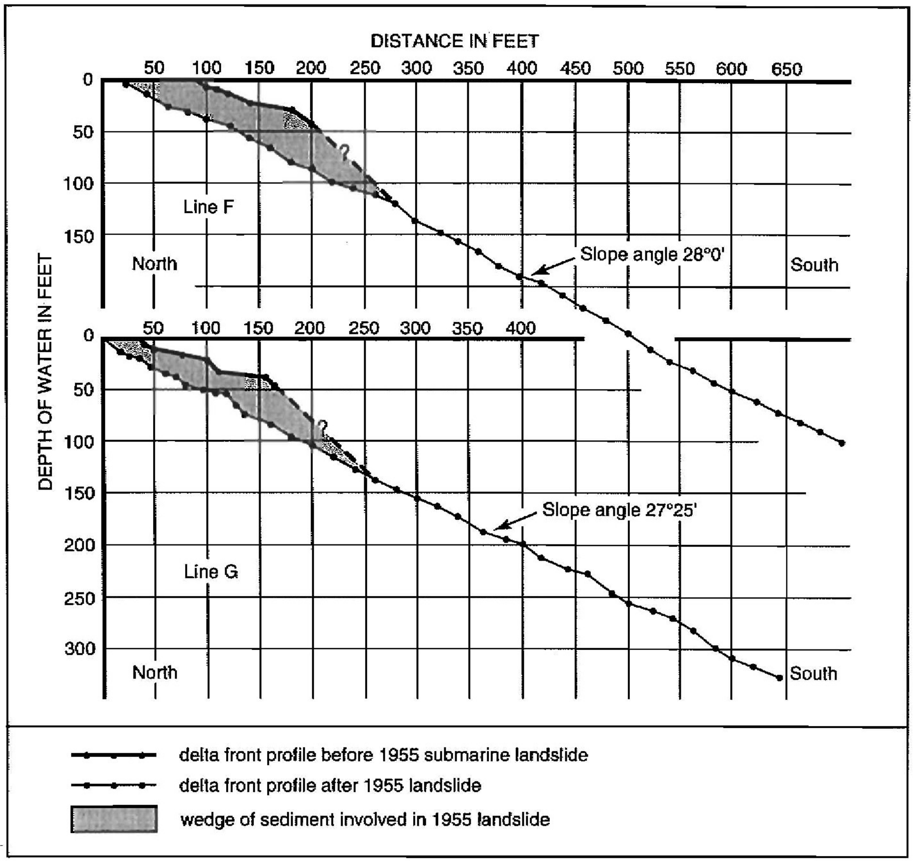

Figure 45. Vertical sections through Woodfibre delta front before and after 1955 slide (modified after Terzaghi, 1956).

Another major catastrophic seepage erosion event in a natural slope is described by Allan (1957) and Armstrong (1984) in the Coquitlam River valley. It occurred in 1952 and was initiated by seepage at the base of a 100 m section of pervious sands and silts at its contact with impervious, partly compacted glaciofluvial gravel. Within 24 hours, approximately 300 000 m^3 of material was washed out into the Coquitlam River blocking it for several days and an amphitheatre-shaped gully complex up to 300 m long had formed in the valley side.

Catastrophic seepage erosion is particularly common in the faces of gravel pits in the Vancouver region (Armstrong, 1984). According to Hungr and Smith (1985) seepage erosion in Coquitlam River gravel pits may take place at a rate of 50 $m{\cdot}a^{-1}$. In 1953 in a pit at Mary Hill, for example, sand and gravel was being mined from beneath an impervious cap of till. A mechanical shovel cut through a silty bed of gravel and tapped a groundwater reservoir which led to excessive seepage. Within 15 hours, 70 000 m^3 of material had flowed from the face.

Landslides in glaciolacustrine deposits

Glaciolacustrine sediments were deposited in the mountain valleys of the Vancouver region when glacier ice in the Fraser Lowland blocked their outlet (e.g., Saunders et al., 1987). The deposits are not widespread but consist of varved silts and clays and are prone to landslides. Extensive retrogressive landslides have occurred in these deposits, for example, on the north side of the Chilliwack River valley, westward from Tamihi Creek (Fig. 1; Armstrong, 1984).

SUBMARINE FAILURES (OUTSIDE FRASER DELTA)

Submarine failures are common in the Vancouver region particularly on delta fronts in both marine and lacustrine environments. The failures largely involve unstable wedges of sediment which form as a result of rapid subaqueous deposition. Instability on the Fraser Delta front is described in Luternauer et al. (1994).

Marine

Submarine failures have been documented in (Fig. 1) Howe Sound on several deltas. In 1955, at Woodfibre, failure of a delta slope at the mouth of Woodfibre Creek damaged warehouses and wharf facilities constructed on the delta. The cost of the damage was in the range of \$500 000 to \$750 000 (Bornhold, 1983) and the pulp mill was forced to shut down

Marine

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Submarine slope failures have also been documented on the delta of Britannia Creek at Britannia Beach on the east side of Howe Sound (Fig. 1; Moore, 1983; Prior and Bornhold, 1986). A history of instability on this delta is described by Moore (1983). In June 1957, for example, a section of the British Columbia Railway was destroyed by a submarine slump on the delta shortly after the grade was constructed.

The Squamish Delta at the head of Howe Sound is of strategic economic importance. According to Hickin (1989) the mean annual sediment flux to Howe Sound from the Squamish River is 1.29 million cubic metres per year and the Squamish Delta is prograding downfiord at a rate of $3.86\ m{\cdot}a^{-1}$. Instability on the Squamish Delta (Fig. 1) in response to this sediment input, detected by side-scan sonar, (Fig. 46) is illustrated by Prior and Bornhold (1984).

Lacustrine

As noted above, submarine failures are not limited to marine environments. Submarine landslide deposits have been detected in several lakes in the Vancouver region. In his study of Lillooet Lake, Gilbert (1975) described mounds of slumped material on the foreset slope of the Lillooet Delta. In Stave Lake, acoustic profiles (Gilbert and Desloges, 1992) indicate extensive slumping in underwater sediments; one extends for 7 km on the west side of the lake, and a second overlying slump deposit occurs over a distance of 3 km in the same region. In Garibaldi Lake, Mathews (1956) noted the occurrence of graded laminae of sand and coarse silt in the deep, flat-bottomed part of the lake 4 km from their source, and concluded that they were deposited by turbidity currents generated by submarine slumps.

EARTHQUAKE-TRIGGERED LANDSLIDES

There has been much discussion on the possibility of a future megathrust earthquake of high magnitude (M≥8) affecting the Vancouver region (e.g., Rogers, 1988). Such an event would have an important impact on slope stability. Given a range in epicentral distance to the Cascadia subduction zone of between 200 and 350 km, the empirical thresholds established by Keefer (1984) suggest that for M≥8 earthquake, disrupted landslides (including rock avalanches and rockfalls) will occur throughout most of the region, particularly west of the Lillooet River-Harrison Lake lineament. As a corollary, if this magnitude event has occurred on the Cascadia Subduction zone in the prehistoric past, it is assumed that a number of seismic triggered landslides would have occurred in the region.

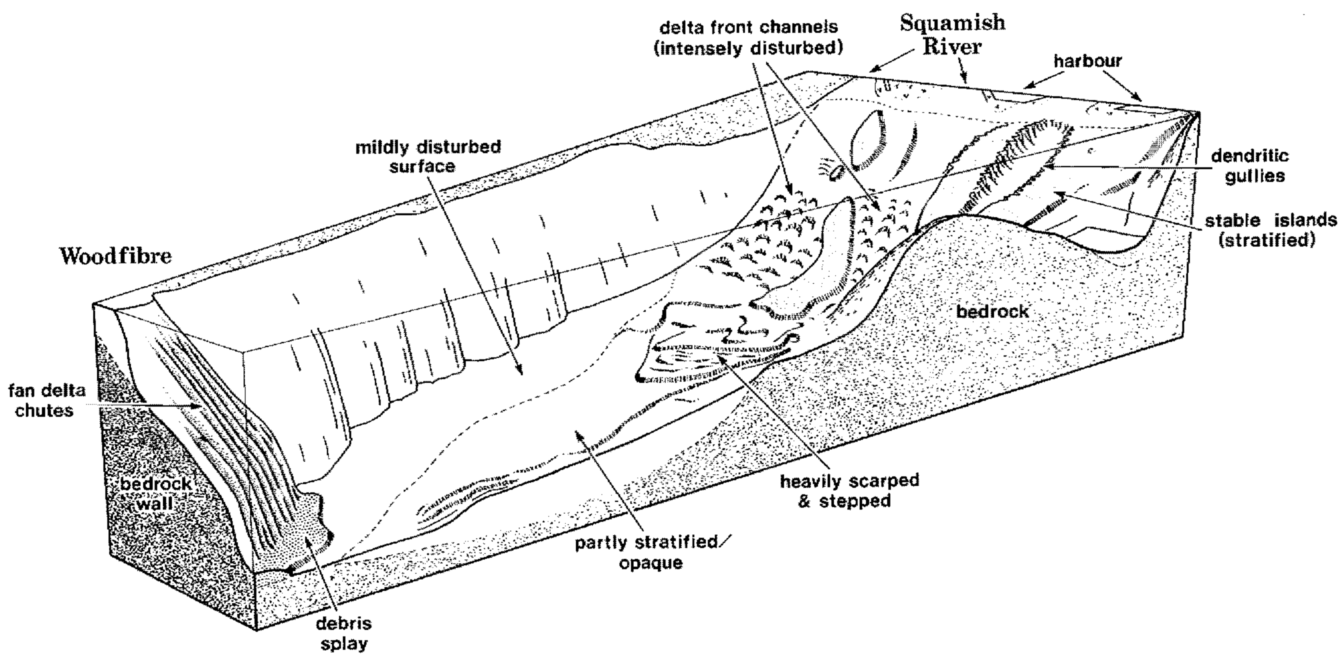

Figure 46. *Submarine geomorphology of Howe Sound from the Squamish delta to the Woodfibre delta showing mass movement features (after Prior and Bornhold, 1984).*

region is within Keefer's (1984) limit for the occurrence of disrupted slides. The earthquake did not trigger the Hope slide which, as noted above was in a state of limiting equilibrium. With reference to the 1946 Vancouver Island earthquake, Rogers (1980) mentions a slump at Matsqui triggered by the earthquake at an epicentral distance of about 250 km. It is also noted that the earthquake triggered several major landslides on Vancouver Island itself (Mathews, 1979) including the 1946 Mount Colonel Foster rock avalanche (Evans, 1989b).

DISCUSSION AND CONCLUSIONS

Landslides in the Vancouver region have had a direct impact on mining camps, residential communities, transportation routes, energy generation and transmission facilities, industrial sites, and on the quality of the forestry and fishery resource base.

The highest impact of landslides in the Vancouver region, in terms of direct cost, cost of mitigation, and loss of life, is by high frequency low-magnitude events (≤100 000 m^3). These events have an annual frequency of about 1:10 and consist of rainfall triggered debris flows and debris avalanches from steep mountain watersheds and escarpments in Pleistocene sediments, catastrophic seepage erosion events, and rockfalls and small rock avalanches.

Canada's second largest landslide disaster (1915 at Jane Camp, Britannia Mine; 56 deaths) and the country's largest known historical rock avalanche (1965 Hope slide) have occurred in the Vancouver region. The Hope slide occurred in two phases separated by 3 hours and occurred at the same location as a prehistoric landslide of similar magnitude. Several large rock avalanche debris accumulations in the region are seen to be the product of multiple events separated by as much as thousands of years.

Two major landslides with volumes in excess of 20 000 000 m^3 have blocked major transportation corridors in the Vancouver region since 1855, viz. the Rubble Creek and Hope events. This indicates an annual frequency of 1:100. Previous to 1855, published and unpublished field data indicate that the same transportation corridors have been blocked by landslides of similar magnitude about ten times since deglaciation, indicating a frequency of 1:1000, assuming no decay effect. Thus the historic frequency appears to an order of magnitude greater than the prehistoric frequency. No explanation of this discrepancy is offered at present.

Noncatastrophic mountain slope deformation has had important direct impact on civil engineering structures, and indirect impact because of uncertainty about their future behaviour.

The Garibaldi Volcanic Belt is highly susceptible to major landslides (≥0.5 million cubic m). The Mount Garibaldi volcanic complex and Mount Cayley volcano (Hickson, 1994) have been the subject to frequent large-scale landslide activity in the Holocene which continues into the present century. At the debris accumulation fans formed by such activity, stratigraphic evidence and radiocarbon dates suggest that they have formed by multiple landslide events.

Pleistocene deposits in the region show a wide variety of landslide types reflecting the stratigraphical, geotechnical, and hydrogeological complexity of the materials. Landslides in these materials are generally triggered by heavy rains.

Submarine failures have occurred on steep delta slopes in both marine and lacustrine environments in response to rapid deposition in a high energy geomorphic environment.

It is noted that even in the absence of a significant historical earthquake, the frequency of landslides and the variety of landslide styles in the Vancouver region is notable. Large prehistoric earthquakes have undoubtedly triggered landslides in the region. Any future large earthquakes affecting the region must be expected to trigger widespread slope movements.

As our knowledge of the distribution of landslides in both space and time increases and our ability to quantify landslide mechanics improves, land use decisions based on this better understanding are increasingly being made. This has the effect of increasing the amount of land sterilized (i.e. taken out of productive use), due to exposure to perceived landslide hazard. In addition, as landslide hazards become better known, retrofitting of existing facilities is increasingly being undertaken, frequently at substantial cost. Both of these responses to an increased knowledge of landslides in the Vancouver region contribute to their indirect cost.

Because of the increasing vulnerability of developed sites associated with rapid economic development, it is concluded that all of the landslide types and processes discussed here will have increasing impact on facilities in the Vancouver region. Much research remains to be done on the nature of their occurrence and the mechanics of their behaviour.

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The geological framework of groundwater in the Greater Vancouver area

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Ricketts, B.D. and Liebscher, H., 1994: The geological framework of groundwater in the Greater Vancouver area; in *Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia, (ed.) J.W.H. Monger; Geological Survey of Canada, Bulletin 481, p. 287-298.*

Abstract: Groundwater in the Vancouver, Fraser Valley and adjoining Coast and Cascade mountain area flows through three main types of material: (1) unconsolidated or semiconsolidated sediment of glacial and interglacial origin in the Fraser Valley and delta, which is less than 2 million years and mainly less than 100 000 years old; the sand and gravel aquifers generally have high porosity (up to 40%) and high hydraulic conductivity; (2) lithified Cretaceous and Tertiary sedimentary rocks, which have low intergranular porosity (<15%) and some fracture porosity; and (3) fractured granitic and metamorphosed sedimentary and volcanic rocks mostly greater than 90 million years old in the Coast and Cascade mountains. The greatest demand for groundwater, comprising more than 44% of water requirements in the Fraser Valley, is from Quaternary deposits and in particular the Sumas Drift (10 000-14 000 BP), which includes the unconfined Abbotsford and Brookswood aquifers. Water quality in some of these aquifers is already compromised by human activities. A sound knowledge of the hydrogeological properties of these, and the many confined aquifers in the region, is critical if expedient and cost effective remediation programs are to be developed.

Résumé : Les eaux souterraines de la région de Vancouver, de la vallée du Fraser et de la chaîne Côtière avoisinante s'écoulent dans trois types principaux de matériaux : (1) des sédiments non consolidés ou semi-consolidés d'origine glaciaire et interglaciaire dans la vallée et le delta du Fraser, qui remontent à moins de deux millions d'années et en général à moins de 100 000 ans; les aquifères de sables et de graviers ont généralement une porosité élevée (jusqu'à 40 %) et une conductivité hydraulique élevée; (2) des roches sédimentaires crétacées et tertiaires lithifiées, qui ont une faible porosité intergranulaire (<15 %) et une certaine porosité de fracture; et (3) des roches granitiques fracturées et des roches sédimentaires et volcaniques métamorphisées qui ont plus de 90 millions d'années dans la chaîne Côtière et la chaîne des Cascades. Le plus fort prélèvement d'eaux souterraines, qui représente plus de 44 % de la demande en eau dans la vallée du Fraser, est effectué dans des dépôts quaternaires, notamment dans le Drift de Sumas qui remonte à entre 10 000 et 14 000 ans et qui comprend les aquifères libres d'Abbotsford et de Brookswood. Les activités humaines ont déjà gâté la qualité de l'eau dans certains de ces aquifères. Il est essentiel de bien connaître les propriétés hydrogéologiques de ces aquifères et des nombreux aquifères artésiens de la région, afin de pouvoir élaborer des programmes correctifs convenables et rentables.

