Chapter 1

QUATERNARY GEOLOGY OF THE CANADIAN CORDILLERA

Tiedemann Glacier, British Columbia. Tiedemann Glacier heads in the highest part of the Coast Mountains and is flanked by large lateral moraines of neoglacial age. The lakes at the bottom of the photo lie between two sets of these moraines. Province of British Columbia BC1413-55.

Chapter 1

QUATERNARY GEOLOGY OF THE CANADIAN CORDILLERA

Compiled by
John J. Clague

SUMMARY

The Canadian Cordillera, the westernmost of the major physiographic and geological regions of Canada, is an area of rugged mountains, plateaus, lowlands, valleys, and seaways. This region extends from the International Boundary on the south to Beaufort Sea on the north and from Pacific Ocean and Alaska on the west to the Interior Plains on the east, a land area in excess of 1 500 000 km^2.

The Canadian Cordillera is located at the edge of the America lithospheric plate and consists of the deformed western margin of the North American craton and a collage of crustal fragments, or terranes, that were accreted to the craton and subsequently fragmented and displaced northward along major strike-slip faults. These processes have given the Cordillera a strong northwest-southeast structural grain and are largely responsible for the present complex distribution of rocks, faults, and other structures in the region.

The distribution of earthquakes, active and recently active faults, and young volcanoes in the Canadian Cordillera is controlled by the motions of the Pacific, America, Juan de Fuca, and Explorer plates and Winona Block which are in contact in the northeast Pacific Ocean west of British Columbia. Most large earthquakes and active faults are associated with offshore plate boundaries beyond the British Columbia continental margin; some, however, occur on the continental shelf and on land. Most Quaternary volcanoes are located in four narrow belts that constitute part of the circum-Pacific "Ring of Fire". Several of these volcanoes have erupted during the Holocene, one less than 150 years ago.

Lithospheric plate interactions are also responsible for recent vertical crustal movements in the Canadian Cordillera. Some outer coastal areas of British Columbia apparently are being uplifted at rates of up to 2 mm/a. In contrast, the inner coast is either stable or subsiding. This pattern continues southward into Washington but is interrupted to the north by considerable uplift in southeastern Alaska and bordering areas of British Columbia and Yukon Territory.

The character of the Cordilleran landscape at the beginning of the Quaternary has been inferred from the distribution of lavas, sediments, and relict surfaces of late Tertiary age, and by estimating late Cenozoic denudation rates from fission-track ages of unroofed plutons and uranium-series dates on speleothems in relict phreatic caves. The evidence indicates that the distribution of high and low ground at the beginning of the Quaternary was probably much as it is today, but relief was much less. The landscape had not yet been extensively modified by glaciation and may have resembled that of northern Yukon Territory today.

Quaternary deposits in the Canadian Cordillera are extremely varied and have complex distributions controlled largely by physiography and glacial history. Most of the surface sediments were deposited during the last glaciation and during postglacial time. Older materials occur at the surface in central and northern Yukon Territory, along the eastern margin of the Cordillera, and on parts of the continental shelf; they also underlie Late Wisconsinan glacial deposits in some parts of British Columbia and southern Yukon Territory.

During Pleistocene nonglacial periods, as at present, sedimentation was concentrated in valleys, lakes, and the sea. Mountains were areas of erosion during nonglacial periods, whereas most plateaus and some lowlands were little affected by either erosion or deposition at these times.

During the early phase of each Pleistocene glaciation, as ice spread from mountains into low-lying areas, streams aggraded their valleys with outwash. Advancing glaciers also impounded large lakes in which fine sediments were deposited. At the climax of each major glaciation, most of British Columbia and southern Yukon Territory was covered by ice. At these times, till was deposited in places at the base of the Cordilleran Ice Sheet and its satellite glaciers. However, mountain areas, fiords, and valleys parallel to the direction of ice flow were subject to intense scour by glaciers, and the older unconsolidated fills of many valleys were thus partly or completely removed. Deglacial intervals, like periods of glacier growth, were times of rapid valley aggradation and extensive drainage change. Glacial lacustrine sediments accumulated in lakes that formed and evolved as the ice sheet decayed. Glacial marine sediments were laid down on the continental shelf and on isostatically depressed lowlands vacated by retreating glaciers and covered by the sea. Many valleys became choked with glacial fluvial and fluvial sediments that were eroded from poorly vegetated, unstable drift deposits mantling upland slopes and valley walls.

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Glaciations were initiated by the expansion of alpine glaciers during periods of global climatic cooling. With continued cooling, these glaciers coalesced to form piedmont complexes and mountain ice sheets. Eventually, piedmont complexes from separate mountain source areas joined to cover most of British Columbia and adjacent areas. Throughout this period, the major mountain systems remained the principal source areas of glaciers, and ice flow continued to be controlled by topography. Occasionally, ice thickened to such an extent (approximately 2500 m) that one or more domes with surface flow radially away from their centres became established over the interior of British Columbia.

Each glacial cycle terminated with rapid climatic amelioration. Deglaciation occurred by complex frontal retreat in peripheral glaciated areas and by downwasting accompanied by widespread stagnation throughout much of the interior. Along the western periphery of the ice sheet, glaciers calved back in contact with eustatically rising seas. In areas of moderate relief, the pattern of deglaciation was more complex, with uplands becoming ice-free first and dividing the ice sheet into a series of valley tongues that decayed in response to local conditions.

Growth and decay of the Cordilleran Ice Sheet triggered isostatic adjustments in the crust and mantle of western Canada. In combination with related eustatic and diastrophic effects, these adjustments produced complex sea level changes along the coast of British Columbia. Gradual growth of glaciers at the beginning of each glacial cycle led to progressive isostatic depression of the land surface. Initially, shorelines may have fallen as a proglacial forebulge migrated through coastal areas and as water was transferred from oceans to expanding ice sheets. However, as glaciers continued to grow, the coastal region began to subside, and the sea eventually rose far above its present level relative to the land in most areas. At the climax of each major glaciation, the entire glaciated Cordillera was isostatically depressed, with areas near the centre of the ice sheet displaced downward more than areas near the periphery. During deglaciation, isostatic uplift in most coastal areas was greater than the coeval eustatic rise, thus the sea fell rapidly relative to the land. Uplift occurred at different times during deglaciation due to diachronous retreat of the Cordilleran Ice Sheet.

Although there probably were glaciers in the high mountains of the Canadian Cordillera during late Tertiary time, the oldest glaciations for which there is reasonable stratigraphic and landform evidence are Pleistocene in age. Westlynn Drift in southwestern British Columbia and drift of the Nansen, Klaza, Shakwak, and "old" glaciations in Yukon Territory predate the last (Sangamonian) interglaciation, some by a considerable margin (i.e., Nansen and "old" glaciations are older than 1 Ma). The Great Glaciation (Glacial Episode 1) in the southern Rocky Mountains is probably Middle Pleistocene in age, and there is some evidence for several older glaciations in this region.

The Muir Point Formation and correlative Highbury Sediments in southwestern British Columbia underlie two drift sheets and are either Sangamonian or pre-Illinoian in age. Westwold Sediments of south-central British Columbia are thought to correlate in part with these units.

Units deposited by the Cordilleran Ice Sheet during the penultimate glaciation (Early Wisconsinan or Illinoian) include Semiahmoo, Dashwood, Muchalat River, and Okanagan Centre drifts in British Columbia, and Reid, Mirror Creek, and (?)Icefield drifts in Yukon Territory. Correlative units deposited by mountain glaciers at the periphery of the ice sheet include: Waterton II, Albertan, and Maycroft tills and associated glacial lacustrine and glacial fluvial sediments in the Rocky Mountains (perhaps also Hummingbird, Baseline, and Early Cordilleran tills in this same region); drift of the "intermediate" glaciation in the Southern Ogilvie Mountains; and drift of the penultimate glaciation on the Queen Charlotte Islands.

A lengthy nonglacial interval, known as the Olympia in British Columbia and the Boutellier in Yukon Territory, began before 60 ka and persisted through the Middle Wisconsinan. In British Columbia, glaciers probably were confined to the major mountain ranges throughout the Olympia Nonglacial Interval, and in general the physical environment was similar to that of postglacial time. Sediments accumulated mainly in large intermontane valleys, on coastal lowlands, and in the sea. Two well defined stratigraphic units of Olympia age are the Cowichan Head Formation of southwestern British Columbia and Bessette Sediments of south-central British Columbia.

The Olympia-Boutellier nonglacial interval ended about 25-29 ka with climatic deterioration and glacier growth at the onset of the last glaciation (locally termed Fraser, McConnell, Macauley, and Kluane glaciations). Glacier growth was slow at first, with ice confined to mountain ranges until 20-25 ka, depending on the locality. The Cordilleran Ice Sheet attained its maximum extent in the south about 14-14.5 ka. Ice sheet growth was accompanied by widespread aggradation of outwash in coastal lowlands and interior valleys and by the formation of proglacial lakes.

Late Wisconsinan drift in many parts of the Canadian Cordillera comprises one till and bounding stratified units (e.g., Kamloops Lake, Gold River, Port McNeill, McConnell, and Macauley drifts). In some areas, however, Late Wisconsinan drift is more complex and includes two or more till units. For example, in Fraser Lowland, three main suites of Fraser Glaciation deposits are recognized: (1) Quadra Sand and Coquitlam Drift deposited during the early part of the Fraser Glaciation; (2) Vashon Drift deposited at the climax of this glaciation; and (3) Capilano Sediments, Fort Langley Formation, and Sumas Drift deposited during deglaciation. Each of these suites contains, among other things, one or more tills.

Late Wisconsinan ice cover in many mountain ranges at the periphery of the Cordilleran Ice Sheet was limited. Most glaciers in the Mackenzie, Wernecke, and Southern Ogilvie mountains terminated inside mountain fronts. Parts of the Queen Charlotte Islands also apparently escaped glaciation during Late Wisconsinan time. In the Rocky Mountains, however, many glaciers flowed onto the Interior Plains and coalesced with the Laurentide Ice Sheet. Drift units of Late Wisconsinan age in peripheral glaciated areas include: Waterton III and IV, Hidden Creek, Bow Valley, Canmore, Eisenhower Junction, Ernst, Lamoral, Jackfish Creek, Marlboro, Obed, and Drystone tills and associated outwash in the Rocky Mountains; the surface drift in the Fort St. John area; and drift of the "last" glaciation in the Southern Ogilvie Mountains.

The southwestern sector of the Late Wisconsinan Cordilleran Ice Sheet began to decay about 14 ka. Parts of *the coastal lowlands of southwestern British Columbia were ice-free by 13 ka, and the ice sheet and most satellite glaciers had completely disappeared by 10 ka or shortly thereafter. During deglaciation, the sea transgressed isostatically depressed lowlands along the British Columbia coast, and large lakes formed behind masses of decaying ice and drift in the interior.*

Late Wisconsinan deglaciation initiated a period of redistribution of glacial materials by fluvial and mass wasting processes that continued into the Holocene. A period of rapid aggradation in many valleys was followed by downcutting as the supply of sediment to streams decreased. By middle or late Holocene time, streams were flowing near their present levels. The main sedimentation sites in the Cordillera during the Holocene have been lake and seafloor basins, fans, and deltas.

Northern Yukon Territory was not glaciated during the Quaternary. The only surficial deposits in most of this region are weathered rock and colluvium on slopes, organic sediments in depressions, and fluvial deposits along streams. However, thick fluvial, lacustrine, and glacial lacustrine sediments underlie Old Crow, Bluefish, and Bell flats. During much of the Pleistocene, as at present, alluvium and lake sediments accumulated in these areas. On at least one occasion, however, the Laurentide Ice Sheet advanced against the Richardson Mountains and blocked easterly drainage at McDougall Pass, thus forming an extensive proglacial lake in the lowlands to the west. Large amounts of glacial lacustrine silt and clay were deposited in this lake.

Quaternary paleoenvironmental information for the Canadian Cordillera comes mainly from studies of fossil pollen, plant macrofossils, paleosols, speleothems, and vertebrate and invertebrate faunal remains. Information for early and middle Quaternary time is sparse, although recent paleoecological studies in northern Yukon Territory have shed new light on the character and evolution of circumpolar floras and faunas during this period. In contrast, there is abundant paleoenvironmental information for Wisconsinan and Holocene time. Environmental conditions during the long Olympia-Boutellier interval were variable in both space and time; the climate at times was colder and at times similar to that of the present. During the Fraser-McConnell Glaciation, very cold and probably arid conditions prevailed at the eastern and northern margins of the Cordilleran Ice Sheet; the vegetation in these areas was tundra. Along the Pacific coast, however, the climate was more moderate, and diverse floras may have persisted in refugial areas.

Major floral changes in the Canadian Cordillera during and immediately following deglaciation are attributable in part to climatic change and in part to different rates of plant migration and plant succession on freshly deglaciated terrain. Nonarboreal plant communities adapted to cold and probably dry conditions were the first to appear. In the southern Cordillera, these plant communities were rapidly replaced by forests as climate ameliorated. In the Yukon, herb tundra was replaced by shrub tundra at this time; later, various tree species expanded into the region in a complex fashion. The general warming trend which accompanied deglaciation continued well into the Holocene, although it may have been interrupted by brief intervals of climatic deterioration. The warm Hypsithermal interval was followed during the latter half of the Holocene by generally cooler and wetter conditions. Vegetation in most areas has changed little in the last few thousand years.

Alpine glaciers in the Canadian Cordillera have fluctuated in response to Holocene climatic change. Cirque moraines of latest Pleistocene or early Holocene age occur in several mountain ranges and were constructed during a minor expansion of glaciers at the end of the Fraser Glaciation or soon thereafter. Glacier advances also occurred between 3.3 ka and 1.9 ka, during the last millennium, and probably between 6 ka and 5 ka. In most areas, glaciers achieved their maximum Holocene extent between about 1500 AD and 1850 AD during the Little Ice Age.

INTRODUCTION

J.J. Clague

The geology of Quaternary deposits in the Canadian Cordillera and various aspects of the landscape of the region attributable to Quaternary events are discussed in this chapter. In Canada in general, and in the Cordilleran region in detail, Quaternary glaciation has profoundly altered the landscape and left a mantle of unconsolidated sediments on bedrock that has significantly affected economic development and other human activities. Repeatedly during the Quaternary Period and perhaps also in late Tertiary time, glaciers enveloped large areas of the Cordillera; only parts of central and northern Yukon Territory, the eastern Mackenzie Mountains, and scattered peaks were never covered by ice. Glaciations were separated by nonglacial or interglacial periods during which the Canadian Cordillera was largely free of ice.