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INTRODUCTION

Groundwater is a hidden, and all too often forgotten resource. About 26% of the population of Canada, and 22% of the population of British Columbia, relies on groundwater for municipal, domestic, and rural use (1981 figures from Environment Canada). In the Fraser Valley, groundwater supplies about 44% of total water needs (1981 figures, Halstead, 1986), of which domestic consumption accounts for about 50%, and farming and industrial activities use the remainder (Dakin, 1991). Groundwater systems in this area range from wells supplying single households or farms (e.g., aquaculture, irrigation), common in rural British Columbia, to high production wells that provide water for large communities (e.g., Langley) and industries such as fish hatcheries.

Despite the obvious reliance on this valuable resource, groundwater quality in many areas has been compromised by some domestic, agricultural, and industrial practices, originating perhaps from the attitude "out of sight, out of mind". Nevertheless, government agencies and technical societies (B.C. professional engineering and groundwater associations) are currently working to make groundwater more 'visible'. As a contribution to that effort, this paper outlines the geological framework of groundwater in the lower Fraser Valley, Vancouver and adjacent coast mountain region.

Groundwater in the greater Vancouver area flows through three main types of material, each being considered separately below (Fig. 1; also Halstead, 1967): unconsolidated or semi-consolidated sediment of glacial and interglacial origin, less than 2 million years old, and predominantly less than 100 000 years old (mainly Fraser Valley and delta); lithified Cretaceous and Tertiary sedimentary rocks; fractured granitic, and metamorphosed sedimentary and volcanic rocks greater than 110 million years old in the Coast Mountains and Cascade Mountains (also Foxworthy et al., 1988; Halstead, 1991; Monger and Journeay, 1994).

GROUNDWATER AS A GEOLOGICAL AGENT

Nearly all geological processes, whether at the Earth's surface or deep in the crust, require water: metal-bearing hydrothermal fluids consist mostly of water; most diagenetic alteration of sediments involve waters in chemical reaction and water as a medium for mass transport of dissolved minerals; and at the surface, sedimentation, erosion, and weathering require precipitated, frozen, or flowing water.

Groundwater is part of the total global water budget and part of the water cycle. Water precipitated on the Earth's surface will infiltrate soil and rock, flow across the land surface, evaporate or freeze, depending on such factors as climate, geology, vegetation, and topography. Groundwater may also be recharged or discharged by streams depending on the relative positions of hydraulic heads in the stream and adjacent aquifer. Local weather and stream flow patterns can seasonally alter these processes. For example, the reach of Fish Trap Creek (Matsqui) near the international border loses groundwater to the aquifer in the winter and gains groundwater from the aquifer (recharges) during the summer (Liebscher et al., 1992).

The residence time of water below the Earth's surface ranges from days in the shallowest aquifers and soils, to hundreds of thousands of years at greater depths. Water that flows through lithified rock and unconsolidated sediment, gains by dissolution or loses by precipitation naturally occurring elements and compounds, the most common being silica, calcium, sodium, magnesium, potassium, iron, carbonate, and sulphate. As a general rule, the longer the time that groundwater resides in an aquifer (or confining unit) the more its chemical signature will change. In the Fraser Valley, the residence time of water in many unconfined aquifers is brief – probably months – and dissolved mineral content is correspondingly low. In contrast, some aquifers in fractured bedrock (e.g., on Gambier and Bowen islands) have elevated dissolved mineral contents, in part because of longer residence times, but also because these fractures were sites of hydrothermal mineralization, commonly depositing lead, zinc, copper, and arsenic sulphides.

GROUNDWATER DATABASES AND AQUIFER MAPPING

Hydrogeological mapping, which portrays groundwater distribution with respect to aquifers and aquitards (confining units), is essential for evaluating any groundwater problem, whether it be estimating of water resources, or development of remediation programs in areas of contamination. Data used to map aquifers and characterize groundwater flow are derived from a variety of sources (Fig. 2), the relative importance of which depend to some extent on the scale of mapping. Geology maps (e.g., Armstrong, 1980a, b; Armstrong and Hicock, 1980a, b) can be used to identify unconfined aquifers (aquifers that are exposed at the surface and contain the water table) or aquifer complexes (many closely spaced aquifers which are not necessarily in hydraulic contact), as well as regions of potential confined aquifers completely saturated aquifers below the surface which are occuring between units of very low permeability, for example clay- and silt-rich sediments. For mapping of aquifers at larger scales, water wells provide critical stratigraphic information, whereas hydraulic data is derived from pumping tests. Many drilling and geotechnical companies voluntarily provide well records and well testing data for curation by the B.C. Ministry of Environment, Lands and Parks. Additional sources of water well data include Environment Canada, the Geological Survey of Canada, and municipalities. Efforts are currently underway to incorporate much of the well water data into digital databases that can be accessed for simple queries or hydrogeological-hydrostratigraphic modelling. In particular, Geographical Information Systems (GIS) provide a powerful computer platform for organizing and analyzing spatial (georeferenced) and relational (e.g., porosity) information (Fig. 2).

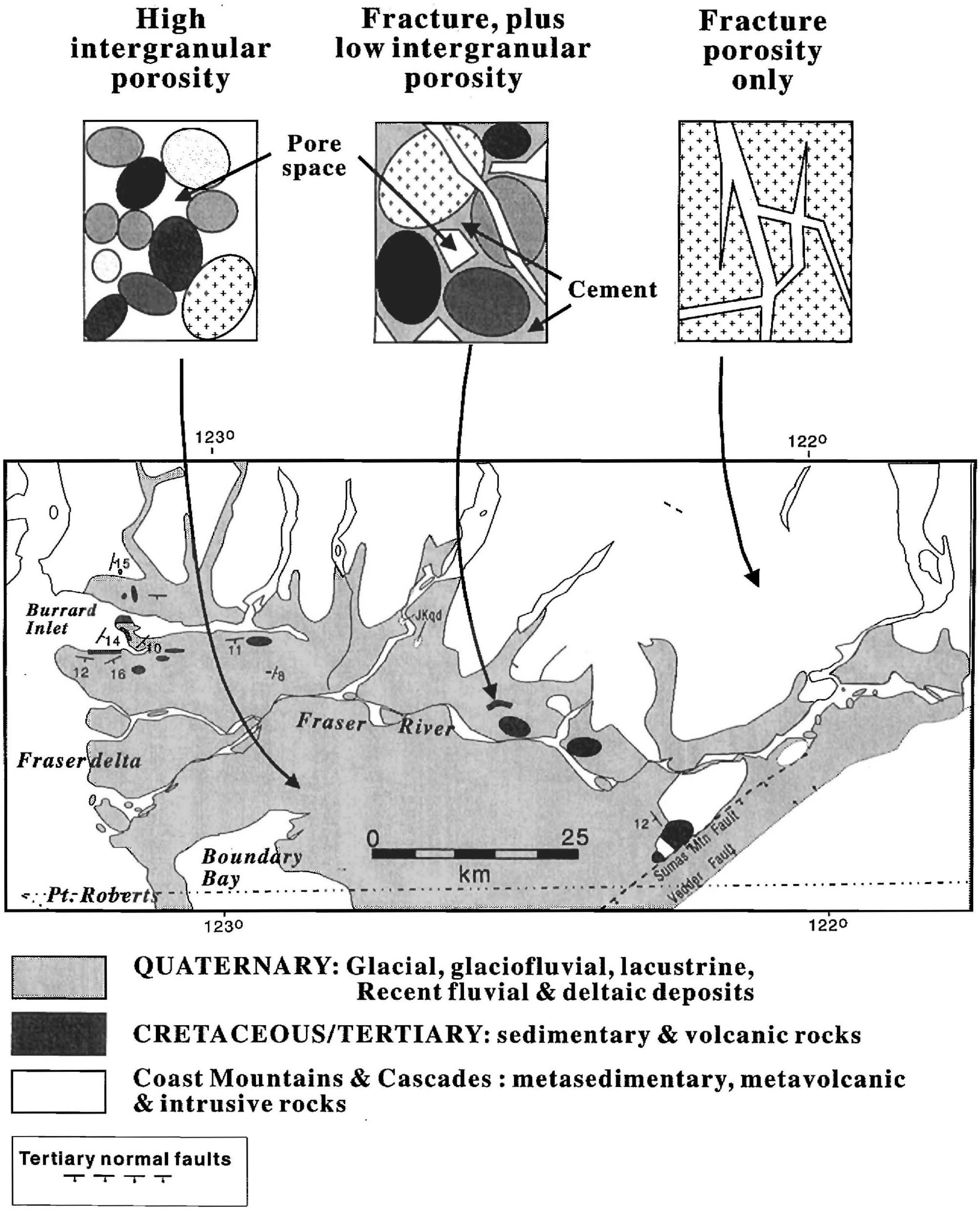


Figure 1. Generalized Vancouver geology, the principle groundwater domains and porosity types (map modified from Mustard and Rouse, 1994).

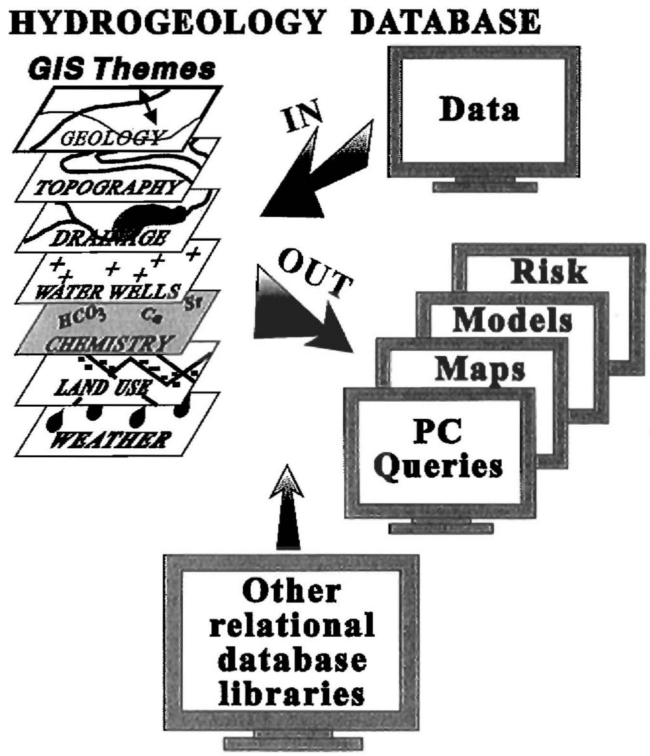


Figure 2. Database structure for groundwater mapping and aquifer characterization, showing typical data input and products.

Data pertaining to water quality and chemical composition is primarily under the purview of the B.C. Ministry of Health and local municipalities. Exceptions are trans-boundary (USA) surface waters and aquifers where data collection and monitoring is also done by Environment Canada.

Groundwater maps contain information on geological structure and lithostratigraphy, hydraulic parameters (e.g., hydraulic conductivity, porosity, and groundwater flow), hydrostratigraphy, and water resources. In lithostratigraphy the basic unit, the formation, is defined by the lithological homogeneity and mappability of a rock body (North American Commission on Stratigraphic Nomenclature, 1983). Formal definition of hydrostratigraphic units, on the other hand seems to be in a state of flux. Maxey (1964) introduced the term 'hydrostratigraphy' to refer to the material properties of a rock body as well as its hydraulic characteristics. However, the Stratigraphic Code does not recognize non-material hydraulic properties in its definitions. Others have included parameters such as permeability and porosity (e.g., Seaber, 1988), in addition to mappability. Defined in this way, the boundaries of a hydrostratigraphic unit (aquifer, aquitard) may not coincide with formal lithostratigraphic boundaries.

Recommendations and guidelines for groundwater mapping in British Columbia have been prepared for the British Columbia Provincial Resources Inventory Committee by Piteau Associates Engineering Ltd. and Turner Groundwater Consultants, 1993, based on guidelines used by UNESCO (1977), the Geological Survey of Canada, and the United States Geological Survey. Attempts are also being made to establish national guidelines for groundwater data management (e.g., Federal-Provincial Working Group on Groundwater, 1991). An aquifer classification system based on the assessment of risk to groundwater contamination and resource depletion has recently been presented (Kreye and Wei, 1994). A regional compilation of unconfined aquifers has also been published for the Fraser River drainage basin in British Columbia (Ricketts et al., 1993).

PRINCIPAL HYDROGEOLOGICAL DOMAINS

The Coast Belt

The geological underpinnings of Vancouver, Fraser Valley and Howe Sound constitute part of the Coast Belt a huge swath of plutonic rocks and subordinate metasedimentary and metavolcanic rocks, that at the surface are manifested as the rugged, high relief Coast Mountains (Roddick, 1965, 1990; Gabrielse et al., 1991). Metamorphic grade in the stratified rocks ranges from greenschist to amphibolite facies. Intergranular permeabilities are correspondingly low; values of 10^{-8} to 10^{-4} darcys are typical for igneous and metamorphic rocks, some 5-10 orders of magnitude less than permeabilities of sands and gravels (Freeze and Cherry, 1979). However, open fractures, resulting from faulting or jointing are abundant and these can serve as conduits for water. Although the orientation of large structures (faults and shear zones on a scale of kilometres in length) can commonly be projected below the surface, trends of smaller fracture sets are less predictable. Therefore, drilling of bedrock water wells in such rocks tends to be a higher risk venture than drilling in most surficial deposits.

Despite the relatively large number of bedrock wells in the Vancouver region, there is very little published information on production rates and other hydraulic parameters. Local reports from Bowen Island indicate a typical range of 4-50 L/min in wells drilled to depths of 60-150 m. Although detailed records are not always kept, most of these wells respond rapidly to seasonal changes in precipitation (and infiltration). Delay times are commonly only 1-3 months between the end of the winter wet season and noticeable drops in static levels and well recovery rates. Furthermore, it is not uncommon for wells separated by as little as 50 m to show no indications of interference during standard 72 hour pumping tests.

More detailed study of wells drilled into fractured granodiorite bedrock on Saanich Peninsula, Vancouver Island, provide an instructive comparison with bedrock wells in the

Vancouver region. Pumped rates in two test wells drilled to 179 m and 67 m were 13 and 16 L/s respectively (Kohut et al., 1983). They found that a 'trough-like' cone of depression (i.e. drawdown during pumping) had a long-axis orientation that bisected the angles between two major joint sets. In this case, the zone of influence extended up to 1.5 km from the wells. In comparison, the more localized areas of drawdown in places like Bowen Island, probably reflect more abrupt changes in bedrock structure, for example where layered metasedimentary rocks occur as pendants within larger granitic intrusions.

Groundwater compositions in fractured bedrock wells can be quite variable, particularly with respect to base metals which commonly are present as sulphides. Although there are presently no active mines in the Vancouver area, ore extraction during the earlier part of this century did take place at Britannia Beach (copper) and Bowen Island (copper, gold, and silver) – (Roddick, 1965). High concentrations of dissolved arsenic have also been detected in a few bedrock wells on Bowen Island.

Cretaceous and Tertiary successions

Indurated, Upper Cretaceous sandstone, conglomerate, and mudstone, exposed in Stanley Park in Vancouver and north of Burrard Inlet, are correlated with the Nanaimo Group (Mustard and Rouse, 1994). The Nanaimo Group, and overlying Tertiary units form a sequence more than 3 km thick that dips south in a broad syncline beneath Vancouver and the Fraser delta, and are exposed again in the Bellingham area (Fig. 3). Beds range in thickness from 1 m in the arkosic sandstones to more than 15 m in conglomerate facies. Bedding is commonly discontinuous. Mustard and Rouse (1994) interpret the local Nanaimo Group occurrences as being mostly of alluvial fan and braided river origin.

No information is available on the porosity and permeability of the Cretaceous rocks, although it might be expected that these parameters would be similar to or less than the average 15% porosity for the overlying Tertiary strata. Some of the Cretaceous conglomerates that are matrix-supported (i.e. high clay content) would be poor candidates for aquifers. Assuming that the abrupt lateral and vertical facies changes observed in

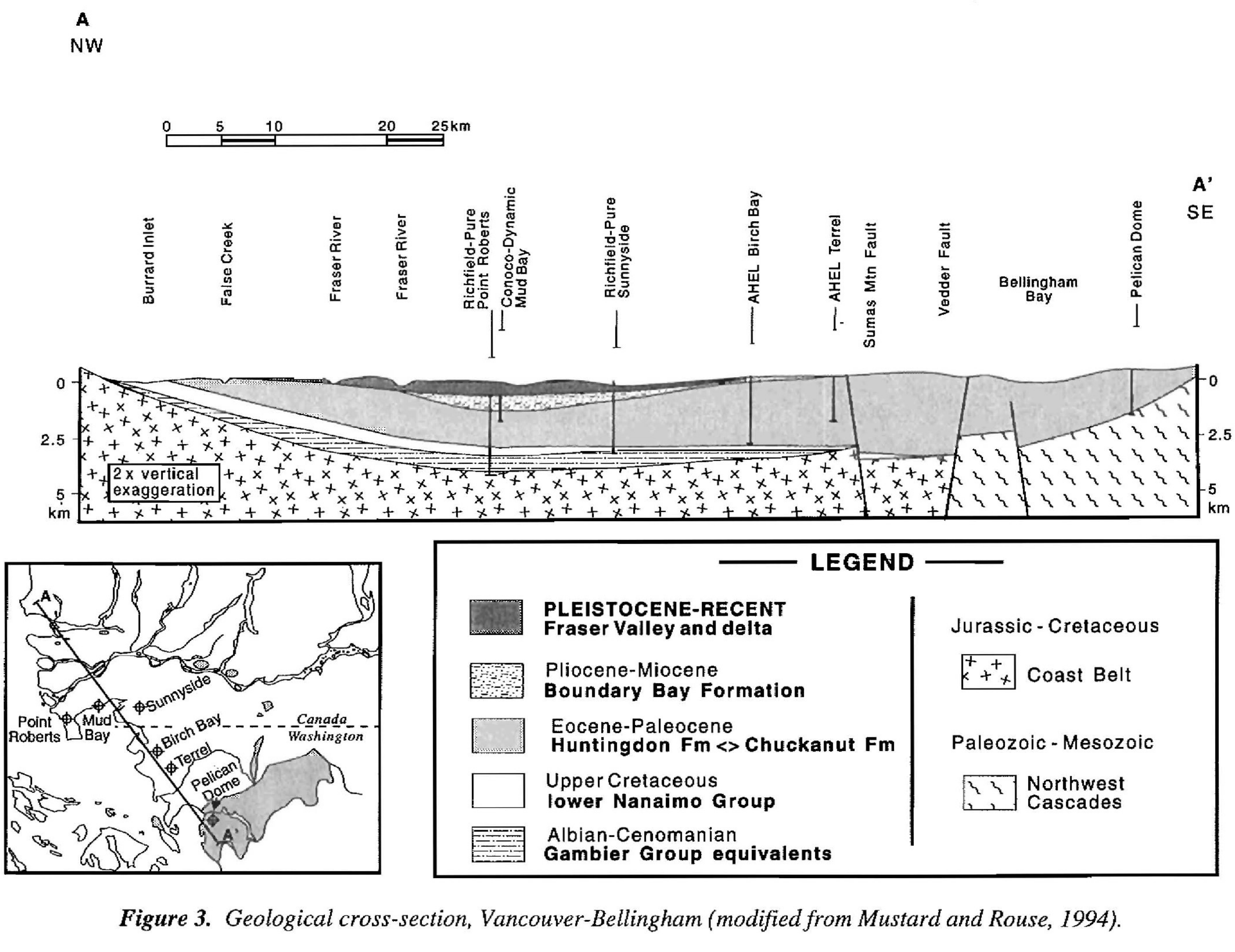


Figure 3. Geological cross-section, Vancouver-Bellingham (modified from Mustard and Rouse, 1994).

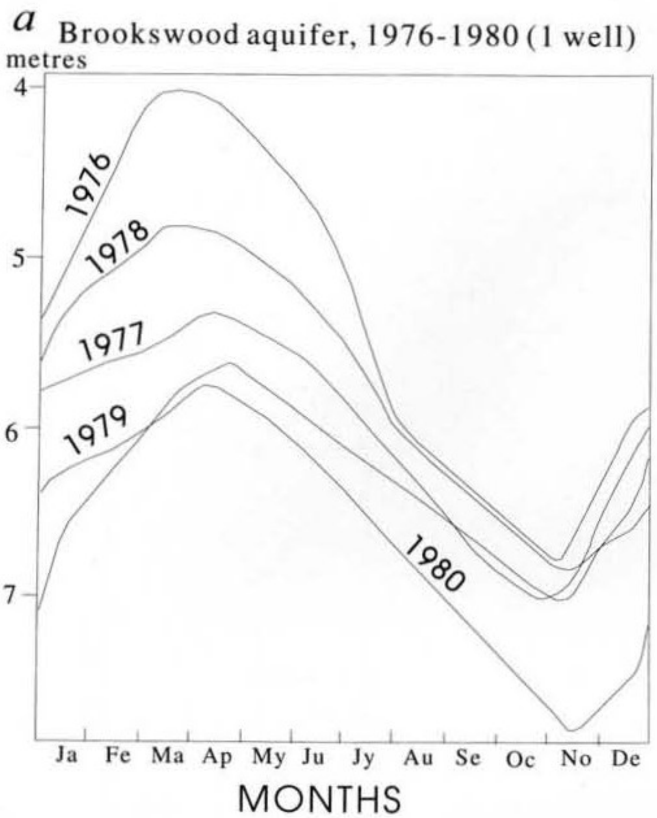
Table 1. Summary of characteristics for the hydrostratigraphic units. Data from Halstead (1986) and Liebscher et al. (1992).

Unit	Lithology	Thickness	Origin	Extent	Dissolved Solids	pH	Aquifer Type
A	Clay, stony clay, silty clay, minor silt and sand, shells. e.g., Fort Langley, Capilano formations	<30 m	Fallout from ice entering the sea	Near surface over most of the region	<120 mg/L	6-8	Aquitard, or isolated aquifers in sands
B	Stony clays with shells. e.g., Fort Langley and Capilano formations	<90 m	Glaciomarine	Langley upland and Serpentine Valley			Aquitard
C	Sand and gravel, minor clay e.g., Sumas Formation	40 m	Glaciofluvial, deltaic	Near surface in Langley and Abbotsford	<120 mg/L to 300 mg/L when locally confined by diamictons	6.5-8.5	Unconfined aquifer
D	Till or diamicton	up to 90 m	Ice contact, moraines	discontinuous throughout the region	<500 mg/L		many small confined aquifers
E	all deposits pre-Fraser Glaciation, sand, silt clay		Mixed marine, estuarine, fluvial	Usually encountered at depths >90 m	750-6000 mg/L		Aquifers and aquitards
F	Tertiary and Cretaceous bedrock			Substrate to Quaternary; exposed in peripheral areas.			Mostly fracture porosity

outcrop are typical of the Nanaimo Group in Vancouver, any potential aquifer would be markedly heterogenous and anisotropic. Fracture porosity may also be important, as is the case for Nanaimo Group rocks on Vancouver Island and the Gulf Islands, where well yields are as high as 4 L/s (Halstead, 1991), but given the drilling depths required to intersect the Cretaceous strata beneath Fraser Valley, their potential as sources of groundwater is low.

Inliers of Tertiary strata unconformably overlying the Nanaimo Group, are scattered throughout the Vancouver, north Fraser River, and Sumas Mountain region (Fig. 1). Two main lithostratigraphic units are present: the Paleocene-Eocene Huntingdon Formation, separated by an unconformity from the overlying Miocene-Pliocene Boundary Bay Formation (Mustard and Rouse, 1994). Exposed Huntingdon Formation strata are characterized by abundant conglomerate, sandstone, and minor mudstone, commonly arranged in fining-upward units ranging from single beds to cycles tens of metres thick. The proportions of these lithologies change in the subsurface, where up to 1500 m of sandstone, mudstone, subordinate conglomerate, and minor coal have been intersected in the Richfield-Pure Point Roberts and Richfield Pure Sunnyside oil wells (Fig. 3). Depositional settings for the Huntingdon Formation are interpreted by Mustard and Rouse (1994) to have been dominated by alluvial-fluvial processes, with alluvial fans, braided and high sinuosity streams represented.

There is little publically available data on the porosity and permeability of the Tertiary rocks. Gordy (1988) however, cites an average 15% porosity for much of the Paleocene-Eocene succession. Aquifers developed in the alluvial fan facies would likely exhibit marked heterogeneity and anisotropy. Sandstone bodies of meandering stream origin, on the other hand, may have more sheet-like geometries and less heterogeneity. However, the fining-upward character of many units implies a 'built in' anisotropy. In general, the groundwater potential for the Paleocene-Eocene succession is low. Except in the few Tertiary outliers, groundwater recharge is probably controlled by vertical leakage from the overlying Pleistocene deposits, or from deeper, regional meteoric flow.



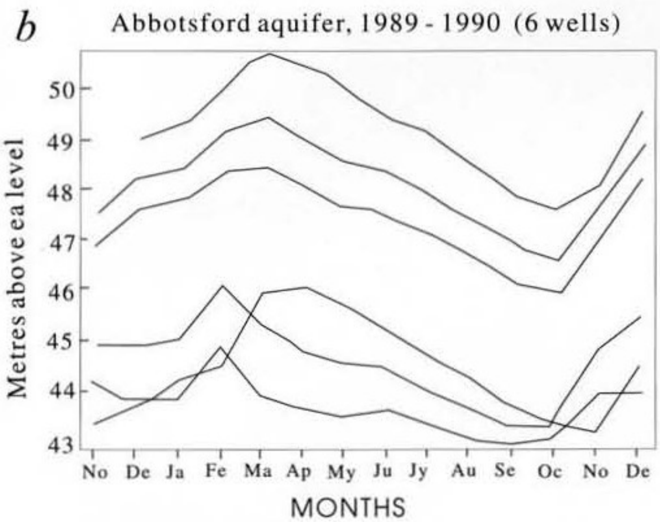


Figure 4. *Hydrographs of water wells in **a)** the Brookswood aquifer (from Halstead, 1986) and **b)** Abbotsford aquifer (from Liebscher et al., 1992), showing seasonal variation in piezometric surfaces (water tables).*

Figure 5. *Fan delta deposits in a gravel pit on Bradner Road, Matsqui, Up to 6 m of foreset beds are overlain by a 1 m topset unit of trough crossbedded and laminated sand. The upper unit includes a clay and diamicton that locally confine the aquifer. Water table is just below the floor of the pit (bottom right). GSC 1994-710*

Up to 1200 m of Miocene-Pliocene sandstone and mudstone, referred to as the Boundary Bay Formation, overlies the Huntingdon Formation south of Fraser River. The Boundary Bay Formation is weakly indurated (Mustard and Rouse, 1994) and likely has higher porosities than the underlying deposits. Although the groundwater potential of the Boundary Bay sandstones is probably greater than that of the older rocks, they too are overlain by Pleistocene glaciogenic and Holocene fluviodeltaic sediments that in places may be 700 m thick, and thus are not economic to exploit.

A great deal of relief exists on the (sub-Quaternary) erosion surface developed on top of Tertiary bedrock in the Fraser Valley (see Hamilton and Ricketts, 1994). This relief must have a profound effect on patterns of regional groundwater flow. Depths to the unconformity range from above sea level on the Tertiary and Cretaceous outliers, to 700 m below sea level between the Fraser River and the International border. Some of the surfaces slope 6°-10° in the Sumas Mountain and Point Roberts areas. Some of these 'paleovalleys' trend north-northeast, and appear to be extensions of modern valleys in the Coast Mountains north of Fraser River (e.g., Indian Arm, Pitt Lake, Stave Lake). This suggests the possibility that the large-scale drainage patterns are at least as old as the earliest glaciation. Although major faults in the Quaternary deposits are unknown, faults or shear zones that disrupt Tertiary and older bedrock could act as potential feeders of deep meteoric water to younger aquifers.

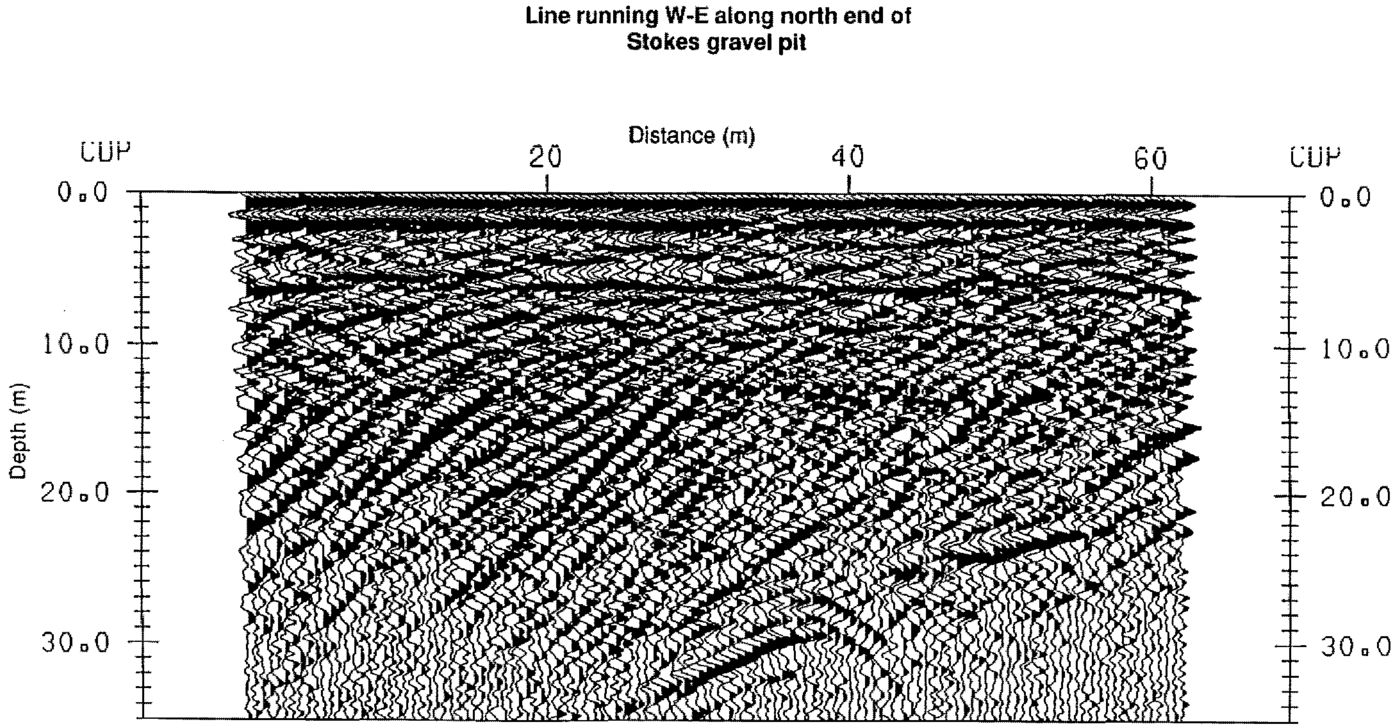


Figure 6. Ground penetrating radar profile of foreset bedded gravel and sand in the Stokes municipal gravel pit, Surrey. A strong horizontal reflection at 5-6 m depth corresponds to the water table. Note the foreset reflectors passing through the water table, particularly on the right side of the profile. Foreset dip is exaggerated in the profile; true dip is about 10°. The two strong reflectors in the upper 1-2 m correspond to direct air and sediment arrivals. From Rea et al. (1994b), Figure 4.

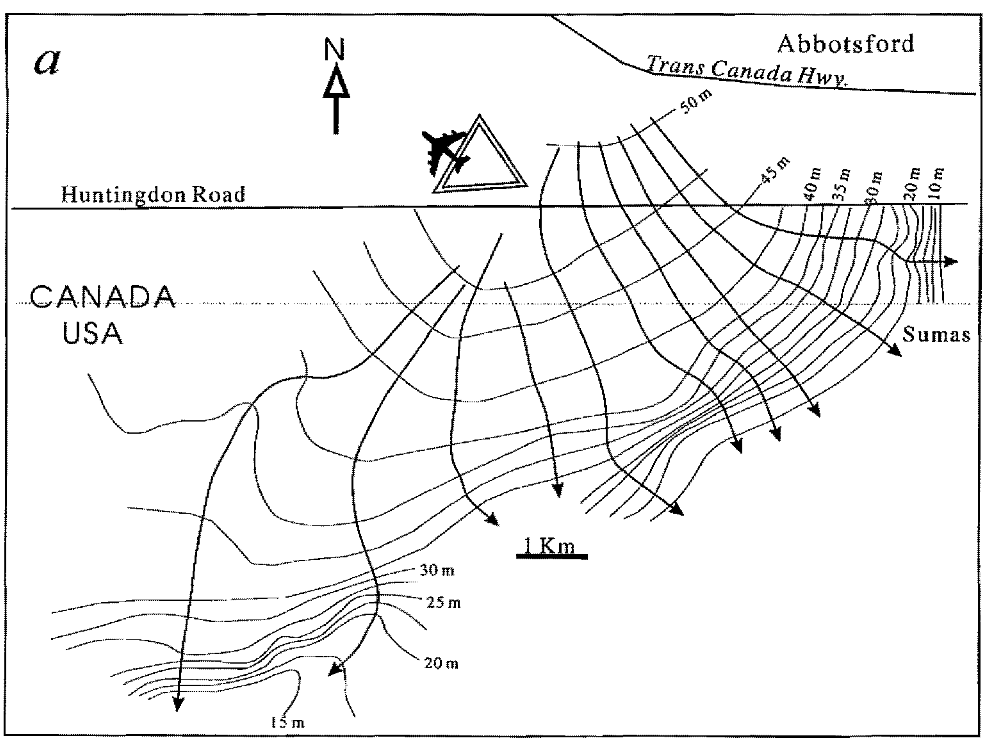


Figure 7.