This chapter highlights the history of the last nonglacial-glacial cycle which is so well documented in the Canadian Cordillera. The relationship between topography, climate, and glacier buildup and decay are discussed, and depositional and erosional responses to major climatic changes are assessed. Special consideration is given to the Quaternary history of Yukon Territory north of the limit of glaciation using evidence obtained from sedimentary deposits in Porcupine River basin. Other significant topics include: (1) the contemporary tectonic setting of the Canadian Cordillera; (2) the evolution of Cordilleran landscapes during the Quaternary; (3) interrelationships of past Cordilleran and Laurentide ice masses; (4) land-sea relationships during the advance and retreat of the last Pleistocene glaciers; (5) the Quaternary paleoecology and paleoclimatology of the Cordillera; and (6) the activity of glaciers during the Holocene.

Description of region

The Canadian Cordillera is part of the great belt of mountains that forms the western margin of North America and South America. Within Canada, this belt is up to 900 km

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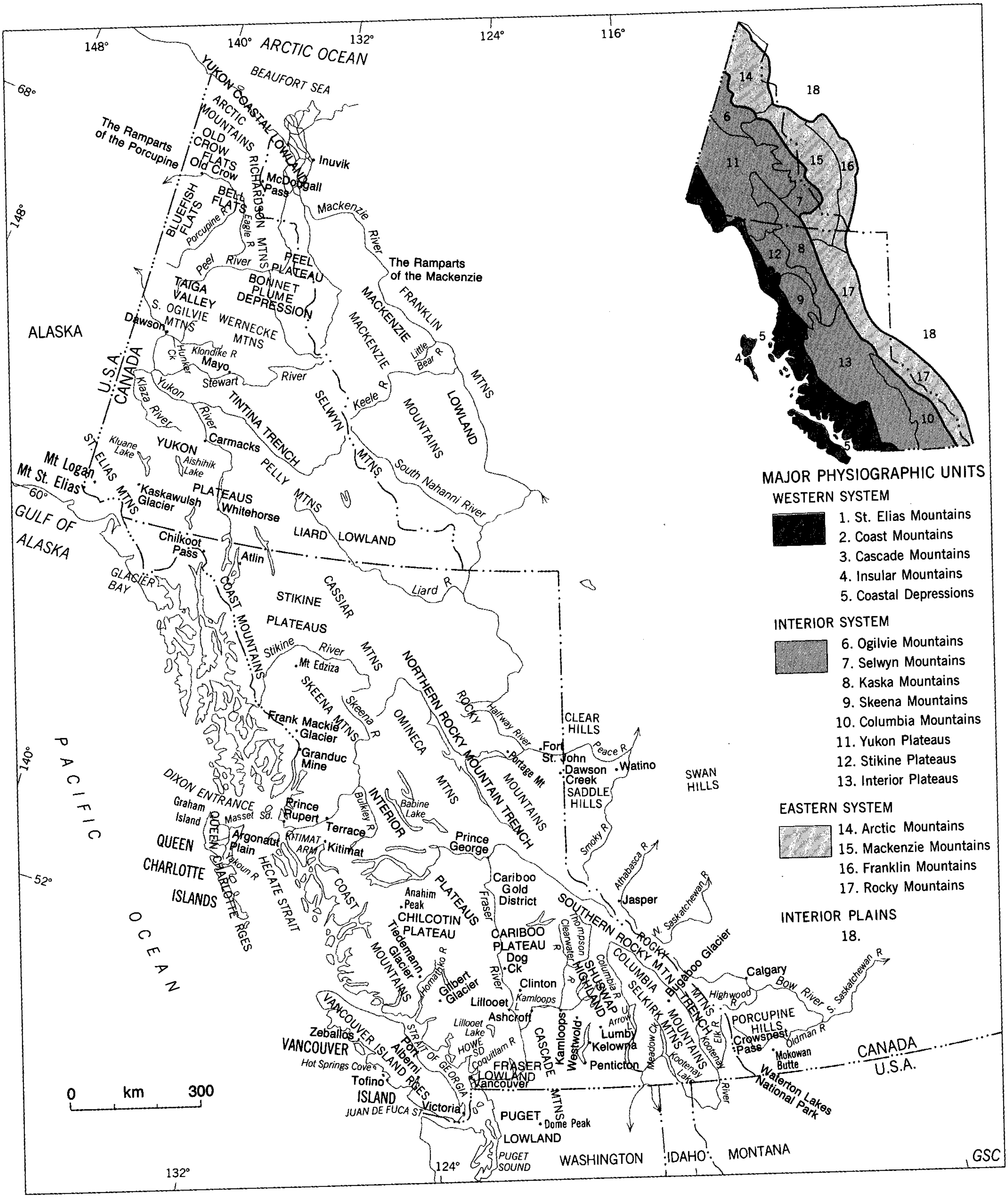

Figure 1.1. Map of the Canadian Cordillera and adjacent regions showing place names cited in the text. Major physiographic units (after Mathews, 1986) are shown on the small-scale inset map.

wide and more than 2500 km long. It extends north from the International Boundary along the forty-ninth parallel and Juan de Fuca Strait to Beaufort Sea, and east from the Pacific Ocean and Alaska to the Interior Plains (Fig. 1.1). The Canadian Cordillera includes most of British Columbia and Yukon Territory, the continental shelf and slope west of British Columbia, and parts of Alberta and District of Mackenzie, an area of about 2 000 000 km^2.

The region is one of extremely diverse topography, geology, climate, and vegetation. Rugged glacier-clad mountains that receive more than 2500 mm of precipitation per year are within sight of broad semiarid valleys. The general form of the Cordillera resembles that of a great wall flanking the Interior Plains and consisting of an elevated platform rimmed by mountain battlements (Bostock, 1948). The eastern battlement of this wall comprises the Rocky, Mackenzie, and Arctic mountains, and the western battlement, the Coast and St. Elias mountains. The platform between the two consists of plateaus, lowlands, valleys, and mountains which form much of the interior of British Columbia and the Yukon. West of the Coast Mountains are the lower ranges of Vancouver Island and the Queen Charlotte Islands, and lowlands bordering the Pacific Ocean.

The high mountains of the Cordillera today support valley glaciers and small ice caps. During major Pleistocene cold periods, glaciers advanced from these mountains to bury most of British Columbia and parts of the Yukon, District of Mackenzie, and Alberta beneath as much as 2.5 km of ice. In the process, the drainage network of western Canada was repeatedly disrupted and rearranged, mountains were sculpted by glaciers, and prodigious quantities of sediment were deposited in valleys and on plateaus, coastal lowlands, and the seafloor. Those parts of the Yukon that completely escaped glaciation have a somewhat muted topography that contrasts sharply with the rugged mountains and fresh drift-covered lowlands of formerly glaciated areas. The landscape of both glaciated and unglaciated areas has been modified since the end of the Pleistocene by fluvial, mass wasting, and other processes, but the resulting changes have been minor in comparison to those induced by repeated growth and decay of Cordilleran glaciers over a period of millions of years.

History of development of Quaternary ideas

Our present understanding of the Quaternary geology of the Canadian Cordillera is based on observations made by earth scientists over about the last 100 years. During this time, ideas concerning the Quaternary have gradually evolved and become more sophisticated.

The first significant studies of the Quaternary of the Canadian Cordillera were made in the late Nineteenth Century by G.M. Dawson. In a series of papers that were remarkable for their time, Dawson (1877, 1878, 1879, 1881, 1886, 1888, 1889, 1890, 1891) established that British Columbia and adjacent areas at one time had been covered by glaciers. Strongly influenced by ideas prevailing elsewhere in North America and Europe at the time, he initially invoked a marine submergence of the Cordilleran interior to account for gravel deposits, or "beaches", high on valley walls, "boulder clay" (till), and bedded silt deposits (subsequently shown by Daly, 1912, to be glacial lacustrine in origin). Dawson abandoned the idea of a marine transgression after it received widespread criticism (Chamberlin, 1894) and devoted more of his energies to understanding the geometry and flow patterns of former ice masses that buried the Cordillera.

In his most comprehensive report on glacial geology, Dawson (1891) suggested that an ice sheet developed in the Canadian Cordillera when mountain glaciers (especially those in the Coast Mountains) expanded and coalesced to form a continuous cover over plateaus and coastal lowlands. Once fully established, this ice sheet had a dome-like central area in northern British Columbia from which ice flowed northwesterly into the Yukon and southeasterly into the United States.

In the late Nineteenth Century, Dawson's "Cordilleran glacier" was the subject of considerable debate among geologists working in British Columbia and Yukon Territory. The existence of such a glacier continued to be questioned into the Twentieth Century (e.g., Tyrrell, 1919). However, careful observations from many parts of the Canadian Cordillera and adjacent areas provided increasing support for an all-encompassing Pleistocene ice sheet (e.g., Willis, 1898; Tyrrell, 1901; Gwillim, 1902; McConnell, 1903a, b; Coleman, 1910; Daly, 1912, 1915; Bretz, 1913; Stewart, 1913; Read, 1921; Johnston, 1926b).

During the early Twentieth Century, attention shifted from questions relating to the existence of a former ice sheet to questions concerning the pattern of ice sheet growth (Kerr, 1934, 1936; Davis and Mathews, 1944) and the sequence of glacial and nonglacial intervals during the Quaternary (e.g., Clapp, 1913; Burwash, 1918; Berry and Johnston, 1922; Johnston, 1923, 1926a; Johnston and Uglow, 1926; Cockfield and Walker, 1933; Bostock, 1934; Kerr, 1934). In addition, a wide variety of other Quaternary phenomena began to attract the attention of scientists. For example, during the late Nineteenth and early Twentieth centuries, the first substantive contributions were made in the following fields in the Canadian Cordillera: Quaternary paleontology and paleoecology (Crickmay, 1925, 1929; Hansen, 1940; Cowan, 1941); recent glacier fluctuations (Munday, 1931, 1936; Wheeler, 1931); Quaternary sea level change (Lamplugh, 1886; Newcombe, 1914; Johnston, 1921a); slope stability (Stanton, 1898; McConnell and Brock, 1904; Daly et al., 1912); the sedimentology and genesis of Quaternary sediments (Johnston, 1921b, 1922; Hanson, 1932; Flint, 1935); and landform development (Peacock, 1935; Beach and Spivak, 1943). Notwithstanding these contributions, there were relatively few scientists working on Quaternary problems in the Canadian Cordillera during this period, and interest was correspondingly low.

The development of Quaternary ideas entered a new era in the middle and late 1940s when an increasing number of geologists developed interests in the surficial geology of the region. Some of these individuals (notably J.E. Armstrong, H.S. Bostock, and W.H. Mathews) recognized that a better understanding of the Quaternary Period could only be achieved if surficial sediments and landforms were investigated in their own right, rather than as adjuncts to bedrock mapping programs, as generally had been the case before. As a result of their efforts, new insights were gained into Quaternary geomorphic change in the Canadian Cordillera and into the pattern and history of glaciation in the region. These new insights came at a time of accelerated economic development and hydrocarbon and mineral exploration, which created a large demand for information on the distribution and character of unconsolidated deposits. This, in turn, led to a rapid growth in Quaternary research by government agencies and universities, a process that has continued unabated to the present. The explosion in Quaternary scientific endeavours in the last four decades has been accompanied by an increase in diversity and sophistication of research. Before World War II, many Quaternary studies were limited to observations of former ice flow directions, upper limits of glaciation, and changes in drainage patterns. In contrast, in recent years there has been increasing emphasis on, among other things, stratigraphy, former sedimentary environments, paleoecology, neotectonics, and hydrogeology. This chapter is a summary and synthesis of information collected in these and other fields.

Organization and authorship

These introductory remarks are followed by an overview of the Canadian Cordillera, which includes brief summaries of the following: bedrock geology; Quaternary tectonic setting; physiography and drainage; climate; Quaternary landscape development; the regional character and distribution of Quaternary deposits; controls on Quaternary deposition and erosion; the character and extent of the Cordilleran Ice Sheet; patterns of ice sheet growth and decay; relationships of Cordilleran and Laurentide ice masses; and Quaternary sea levels. Following this regional overview are sections concerned with the Quaternary stratigraphy, history, and paleoecology of various parts of the Canadian Cordillera. These are followed, in turn, by a section on Holocene glacier fluctuations and by one on the economic implications of Quaternary deposits and contemporary geologic processes.

The authorship of each major section is indicated after the corresponding section heading. Clague edited all contributions to ensure uniformity of style and completeness of coverage.

BEDROCK GEOLOGY

J.J. Clague

The Canadian Cordillera comprises part of the deformed western margin of the North American craton and a collage of far-travelled crustal fragments or "terranes"; the latter have become accreted to the craton in response to oblique convergence of North American and Pacific Ocean lithosphere during Mesozoic and Cenozoic time (Coney et al., 1980; Jones et al., 1982a, b; Monger et al., 1982; Chamberlain and Lambert, 1985). Each terrane is separated from its neighbours by major faults, intrusions, or a cover of younger rocks, and comprises a unique assemblage of rocks that formed in a particular tectonic setting (Tipper et al., 1981). By analogy with modern examples, most terranes in the Cordillera can be interpreted to be relicts of island arcs, oceanic plateaus and islands, continental margin fragments, and complex accretionary terranes, the latter including melange belts, ophiolite fragments, and thrust-faulted forearc provinces (Silver and Smith, 1983).

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Accretion and related faulting have given the Canadian Cordillera a strong northwest-southeast structural grain. The region comprises five major northwest-trending geological belts, each with its own distinctive stratigraphy, metamorphism, plutonism, volcanism, and structure (Fig. 1.2, 1.3; Tipper et al., 1981; Monger et al., 1982; Monger and Berg, 1984; Gabrielse and Yorath, in press):

1. The Foreland Belt borders undeformed rocks of the Interior Platform on the east and consists of a northeasterly thinning wedge of middle Proterozoic to Upper Jurassic miogeoclinal and platform carbonates and craton-derived clastics, and Upper Jurassic to Paleogene exogeoclinal clastics derived from the Cordillera. This wedge is foreshortened about 200 km along a series of west-dipping thrust faults and northwest-trending folds that developed during the late Mesozoic and early Cenozoic.
2. The Omineca Belt straddles the boundary between ancestral North America and allochthonous terranes to the west, and is a major tectonic welt characterized by deformation, regional metamorphism, and granitic plutonism. The eastern part of the Omineca Belt comprises middle Proterozoic to middle Paleozoic miogeoclinal rocks, whereas the western part consists of accreted Paleozoic and lower Mesozoic volcanic and sedimentary rocks. Granitic plutons are common throughout the belt and are mainly of Jurassic and Cretaceous age. The intense metamorphism and complex structure of the Omineca Belt were produced when a large composite terrane collided with and became accreted to continental crust to the east, probably in Early and Middle Jurassic time.
3. The Intermontane Belt is a composite of three major exotic terranes formed of upper Paleozoic to middle Mesozoic, allochthonous, marine volcanic and sedimentary rocks. These rocks are superposed by autochthonous Jurassic and Cretaceous clastic wedges and generally flat-lying Tertiary volcanic and sedimentary rocks. The belt also contains granitic intrusions which are comagmatic with the various volcanics. The three major terranes of the Intermontane Belt amalgamated by the Middle Jurassic prior to colliding with continental crust to the east.
4. The Coast Belt, west of the Intermontane Belt, is the second large metamorphic and plutonic welt in the Canadian Cordillera. It consists of Jurassic to Tertiary granitic rocks and variably metamorphosed sedimentary and volcanic strata ranging in age from Paleozoic to early Tertiary. The Coast Belt formed mainly during Cretaceous and early Tertiary time, probably as a result of the collision of amalgamated terranes of the Intermontane Belt with a large composite terrane to the west (the Insular Belt). Compressional thickening and tectonic overlap resulting from this collision have been invoked to explain 5 to 25 km of uplift and erosion that occurred along the axis of the belt in Cenozoic time.
5. The Insular Belt, like the Intermontane Belt, is a composite allochthonous terrane. It comprises Upper Cambrian to Tertiary volcanic and sedimentary rocks, and granitics in part comagmatic with the volcanics. Its major elements probably amalgamated in Middle Jurassic time and collided with terranes of the Intermontane Belt during the Cretaceous. Accretionary

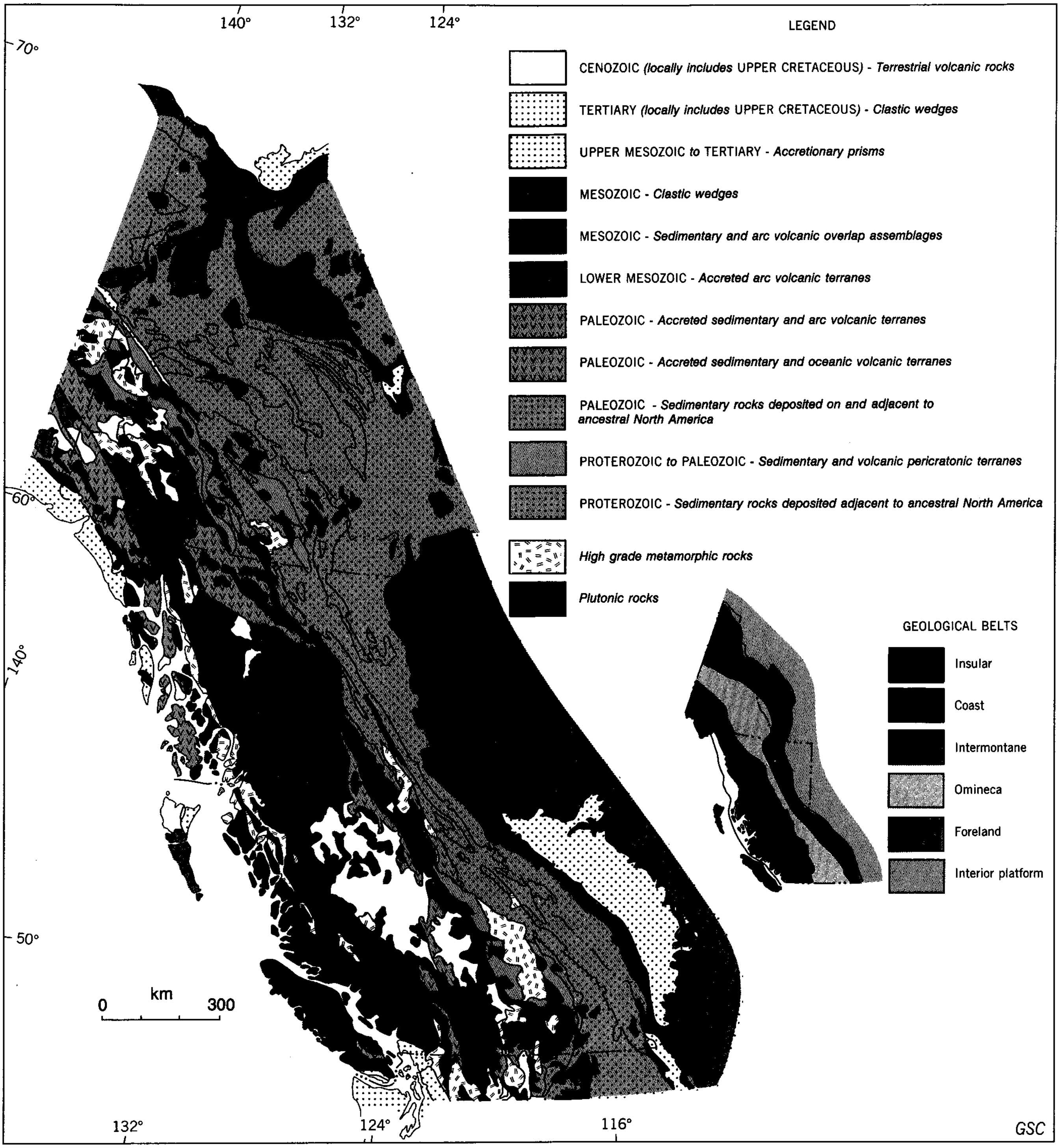

Figure 1.2. Simplified bedrock geology map of the Canadian Cordillera (adapted from Tipper et al., 1981).

prisms at the western margin of the Insular Belt, however, were added to the continent later, and, in fact, the Yakutat terrane is still accreting as the Pacific plate underthrusts Alaska.

Accretion of allochthonous terranes to North America was accompanied by compression and by northwestward displacements of accreted rocks. Displacements are manifested, in part, in the great right-lateral strike-slip faults of the region, such as the Tintina in Yukon Territory, active in Late Cretaceous-early Tertiary time, the Denali in Alaska and the Yukon, active since middle Tertiary time, and the offshore Queen Charlotte-Fairweather system which is

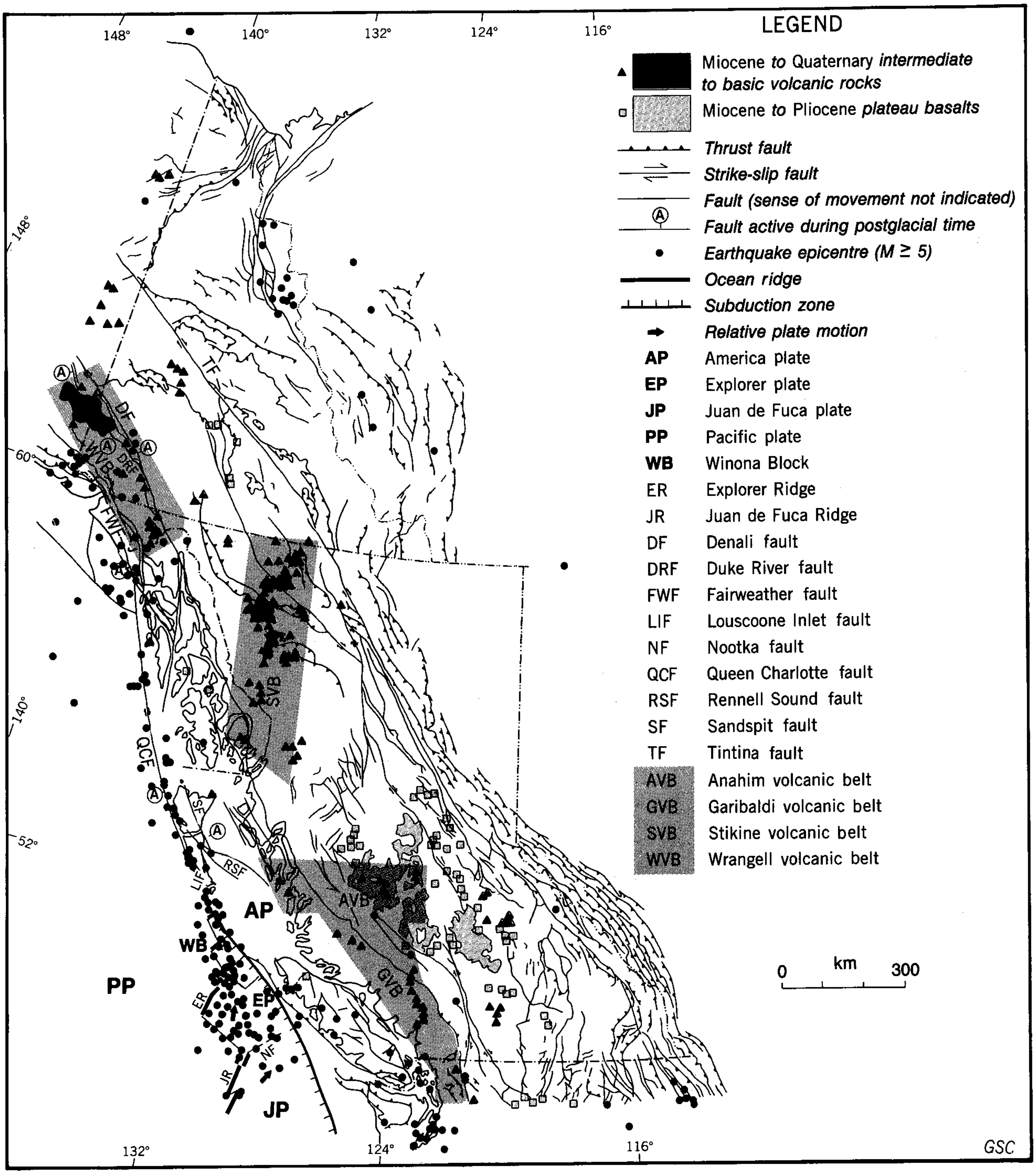

Figure 1.3. Major faults, plate boundaries, epicentres of large historic earthquakes (to December 1980) and young volcanic rocks in the Canadian Cordillera. Sources of information: Riddihough and Hyndman (1976), Souther (1976), Tipper et al. (1981), Geological Survey of Canada (unpublished).

active today. In the southeastern Canadian Cordillera, early Tertiary displacements were accompanied by crustal extension and uplift along listric normal faults, resulting in tectonic denudation of cover rocks above metamorphic core complexes.

QUATERNARY TECTONIC SETTING

J.J. Clague

The present tectonic regime of western North America is controlled mainly by the motions of the Pacific, America, and Juan de Fuca plates (Fig. 1.3; Atwater, 1970). In addition, there are two small lithospheric blocks (Explorer plate, Winona Block) at the north end of the Juan de Fuca plate; these may be moving as independent units (Riddihough, 1977). The Pacific and America plates and Winona Block intersect off the north end of Vancouver Island (Riddihough et al., 1980). North of this triple junction, the Queen Charlotte-Fairweather transform fault separates the Pacific and America plates. Displacements along this fault are right-lateral and average about 5.5 cm/a (Chase and Tiffin, 1972; Keen and Hyndman, 1979). South of the triple junction, a system of spreading ridges and short transform faults forms the boundary between the Pacific plate on the west and the Juan de Fuca and Explorer plates and Winona Block on the east (Barr and Chase, 1974; Riddihough et al., 1983). Spreading rates on Juan de Fuca and Explorer ridges range from about 4 to 6 cm/a (Keen and Hyndman, 1979). The Juan de Fuca and Explorer plates are separated by the Nootka transform fault which extends in a northeasterly direction from Juan de Fuca Ridge to the continental shelf of north-central Vancouver Island (Hyndman et al., 1979). Motion along this fault probably is left-lateral at a rate of about 2 cm/a.

The boundary between the America and Juan de Fuca plates is thought to be a zone of convergence or subduction. It is well established that subduction has occurred along the coasts of British Columbia, Washington, and Oregon during the past few million years (Riddihough and Hyndman, 1976), but there has been some debate as to whether or not it is continuing at present (Crosson, 1972). The doubt arises primarily from the absence of a deep marginal trench characteristic of most active subduction zones, the lack of deep or thrust-type earthquakes in an eastward-dipping Benioff zone, and the relatively low level of historic volcanism in the mountains bordering the Pacific Ocean. Riddihough and Hyndman (1976) reviewed relevant geological and geophysical data bearing on this problem and concluded that subduction probably is continuing south of 50°N, although at a very low rate in southwestern British Columbia.

The distribution of earthquakes, active and recently active faults, and Quaternary volcanoes in the Canadian Cordillera (Fig. 1.3) is related to the plate tectonic regime outlined above. The main earthquake areas are (1) the Queen Charlotte-Fairweather fault, (2) the offshore spreading ridge and associated fracture zones, (3) the southern Strait of Georgia-Puget Sound region, (4) the St. Elias Mountains, and (5) part of eastern Yukon Territory and westernmost District of Mackenzie (Basham et al., 1977; Milne et al., 1978). The first two areas are seismically active because they are plate boundaries. Focal mechanisms and fault-plane solutions of earthquakes in these two areas are in good agreement with postulated relative plate motions (Hyndman and Weichert, 1983). The southern Strait of Georgia-Puget Sound seismic area may correspond to the zone beneath which the Juan de Fuca plate remains essentially coherent as it is subducted beneath North America (Keen and Hyndman, 1979). Seismicity in this area may be related to lateral compression or overlapping in the sinking plate due to the major change in trend of the North American continental margin at this latitude. Earthquake activity in the Strait of Georgia decreases northward and is very low north of the presumed Explorer-America plate boundary at about 51°N. Seismicity in the St. Elias Mountains probably is linked in some way to convergence of the Pacific and America plates in the northern Gulf of Alaska. The relationship of seismicity in eastern Yukon Territory and western District of Mackenzie to the present-day tectonic regime is unknown, and the same is true for scattered earthquakes that occur elsewhere in the Cordillera, for example in south-central and southeastern British Columbia.

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1989: Quaternary tectonic setting (Canadian Cordillera); in Chapter 1 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

Many faults in the Canadian Cordillera are presently active or have been active sometime during the Holocene. On the basis of high seismicity, the Queen Charlotte fault and some of the transform faults connecting segments of Juan de Fuca and Explorer ridges are known to be active at present (Milne et al., 1978). Other major, predominantly strike-slip faults for which Holocene and recent movements are known or suspected include the Denali, Duke River, and associated faults in southwestern Yukon Territory (Clague, 1979; Horner, 1983), and the Sandspit, Rennell Sound, and Louscoone Inlet faults on the Queen Charlotte Islands (Sutherland-Brown, 1968; Young, 1981). Other major faults also may be active, but studies required to document recent displacements have not been undertaken. Finally, there probably are many small active faults in the Cordillera that have not been recognized by geologists. Hamilton and Luternauer (1983), however, identified numerous small Holocene faults of possible tectonic origin on the seafloor of the southern and central Strait of Georgia, and Eisbacher (1977) described fresh fault scarps in the Mackenzie Mountains.