*Groundwater surface contour maps for: **a)** Abbotsford aquifer. All water well data has been corrected for seasonal fluctuations in static level, using static level data from observation wells. **b)** Brookswood aquifer. Two water table maps constructed from water well data at two different times of the year, corresponde to winter recharge (January), and the summer dry period (July-August-September). Static level data was taken from wells drilled between about 1960 and 1985. Although the general pattern of quasi-radial groundwater flow does not change significantly during the year, actual static levels fluctuate about 3 m (10 feet). Contours in feet above sea level. British Columbia Ministry of the Environment CMOE 1:5000 water well location map numbers are indicated, for example 007.2.3.*

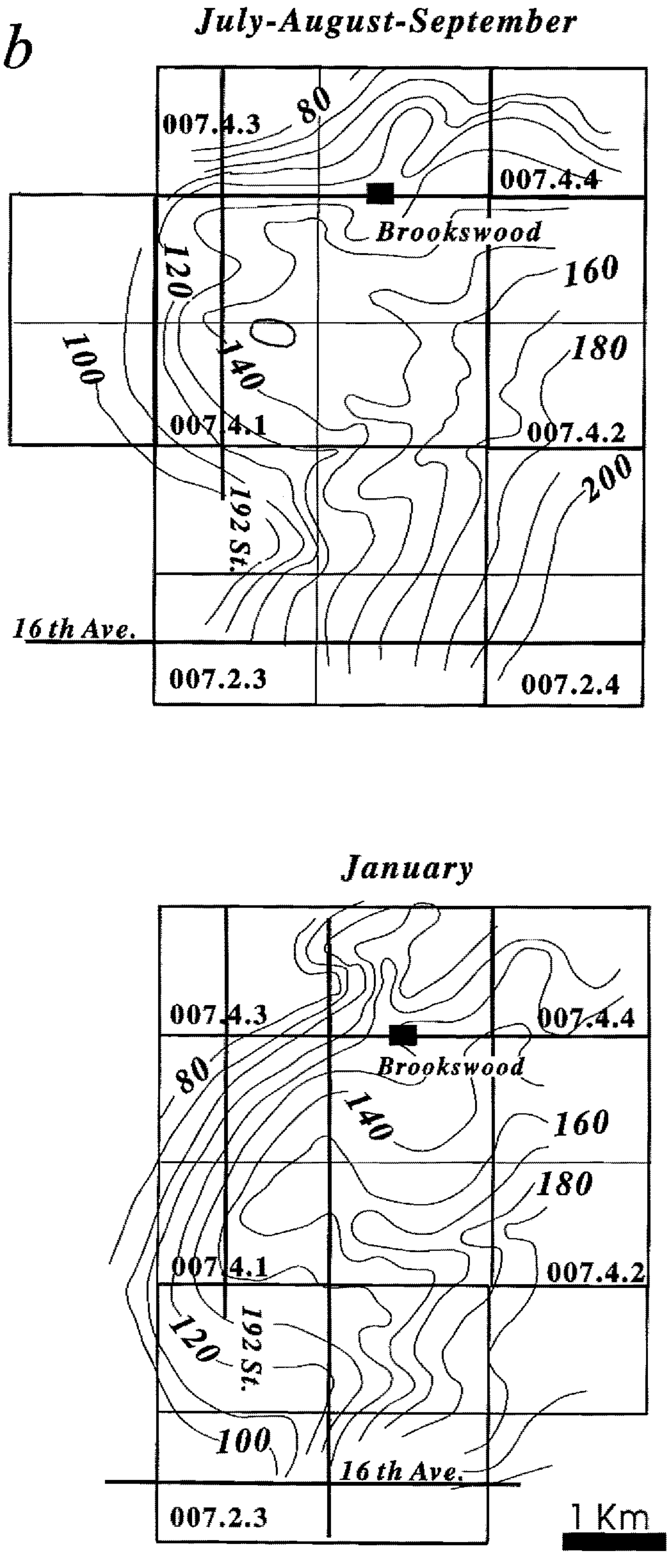


Figure 7. (cont.)

The Quaternary succession

Most of the groundwater in the greater Vancouver region is derived from extensive Pleistocene and Holocene deposits in the Fraser Valley and from valley fills in the adjacent Coast Mountains (Halstead, 1977). The knowledge of the lithostratigraphy of the Quaternary succession has evolved over many years of study to a fairly sophisticated degree, through mapping of surficial deposits (Armstrong, 1980a, b; Armstrong and Hicock, 1980a, b), and stratigraphic studies (Halstead, 1960; Clague, 1976, 1991; Armstrong and Clague, 1977; Clague and Luternauer, 1982). Extension of this lithostratigraphic scheme into the subsurface is difficult except for some areas of shallow post-Vashon units (for a summary of formal stratigraphy see Armstrong, 1984). However, considerable progress in this direction has been made by E.C. Halstead using information from a number of deep test wells (Halstead, 1957, 1959, 1961, 1978).

At least three major glacial and interglacial periods are recorded in the Fraser Valley. The Westlynn Glaciation, of possible Middle or Early Pleistocene age, is the oldest known glacial event. The Semiahmoo Glaciation is probably Early Wisconsin in age and is too old for ^{14}C dating. The youngest glaciation, known as the Fraser Glaciation, occurred between approximately 26 000 and 10 000 BP. Exposures of deposits older than Fraser Glaciation are confined to a few coastal cliffs and incised stream valleys; all of the surface exposures in the Fraser Valley are Late Wisconsin and Holocene in age.

Perhaps the most characteristic attribute of all the Quaternary deposits is their marked lateral and vertical lithological complexity at all scales of mapping. Herein lies one of the main difficulties for extending the lithostratigraphic scheme into the subsurface. This complexity is a direct result of the interactions between sedimentation and erosion during advance and retreat of the ice sheets, the concomitant retreat and advance of the seas, and the isostatic effects of ice loading and unloading that resulted in subsidence and uplift respectively. As a consequence, all of the deposits, whether of glacial or interglacial origin, contain units of sufficiently high porosity and hydraulic conductivity to qualify as aquifers; more than 200 aquifers have been identified in the region (Liebscher et al., 1992).

Armstrong (1984) identified three main types of Quaternary deposit: waterlain or bedload deposits typically formed in glaciofluvial or glaciomarine settings (e.g., deltas and outwash fans); diamictons deposited directly by glacial ice; and glaciomarine and glaciolacustrine deposits formed by fallout of sediment from ice into water. The largest aquifers in the Fraser Valley result from fluvial and glaciofluvial deposition. Porous units of more limited thickness and areal extent also are common within relatively impermeable diamictons.

Because of the stratigraphic and lithological variability of Quaternary deposits in the Fraser Lowland, Halstead (1986) subdivided the succession into 5 hydrostratigraphic units, that are different from the formal lithostratigraphic units (summarized in Table 1). These units were defined on the basis of lithology, permeability, and porosity, and on subordinate factors such as origin (marine, fluvial), stratigraphic position, and to some extent on aquifer type (e.g., water table aquifers in unit C). To some extent, this scheme mixes lithostratigraphic criteria and hydraulic properties. For example hydrostratigraphic unit A was largely used for glaciomarine clays and stoney clays in Fort Langley and younger formations. However, there is no reason to suppose that type A hydrostratigraphic units are not present in deeper parts of the Quaternary succession, although determination of their origin might be difficult from borehole data alone. Unit E on the other hand, corresponding to sediments older than the Fraser Glaciation, contains lithologies identical to other 'Fraser' units; it is distinguished more by age and in places higher dissolved solid content (presumably because of longer residence times). In general, hydrostratigraphic units should be defined on lithological and hydraulic properties alone, so they can be applied to any part of the succession.

Unconfined aquifers in Sumas Drift

Aquifers of hydrostratigraphic unit C are the most prolific and most utilized in the Fraser Valley. The Brookswood and Abbotsford aquifers are the largest unconfined aquifers in the region and serve as good examples of groundwater systems whose integrity has been compromised by anthropogenic activities. Abbotsford aquifer is the largest of these, covering an area of about 200 km^2 (about 100 km^2 in British Columbia and 100 km^2 in Washington State). The Brookswood aquifer about covers 50 km^2.

Both aquifers occur within Sumas Drift (Armstrong, 1981), which was deposited during a brief glacial readvance near the end of the Fraser Glaciation. Although dominated by sand and gravel, the aquifers have complex stratigraphy and sedimentology. This is apparent in gravel pit exposures and closely spaced drill records where stratigraphic sections exhibit substantial differences in grain size, compaction, and the presence of nonpermeable lenses of lodgement and flow till or stony clay.

Both Brookswood and Abbotsford aquifers are recharged by a combination of direct precipitation, infiltration from streams (seasonally dependent), and surface runoff from adjacent uplands. Hydrographs from the two aquifers show a close correspondence between winter precipitation and high static levels in observation wells (Fig. 4). Well production rates up to 30 L/s are common, and reach 150 L/s in some community wells. High capacity industrial wells like those used by the Fraser Valley Trout Hatchery in the Abbotsford aquifer, that pump more than 4 000 000 m^3/a, result in cones of depression extending in excess of a kilometre from well heads. One estimate of groundwater budgets in the Abbotsford aquifer, taking into account precipitation, evaporation, evapotranspiration, and surface runoff, indicates that groundwater extraction in 1985 was less than the total water recharged to the aquifer (Kohut, 1987).

Common sedimentological features in the Sumas deposits are fan deltas (several metres thick), characterized by large-scale, sand and gravel foresets that dip up to 20°, and cross-bedded top sets (Fig. 5). Some foresets are draped by mud veneers. Fluvial channels, mostly of braided glacial outwash origin, are filled with coarse, close-packed gravel, multiple sets of dune-like crossbeds representing within-channel bars, or mud and silt (e.g., abandoned channel fills). Large foresets of sand and gravel, extending more than 200 m laterally in the Brookswood area, have also been identified in the subsurface by ground penetrating radar (Fig. 6; Rea et al., 1994a, b). Some shallow-dipping sand-mud foreset couplets contain lenticular and flaser ripple bedding, providing some evidence of weak tidal currents. This combination of sedimentary structures may represent deposition in an estuarine environment (Ricketts and Jackson, 1994).

Groundwater flow in the Abbotsford aquifer has a radial pattern (Fig. 7a). Flow directions are reasonably well determined south of the airport in Abbotsford aquifer, but are poorly known north farther north. The piezometric surface map for Abbotsford aquifer (Liebscher et al., 1992) has been corrected for seasonal fluctuations in the water table, which (in a small number of observation wells) can be as high as 3-4 m. For comparison, the Brookswood aquifer maps have been constructed by taking static levels in wells for specific time periods, July-August-September in one and January the other (Fig. 7b), from drill records covering several years. Variations in static levels here can be great as 5 m according to hydrographs of observation wells in Surrey (Halstead, 1986). Generally radial flow patterns extend from the south-western sector to the northwest for the Brookswood aquifer. The western and southwestern areas are bordered by clay uplands (Fort Langley Formation) and are areas of recharge.

SUMMARY

Understanding geological processes is the key to understanding the fate of groundwater. The flow of groundwater through fractured rock or nonlithified sediment is governed by many factors including: the physical boundaries of an aquifer or aquitard, the stratigraphy, elements of fabric such as porosity, permeability, and sedimentary structures, and hydraulic head. All of these factors are mappable quantities. Likewise, an understanding of the fate of human-derived contamination in groundwater systems is strongly dependent on the geological factors controlling groundwater flow. In the Fraser Valley, the map limits of most unconfined aquifers are reasonably well known. This is not the case for many of the confined aquifers in the region. In fact there are very few published maps or models in the Fraser Valley area showing potentiometric surfaces, groundwater flow, or (natural) chemical gradients.

With the domestic, industrial, and agricultural demands for groundwater increasing, and concomitant levels of contamination rising, there is an obvious need for aquifer characterization, and groundwater mapping and modelling at both regional scales and at the scale of single aquifers. Federal, provincial, and municipal agencies have begun to address these problems. Environment Canada has a continuing program for evaluating nitrate and pesticide contamination in the trans-boundary Abbotsford aquifer (e.g., Liebscher et al., 1992). The B.C. Ministry of Environment, Lands and Parks also has begun to assess the risk contamination of of groundwater human impact (e.g., Kreye and Wei, 1994). Regional mapping and aquifer characterization in the Fraser Valley is also being undertaken by the Geological Survey of Canada. All of these projects will increase our knowledge of groundwater resources in Vancouver, and provide the basis for expedient and cost-effective remediation programs when the need arises.

ACKNOWLEDGMENTS

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A preliminary directory of trace element databases available in the Vancouver map area

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Delaney, T.A. and Turner, R.J.W., 1994: A preliminary directory of trace element databases available in the Vancouver map area; in Geology and Geological Hazards of the Vancouver Region, Southwestern British Columbia, (ed.) J.W.H. Monger; Geological Survey of Canada, Bulletin 481, p. 299-316.

Abstract: A preliminary directory of trace element data from rocks, soils, vegetation and stream, lake, and marine waters and sediments, as well as from water supply and sewerage systems is compiled for the Vancouver NTS map area (92G). The elements include Ag, Al, As, Au, Ba, Be, Bi, Cd, Co, Cr, Cu, F, Fe, Hg, Mg, Mn, Mo, Ni, Pb, Sb, Se, Sn, Sr, U, V, W, and Zn. Included in the directory are data sets collected for geochemical characterization of (1) natural background concentrations (background studies); (2) baseline concentrations in areas of possible contamination (baseline studies); (3) areas of known contamination (impact studies); and (4) variation over time (monitoring studies). Available data are concentrated in the immediate area of metropolitan Vancouver, the Lower Fraser River valley and, to a lesser extent, Howe Sound and the Strait of Georgia.

Résumé : On a compilé pour le secteur de la carte de Vancouver (SNRC 92G) un répertoire préliminaire des éléments traces dans les roches, les sols, la végétation, les sédiments et eaux fluviatiles, lacustres et marins, et les réseaux d'adduction d'eau et d'égouts. Les éléments comprennent Ag, Al, As, Au, Ba, Be, Bi, Cd, Co, Cr, Cu, F, Fe, Hg, Mg, Mn, Mo, Ni, Pb, Sb, Se, Sn, Sr, U, V, W et Zn. On a inclu dans le répertoire des ensembles de données recueillis en vue de la caractérisation géochimique (1) des concentrations naturelles (études de base); (2) des concentrations de base dans les secteurs de contamination possible (études préliminaires); (3) des secteurs de contamination connue (études d'impact); et (4) de la variation en fonction du temps (études de contrôle). Les données disponibles sont concentrées dans la région métropolitaine de Vancouver, la basse vallée du Fraser et, dans une moindre mesure, le détroit de Howe et le détroit de Georgia.

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INTRODUCTION

Widespread introduction of trace elements to air, land, water, and vegetation by human activity is one of the most important environmental issues facing the world today (Moore, 1991). Trace element pollution is greatest near densely populated urban centres, as pollution control technologies are rarely adequate to clean municipal and industrial wastes of harmful contaminants prior to their release into the environment. Despite growing awareness of their toxicity to humans and other organisms, huge quantities of these contaminants are discharged annually.

Because high concentrations of certain elements in the surficial environment may damage ecosystems and pose human health risks, it is important to determine natural background levels of these elements and to monitor the environment for possible contamination. In the Vancouver region, most geochemical studies fall under the mandate of a host of municipal, regional district, provincial, and federal agencies. Such an abundance of data collectors hinders integration and use of this geochemical information. Therefore, the initial step toward understanding the effects of trace elements is to provide an inventory of what is known about their distribution in the local environment.

The purpose of this directory is to make available trace element geochemical data on rocks, sediments (lake, stream, and marine), waters (lake, stream, and marine), soil, and vegetation for the Vancouver map area (NTS 92G). Elements in this compilation include Ag, Al, As, Au, Ba, Be, Bi, Cd, Co, Cr, Cu, F, Fe, Hg, Mg, Mn, Mo, Ni, Pb, Sb, Se, Sn, Sr, U, V, W, and Zn. The term "trace element" is used loosely in this paper to include elements such as aluminum, iron, and manganese even though these elements are generally too abundant to be described as occurring in trace quantities. It is important, however, to provide a complete list of elements included in each study.

This preliminary report is part of a longer term project to make an inventory of geochemical data in the Georgia Basin area and to assess where additional work is required. A brief discussion of the geochemical landscape of the Vancouver area is followed by a description and discussion of the directory.