Most Quaternary volcanoes in the Canadian Cordillera are located in four linear belts (Fig. 1.3; Souther, 1970, 1977). The southernmost, or Garibaldi, belt is the northern extension of a well defined chain of late Cenozoic volcanoes in northern California, Oregon, and Washington. It contains more than 30 Quaternary cones and domes of andesite, dacite, and minor basalt, which probably formed in response to subduction of the Juan de Fuca and Explorer plates. The Anahim volcanic belt, which extends in an east-west direction across western British Columbia at about 52°N, includes about 30 Quaternary volcanoes of mainly basaltic composition. This belt may be the product of progressive movement of the America plate over one or more hot spots in the mantle (Bevier et al., 1979; Rogers, 1981; Souther, 1986; Souther et al., 1987). The Stikine volcanic belt of northwestern British Columbia and southeastern Alaska contains more than 50 Quaternary eruptive centres, at least one of which is younger than 150 years old. The volcanoes are mainly basaltic in composition like those of the Anahim belt. Souther (1977) suggested that this belt may have formed in response to tensional stresses related to right-lateral shear between the Pacific and America plates. However, the position of the Stikine belt far to the east of the Pacific-America plate boundary and its termination near 60°N are difficult to explain in the context of the present plate tectonic regime. The northernmost, or Wrangell, volcanic belt defines a broad arc of calc-alkaline eruptives that extends through the St. Elias Mountains from southwestern Yukon Territory to central Alaska. Although the Canadian part of the belt is dominated by Tertiary lavas, the Alaskan part includes numerous Quaternary volcanoes which probably have formed in response to subduction of the Pacific plate at Aleutian Trench (Van Wormer et al., 1974).

Finally, lithospheric plate interactions are also responsible for recent vertical crustal movements in the Canadian Cordillera. Evidence from tidal records and geodetic releveling indicates that there is a consistent pattern of uplift in outer coastal areas of British Columbia (up to 2 mm/a) and subsidence or no change on the inner coast (0-2 mm/a); the zero-uplift contour or "hinge line" runs through Hecate Strait and either the Strait of Georgia or Vancouver Island (Clague et al., 1982b; Riddihough, 1982). This pattern continues southward into Washington (Ando and Balazs, 1979), but is interrupted to the north by considerable uplift in southeastern Alaska and bordering areas of British Columbia and Yukon Territory (Hicks and Shofnos, 1965; Vaníček and Nagy, 1980, 1981). Parts of the rugged St. Elias Mountains and northern Coast Mountains are presently rising at rates of several millimetres to a few centimetres per year. Contemporary uplift also has been postulated for other areas. For example, an area in southeastern British Columbia, purportedly above a mantle hot spot, may be rising at a rate of several millimetres per year (Vaníček and Nagy, 1980; Riddihough, 1982). The pattern of contemporary uplift and subsidence in the Cordillera indicates that the movements are mainly tectonic in origin (Clague et al., 1982b; Riddihough, 1982). However, some uplift in heavily glacierized areas such as the St. Elias Mountains (e.g., at Glacier Bay) may be an isostatic response to recent deglaciation (Hicks and Shofnos, 1965).

PHYSIOGRAPHY AND DRAINAGE

J.J. Clague

The Canadian Cordillera comprises four main landscape elements: mountains (including "highlands"), plateaus, lowlands, and valleys (including fiords) (Fig. 1.4 to 1.7). Local physiographic units consisting of one or more of these landscape elements may be grouped into three major "systems" that extend the length of British Columbia and Yukon Territory: the Western System, Interior System, and Eastern System (Fig. 1.1; Bostock, 1948, 1970; Holland, 1964).

The Western System includes: (1) the great mountain ranges of southwestern Yukon Territory (St. Elias Mountains) and western British Columbia (Coast, Cascade, and Insular mountains); (2) narrow strips of coastal lowland bordering the British Columbia coast; (3) intramontane valleys and fiords; and (4) the seafloor of the British Columbia continental margin. The Western System also contains the largest icefields in the Cordillera and the highest peaks in Canada, including Mount Logan (ca. 5951 m) and Mount St. Elias (5489 m) (Fig. 1.8).

The Interior System is dominated by the large plateaus of interior British Columbia and Yukon Territory and by bordering mountain ranges. The main plateaus are the Yukon, Stikine, and Interior plateaus. Major mountain systems include the Ogilvie, Selwyn, Pelly, Cassiar, Omineca, Skeena, and Columbia mountains (some of these comprise two or more named ranges). The boundary between the Interior and Eastern systems in British Columbia is Rocky Mountain Trench, a remarkable topographic and structural feature that extends, with one major gap, over 1400 km in a northwesterly direction from the forty-ninth parallel to near the British Columbia-Yukon border. Tintina Trench, a similar structurally controlled valley, crosses the Interior System of Yukon Territory along the same trend.

The Eastern System is a region of rugged mountains, plateaus, lowlands, and valleys bordering the Interior Plains on the east. Its main elements are the Arctic, Mackenzie, Franklin, and Rocky mountains, and Peel Plateau.

Parts of the watersheds of the Pacific Ocean, Bering Sea, Arctic Ocean, and Hudson Bay occur within the Canadian Cordillera, and most of the large rivers of western Canada flow through or head in this region (Fig. 1.1). Fraser River and Columbia River drain most of the southern half of British Columbia and empty into the Pacific Ocean. Skeena River and Stikine River drain much of northwestern British Columbia and also flow to the Pacific. Saskatchewan River, which heads in the southern Rocky Mountains of Alberta, is within the Hudson Bay watershed. Athabasca River to the north also heads in the Rocky Mountains, but is part of the Mackenzie River system emptying into the Arctic Ocean. Most of north-central and northeastern British Columbia is drained by Peace and Liard rivers which are major tributaries of Mackenzie River. Western District of Mackenzie and parts of northern Yukon Territory are the sources of several streams which also empty into the Mackenzie. Mackenzie River itself is outside the Cordillera, except for a 500 km reach northwest of the mouth of Liard River (Fig. 1.1). Much of Yukon Territory is drained by Yukon River and its tributaries; this large river flows across Alaska into Bering Sea.

CLIMATE

J.M. Ryder

The climate of the Canadian Cordillera is controlled primarily by: (1) the location of the region in middle to high latitudes along a continental margin adjacent to the Pacific Ocean and (2) topography, with several mountain belts trending parallel to the coastline. The dominant atmospheric movement in this region is eastward, with moist air masses and cyclonic storms generated over the Pacific

Clague, J.J.
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Ryder, J.M.
1989: Climate (Canadian Cordillera); in Chapter 1 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

Ocean moving inland across successive mountain ranges, plateaus, and valleys (Hare and Hay, 1974; Hare and Thomas, 1974).

The coastal region, west of the crest of the Coast, Cascade, and St. Elias mountains, is dominated by warm moist Pacific air. Much of this region receives more than 2500 mm of precipitation per year and thus is by far the wettest part of Canada (Hare and Hay, 1974). Winter precipitation along the coast results from a succession of frontal systems associated with cyclonic storms in the Gulf of Alaska; frontal precipitation is enhanced orographically so that amounts increase markedly with elevation. Snowfall accounts for only a small fraction of precipitation near sea level, but thick snowpacks accumulate during winter in mountains adjacent to the coast. In summer, the coastal region comes under the influence of a large anticyclone and there are spells of fair weather with infrequent convective storms. Air temperatures vary much less on a daily and seasonal basis than farther inland due to the moderating influence of maritime air masses. Winters are mild; mean January temperatures range from about 1°C to 5°C, decreasing rapidly inland from the open coast (Atmospheric Environment Service, n.d.). Along the coast, the mean annual temperature range (i.e., the difference between the mean temperatures of the warmest and coldest months) is only 10-15°C, the lowest in Canada.

The southern interior region, which extends eastward from the crest of the Coast and Cascade mountains to the

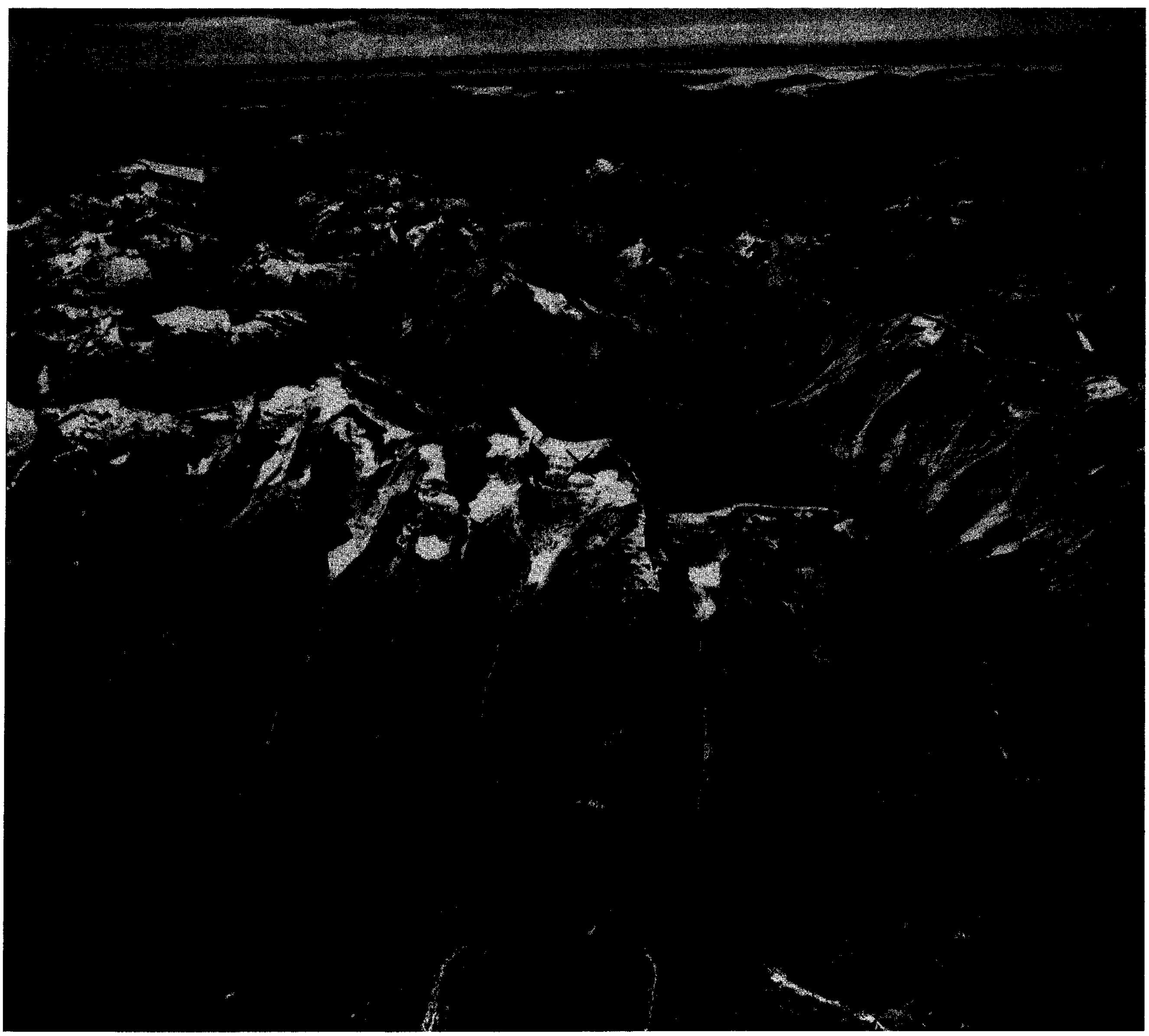

Figure 1.4. Coast Mountains west of Kitimat, British Columbia. Province of British Columbia BC528-21.

Interior Plains and northward to about 58°N, is characterized by strong climatic contrasts. Temperatures and humidity are controlled partly by modified maritime air masses which have lost much of their moisture during their passage across the coastal mountains. However, incursions of continental arctic air in winter and continental tropical air in summer produce extreme and variable conditions. Precipitation is more evenly distributed throughout the year in this region than on the coast, although the proportion of snow to rain is greater. Local rain shadows and orographic effects determine the regional distribution of precipitation. Some valleys in the south receive less than 250 mm of precipitation per year, whereas valleys in the northern part of this region receive 400-600 mm (Atmospheric Environment Service, n.d.). Greater amounts of precipitation fall on plateaus (700 mm/a or more) and in the Columbia and Rocky mountains (up to 1500-2000 mm/a on the western flanks). In general, air temperatures are warmest in the south and decrease both northward and with increasing elevation. Mean valley floor temperatures in

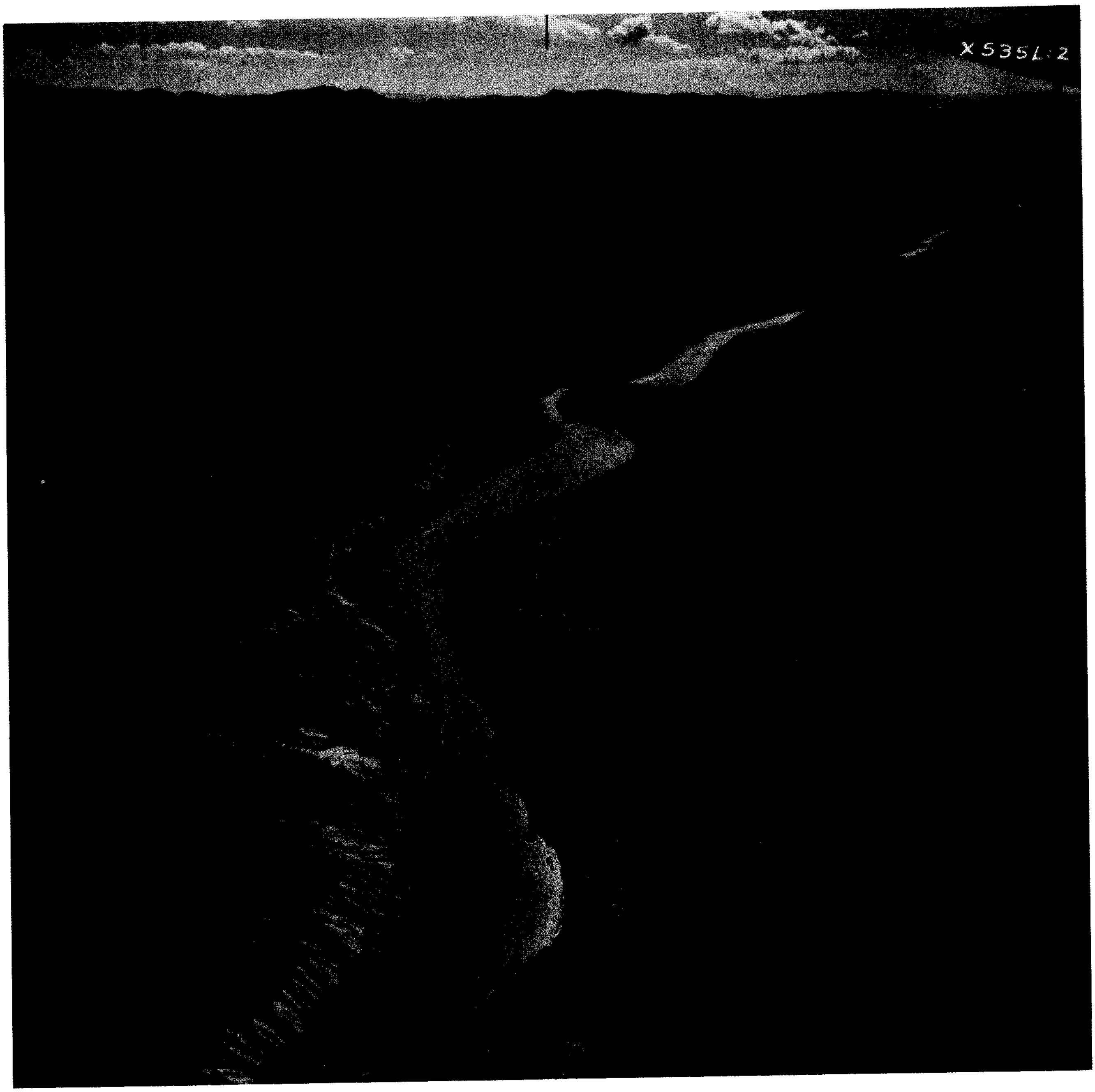

Figure 1.5. Fraser River valley west of Clinton, British Columbia. Province of British Columbia BC1087-46.