GEOCHEMICAL LANDSCAPE OF THE VANCOUVER MAP AREA

The Vancouver map area includes five important physiographic elements: (1) the southernmost part of the Coast Mountains; (2) the Fraser Lowland south to the U.S. border; (3) the northern end of the Cascade Mountains; (4) the central Strait of Georgia; and (5) a small portion of eastern Vancouver Island (Fig. 1). The Fraser Lowland is comprised of a broad plain and rolling uplands of fluvial and glacial deposits flanking the lower reaches of the Fraser River. The Cascade Mountains bound these lowlands to the southeast, while to the south the lowland extends across the Canada-U.S. international border, which is the southern boundary of the map area. A large delta and estuary have formed where the Fraser River empties into the Strait of Georgia. The southern Coast Mountains cover the northern half of the map area, rising abruptly from the lowland and Strait to glacier-covered summits over 1500 m high. Deep glaciated valleys cut the Coast Range; to the east they contain large lakes, whereas on the coast they are flooded by the ocean to form steep-walled inlets and fiords.

Moderate temperatures, high amounts of winter precipitation (1000-5000 $mm{\cdot}a^{-1}$) and dry summers characterize the climate of the area (Hay and Oke, 1973). Rivers are fed by snow melt runoff that peaks in May and June, by glacial meltwaters which peak in July and August, and by autumn rainstorms that cause floods. Snowfall is rare in the Fraser Lowland but 8-15 m accumulates annually in the mountains to the north and southeast.

Over one half of the population of British Columbia lives in the Fraser Lowland. Metropolitan Vancouver currently surrounds Burrard Inlet, extends part way up the mountains, and spreads south and east across the Fraser Lowlands. Areal coverage of the city will likely expand in the future as the population is projected to increase by 600 000 over the next 20 years (Environment Canada, 1992).

Geology

The Vancouver map area is underlain by six major types of rock and sediment (Roddick and Woodsworth, 1979; Armstrong, 1984; Monger, 1990, 1991). (1) Granitic to dioritic intrusive rocks comprise the dominant rock unit in the Coast Mountains. (2) Metamorphosed volcanic and sedimentary rocks occur in the Cascade Mountains and to a lesser extent in the Coast Mountains. (3) Unmetamorphosed sedimentary and local volcanic rocks of Cretaceous and Tertiary age are exposed along the northern margin of the Fraser Lowlands. (4) Quaternary mafic to intermediate composition volcanic rocks are widespread in the Mount Garibaldi area. (5) Quaternary unconsolidated or partially consolidated glacial sediments were deposited throughout the map area. (6) Postglacial estuarine and marine sediments are derived from the Fraser River and other lesser river and stream systems.

The bedrock is the primary natural source of trace elements to the environment. The breakdown of rocks due to weathering releases a variety of elements, either as dissolved ions or in particulate form, into soils and streams that eventually make their way to the ocean (Fig. 2). Although natural concentrations of trace elements in rocks are generally low, higher concentrations have been created locally by past geological processes. In the Coast Mountains, exposed metamorphosed volcanic rocks contain numerous natural metal sulphide occurrences including the giant Fe-Cu-Zn deposit at Britannia Beach on Howe Sound, Fe-Cu-Zn deposits in the Indian River and Harrison Lake areas, Pb-Zn-Ag deposits northwest of Squamish, and a low grade Cu-Mo deposit on Gambier Island in Howe Sound (MINFILE, 1990). Weathering naturally erodes such deposits and this may cause anomalous metal concentrations in nearby streams. Mining operations can intensify these processes by exposing larger amounts of rock to weathering processes.

Soils

Soils in the map area vary from poorly developed regosols on steep slopes prone to mass wasting, to highly developed podzols in well drained forests, to gleysols in poorly drained areas (Luttmerding, 1981b). Intense decomposition and leaching of organic and mineral matter in the upper layers of podzol releases organic acids that mobilize metals and result in the depletion of metals in upper soil horizons. Mobilized metals are precipitated with secondary clay minerals, iron and manganese hydrous oxides, and organic matter in the underlying soil layer. Metal contents can also be elevated in the organic litter at the surface due to concentration in plants. Mobility of metals in soils is dependent mainly on Eh, pH, and the availability of inorganic and organic complexes (Levinson, 1974).

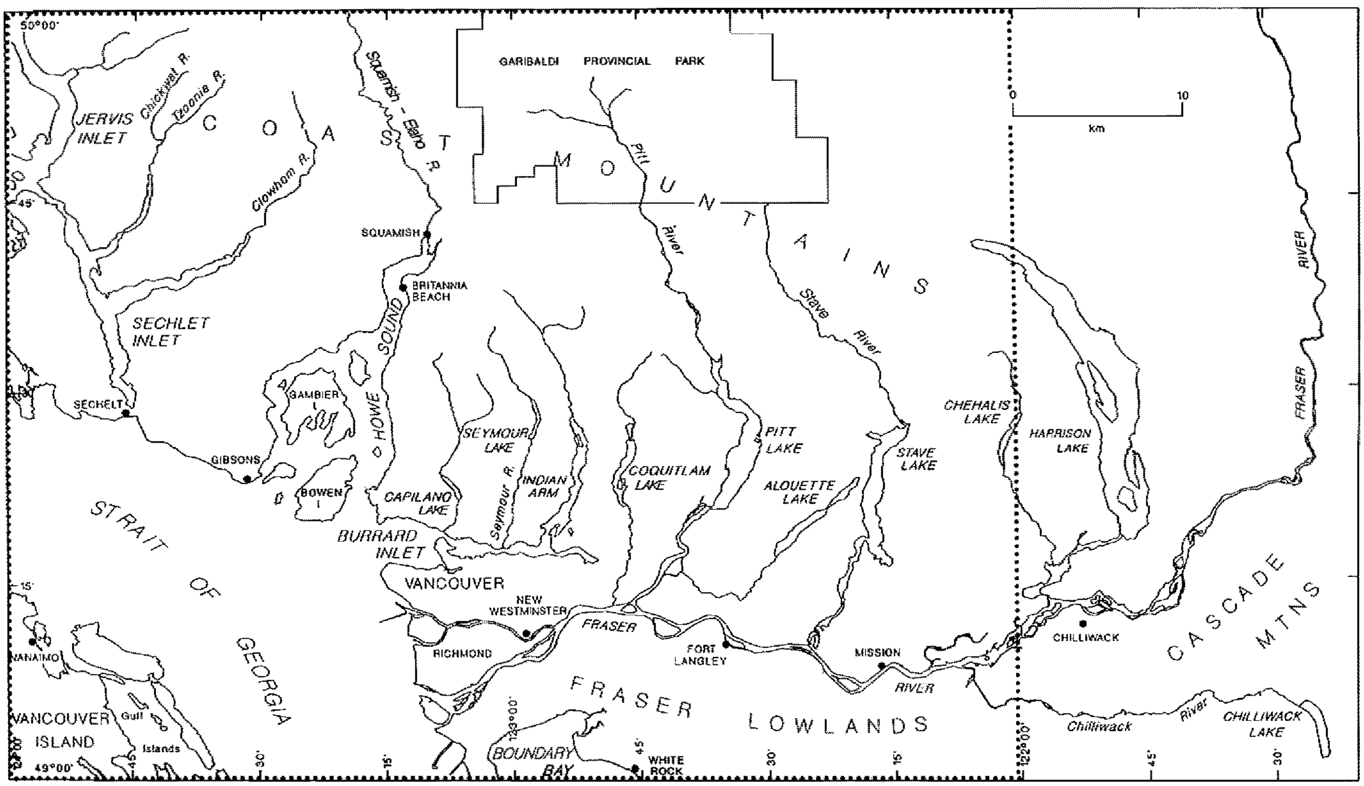

Figure 1. Map of the Vancouver area showing the boundary (dotted line) of the Vancouver (NTS 92G) map area.

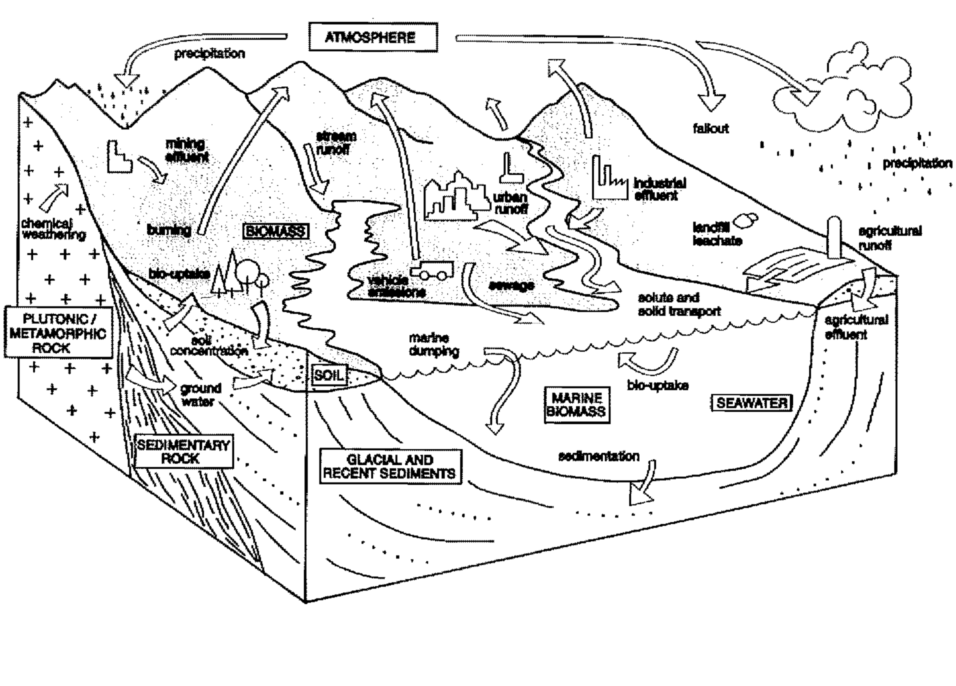

Figure 2.

The geochemical landscape of the Vancouver map area illustrating the cycling of metals between landscape components.

Erosion, acidification, and compaction of soils and subsidence of organic soils are major sources of soil degradation whereas trace element contamination is generally a lesser problem, except in urban areas (Environment Canada, 1992). Loading of trace elements occurs from landfill leachates, land application of waste treatment sludge, and more broadly through deposition of lead from automobile exhaust and other sources of air pollution (Fig. 2). Soils in the Vancouver region are vulnerable to acidic deposition from the atmosphere relative to other parts of British Columbia because of their low capacity for neutralizing acids (Environment Canada, 1993).

Vegetation

A number of biogeoclimatic zones occur within the Vancouver map area, controlled by elevation and distance from the moderating effects of the ocean. Zones vary from wet maritime forest to subalpine forests to alpine tundra (Meidinger and Pojar, 1991). Low coastal areas northwest of Vancouver and in the Gulf Islands lie within the influence of the rain shadow of Vancouver Island and support a drier maritime forest.

Trace elements are incorporated into plants primarily through the root systems and distributed to other parts of the plant depending on the plant species (Levinson, 1974). As the leaves and other plant organs fall to the ground and decay, soluble components enter surface or ground water (Fig. 2). Insoluble constituents are incorporated into the soil as organic matter. Thus, both vegetation and soils can act as sinks for metals.

Streams

Most streams in the Vancouver map area are located in mountain valleys and flow southward into the Fraser River or coastal inlets. Surface waters in the coastal area of southwestern British Columbia have a particularly low buffering capacity making them susceptible to acidification (Environment Canada, 1993). This is due to the predominance of granitic bedrock which is resistant to chemical weathering and produces surface waters with low alkalinity. The pH of streams and lakes can be lowered by acidic precipitation due to short-and long-range transport of airborne contaminants (e.g., sulphur and nitrogen oxides) from natural gas processing plants, pulp mills, oil refineries, power plants, and automobiles in the Vancouver metropolitan area (Fig. 2).

Trace elements in streams are derived primarily from bedrock, sediments, and soils, and are transported as dissolved species, adsorbed onto iron and manganese hydrous oxide coatings on detrital grains, or as particulates (Grieve and Fletcher, 1976). Mining activities can cause loading of metals into nearby drainages. For example, during early development of the Britannia Mine, mine tailings were discharged directly into Britannia Creek (Drysdale, 1990), and metal-bearing acid mine drainage continues to enter the creek today (BCMEMPR, 1991).

Fraser River and estuary

The Fraser River system drains much of the southern and central interior of British Columbia and most of the sediment in the lower Fraser River is derived from tributary drainages upstream of the Vancouver map area. Below New Westminster, the river is influenced by tidal fluctuations and the intrusion of a wedge of dense saline water along the river channel floor which significantly retards the flushing action of the river. The front of the Fraser delta, which receives silt and sand from the river is advancing westward at a rate of about 4.5 $m \cdot a^{-1}$ (Stewart and Tassone, 1989).

Several factors influence water quality in the lower Fraser River: seasonal change in discharge, sediment load, tidal movement, and intrusion of salt water from the Strait of Georgia; natural inputs of nutrients and metals; and inputs from municipal, industrial, domestic and agricultural activities. Swain and Walton (1988) noted higher trace element concentrations associated with the silt and clay fractions of Fraser River sediment. This fine grained fraction enriched in trace elements tends to be deposited on distal marine portions of the Fraser delta (Grieve and Fletcher, 1976). The high rate of sediment deposition on submarine slopes of the delta tends to dilute trace element concentrations. With time, trace element-enriched sediments may be separated from the benthic ecosystem by continued sedimentation and burial.

Major anthropogenic sources of trace elements to the lower Fraser River include domestic discharges, industrial effluent, and urban runoff (Environment Canada, 1992). Smaller inputs include accidental spills, contaminated groundwater flows, leachates from landfills, and discharges from boats. Contamination of the river occurs not only in the Vancouver region but upstream as well, and the amount of material introduced has increased as the population in central British Columbia has grown. Between 1965 and 1985, input of municipal effluents into the Fraser River and its tributaries upstream from Vancouver increased 4 times whereas industrial effluents increased 429 times (Environment Canada, 1992).

Marine environment

Marine areas within the Vancouver map area include the central Strait of Georgia, the coastal inlets of Howe Sound and Sechelt Inlet, and heavily industrialized Burrard Inlet. Influx of trace elements to marine systems from both natural and anthropogenic sources occurs via stream and river water and sediment, runoff directly from land, precipitation, and atmospheric fallout (Fig. 2). Marine sediment can be a major sink for some trace elements in marine systems if contaminated material is buried by other sediment inputs, isolating them from the marine ecosystem. Burial is most rapid in areas with high sedimentation rates such as near the mouths of the Fraser and Squamish rivers and can be an important process for removing from marine waters elements such as lead which strongly bind to particles. In contrast, elements such as zinc and copper that weakly bind to particles, and elements that remain dissolved such as cadmium, stay in the marine watermass for much longer before being sequestered in sediments. Locally, recycling of buried sediments into the marine biomass and overlying waters can occur through wave reworking in shallow water areas, slope failure and resuspension of sediment, and bioturbation by benthic infauna. Dredging can also remobilize metals in the water column by disturbing contaminated material (Lay, 1993).

Many trace elements are concentrated in the fine clay fraction of sediment by adsorption onto clay minerals, oxides, and organic matter (Grieve and Fletcher, 1976). The biological availability of metals in sediments depends on the chemical speciation of the metal, and the chemical and mineral composition, crystallinity, and surface area of the sediment (Demayo et al., 1978). The reducing nature of marine sediments tends to fix many metals as insoluble metal sulphides (e.g., Fe, Cu, Pb, Zn). Manganese and iron, in the form of high surface area oxides and oxyhydroxides, tend to concentrate at the redox boundary at or just below the seafloor where they scavenge and therefore concentrate other metals.

Human activity contributes trace elements directly to marine environments through sewage outfalls, ocean dumping, sulphide mine tailings, and port and harbour activities. Pulp and paper mills are no longer a major contributor of metals into the marine environment since the use of both mercury and zinc in paper production was discontinued in 1960 and 1974, respectively (Kay, 1989).

Sources of trace elements into Burrard Inlet and the Fraser River include past and present industrial discharge, aerial emissions, marinas, marine traffic, urban runoff, ship building and repair facilities, ore concentrate loading docks, and sewer discharge. In the Vancouver area, there are four sewage outfall locations: Iona Island, Annacis and Lulu islands in the Fraser Estuary, and in Burrard Inlet near the Lion's Gate Bridge. Each of these sites are monitored for metals and metal concentrations by the Greater Vancouver Regional District (Cain and Swain, 1980; Fanning et al., 1989). Water, sediment, and biota near outfalls is generally elevated in metal content (Bawden et al., 1973; Parsons et al., 1973; Grieve and Fletcher, 1976). Secondary sewage treatment facilities at both Annacis and Lulu islands remove most trace metals from the effluent (Environment Canada, 1992). The highest metal loadings occur near the Iona Island outfall as it has only a primary treatment facility. Daily discharge of metals at the Iona outfall is estimated to be 0.6 kg cadmium, 14.7 kg lead, 50.5 kg copper, and 45.9 kg zinc (Kay, 1989).