January range from −3°C to −6°C in the south to −10°C to −17°C in the north; corresponding mean July temperatures are 18-21°C and 14-17°C (Atmospheric Environment Service, n.d.). The annual temperature range in the southern interior climatic region is about 26°C, twice that of the coast.

The northern Cordilleran region, which includes British Columbia north of 58°N, Yukon Territory, and western District of Mackenzie, experiences a climate that is similar to other high latitude boreal forest regions of Canada. Winters are long (October-April) and are dominated by extremely cold dry continental arctic air. Precipitation amounts are low (300-400 mm/a on plateaus and in valleys and lowlands) due to the position of this region on the lee side of the St. Elias and Coast mountains and due to the low water-vapour holding capacity of the cold air masses (Kendrew and Kerr, 1955; Atmospheric Environment Service, 1982). There is no pronounced wet or dry season, although spring is relatively dry and many localities have modest summer precipitation maxima.

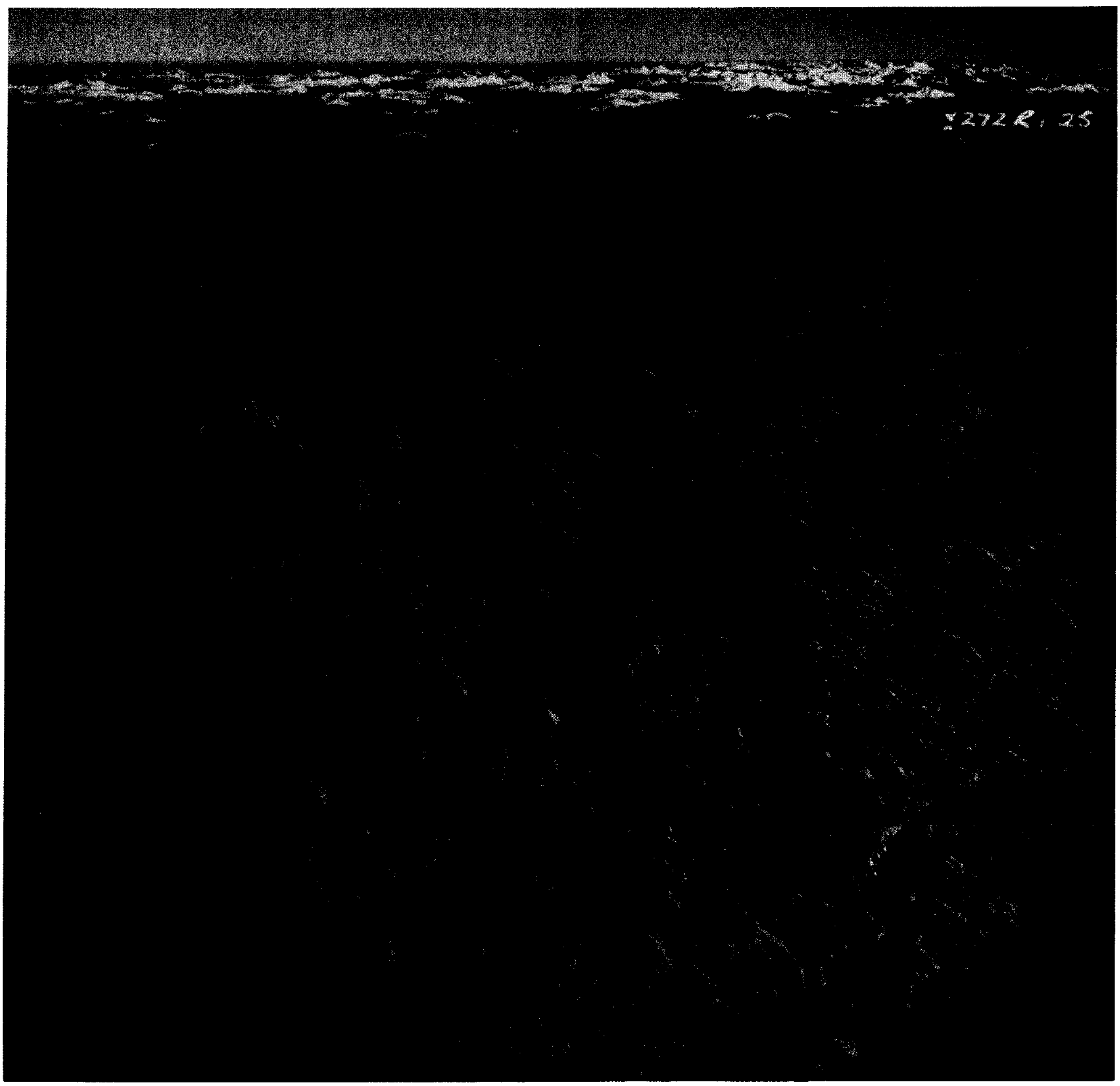

Figure 1.6. Drumlinized plateau northeast of Prince George, British Columbia; view northeast towards the Rocky Mountains. Province of British Columbia BC761-71.

Permafrost is continuous in northern Yukon Territory and is common although discontinuous elsewhere in the Yukon and in the mountainous part of District of Mackenzie. In British Columbia, permafrost has a patchy distribution above 1200 m elevation in the north and above 2100 m in the south (Brown, 1967; Hare and Thomas, 1974).

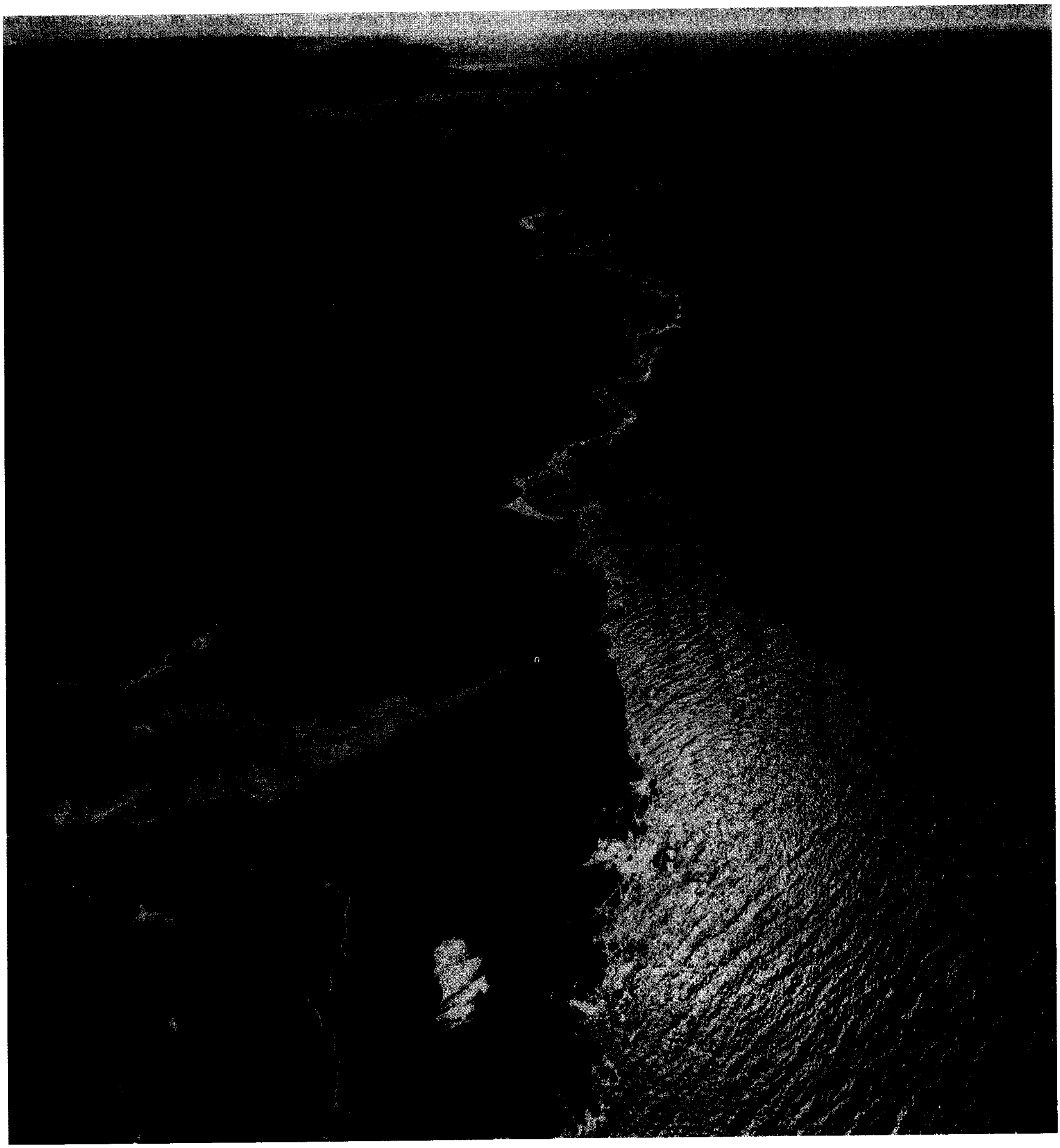

Figure 1.7. Strandflat and mountain fringe, western Vancouver Island. Province of British Columbia BC666-93.

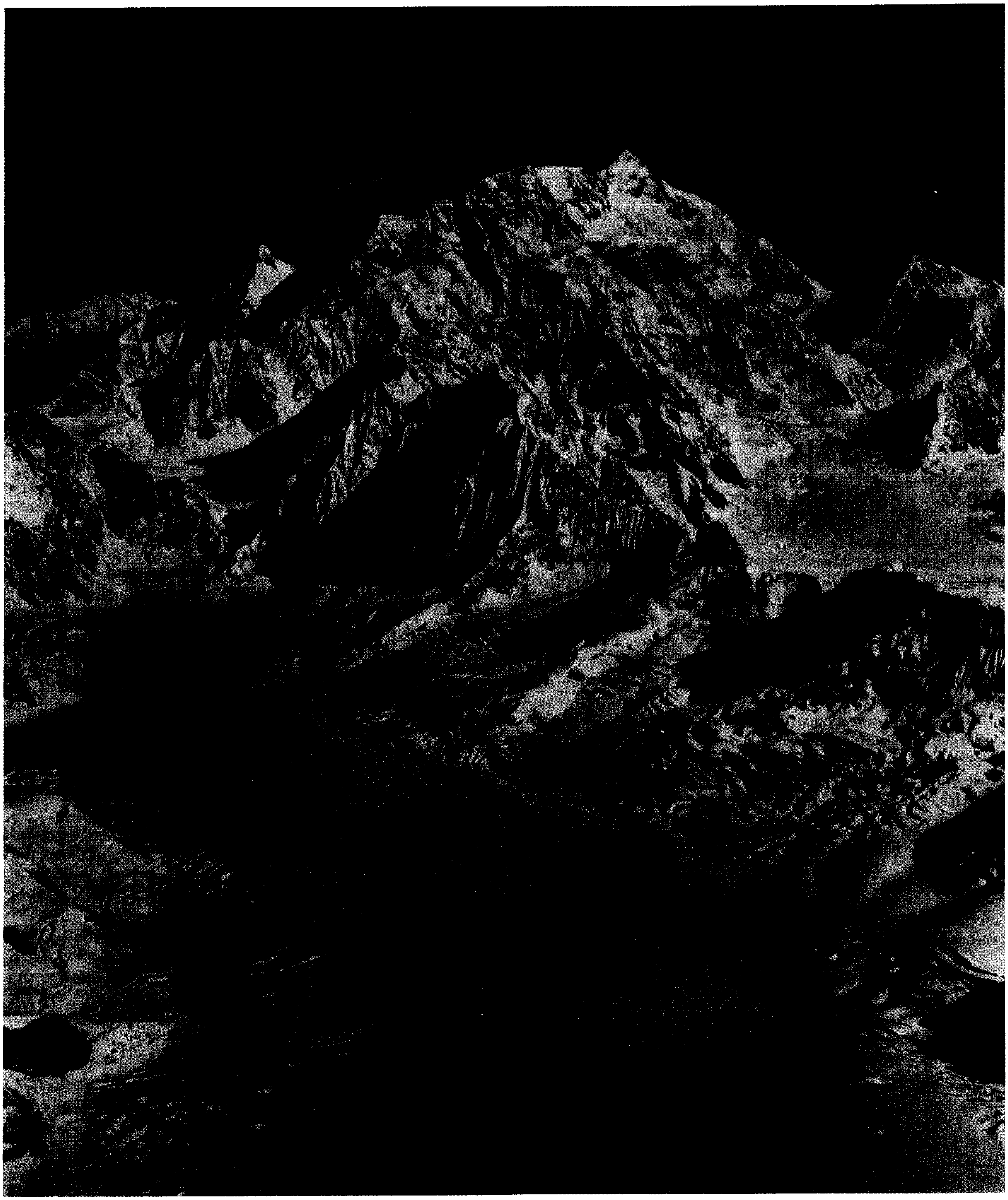

Figure 1.8. Mount St. Elias, the second highest mountain in Canada. Courtesy of Austin Post.