Of the six major ocean dump sites in the Vancouver map area, the largest quantities of material are deposited at a site just west of Point Grey, and at the Sandshead site offshore from Lulu Island at the mouth of the Fraser River. Material disposed is primarily dredge spoils (mostly sand and gravel) and wood wastes usually dredged near forest operations (Kay, 1989). Allowable levels of contaminants for material for ocean disposal are set by the Canadian Environmental Protection Act (CEPA) and Environment Canada is responsible for enforcement of these guidelines. Monitoring of dump sites indicates that most material does not adversely effect the marine biota (Packman, 1980; Brothers et al., 1985).

The upper part of Howe Sound adjacent to the Britannia Mine is a site of significant metal contamination. The mine was open from the early 1920s to 1975 during which time mine tailings rich in iron, copper, zinc, lead, and silver sulphides were disposed of in Britannia Creek and the near shore areas of Howe Sound. Acidic waters still drain from the mine (Drysdale, 1990). This metal loading has resulted in elevated copper and zinc contents in the water column (van Aggelen and Moore, 1986) and in the biota (Percival et al., 1992; McDaniel, 1973), and severely damages benthic and intertidal life. Drysdale (1990) found that although concentrations of iron, copper, lead, zinc, and barium in marine sediments remain elevated, levels in the shallowest sediments have decreased in the 13 years since cessation of tailings disposal. This decrease in metal content reflects the continuing burial of the tailings by sediment from the Squamish River to the north.

Atmospheric environment

In the Fraser Valley, airflow and the dispersion of airborne pollutants is controlled by physiography and seasonal weather variations. Rainfall tends to reduce air pollution by transferring particulates from the air to the land and water. Strong winds, though uncommon in the valley, are related to low pressure systems characteristic of cold fronts or winter storms, and tend to flush the valley of pollutants. During the summer, temperature differentials between land and sea produce light breezes that also help clear the air. However, high pressure systems, which occur in the spring, fall, and summer produce weak pressure gradients, so that vertical mixing occurs only within 100 to 500 m of the ground surface. More stable air above traps the underlying polluted air against the mountains. Such inversions, which are most common in the fall and winter, can last weeks and greatly reduce air quality.

Major sources of air pollution include motor vehicles, petroleum and forest industries, municipal incinerators, home heating, and outdoor fires (Fig. 2). Motor vehicles account for nearly 85% of total air pollution loadings in the lower Fraser Valley area and have been the primary source of suspended lead particulates (Environment Canada, 1992). Levels of suspended lead particulates have decreased with the sharp decline in use of leaded gasoline.

There is one mass-burn waste incinerator in the Vancouver area which has been operated since 1988. Output from the incinerator is monitored for antimony, arsenic, cadmium, cobalt, chromium, copper, lead, nickel, selenium, vanadium, and zinc on a quarterly basis (Greater Vancouver Regional District, 1992c). From 1988 through 1991, only mercury exceeded the recommended emission limits set for the incinerator by the B.C. Ministry of the Environment, Lands and Parks. Mercury emissions have since been reduced to less than one third the allowable limit by addition of a mercury abatement system (Greater Vancouver Regional District, 1992c). To further assess the impact of the incinerator, the Greater Vancouver Regional District monitored metal contents in air, soils, and vegetation in the surrounding area prior to and after operation of the incinerator. Trace element concentrations showed no increase between 1988 and 1990 as a result of operation of the incinerator (Greater Vancouver Regional District, 1992a, b).

DESCRIPTION OF THE DIRECTORY

This directory consists primarily of information from municipal, provincial, and federal government agencies including the Greater Vancouver Regional District (GVRD), the City of Vancouver Engineering Department, the Environmental Protection and Water Quality Branches of the B.C. Ministry of Environment (BCMEMPR), Lands and Parks (BCELP), the Geological Survey Branch of the B.C. Ministry of Energy, Mines and Petroleum Resources, the Institute of Ocean Sciences (IOS) of the Department of Fisheries and Oceans, Environment Canada, and the Geological Survey of Canada (Table 1). Information was also collected from the Burrard Inlet Environmental Action Program (BIEAP), the Fraser River Estuary Monitoring Program (FREMP), and Westwater Research Centre, co-operative organizations focusing on environmental management within the Vancouver area. Additional databases were taken from references found in the Bibliography and Index of Geology (GEOREF) and the Enviro/Energyline abstracts, both CD-ROM electronic databases, the Decade of North American Geology "Geology of Canada" volumes copublished by the Geological Society of America and the Geological Survey of Canada, and from "A Bibliography of Scientific Information on Fraser River Basin Environmental Quality" (Missler, 1992). Information collected privately by industry was considered to be beyond the scope of this study. It is possible that the database can be expanded in the future to include this information.

An early draft of this report was sent to personnel at the organizations listed above, with the exception of BIEAP, to insure the accuracy of the information presented and to obtain information on additional databases. A copy was also sent to officials at the Coast Garibaldi Health Unit in Gibsons British Columbia and the Vancouver Health Department, and to one environmental consulting firm, Steffen, Robertson & Kirsten (Canada) Inc. Comments were received from all organizations solicited and incorporated in the database.

Geochemical trace element databases listed in Table 1 are organized by sample medium: stream, lake or city water; stream or lake sediment; marine water; marine sediment; soils; vegetation; rock; and sewage. The first column of Table 1, labelled "Map no.", refers to Figures 3 and 4 on which the geographic extent of each database is outlined. Other columns include the source of the data, purpose of sampling, date of sampling, sample type, areal coverage of the studies, number of sample sites, sampling density and frequency, analytical laboratory, trace element analyses, computer

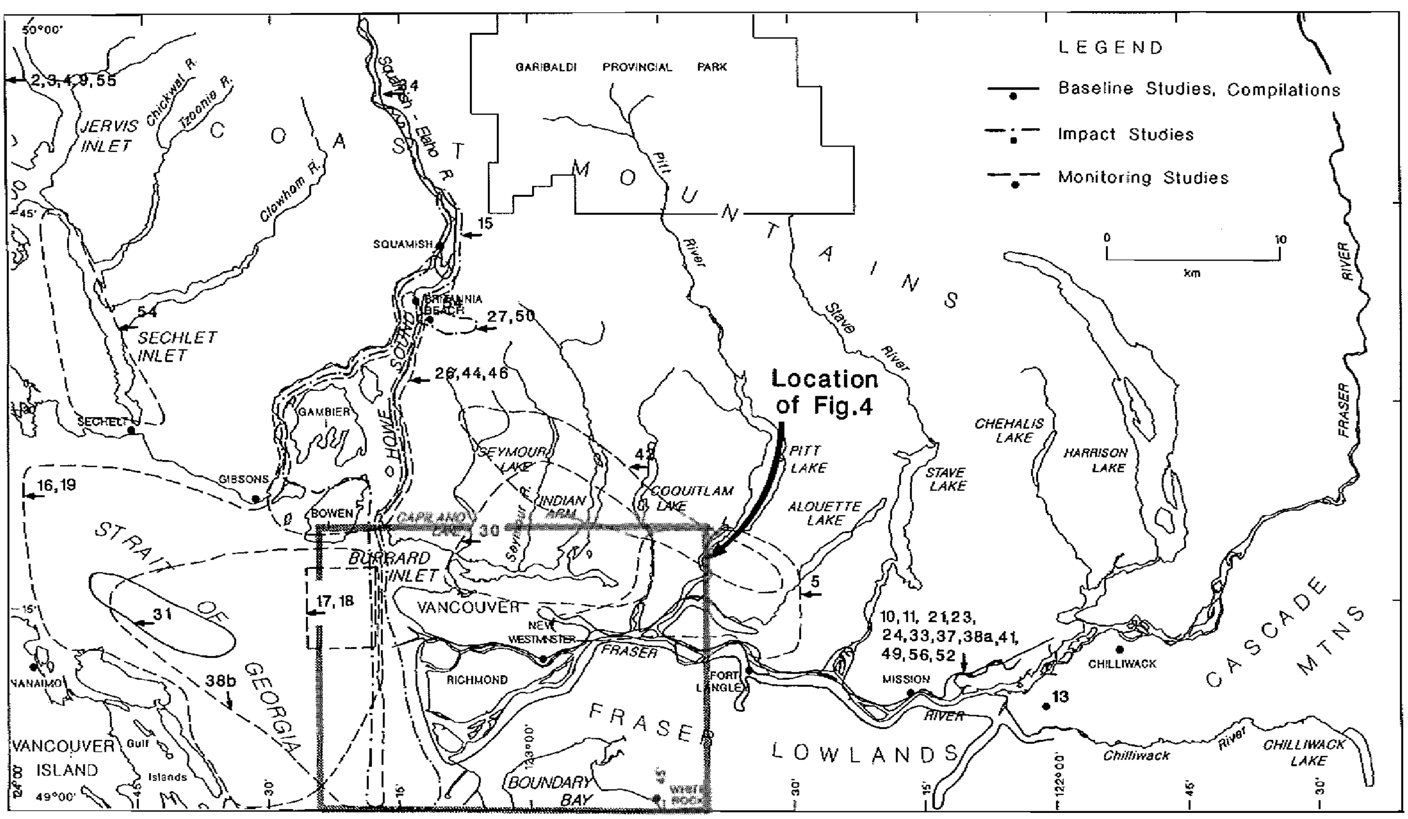

Figure 3. Location of geochemical data sets described in Table 1. Geochemical data sets in the area immediate to Vancouver are shown on Figure 4.

format in which the data is available and the references containing or discussing the data. Addresses of agencies that supplied data are listed in Table 3 and acronyms of analytical laboratories are listed in Table 4.

Table 2 summarizes trace element sampling activities in the Vancouver map sheet, allowing the reader to quickly evaluate which organizations are involved in different types of studies and what sampling media are being used. The table is organized according to the purpose of the study: characterization of natural background metal concentrations (background); determination of baseline metal concentrations (baseline); impact assessment (impact); or monitoring of metal concentrations over time (monitoring). The second column lists the types of samples being collected and column three lists the organizations that are involved in each type of sampling. The general status of trace element geochemical sampling within the Vancouver map area is discussed below.

Background studies

In this report, the term "background study " refers to sampling programs designed to determine natural levels of trace elements. Such sampling is conducted using media not directly impacted by human activity. Although all areas are effected to some degree by atmospheric loadings of pollution, background studies make an effort to avoid major sources of human contamination. It is important to note that natural concentrations of trace elements are highly variable due to the variety of geological processes by which they can be concentrated. Elevated levels do not require an anthropogenic source. It is necessary, therefore, to determine natural background concentrations of an element before the impact of human activities can be assessed.

Background metal concentration data throughout the entire Vancouver map area is only available for stream sediments and stream and lake waters (Table 1, Fig. 3). A

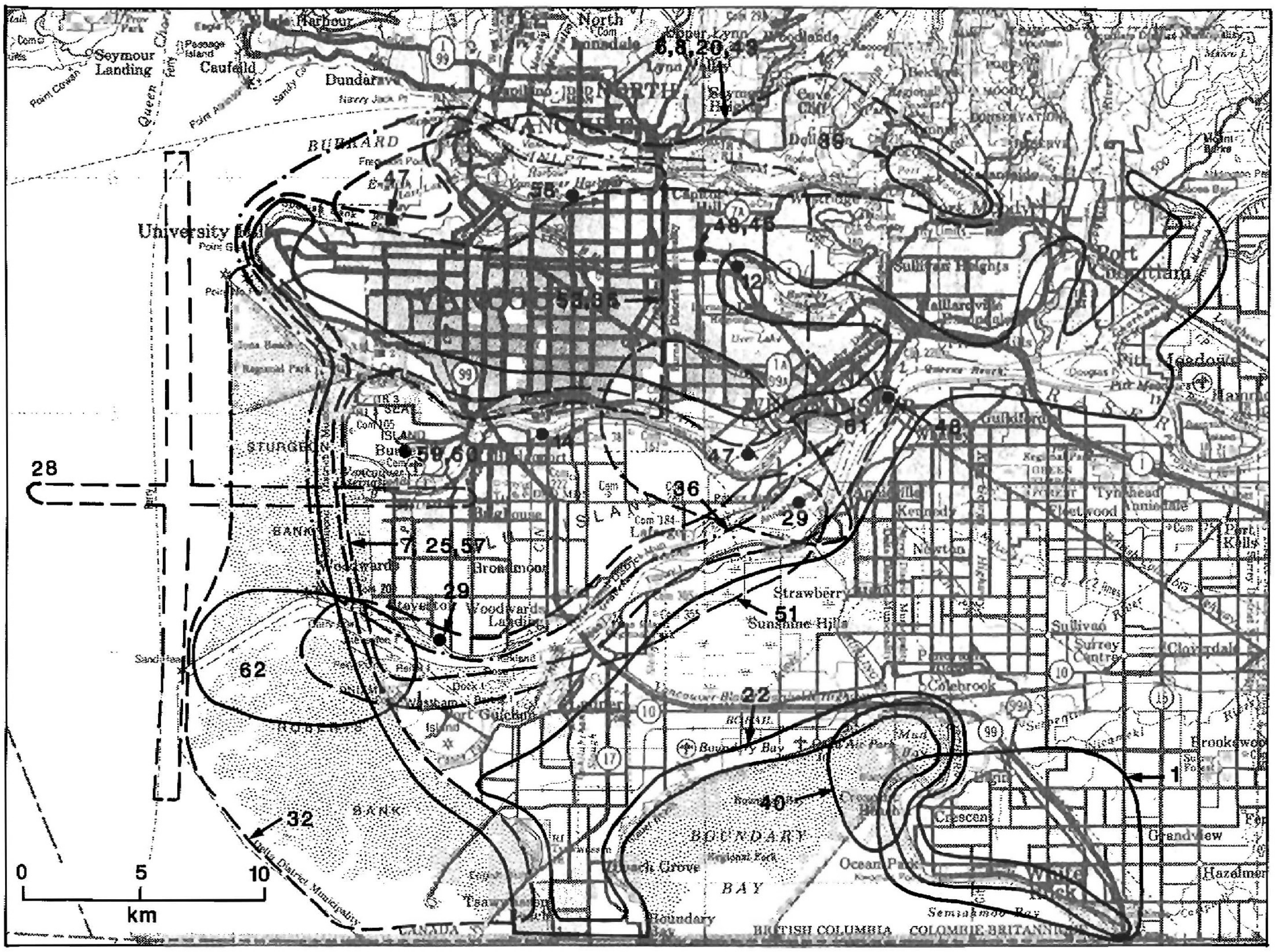

Figure 4. Location map of geochemical data sets in the area immediate to Vancouver area. See Figure 3 for legend.

Table 1. Summary of trace element data sources for the ancouver map sheet (NTS 92G).

Map No.[1]	Source[2]	Purpose of study[3]	Date of sampling	Sample type[4]	Areal coverage	Number of sites	Sample density[5]	Sample frequency[6]	Analytical[7] Laboratory	Analyses	Computer Format[8]	References
STREAM / LAKE / CITY WATER												
2	BCMEMPR	background	June-Aug.1989	water	92G	922	1/10.8 km^2	na	Barringer	F, U, pH	ASCII	Matysek et al., 1990 Gravel et al., 1990
26	GSC	baseline	July 1991	water	Howe Sound	23	na	na	GSC	As, Bi, Cu, Fe, Mg, Mn, Sb, Zn		Percival et al., 1992
10	BCELP	baseline	1975	water	Fraser River	16	na	na	BCELP	Cd, Cr, Cu, Fe, Hg, Mg, Mn, Ni, Pb, Zn		Clark et al., 1980
22	FREMP	baseline	July-Oct.1978	sewer	Fraser River	32	na	1/2 months	BCELP	Al, As, Cr, Cu, Fe, Pb, Mn, Hg, Ni, Zn		Clark et al., 1981
37	WEST	baseline	July-Aug.1972	water	Fraser River and tributaries	46	~1/5 km^2	na	BCELP	Ag, Cd, Co, Cu, Fe, Pb, Mn, Mo, Ni, Zn		Benedict et al., 1973
41	UBC, Chem.	baseline	Jan-May 1979	water	Fraser River estuary	14	1/3 km^2	2/17 months	UBC, Chem.	Al, Mn		de Mora, 1981
1	UBC, Geol.	baseline	Jan-Feb.1971	supply	White Rock & vicinity	69	na	na	UBC, Geol.	Cd, Co, Cu, Fe, Ni, Pb, Zn		Folk, 1969
50	SRK	baseline	Sept-Dec.1990	water	Britannia Creek	11	na	na	ASL	Cu, Fe, Zn		BCMEMPR, 1991
58	EC	baseline	Jan-Mar.1993	sewer	downtown Vancouver	1	na	7/year	Cantest	Ag, Al, As, Ba, Be, Bi, Ca, Cd, Co, Cr, Cu, Fe, Ga, Hg, Mg, Mn, Mo, Na, Ni, P, Pb, Sb, Sc, Sr, Ti, Tl, V, W, Zn	EXCEL	A. David, EC, North Vancouver, B.C., pers. comm., 1993
27	GSC	impact	July 1991	water	Brittania Mine	20	na	na	GSC	As, Bi, Cu, Fe, Mg, Mn, Sb, Zn		Percival et al., 1992
15	BCELP	impact	1983-1986	water	Brittania Mine	17	na	yearly	BCELP	Al, Ca, Cd, Co, Cu, Fe, Mg, Mn, Ni, Pb, Zn		Moore, 1985
14	BCELP	impact	Aug.1989-Jan.1990	water	Richmond	11	<1/km^2	na	Zenon	Ba, Cd, Co, Cr, Cu, Fe, Mg, Mn, Mo, Ni, Pb, V, Zn		Metalex, 1990
36	WEST	impact	June-Nov.1974	sewer	Vancouver	12	na	1/6 months	BCELP	Ag, Cd, Co, Cu, Fe, Hg, Mn, Ni, Pb, Zn		Koch et al., 1977
13	BCELP	impact	Oct.-Nov.1992	water	near Chilliwack	17	na	na	Zenon	Ag, Al, As, Ba, Be, Bi, Cd, Co, Cr, Cu, Fe, Mg, Mo, Ni, Pb		Freyman, 1993
33	WEST	base/mon	Feb.-May 1973	water	Fraser River and tributaries	9	na	17.00	BCELP	Cd, Cu, Fe, Hg, Mn, Pb, Zn		Hall et al., 1974
	BCELP	monitoring	1982-1983	sewer	Burnaby	2	na	~1/2 months	BCME	Al, Cd, Cu, Fe, Pb, Zn		Lawson et al., 1985
5	BCELP	monitoring	Sept.-Nov.1992	water	north shore, Vancouver	19	~1/10 km^2	na	Zenon/ASL	Cr, Cu, Hg, Pb, Zn		Swain, 1989
30	GVRD	monitoring	1954-1992	water	Capilano, Seymour, Coquitlam lakes and Vancouver	~24	na	weekly, bimonthly	GVRD	Ag, Al, As, Ba, Ca, Cd, Cr, Cu, F, Fe, Hg, K, Mg, Mn, Na, Ni, P, Pb, Sb, Se		GVRD, information sheet
51	GVRD	monitoring	1992-1993	water	south arm, Fraser River	5	na	alternate months	GVRD	Ca, Cd, Co, Cr, Fe, Mg, Pb, Zn	SEAM	G. Marsh, pers. comm., 1993
7	BCELP	monitoring	1985	water	Fraser River	23	na	na	Zenon	Cu, Pb, Zn		Swain, 1985
23	FREMP	monitoring	Jan.1987, Nov.1987	water	Fraser River and tributaries	18	na	na	BCMEP	Al, Ba, Cr, Cu, Fe, Hg, Mn, Ni, Pb, Sr, V, Zn		Swain and Walton, 1988
38a	UBC, Chem	monitoring	May 1973-May 1974	water	Fraser River	4	na	~1/4 months	UBC, Chem.	Cu, Zn		Thomas, 1975
35	COV	monitoring	1984-present	supply	Vancouver city	5	na	monthly	CAL			COV 1993
49	FREMP	monitoring	1992-1993	water	Vancouver area	10	na	biweekly/quarterly	Zenon/ASL	As, Cd, Cr, Cu, Hg, Ni, Pb, Zn	Foxpro	E. Greer, FREMP, New Westminister, B.C., pers. comm., 1993
54	CGHU	monitoring	1989-present	water	Howe Sound; Sunshine Coast; Egmont	6	na	yearly	Zenon	Al, As, Ba, Be, Bi, B, Cd, Ca, Co, Cu, Fe, K, Mg, MN, Mo, Ni, P, Pb, Sb, Se, Si, Sn, Sr, S, Te, Ti, Tl, V, Zn, Zr		R.D. Weston, Chief Environmental Health Officer, CGHU, Gibsons, B.C., pers. comm., 1993
59	VIAA	monitoring	Nov.91-present	water	Vancouver International Airport	4	na	occasionally	AERCO	various metals		A. Murray, VIAA, Richmond, B.C., pers. comm., 1993

Table 1. (cont.)

Map No.[1]	Source[2]	Purpose of study[3]	Date of sampling	Sample type[4]	Areal coverage	Number of sites	Sample density[5]	Sample frequency[6]	Analytical[7] Laboratory	Analyses	Computer Format[8]	References
60	VIAA	monitoring	1991-1992	water	Vancouver International Airport	12	na	quarterly	ASL	Ba, Ca, Cd, Co, Cr, Cu, Fe, Hg, K, Mg, Mn, Na, Pb, Se		A. Murray, VIAA, Richmond, B.C., pers. comm., 1993
21	FREMP	compilation	1970-1978	water	Fraser River	~954	na	na	various	As, Cd, Cr, Cu, Fe, Hg, Mn, Mo, Ni, Pb, Zn		Drinnan and Clark, 1980
55	EC	compilation	1979-present	water	B.C.	various	na	various	various	various	ORACLE	J. Strough, pers. comm., 1993
57	BCELP	compilation	1990-1992	water	Fraser River	various	na	various	various	various	EXCEL	D. Walton, pers. comm., 1993
STREAM / LAKE SEDIMENT												
2	BCMEMPR	background	June-Aug.1989	stream	92G	922	1/10.8 km^2	na	Barringer	Ag, As, Au, Bi, Cd, Cr, Co, Cu, Fe, Hg, Mn, Mo,	ASCII	Matysek et al., 1990
9	BCELP	background	1982-1987	lake	92G	19 lakes	na	na	BCELP	Al, As, Ba, Be, Ca, Cd, Cr, Co, Cr, Cu, Fe, Hg, Mg Mn, Mo, Ni, Pb, Se, Sn, Sr, Ti, V, Zn		Rieberger, 1992
41	UBC, Chem.	baseline	Jan.-May 1979	stream	Fraser River Estuary	14	1/3 km^2	na	UBC, Chem.	Al, Fe, Mn		de Mora, 1981
10	BCELP	baseline	1975	stream	Fraser River	6	na	na	BCELP	Cr, Fe, Mn, Pb		Clark et al., 1980
37	WEST	baseline	July-Aug.1972	stream	Fraser River and tributaries	46	~1 ? 5 km^2	na	BCELP	Ag, Co, Cr, Cu, Fe, Pb, Mn, Mo, Ni		Benedict et al., 1973
52	SFU, Geog.	baseline	1992	stream	Fraser River		na	na	SFU, Chem/ UBC, Geol.	Ca, Cr, Cu, Fe, K, Mn, Na, Ni, Pb, Rb, Sr Ti, V, Y, Zn, Zr		Robertson, 1993 T. Hicken, SFU pers. comm., 1993
62	UBC, Ocean.	baseline	1991-1992	stream	Fraser River, delta	64	na	na	ASL/UBC, Ocean	Cd, Fe, Hg, Pb		Robertson, 1993 T. Feeney, UBC, Ocean., pers. comm., 1993
45	UBC, Civil.	baseline	May-Dec.1993	stream	Brunette watershed	~120	na	na	UBC, Civil, Chemex	Cd, Cu, Zn, Pb - 32 element ICP		J. Smith, UBC, Civil,\, pers. comm., 1993
14	BCELP	impact	Oct.1989-Jan.1990	stream	Richmond	11	<1/km^2	na	Zenon	Al, As, Ba, Cd, Co, Cr, Cu, Fe, Hg, Mg, Mn, Mo, Ni, Pb, Se, Sn, V, Zn,		Metalex, 1990
13	BCELP	impact	Oct.-Nov,1992	stream	near Chilliwack	9	na	na	Zenon	Al, As, Ba, Cd, Co, Cr, Cu, Fe, Mg, Mn, Mo, Ni, Pb, Se, Sr, V, Zn		Freyman, 1993
12	BCELP	monitoring	1982-1983	stream	Burnaby	1	na	na	BCELP	Al, As, Ba, Be, Cd, Co, Cr, Cu, Fe, Hg, Mg, Mn, Mo, Pb, Se, Sn, Sr, Te, Ti, Tl, V, Zn		Lawson et al., 1985
5	BCELP	monitoring	Sept.-Nov.1987	stream/lake	North Shore, Vancouver,	5	na	na	Zenon/ASL	Cr, Cu, Hg, Pb, Zn		Swain, 1989
24	FREMP	monitoring	June 1990	stream	Fraser River and Boundary Bay	4	na	12/month	ASL	As, Cd, Cr, Cu, Fe, Hg, Mn, Mg, Mo, Ni, Pb, Zn		Swain and Walton, 1991
25	FREMP	monitoring	Aug. 1989	stream	Fraser River and Estuary	6	na	5/month	ASL	As, Cd, Cr, Cu, Fe, Hg, Mn, Mg, Mo, Ni, Pb, Zn		Swain and Walton, 1990
7	BCELP	monitoring	1985	stream	Fraser River	5	na	1/year	Zenon	Cu, Pb, Zn		Swain, 1985
32	UBC, Geol.	monitoring	Feb.-June 1974	stream	Frasor River	691	~1/km^2	~1/month	UBC, Geol.	Co, Cu, Fe, Mn, Ni, Pb, Zn		Grieve, 1977
48	UVIC, Biochem.	monitoring	May-July 1982?	stream	Fraser River, Still Creek	3	na	2/year	UVIV, Biochm UVIC, Biochem.	Cu, Ni, Zn		Geesey et al., 1984
56	FREMP	compilation	1987-1992	stream	Fraser River	na	na	na	various	various		Anderson, 1993
MARINE WATER												
47	IOS	baseline	1978-1979	water	English Bay	18 (1 in 92G)	na	na	IOS	Pb (total, dissolved and isotopes)		Stukas and Wong, 1981
41	UBC, Chem.	baseline	Jan.-May 1979	water	Strait of Georgia	26	1/10 km^2	2/17 months	UBC, Chem.	Al, Fe, Mn		de Mora, 1981
44	UBC, Ocean	impact	April 1988	water	Howe Sound	2 cores	~23/core	na	UBC, Ocean	Cu, Fe, Mn, Pb, Zn		Drysdale, 1990 Drysdale and Pederson, 1992
26	GSC	impact	July 1991	water	Howe Sound	4	na	na	GSC	As, Bi, Cu, Fe, K, Na, Mg, Mn, Sb, Zn		Percival et al., 1992

Table 1. (cont.)

Map No.[1]	Source[2]	Purpose of study[3]	Date of sampling	Sample type[4]	Areal coverage	Number of sites	Sample density[5]	Sample frequency[6]	Analytical[7] Laboratory	Analyses	Computer Format[8]	References
15	BCELP	impact	1983-1986	water	Brittania Mine	10	na	yearly	BCELP	Al, Ca, Cd, Co, Cu, Fe, Mg, Mn, Ni, Pb, Zn		Moore, 1985 van Aggelen and Moore, 1986
28	GVRD	monitoring	1987	water/ effluent	Strait of Georgia	8	na	na	ASL	Ag, Al, As, Cd, Cr, Cu, Fe, Mo, Ni, Pb, Zn		Fanning et al., 1989
29	GVRD	monitoring	1973-present	effluent	Annacis & Lulu Islands	52 to 11	na	na	ASL	Al, Ag, As, Cd, Cr, Cu, Fe, Mn, Mo, Hg, Ni, Zn		Cain and Swain, 1980 S. Berthold, GVRD, Vancouver, B.C., pers. comm., 1993
6	bchelp	monitoring	Aug.-Sept. 1992	wayter	Burrard Inlet	70	na	na	Zenon/ASL	Cr, Cu, Fe, Hg, Ni, Pb, Zn		Nijman and Swain, 1990
38b	UBC, Chem	monitoring	May 1973-May 1974	water	Strait of Georgia	16	~1/20 km^2	~1/2 months	UBC, Chem	Cu, Zn		Thomas, 1975
MARINE SEDIMENT												
41	UBC, Chem.	baseline	Jan.-May 1979	core/sediment	Strait of Georgia	22	1/10 km^2	na	UBC, Chem.	Al, Mn		de Mora, 1981
44	UBC, Chem.	baseline	May 1987	sediment	Howe Sound	93	~1/km^2	na	UBC, Ocean.	Ba, Co, Cr, Cu, Mn, Ni, Pb, Sr, V, Y, Zn, Zr		Drysdale, 1990
34	IOS	baseline	Nov.1976-Sept.1977	core	Howe Sound	29 cores	~10/core	na	Chemex	Hg		Macdonald and Wong, 1977
46	IOS	baseline	1972-73, 1976-77	sediment/ core	Howe Sound	~108	~2/km^2	na	IOS	Hg		Thompson et al., 1980
26	GSC	baseline	July 1991	sediment	Howe Sound	13	na	na	Bondar-Clegg	Ag, As, Ba, Cd, Co, Cr, Cu, Mo, Ni, Pb, Sb,		Percival et al., 1992
31	IOS	baseline	Sept. 1980	core	Ballenas Basin	4 cores	~15-20/core	na	IOS	Cd, Cu, Fe, Mn, Pb, Zn	Lotus 123	Macdonald et al., 1991
39	UBC, Geol.	baseline	Sept. 1973	core	Port Moody Inlet	70	~4/km^2	na	UBC, Geol.	Ag, Cd, Co, Cu, Fe, Mn, Ni, Pb, Zn		Bourne, 1974
40	UBC, Geol.	baseline	Sept.-Dec.1960	sediment	Mud Bay, Cresent Beach	113	~4/km^2	na	UBC, Geol.	Cu, Fe, Mo, Pb, Zn		Northcote, 1961
43	UBC, Geol.	baseline	Sept. 1973	sediment	False Creek	63	na	na	UBC, Geol.	Ag, Cd, Co, Cu, Fe, Mn, Ni, Pb, Zn		Whiticar, 1974
8	BCELP	monitoring	1974 to 1990	sediment	Burrard Inlet	18	na	~1/2 years	Zenon	Cu, Cr, Hg, Pb, Zn		BCELP, 1990
6	BCELP	monitoring	Aug.-Sept.1992	sediment	Burrard Inlet	7	na	~1/2 weeks	Zenon/ASL	As, Cd, Cu, Hg, Ni, Pb, Zn		Nijman and Swain, 1990
20	EC	monitoring	May 1985-Sept.1986 Oct.1987	sediment	Vancouver Harbour	73	~1/7km^2	1/2 years	ECWVL	Al, As, Ba, Cd, Co, Cr, Cu, Fe, Hg, Mg, Mn, Mo,Mn, Ni, Pb, Pb, Sn, Sr, V, Zn		Thomas and Goyette, 1989 Goyette and Boyd, 1989
32	UBC, Geol	monitoring	Feb.-Oct.1974	sediment	mouth of Fraser River	~280	1/km^2	~1/month	UBC, Geol.	Co, Cu, Fe, Mn, Ni, Pb, Zn		Grieve, 1977
17, 18	EC	monitoring	1975, 1978, 1980-1984, 1988	sediment	Point Grey/Strait of Georgia	50	1/km^2	~1/year	UBC/F&O, Geol.	Co, Cu, Fe, Mn, Ni, Pb, Zn		Packman, 1980 Brothers et al., 1985 Brothers, 1990b
28	GVRD	monitoring	1989	sediment	Strait of Georgia	8	na	na	ASL	As, Cd, Cr, Cu, Hg, Pb, Zn		Chapman et al., 1987
28	GVRD	monitoring	1987	sediment	Strait of Georgia	3	na	na	Cantest	As, Cd, Cr, Cu, Hg, Pb, Zn		Fanning et al., 1989
28	GVRD	monitoring	1990, 1991	sediment	Strait of Georgia	11	na	na	ASL	As, Cd, Cr, Cu, Hg, Pb, Zn		Chapman et al., 1991
19	EC	monitoring	1987	sediment	Squamish Estuary	35	na	na	ECWVL	Cd, Cu, Hg, Pb, Zn		Brothers, 1990a
16	EC	monitoring	Jan.1980-Dec.1987	sediment	Squamish Estuary/ Strait of Georgia	~118	na	~1/1-3 years	ECWVL	Cd, Cu, Hg, Pb, Zn		Sullivan, 1987
SOILS												
11	BCEP	background	1950 to 1972	soil	lower Fraser Valley	variable	na	na	BCELP	Cu, Fe, Zn		Luttmerding, 1981a,b
14	BCEP	impact	Oct.1989-Jan.1990	soil	Richmond	19	<1/km^2	na	Zenon	Al, As, Ba, Ca, Cd, Co, Cr, Cu, Fe, Mg, Mn, Mo, Ni, Pb, Se, Sr, V, Zn		METALEX, 1990
42	FC	monitoring	Aug.-Sept.1986, 1990	soil	north shore of Fraser River	6	na	annually	NFCL	Cu, Mn, Pb, Zn	ASCII	Palmer, 1986 Hall and Addison, 1991
60	GVRD	monitoring	1987-1990	soil	Burnaby	8	na	annually	GVRD	As, Cd, Hg, Ni, Pb, Se		GVRD, 1992b

Table 1. (cont.)