DEVELOPMENT OF CORDILLERAN LANDSCAPES DURING THE QUATERNARY

W.H. Mathews

An understanding of landscape evolution in the Canadian Cordillera involves two more or less interrelated questions: (1) what was the landscape like at the beginning of the Quaternary (or how did it differ from that of the present) and (2) what geomorphic processes were effective during the Quaternary in bringing about physiographic change.

The landscape at the beginning of the Quaternary

Data to answer the first question are limited because of the scarcity of well dated deposits from this part of geological time. Nevertheless, some generalizations can be made regarding the physiography of the Canadian Cordillera at the start of the Quaternary Period. These generalizations are based largely on observations of the distribution and character of late Tertiary sedimentary and volcanic rocks and landforms.

The Cordilleran landscape at the beginning of the Quaternary was dominated by mountainous uplands, dissected plateaus, and alluvial plains; it thus probably resembled the landscape of present-day northern Yukon Territory. The distribution of high and low ground was probably much as it is today, but the relief was almost certainly significantly less. Valleys, particularly those of the interior, were not cut to anything like their present depths nor were they yet extensively modified by glaciation. The streams that occupied them may have differed in size and even in direction of flow from those now existing. The mountains were almost certainly lower than their modern counterparts, and their glacial modification, though probably present, may have been much more restricted than now in both distribution and degree. The coastline was probably less indented than at present, looking perhaps somewhat like the northeast shores of the Queen Charlotte Islands or those bordering the northern Gulf of Alaska today. The possibilities of sea level changes associated with Pliocene glaciations in Antarctica could have led to incision of river mouths followed by drowning to form estuaries such as Masset Sound on the Queen Charlotte Islands. It is questionable whether any local Pliocene glacier reached the sea to create the fiords which are such a distinctive feature of the present west coast. The existence at that time of fiord lakes in the interior valleys is also in doubt.

Geomorphic processes

Major geomorphic processes that have shaped the surface of the Cordillera during Quaternary time include diastrophism, volcanism, subaerial erosion and deposition, and glaciation.

Mathews, W.H.
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Diastrophism

Diastrophism has resulted in uplift and subsidence of various parts of the Cordillera during the Quaternary, as illustrated by the following examples:

1. Miocene lavas in the northeastern St. Elias Mountains and southern Coast Mountains have been tilted, faulted, and uplifted (Tipper, 1963; Souther and Stanciu, 1975). On the basis of apatite fission-track dates, Parrish (1981) concluded that there has been sustained uplift and denudation of Mount Logan in the St. Elias Mountains averaging about 0.3 m/ka for the past 15 Ma. Somewhat slower uplift, generally >0.2 m/ka, is recorded in the southern Coast Mountains over the last 10 Ma (Fig. 1.9; Parrish, 1983). Tilting and deep incision of Pliocene and early Quaternary lavas along Fraser River between 51°N and 52°N suggest that this uplift has continued in the Quaternary.
2. Dissected gravel-capped pediplains are locally present over the folded and faulted rocks of the eastern foothills of the Cordillera from Montana north to the Richardson Mountains (Alden, 1932; Ross, 1959; Roed, 1975; Rampton, 1982); some also occur on the Interior Plains to the east. These relict surfaces range in age from Early Oligocene to late Tertiary (Russell, 1957; Russell and Churcher, 1972; Mathews, 1978; Rampton, 1982). They were uplifted and incised during middle to late Tertiary and Quaternary time as the mountains to the west rose.
3. Additional evidence for relatively recent uplift of the Rocky Mountains comes from a study of speleothems in relict phreatic caves preserved in valley walls. Uranium-series dates from such speleothems provide information on rates of valley deepening; data from the southern Rockies (Ford et al., 1981) indicate downcutting at rates of >0.04-2.07 m/ka over the past several hundred thousand years, and by extrapolation suggest that the mean relief of these mountains, some 1340 m, could have developed over a span of 0.65-33.4 Ma, most probably 1.2-12 Ma.
4. In the Klondike area of central Yukon Territory, the "White Channel gravels", assigned a Pliocene and early Pleistocene age (Hughes et al., 1972; Naeser et al., 1982), mark a series of river courses since incised to depths of about 100 m. Hughes et al. (1972) concluded that tectonic tilting contributed to the aggradation of these and associated Yukon River gravels which are now 150 m higher downstream from Dawson than at the mouth of Stewart River, 150 km upstream.
5. Tectonic subsidence or faulting has been offered as the explanation for Old Crow, Bluefish, and Bell flats in northern Yukon Territory (Hughes, 1972), and for the Strait of Georgia (Mathews, 1972) and Hecate Strait (Yorath and Hyndman, 1983). Downwarping of late Miocene strata beneath the floor of Hecate Strait (Fig. 1.9) indicates that much of the subsidence there is Plio-Pleistocene in age.

Volcanism

Volcanism, which generated widespread basaltic lavas in the British Columbia interior in the late Tertiary, continued on a reduced scale during the Quaternary. Most Quaternary volcanoes in the Canadian Cordillera are localized in four relatively narrow belts (Fig. 1.3; see also Quaternary tectonic setting). The eruptive products are mainly basalts, except in the Garibaldi belt in the south where andesites and dacites of calc-alkaline affiliation are dominant. The most common volcanic landforms are steep-sided pyroclastic cones, large eroded stratovolcanoes, and intra-valley flows. Anomalous landforms at several centres have been ascribed to eruption under, through, or onto glacier ice or into glacier-dammed lakes. A flat-topped version, the "tuya" (Mathews, 1947), consists of a cap of gently sloping, subaerially cooled lava flows crowning a pile of outward-dipping brecciated basalt and/or pillow lava laid down within a glacier-dammed lake. Another form, most closely resembling a giant sawn-off tree stump, is believed to be a slightly modified plug that was intruded through the Cordilleran Ice Sheet (Mathews, 1951b). Yet other forms, which are esker-like in shape, probably were produced when lava solidified while flowing through trenches or tunnels in the ice (Mathews, 1958).

Subaerial erosion and deposition

Subaerial erosion accompanying and following late Cenozoic uplift did much to develop the major valleys of the Cordillera. Fluvial incision, particularly in its early stages, could be expected to form V-shaped valleys. Subsequent occupation by valley glaciers could then lead to their widening and deepening. Although the relative importance of these two processes in glaciated parts of the Cordillera cannot be completely evaluated, the common localization of valleys along narrow fracture systems or belts of nonresistant rocks is more consistent with the selective action of streams than with the less precise scouring of glaciers. Accordingly, most Cordilleran valleys probably were initiated by steams prior to glaciation (some exceptions are noted in the next section). In unglaciated northern Yukon Territory, streams alone controlled erosion. However, even here periodic glacial activity to the east and south could have influenced stream behaviour by adding to the sediment load and encouraging aggradation of such streams as Klondike and Yukon rivers (Hughes et al., 1972).

Though streams may have been instrumental in cutting the major valleys of the Cordillera, they did not necessarily flow along them in the same direction as their modern counterparts. Indeed, stream diversions and reversals took place frequently during the Quaternary. Many diversions resulted from the blockage of stream courses by glaciers, but damming and diversion of waters by glacial and fluvial deposits or more locally by lava flows or landslides were also important in rearranging the drainage pattern.

Mass movement processes have contributed to the denudation of mountains and the widening of valleys during the Quaternary by transferring material from oversteepened mountainsides to sites where the debris could be carried away by streams. Weathering and soil formation accompanied by removal of the alteration products have been less important in this respect. Periglacial processes, such as solifluction, cryoplanation, and thermokarst development, limited today to high latitudes and high altitudes, were more widespread during cooler nonglacial stages of the Pleistocene and during early parts of glacial cycles. Such processes were probably more important at inland sites, with a relatively dry continental climate, than near the coast where a blanket of winter snow or permanent glacier ice would inhibit periglacial activity.

A strandflat (i.e., low coastal rock platform abutting against a sharply rising mountain slope) is recognizable along the Pacific coast from Vancouver Island northward

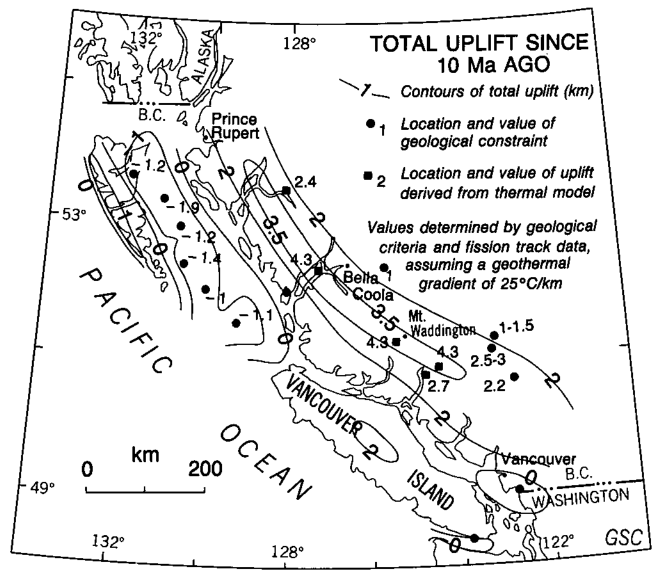

Figure 1.9. Total uplift of western British Columbia during the last 10 Ma (adapted from Parrish, 1983, Fig. 11).

(Fig. 1.7; Holland, 1964). It extends as much as 10 km inland and to an elevation of about 150 m; locally it comprises swarms of small islets and rocky reefs more or less modified by wave action. The origin of the strandflat is debated, but accelerated Pleistocene frost weathering in the presence of abundant moisture and poor drainage likely contributed to its formation (Holtedahl, 1960); marine planation and glaciation seem to have played a lesser role.

Subaerial depositional processes have had a significant, although localized, impact on the Cordilleran landscape. Deltas, floodplains, and alluvial fans are the most important features produced by fluvial deposition; landslides and colluvial fans are the main products of mass wasting.

Glaciation

Glaciers first formed in the northern Cordillera more than 9 Ma ago (Denton and Armstrong, 1969) and since then have profoundly altered the landscape of the region. In the high mountains, the gathering ground for the Cordilleran Ice Sheet and satellite glaciers, where summits and ridges stood above névé surfaces, classic alpine landforms were created, including cirques and overdeepened valley heads, horns, and comb ridges. In the major valleys, especially along the Pacific shore, the effects of glaciation are even more conspicuous. The coastline is highly indented by fiords extending as much as 150 km inland and to water depths of as much as 755 m (Peacock, 1935; Pickard, 1961). Most fiords have reaches with relatively flat floors stemming from partial infilling by sediments up to several hundred metres thick. Nevertheless, longitudinal profiles at the sea-sediment interface are generally highly irregular, and even more so on the sediment-bedrock interface. The major fiords trend more or less perpendicular to the coastline, but oblique and cross channels linking these fiords to one another are common.

The continental shelf between Vancouver Island and the Queen Charlotte Islands is crossed by three major streamlined U-shaped troughs 20-30 km wide and 100-300 m deep (Luternauer and Murray, 1983). These share many of the characteristics of the Laurentian Channel west and south of Newfoundland and others off the Labrador coast. The similarity in pattern and scale of these troughs to those carrying fast-moving ice streams within the glacier apron draining westerly to the Ross Ice Shelf, Antarctica (Rose, 1979) supports the idea that they owe their origin to glacial scour. Juan de Fuca Strait, the Strait of Georgia, and Dixon Entrance also have been extensively modified by streams of ice at the western margin of the Cordilleran Ice Sheet.

East of the axis of the Coast and St. Elias mountains, the erosional effects of glaciation, though not as spectacular as on the coast, are still evident. Deep elongate lakes take the place of the coastal inlets. These fiord lakes are most common at the margins of major mountain systems, are up to 125 km long, and are known to have depths of as much as 418 m. Most are dammed at their outlets by either glacial deposits or alluvial fans.

Some valleys in the Cordillera, such as those of Homathko River crossing the southern Coast Mountains, Kootenay River between Kootenay Lake and Columbia River, and Mackenzie River near The Ramparts (Mackay and Mathews, 1973), show no signs of channel fills predating the last glaciation. It is likely that these valleys mark the sites of former bedrock saddles which have been deepened by glacial scour, aided perhaps by meltwater streams, so that in late glacial time they could be occupied by through-going rivers which had formerly discharged by other routes.

Smaller-scale glacial landforms, almost all of which were produced during the last (Late Wisconsinan) glaciation, are conspicuous in many parts of the Canadian Cordillera. Drumlins, flutings, and other streamlined forms are products of subglacial erosion and/or deposition and are found on perhaps 50% of the plateau surface of the Cordilleran interior (Armstrong and Tipper, 1948; Tipper 1971a, b). In contrast, morainal ridges are rare except in front of modern glaciers. Glacial fluvial landforms, including abandoned meltwater channels, compound eskers, kames, kame terraces, outwash plains, and valley trains, are striking features of the landscape and record the decay of the Cordilleran Ice Sheet at the close of the Pleistocene. Incised depositional plains of glacial lacustrine and glacial marine origin are found in some interior valleys and along the coast, respectively, but are subdued and inconspicuous except in relatively arid grassland areas.

CHARACTER AND DISTRIBUTION OF QUATERNARY DEPOSITS

J.J. Clague

Quaternary deposits in the Canadian Cordillera are extremely varied in character and have complex distributions controlled largely by physiography and glacial history. Most of the surface sediments were deposited during the last glaciation and during postglacial time. Older materials occur at the surface in central and northern Yukon Territory, District of Mackenzie, the eastern Rocky Mountains, and on parts of the continental shelf; they also underlie Late Wisconsinan glacial deposits in some valleys and lowlands in British Columbia and the southern Yukon. Surficial geology maps prepared by the Geological Survey of Canada, British Columbia Ministry of Environment, and Alberta Research Council show the distribution of surface Quaternary deposits and landforms for many parts of the Canadian Cordillera. Map 1704A shows the areas covered by these maps and gives a list of references.