Map No.[1]	Source[2]	Purpose of study[3]	Date of sampling	Sample type[4]	Areal coverage	Number of sites	Sample density[5]	Sample frequency[6]	Analytical[7] Laboratory	Analyses	Computer Format[8]	References
VEGETATION												
26	GSC	baseline	July 1991	vegetation	Howe Sound	36	na	na	ACT/Min-En	As, Ba, Cu, Fe, Ni, Pb, U, Zn		Percival et al., 1992
14	BCEP	impact	Oct.1989-Jan.1990	vegetation	Richmond	19	<1/km^2	na	Zenon	Al, As, Ba, Ca, Cd, Co, Cr, Cu, Fe, Mg, Mn, Mo, Ni, Pb, Se, Sr, V, Zn		Metalex, 1990
42	FC	monitoring	Aug.-Sept.1986	vegetation	north shore of Fraser River	6	na	annually	NFCL	Cu, Mn, Pb, Zn	ASCII	Palmer, 1986 Hall and Addison, 1991
61	GVRD	monitoring	1987-1990	vegetation	Burnaby	8	na	annually	GVRD	As, Cd, Hg, Ni, Pb, Se		GVRD, 1992b
ROCK												
3	BCMEMPR	compilation		rock	92G	na	na	na	various	various	Foxpro	Minfile, 1990
4	BCMEMPR	compilation		rock	92G	na	na	na	various	various	ASCII	ARIS, 1991
SEWERAGE SLUDGE												
53	GRVD	monitoring	1985 to present	sludge	GVRD	4	na	monthly, quarterly	GVRD	As, Cd, Cr, Cu, Fe, Hg, Ni, Mn, Mo, Pb, Se, Zn	EXCEL	Peddie et al., 1992

na = not applicable; + = computer formats only listed where data is available in electronic format.

EXPLANATION OF TABLE HEADINGS

[1] Map No. The area sampled is identified on Fig. 3, 4 by this number.

[2] Source The organization that collected the data and/or where the data can be obtained. Explanations of acronyms for the organization names are listed below:

AC	Agriculture Canada
BCELP	B.C. Ministry of Environment, Lands and Parks
BCMEMPR	B.C. Ministry of Energy, Mines and Petroleum Resources
COV	City of Vancouver
CGHU	Coast Garibaldi Health Unit
EC	Environment Canada
FREMP	Fraser River Estuary Management Program
GSC	Geological Survey of Canada
GVRD	Greater Vancouver Regional District
IOS	Institute of Ocean Science
FC	Forestry Canada, Pacific and Yukon Region
SFU	Simon Fraser University
	Chem. - Department of Chemistry
	Geog. - Department of Geography
SRK	Steffen, Robertson & Kirsten, (B.C.), Inc.
UBC	University of British Columbia
	Civil. - Department of Civil Engineering
	Chem. - Department of Chemistry
	Geol. - Department of Geological Sciences
	Ocean. - Department of Oceanography
UVIC	Universty of Victoria
	Biochem. - Department of Biochemistry
VCHD	Vancouver City Health Department
VIAA	Vancouver International Airport Authority
WEST	Westwater Research Centre

[3] Purpose of study

(1) Background. A program designed to determine natural background concentrations of elements.
(2) Baseline. Programs designed to obtain trace element levels in an area of known human activity.
(3) Impact. Programs that assess the impact and/or extent of contamination at a particular site.
(4) Monitoring. Repeated sampling at one or more stations to determine change with time.
(5) Compilation. Compilations of assorted reports containing trace element data.

[4] Sample type

Sediment	Lake, stream or marine sediment.
Core	Marine sediment collected using a coring device.
Water	Lake, stream or marine water.
Soil	Soil material.
Vegetation	Plant material.
Effluent	Samples from receiving environment near sewage outfall.
Sewer	Samples from city sewer systems.
Supply	Samples from city water supply systems.

[5] Sampling density Density of sampling sites is given except where only a few samples were collected.

[6] Sampling frequency Frequency of sample collection in monitoring programs.

[7] Laboratory Analytical laboratory that conducted the analyses. Acroynyms for laboratories are listed in Table 4.

[8] Computer format Electronic data format is listed where available. A paper copy or all data is available either in a publication listed in the references or can be obtained from the source organization.

regional geochemical study by the B.C. Ministry of Energy, Mines and Petroleum Resources (BCMEMPR) consists of 922 stream sediment samples analyzed for 20 elements (Matysek et al., 1990; Fig. 3, map no. 2). Stream waters collected from the same sites were analyzed for F, U, and pH. These data are collected to identify areas with anomalous metal concentrations as a guide for mineral explorationists. Rieberger (1992) conducted a survey of lake sediment throughout British Columbia but only 19 lakes were sampled in the Vancouver map area (Fig. 3, map no. 9).

Baseline studies

"Baseline studies" are considered in this report to be those conducted in areas of human activity where possible contamination may have elevated metal concentrations above natural background levels. For example, vegetation and stream and marine water and sediment have been sampled for trace elements in and around Howe Sound to determine the extent of contamination from the Britannia Mine (Percival et al., 1992; Fig. 3, map no. 26). Several studies have determined

Table 2. Summary of trace element sampling in the Vancouver map area (NTS 92G).

Type of study	Type of sample		Organizations
Background study	Sediment	fresh	BCMEMPR
		marine	
	Water	fresh	BCMEMPR
		marine	
	Soils		BCELP
	Vegetation		
Baseline study	Sediment	fresh	BCELP, GSC, UBC, WEST, SFU
		marine	IOS, UBC
	Water	fresh	BCELP, FREMP, GSC, WEST, UBC
		marine	IOS, UBC
		sewer	BCELP, EC
	Soils		BCELP+
	Vegetation		GSC
Impact study	Rock		
	Sediment	fresh	BCELP
		marine	
	Water	fresh	BCELP, GSC, WEST, SRK
		marine	BCELP, GSC, UBC
	Soils		BCELP
	Vegetation		BCELP
Monitoring/ objectives	Sediment	fresh	BCELP, FREMP, UBC, UVIC
		marine	BCELP, EC, GVRD*, UBC
	Water	fresh	BCELP, FREMP, GVRD*, UBC, COV, CGHU, VIAA, EC
		marine	GVRD*
		effluent	GVRD*
	Soils		FC, GVRD
	Vegetation		FC, GVRD
	Sludge		GVRD
Compilation	Rock	fresh	BCMEMPR
	Water		BCELP, EC, FREMP
	Sediment		FREMP

+ = compilation. * = on-going sampling.

Table 3. List of geochemical information sources.

B.C. Ministry of Environment, Lands and Parks (BCELP) Environmental Protection Branch 15326 103A Avenue Surrey, B.C. V3R 7A2 Tel: (604) 582-5200 Fax: (604) 584-9751 Water Quality Branch 3rd Floor 765 Broughton Street Victoria, B.C. V8V 1X4 Tel: (604) 356-8298 Fax: (604) 387-9500	Fraser River Estuary Management Program (FREMP) Suite 301 960 Quayside New Westminster, B.C. V3M 6G2 Tel: (604) 525-1047 Fax: (604) 525-3005
B.C. Ministry of Energy, Mines and Petroleum Resources (BCMEMPR) Geological Survey Branch Parliament Buildings Victoria, B.C. V8V 1X4 Tel: (604) 952-0372 Fax: (604) 952-0371	Geological Survey of Canada (GSC) 601 Booth Street Ottawa, Ontario K1A 0E8 Tel: (613) 992-2828 Fax: (613) 996-9990
Burrard Inlet Environmental Action Program Suite 803 510 West Hastings Street Vancouver, B.C. V6B 1L8 Tel: (604) 775-5195 Fax: (604) 775-5198	Greater Vancouver Regional District (GVRD) Quality Control Division 4330 Kingsway Burnaby, B.C. V5H 4G8 Tel: (604) 451-6000 Fax: (604) 451-6019
City of Vancouver (COV) Engineering Department Waterworks Design Branch 1770 West 7th Avenue Vancouver, B.C. V6J 1Y6 Tel: (604) 736-2866 Fax: (604) 736-8651	Institute of Ocean Sciences (IOS) P.O. Box 6000 Sidney, B.C. V8L 4B2 Tel: (604) 363-6409 Fax: (604) 363-6807
Environment Canada, Environmental Protection (EC) Pacific and Yukon Region Chemicals Evaluation Division 224 West Esplanade North Vancouver, B.C. V7M 3H7 Tel: (604) 666-6711 Fax: (604) 666-6281	Simon Fraser University (SFU) Department of Geography Simon Fraser University Burnaby, B.C. V5A 1S6 Tel: (604) 291-3321 Fax: (604) 291-5841
Forestry Canada (FC) Pacific and Yukon Region Pacific Forestry Centre 506 West Burnside Road Victoria, B.C. V8Z 1M5 Tel: (604) 363-0600 Fax: (604) 363-0775 (604) 363-0797	Steffen, Robertson & Kirsten (B.C.) Inc. (SRK) #800 - 580 Hornby Street Vancouver, B.C. V6C 3B6 Tel: (604) 681-4196 Fax: (604) 687-5532
	University of British Columbia (UBC) Vancouver, B.C. V6T 1Z2 Tel: (604) 822-6375
	University of Victoria Victoria, B.C. V8W 3P2 Tel: (604) 721-7211
	Westwater Research Centre (WEST) 1933 West Mall Annex University of British Columbia Vancouver, B.C. V6T 1Z2 Tel: (604) 822-4956 Fax: (604) 822-5357

Table 4. List of analytical laboratories.

ASL	Analytical Services Laboratories, Vancouver, B.C.
ACT	Activation Laboratories Ltd., Ancaste, Ontario
AERCO	AERCO, Ottawa, Ontario
Barringer	Barringer Labs Ltd., Calgary, Alberta
Bondar-Clegg	Bondar-Clegg & Co. Ltd., Ottawa, Ontario
BCELP	B.C. Ministry of Environment, Lands and Parks, Vancouver, B.C.
Cantest	Cantest, Vancouver, B.C.
CAL	City Analyst, Health Department, City of Vancouver, B.C.
Chemex	Chemex Labs Ltd., North Vancouver, B.C.
F&O	Department of Fisheries and Oceans, West Vancouver, B.C.
GSC	Geological Survey of Canada, Ottawa, Ontario
UBC Geol.	Department of Geological Sciences, UBC, Vancouver, B.C.
GVRD	Greater Vancouver Regional District, Burnaby, B.C.
IOS	Institute of Ocean Sciences, Vancouver, B.C.
Min-En	Min-En Laboratories, Vancouver, B.C.
NFCL	Northern Forestry Centre, Edmonton, Alberta
UBC Ocean.	Department of Oceanography, UBC, Vancouver, B.C.
ECWVL	Environment Canada, Pacific and Yukon Region, West Vancouver, B.C.
SFU Chem.	Department of Chemistry, Simon Fraser University, Burnaby, B.C.
Zenon	Zenon Environmental Inc., Vancouver, B.C.
UVIC Biochem.	Biochemistry Department, University of Victoria

baseline trace element concentrations for the Fraser River (Benedict et al., 1973; Hall et al., 1974; Clark et al., 1980; Fig. 3, map no. 17, 33 and 10, respectively). Luttmerding (1981c) compiled baseline surveys on soils of the Langley-Vancouver area, some of which involved analyses for trace elements (Fig. 3, map no. 11). Thus, these studies summarize the effects of previous contamination and provide a basis for monitoring additional input of contaminants.

Impact studies

Impact studies involve sampling of a specific area of contamination to determine the extent and amount of material released. Recommendations regarding remediation are often based on such studies. Although most impact studies are conducted by consultants on contract either to individual companies or government agencies, the Greater Vancouver Regional District, Environment Canada, and the B.C. Ministry of the Environment also conduct site specific studies. For example, the B.C. Environmental Protection Branch (BCELP) sampled runoff from two gas stations near Chilliwack to assess the ability of a wetland built near one of the stations to decompose petroleum contaminants in the water (Freyman, 1993; Fig. 3, map no. 13). Impact studies were also conducted at the Britannia Mine and in Howe Sound adjacent to Britannia Creek by the BCEP (Moore, 1985; van Aggelen and Moore, 1986; Fig. 3, map no. 15) and by the Geological Survey of Canada in conjunction with their baseline survey of Howe Sound (Percival et al., 1992; Fig. 3, map no. 26).

Impact studies provide excellent opportunities to examine transport mechanisms of trace elements within a particular environment. For example, by surveying the mercury distribution in marine sediments near a chlor-alkali plant at Squamish on Howe Sound, researchers found mercury to be associated with the fine grained and organic-rich component of the sediment (Thompson et al., 1980; Fig. 3, map no. 46). Similarly, at the Britannia Mine, Drysdale (1990) found that there was no net diffusion of metals from sulphide tailings into overlying marine waters (Fig. 3, map no. 44).

Monitoring/objectives

Monitoring studies attempt to identify temporal changes in element concentrations within a particular medium. These studies often focus on natural element fluctuations, such as occur during the course of spring runoff in a stream. The Environmental Protection Branch of the B.C. Ministry of Environment, for example, samples particular streams over a three year period to determine background levels of metals and fluctuations. Results are then used to set water and sediment quality objectives for similar environments throughout the province (Swain, 1985, 1989; Nijman and Swain, 1990; Fig. 3, map no. 7, 5, and 6, respectively).

Monitoring is also conducted to determine the effect of human activity on the environment. To determine the effects of the opening of a new incinerator in Burnaby, the Greater Vancouver Regional District monitored trace metal levels in surrounding soil, vegetation, and air beginning prior and extending after startup of the incinerator. No increase in metal loadings were detected in any of these media (Greater Vancouver Regional District, 1992a, b, c; Fig. 4, map no. 57). Starting in 1986, the B.C. Ministry of Environment monitored metals in the Fraser River based on a 5 year agreement with the Fraser River Harbour Commission. Water, sediment, and biota were sampled for trace elements and the data compiled and reported by the Fraser River Estuary Monitoring Program (e.g., Swain and Walton, 1988, 1990, 1991; Fig. 3, map no. 23, 24, and 25, respectively).

The Greater Vancouver Regional District monitors drinking water in the Vancouver area throughout the year. Water from the Seymour, Capilano and Coquitlam watersheds are sampled for trace elements both before and after chlorination treatment (GVRD information sheets; Fig. 4, map no. 30). In addition, both the Greater Vancouver Regional District and the Vancouver City Engineering Department analyze water samples taken from within the city limits (City of Vancouver information sheet; Fig. 4, map no. 35).

Like impact assessments, monitoring studies often provide opportunities to examine the processes involved in trace element transport. In the Fraser River estuary for example, Grieve (1977) found significant metal concentrations associated with Mn and Fe hydrous oxide coatings on mineral grains. The study also found that formation of a portion of these coatings and their scavenging of metals may occur while grains are suspended in the estuary (Fig. 3, map no. 32).

Compilations

There are several compilations of information related to trace elements in British Columbia. The B.C. Ministry of Energy, Mines and Petroleum Resources compiles information on metals found at individual mineral occurrences in the MINFILE (MINFILE, 1990; Fig. 3, map no. 1) database and at prospects and mines in the Assessment Report Indexing System (ARIS, 1991; Fig. 3, map no. 4). Actual analytical data are not included, however. Additional whole rock geochemistry of major and minor elements can be found in provincial and federal geological survey mapping studies though major element analyses are more commonly reported than trace element data. Data on Fraser River water collected from 1970 to 1978 was compiled by the Fraser River Estuary Monitoring Program (Drinnan and Clark, 1980; Fig. 3, map no. 21). Later monitoring of river waters was based on this information as well as on baseline studies.

SUMMARY

Some general conclusions about the status of trace element geochemical studies in the Vancouver map area can be made from this preliminary compilation. As mentioned above, stream sediments and waters and, to a lesser extent, lake waters have been sampled throughout the map sheet. Similar databases for marine sediment, stream water, soils, and vegetation would be valuable as a means of determining background and/or baseline element concentrations and for monitoring pollution trends.

When determining the extent of contamination at a particular site or trace element fluctuations in the natural environment, most agencies focus primarily on measuring trace element concentrations in a particular medium. Little work has been devoted to geochemical and biogeochemical processes affecting partitioning, distribution, and mobility of trace elements. Such research would assist our understanding of how elements travel through the geochemical landscape and how best to minimize effects of human contamination and remediate contaminated sites.

Finally, co-operative organizations such as the Fraser River Estuary Monitoring Program (FREMP) and the Burrard Inlet Environmental Action Program (BIEAP) have been developed to ensure that resources are focused on specific goals and efforts are not duplicated. However, the difficulty even these integrated projects confront is the complexity and interdependence of the geochemical landscape. For example, a particular water mass such as Burrard Inlet receives a multitude of geochemical inputs via stream, sewer, and atmospheric fluxes. Studying all of the components may be outside the mandate area of a particular program. Such problems only reinforce the need for greater integration of efforts by the various agencies and other organizations responding to issues of environmental quality

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