Morainal deposits (till)

In the Canadian Cordillera, sediments deposited directly from glacier ice in subglacial and supraglacial settings typically are poorly sorted, massive to weakly stratified, and very stony; they have matrixes of mixed sand, silt, and clay (Fig. 1.10). Morainal sediments vary in character both locally and regionally because of differences in source materials, complexities in the pattern of glacier flow, the presence of diverse depositional environments, and the effects of water and gravity during deposition (Keser, 1970; Scott, 1976). In general, there is a direct relationship between local bedrock type and till composition and texture. Till derived from volcanic rocks, carbonates, mudstone, shale, or slate typically has a matrix rich in silt and clay; in contrast, till derived from granitic rocks, schist, gneiss, sandstone, or conglomerate has a sandy matrix. Some till is composed of particles eroded from older Quaternary sediments and commonly exhibits the textural and compositional properties of these materials. For example, in the Strait of Georgia region of southwestern British Columbia, the surface till is very sandy and contains abundant, well rounded granitic pebbles and cobbles. It was deposited by glaciers that flowed across sandy outwash rich in granitic detritus.

Till in high-relief areas generally is more gravelly and has less silt and clay than till in lowlands and on plateaus. Some coarse till on steep mountain slopes is similar to poorly sorted glacial fluvial and colluvial sediments and, in fact, there is a complete spectrum of materials ranging from till to glacial fluvial gravel on one hand, and from till to colluvium on the other.

Till probably is the most extensive of all surficial materials in the Canadian Cordillera. It covers most plateaus in British Columbia and southern Yukon Territory and also is common in some lowlands. In mountain areas, it occurs on the floors of some valleys and on adjacent slopes. At high elevations and on steep slopes, however, till generally is either thin and patchy, or completely absent.

Glacial fluvial deposits

Glacial fluvial sediments include both ice contact and outwash materials deposited in subaerial environments and in lakes and the sea. Ice contact fluvial materials consist of gravel and/or sand which exhibit variable sorting and stratification. Where present, bedding commonly is distorted and broken due to melt of associated ice during and shortly after deposition. Ice contact sediments in the Canadian

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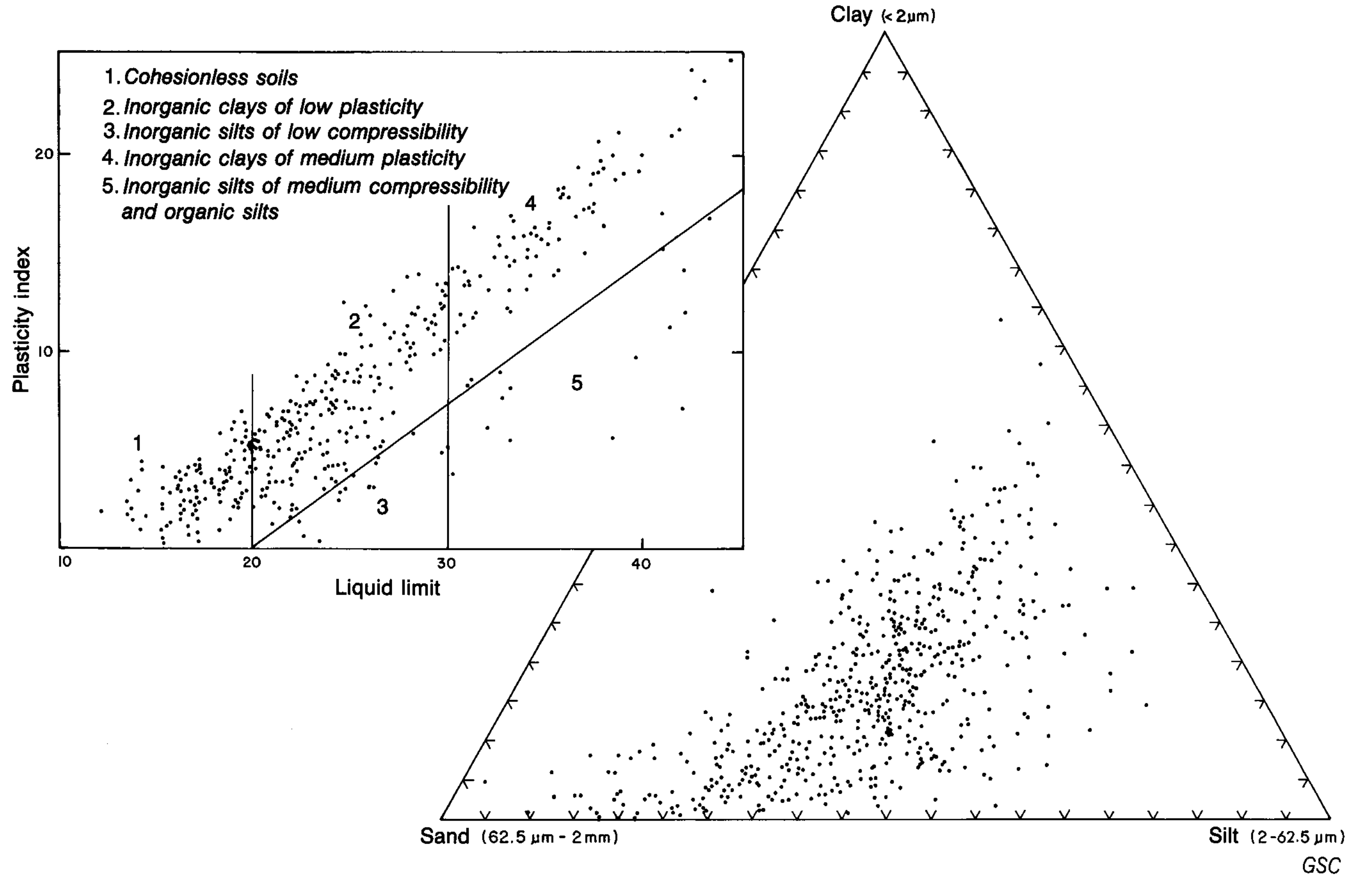

Figure 1.10. Particle-size distribution (<2 mm) and engineering properties of representative till samples from the Canadian Cordillera. Plasticity chart and nomenclature after Casagrande (1932). Data, in part, from Fyles (1963), Clague (1973), Jackson (1977, 1987), Howes (1981b), Ryder (1981a, 1985), Clague and Luternauer (1983), and Klassen (1987); unpublished data provided by J.J. Clague, D.E. Howes, R.W. Klassen, L. Lacelle, J.M. Ryder, and B. Thomson.

Cordillera are preserved in kames, kame terraces, knob-and-kettle topography, and eskers, and also occur as thin discontinuous sheets of gravel and sand overlying till. These sediments are widely distributed, except in rugged alpine terrain and, of course, in areas that escaped glaciation. Major esker and kame complexes and associated large meltwater channels are present on the Yukon and Interior plateaus and in Liard Lowland (Armstrong and Tipper, 1948; Prest et al., 1968; Tipper, 1971a, b).

Outwash consists of stratified, well sorted gravel and sand and occurs as valley trains, plains, terraces, and deltas. These materials generally lack deformation structures characteristic of most ice contact deposits. In the Canadian Cordillera, large amounts of outwash were deposited at the close of the last glaciation on proglacial floodplains and deltas and in subaqueous fans that prograded outward from glaciers terminating in the sea or in lakes. Outwash is common in most large valleys and lowlands and also is present locally on the British Columbia continental shelf. Late Pleistocene valley trains, outwash plains, and deltas were incised by streams during the Holocene, and thus now occur as terraces above present-day floodplains. In contrast, active outwash trains extending downvalley from large glaciers in the high mountains of the Cordillera are graded to existing streams.

Glacial fluvial sediments also occur beneath other types of Quaternary sediments and have little or no surface expression. Some of these subsurface glacial fluvial materials predate the Late Wisconsinan glacial maximum.

Glacial lacustrine deposits

Glacial lacustrine deposits consist mainly of fine sand, silt, and clay carried into glacier-dammed lakes by meltwater streams and deposited from overflows, interflows, and underflows; coarser sediments generally are restricted to subaqueous outwash fans, deltas, and beaches. Most glacial lacustrine sediments are well stratified, and many are laminated or varved, suggesting deposition under relatively stable conditions; others have irregular or disturbed stratification and apparently were deposited as ice contact sediments.

These materials are sufficiently thick in many areas to completely mask the morphology of underlying sediments or bedrock; in these areas the old lake bed forms a level or undulating surface dissected by streams. Thick glacial lacustrine sediments deposited in contact with ice may slump or settle as the ice melts, giving rise to irregular or kettled topography.

Glacial lacustrine sediments deposited at the close of the last glaciation are common in several large valleys in British Columbia and southern Yukon Territory (Mathews, 1944; Kindle, 1952; Prest et al., 1968; Fulton, 1969; Tipper, 1971a; Clague, 1987). Similar sediments accumulated in lakes in unglaciated northern Yukon Territory when the Laurentide Ice Sheet advanced to the front of the Richardson Mountains and dammed east-flowing streams (Hughes, 1972; Morlan, 1979; Hughes et al., 1981). Glacial lacustrine silt and clay also are found in many places beneath till; these sediments probably accumulated in lakes impounded by advancing Late Wisconsinan glaciers.

Glacial marine deposits

Glacial marine sediments in the Canadian Cordillera were deposited on the sea floor in front of temperate tidewater glaciers. They consist of material that was introduced into the sea by meltwater streams, by flows of supraglacial debris directly from ice, and by the melting of icebergs. Fine detritus settled out of suspension from overflows, interflows, and underflows; some was redeposited by mass movement processes. Some coarse material was dropped from icebergs and the calving fronts of debris-laden glaciers; some was deposited at the mouths of meltwater streams.

The most common type of glacial marine sediment in the Cordillera is massive to stratified mud. Most of this material contains less than 5% gravel-size material, and some is stone-free. In contrast, some mud is very gravelly and is not easily distinguished from till. Stratified gravel and sand of glacial marine origin are restricted to deltas, subaqueous outwash fans, and beaches. These sediments are fairly well sorted, except where laid down directly against ice.

Glacial marine sediments occur locally on the coastal lowlands of western British Columbia (Fyles, 1963; Armstrong, 1981; Clague, 1981, 1985) and on the adjacent continental shelf (Luternauer and Murray, 1983). Most of the sediments were deposited at the close of the last glaciation as the Cordilleran Ice Sheet decayed and retreated east and north in contact with the sea. At that time, coastal lowlands in most areas were submerged because the crust was isostatically depressed by the decaying ice sheet. Glacial marine sedimentation ceased when glaciers disappeared from coastal areas. This was accompanied by the emergence of coastal lowlands due to isostatic rebound. Older glacial marine sediments, deposited during one or more glaciations preceding the Late Wisconsinan, have been identified on eastern Vancouver Island and in Fraser Lowland (Fyles, 1963; Armstrong, 1981; Howes, 1981a; Hicock and Armstrong, 1983).

Colluvial deposits

Sediments produced by mass movement processes are common at the surface throughout the Canadian Cordillera and also occur locally in the subsurface beneath Late Wisconsinan drift. There are four main types of colluvium in this region: (1) Landslide deposits. Deposits produced by landslides range from fine textured masses of displaced glacial lacustrine and glacial marine sediments to coarse blocky accumulations of broken rock. Their physical characteristics are determined by the geology of the source materials, the topography of the failure site, and the type of landslide motion (e.g., fall, topple, slide, spread, flow; Varnes, 1978). Most landslides have irregular hummocky surfaces, although highly fluid, channelized flows of debris and mud characteristically build cones or fans similar to alluvial fans. Although widely distributed in the Canadian Cordillera, landslide deposits cover a very small part of the total surface area (<1%). (2) Talus. Cones, aprons, and thin mantles of broken rock occur below steep bedrock slopes in all mountain areas. Some thick talus accumulations have glacier-like forms resulting from the deformation of interstitial ice or the flow of subjacent glacier ice (i.e., rock glaciers). (3) Colluviated drift. Sheet erosion and creep have gradually transformed some Late Wisconsinan glacial deposits into colluvium. Thin mantles of diamicton and poorly sorted gravel formed in this way are common on moderate and steep slopes in most parts of the Cordillera. The role, if any, of freeze-thaw activity in the genesis of these sediments is unknown. (4) Solifluction deposits. Masses of weathered rock and unconsolidated sediments which have flowed slowly downhill in response to cyclic freezing and thawing are common both in alpine areas and at lower elevations on slopes with impervious (commonly frozen) substrates.

Fluvial (alluvial) deposits

Stratified sediments deposited by streams are associated with Holocene and contemporary channels, floodplains, terraces, fans, and deltas, and also occur beneath Late Wisconsinan drift.

Streams with steep gradients and high sediment loads tend to have shallow braided channels and gravelly floodplains. They are most common in mountain areas. Streams with low gradients and low to intermediate sediment loads typically have sinuous or meandering channels. Active channels of low gradient streams commonly are floored by sand and/or fine gravel, whereas their floodplains are underlain mainly by sand and silt deposited during floods. Streams of this sort occur in broad, low gradient valleys, mostly outside mountain ranges.

Fluvial terraces are underlain by gravel similar to that found on floodplains of braided streams (note: a thin veneer of sand and silt commonly overlies the gravel). They formed during the Holocene when streams incised Pleistocene valley fills.

Alluvial fans consist of gravel, sand, and minor diamicton. Sediments generally are coarser at the apex of a fan than at the toe. Many alluvial fans are inactive relict features that formed during and shortly after deglaciation at the close of the Pleistocene, whereas others have accumulated sediment throughout the Holocene and remain active today.

Deltas are gently sloping alluvial surfaces built out into lakes or the sea by streams. Deposition occurs mainly on subaqueous foreslopes that drop off from the delta top to the lake or sea floor beyond. The texture of deltaic sediments depends to a large extent on the gradient of the source stream and the calibre of the material carried by that stream. Deltas of steep mountain torrents consist of coarse gravel, whereas those of large, low gradient streams such as the Fraser, Columbia, and Skeena rivers consist mainly of sand and silt. Fine deltaic sediments are overlain by coarser alluvium at the mouths of many rivers and for some distance upstream. Some active deltas in the Canadian Cordillera began to grow during deglaciation at the close of the Pleistocene; others were initiated during the Holocene. In addition to active deltas, there are terraced relict deltas that were built into former lakes and the sea and later incised by streams.

Lacustrine and marine deposits

Lacustrine and marine sediments resemble their glacial counterparts (glacial lacustrine and glacial marine sediments), but unlike them, accumulated in environments free of glacier ice. These nonglacial materials include offshore silt and clay deposited from suspension and locally redistributed by sediment gravity flows, and nearshore gravel and sand formed by wave and current action. In addition, coarse lag deposits occur in areas of strong bottom currents on the British Columbia continental shelf. Holocene lacustrine sediments are thickest and most extensive in large lakes with high sediment influx, for example, Upper Arrow, Lillooet, and Kamloops lakes in British Columbia (Fulton and Pullen, 1969; Gilbert, 1975; Pharo and Carmacks, 1979) and Kluane Lake in Yukon Territory (Terrain Analysis and Mapping Services Ltd., 1978). Marine sediments are widespread on the seafloor off British Columbia (e.g., Pickard, 1961; Carter, 1973; Clague, 1977b; Bornhold, 1983; Luternauer and Murray, 1983), and also occur in a narrow fringe inland from the present shore in a few areas, for example on parts of the Queen Charlotte Islands (Clague et al., 1982a, b). Lacustrine and marine sediments predating the last glaciation are also present in the Cordillera, but they typically occur beneath younger sediments, consequently their extent is not well known.

Other Quaternary deposits

Organic materials in the Canadian Cordillera include peat, gyttja, and marl. Deposits of these materials are generally less than 5 m thick and are found in shallow closed depressions, at the margins of shallow lakes, and on gently to moderately sloping, poorly drained surfaces. They occur in favourable topographic situations throughout the Cordillera, but attain their greatest extent in areas of low relief and high rainfall and in low-lying areas underlain by permafrost. Organic deposits record past vegetation changes and thus are an important source of paleoenvironmental information (see Paleoecology and paleoclimatology).

Eolian sediments, mainly sand and silt, form localized dunes and thin mantles on older deposits. In addition, fine wind blown particles are present in the surface soil throughout the Cordillera. Thick eolian sediments are restricted to valleys containing sandy floodplain sediments. Most surface eolian sediments were deposited at the close of the Pleistocene when large outwash plains and valley trains were active. Today, significant eolian sedimentation is restricted to small widely scattered sites.

Volcanic materials, mainly lavas and coarse pyroclastics of basaltic, andesitic, and dacitic composition, are found in more than 100 Quaternary eruptive centres in British Columbia and southwestern Yukon Territory (Fig. 1.3; Souther, 1970, 1977). In addition, layers of tephra, consisting of sand-, silt-, and clay-size particles, occur within nonvolcanic sedimentary deposits in many areas. Although volumetrically small, these tephras are invaluable for correlation purposes because they are widespread and of known age (Table 1.1, Fig. 1.11). Those in the southern Cordillera are the products of volcanic eruptions in western Oregon, western Washington, and southwestern British Columbia, whereas those in the Yukon were erupted from Alaskan volcanoes.

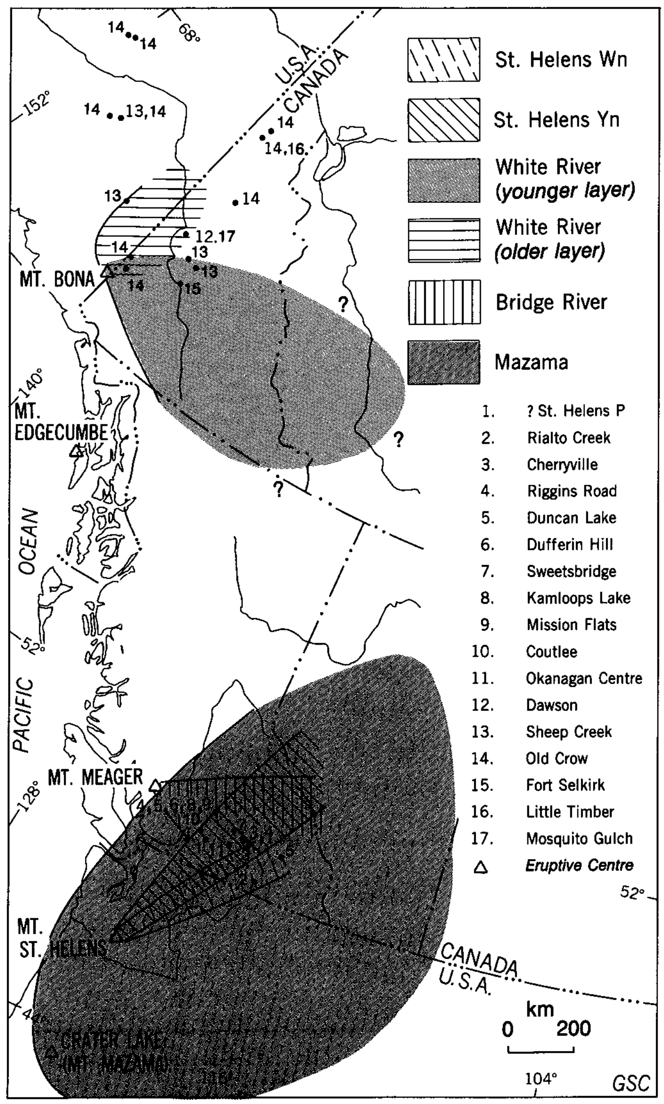

Figure 1.11. Distribution of Quaternary tephra deposits found in the Canadian Cordillera. Sources of information: Bostock (1952), Nasmith et al. (1967), Rampton (1969), Westgate et al. (1970, 1985), Fulton (1971), Hughes et al. (1972), Mullineaux et al. (1975), Westgate and Fulton (1975), Smith et al. (1977), Westgate (1977, 1982), Mathewes and Westgate (1980), Naeser et al. (1982). Tephras derived from Mount Edgecumbe are present locally in the northern Coast Mountains near the Alaska-British Columbia boundary.

Table 1.1. Documented Quaternary tephras in the Canadian Cordillera

Tephra	Age (ka BP)[1]	Source	Known occurrence in Canadian Cordillera	References
HOLOCENE				
St. Helens Wn	0.508[2]	Mt. St. Helens, Washington	south-central B.C.	Fulton, 1971; Smith et al., 1977; Yamaguchi, 1983
White River (younger layer)	1.2	Mt. Bona, Alaska	southern Yukon, western District of Mackenzie	Bostock, 1952; Lerbekmo and Campbell, 1969; Hughes et al., 1972; Lerbekmo et al., 1975
White River (older layer)	>1.5 <1.9	Mt. Bona, Alaska	west-central Yukon	Bostock, 1952; Lerbekmo and Campbell, 1969; Hughes et al., 1972; Lerbekmo et al., 1975
Bridge River[3]	2.4	Mt. Meager, B.C.	southern B.C., southwestern Alberta	Nasmith et al., 1967; Westgate and Dreimanis 1967; Westgate et al., 1970; Fulton, 1971; Westgate, 1977; Mathewes and Westgate, 1980
St. Helens Yn	3.4	Mt. St. Helens, Washington	southern B.C., southwestern Alberta	Westgate and Dreimanis, 1967; Westgate et al., 1970; Fulton, 1971; Mullineaux et al., 1975; Westgate, 1977
Mazama	6.8	Crater Lake, Oregon	southern B.C., southwestern Alberta	Powers and Wilcox, 1964; Fryxell, 1965; Westgate and Dreimanis, 1967; Westgate et al., 1970; Fulton, 1971; Bacon, 1983
PLEISTOCENE				
Rialto Creek	20	Mt. St. Helens, Washington?	south-central B.C.	Westgate and Fulton, 1975
Cherryville	25	Mt. St. Helens, Washington?	south-central B.C.	Westgate and Fulton, 1975
Riggins Road	30 ±	Mt. St. Helens, Washington?	south-central B.C.	Westgate and Fulton, 1975
Duncan Lake	34	Mt. St. Helens, Washington?	south-central B.C.	Westgate and Fulton, 1975
Dufferin Hill[4]	?	?	south-central B.C.	Westgate and Fulton, 1975
Sweetsbridge[4]	>22	?	south-central B.C.	Westgate and Fulton, 1975
Kamloops Lake	34 +[5]	?	south-central B.C.	Westgate and Fulton, 1975
Mission Flats	>35	Mt. St. Helens, Washington	south-central B.C.	Westgate and Fulton, 1975
Okanagan Centre	?[6]	?	south-central B.C.	Westgate and Fulton, 1975
Dawson	<52[7,8]	?	west-central Yukon	Naeser et al., 1982
Sheep Creek	80 ±[8,9]	Wrangell Mountains?	central Yukon	Schweger and Matthews, 1985
Old Crow	>80 ≤ 130[8,10]	Aleutians?	western Yukon	Naeser et al., 1982; Westgate, 1982; Westgate et al., 1983, 1985; Schweger and Matthews 1985; Wintle and Westgate 1986; Berger, 1987
Surprise Creek	?	?	northern Yukon	Matthews, 1987
Coutlee	>670[8,11]	?	south-central B.C.	Westgate and Fulton, 1975; Berger, 1985
Fort Selkirk	>840 <940[7,8]	?	south-central Yukon	Naeser et al., 1982
Little Timber	>1200[7,8]	?	northern Yukon	Matthews, 1987
Mosquito Gulch	1220 ±[7,8]	?	west-central Yukon	Naeser et al., 1982

1 All ages are based on radiocarbon dates unless otherwise indicated.
2 Age based on dendrochronology (1988 datum).
3 Bridge River tephra at Otter Creek bog was thought by Westgate (1977) to date about 1.9-2.0 ka and thus to be distinct from the slightly older Bridge River tephra occurring more widely in south-central British Columbia. However, more recent work (Mathewes and Westgate, 1980) has raised the possibility that the former is, in fact, the 2.4 ka tephra.
4 This tephra is perhaps close in age to the Duncan Lake tephra and, in fact, the two may be equivalent.
5 Kamloops Lake tephra is only slightly older than Duncan Lake tephra.
6 Okanagan Centre tephra probably was deposited sometime between the last glaciation (Late Wisconsinan) and the penultimate glaciation (Early Wisconsinan or Illinoian).
7 Fission-track age.
8 Age is provisional and subject to revision or refinement as a result of ongoing work.
9 Age is based on radiocarbon, thorium-uranium, and protactinium-uranium dates and paleomagnetic data.
10 Age is based on fission-track, thermoluminescence, and radiocarbon dates and paleomagnetic data.
11 Thermoluminescence age.

CONTROLS ON QUATERNARY DEPOSITION AND EROSION

J.J. Clague

Physiography probably has been the most important factor affecting deposition and erosion in the Canadian Cordillera during Quaternary time. During nonglacial periods, sedimentation occurred mainly in valleys, lakes, and the sea. The principal subaerial depositional sites were floodplains and fans. Deltas were constructed where streams emptied into lakes and the sea. Large amounts of fine sediment were deposited from suspension and turbidity flows in lakes and fiords, on parts of the British Columbia continental shelf and slope, and on North Pacific abyssal plains and basins.

Mountains were areas of erosion during nonglacial periods. Glacial, periglacial, and fluvial processes and mass wasting contributed to the denudation of mountain slopes. Sediment produced by these denudational processes initially may have been transported only a short distance, for example to a position lower on the slope or to the adjacent valley floor. Ultimately, however, much of the sediment was reentrained and carried to lowland floodplains, lakes, and the sea. Quaternary deposits and bedrock outside mountain areas also were eroded during nonglacial periods. At these times, many streams flowed in valleys incised into Quaternary deposits; escarpments in these materials supplied large amounts of sediment to the streams. Similarly, bluffs of unconsolidated deposits along the British Columbia coast were eroded by currents and waves and thus contributed material directly to the sea.

Large areas of the Canadian Cordillera were relatively stable during nonglacial periods. For example, much of the plateau surfaces in British Columbia and southern Yukon Territory experienced no significant erosion or deposition for tens of thousands of years between glaciations. Support for this supposition is provided by the fact that unmodified glacial landforms dating back to the close of the Pleistocene are widespread on low-relief surfaces in the interior of the Cordillera.

Denudation during Quaternary nonglacial periods also has been affected by climate. Rates of mass wasting and fluvial erosion at these times varied in relation to precipitation, air temperature, and vegetation. The mechanical breakdown of rock due to repeated freezing and thawing was most effective in humid areas where air temperatures regularly fluctuated around 0°C. In regions where mean annual temperature was at or below freezing (i.e., most of Yukon Territory, District of Mackenzie, and the high mountains of British Columbia and Alberta), there was significant downslope movement of sediment and shattered rock due to solifluction and shallow slope failures involving water-saturated material overlying frozen ground. Such denudation was much reduced or did not take place at all in areas lacking permafrost or seasonally frozen ground.

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The pattern of erosion and deposition during glaciations is quite different from that outlined above. Climatic deterioration at the onset of each glaciation led to increased sediment production in the mountains of the Cordillera. As glaciers advanced out of mountain areas, many interior and coastal valleys became aggraded with outwash, and thick deltaic and marine sediments accumulated locally in offshore areas. At the same time, the nonglacial drainage system was disrupted and rearranged. Lakes were impounded by advancing glaciers, and thick deposits of sand, silt, and clay accumulated in them. Those within the limits of glaciation eventually were overridden and obliterated, and their fills either eroded or buried by drift. In contrast, in northern Yukon Territory large lakes dammed by the Laurentide Ice Sheet were not subsequently overridden by glaciers, and thus their sedimentary fills are better preserved than those farther south. Advancing glaciers also forced streams out of their nonglacial valleys. Swollen with meltwater and laden with sediment, these streams followed complex paths along constantly shifting glacier margins. This situation contrasts with that in unglaciated Yukon Territory where streams generally maintained their courses during each glaciation, although they carried large amounts of meltwater and additional sediment.

At climaxes of major glaciations, a variety of sediments continued to accumulate in unglaciated parts of the Yukon, District of Mackenzie, and offshore British Columbia, and till was deposited locally beneath the Cordilleran Ice Sheet and satellite glaciers. However, mountains, fiords, and valleys parallel to the direction of ice flow were subject to intense scour by glaciers, and the unconsolidated fills of many valleys were partly or completely removed. The eroded materials, in part, were redeposited as till and, in part, carried englacially to the periphery of the ice sheet and laid down in ice contact, proglacial, and extraglacial environments.

Deglacial intervals, like periods of glacier growth, were times of rapid valley aggradation and extensive drainage change. Large glacial lakes formed and evolved as the Cordilleran Ice Sheet downwasted and separated into valley tongues that retreated in response to local conditions. In coastal areas, thick glacial marine sediments were laid down on isostatically depressed lowlands vacated by retreating glaciers and covered by the sea. Similar sediments also accumulated on the adjacent continental shelf. Throughout the glaciated Cordillera, unconsolidated sediments (mainly till) were eroded from uplands and valley walls and transported to lower elevations where they were redeposited on floodplains and fans and in lakes and the sea. Streams at first rapidly aggraded their valleys because they were unable to cope with the large amounts of sediment made available during deglaciation (Church and Ryder, 1972). However, the supply of sediment decreased as slopes stabilized and vegetation became established. Together with a fall in base level due to isostatic uplift, this reduction in sediment supply forced streams to deeply incise their late glacial floodplains, and the drainage network thus became locked into a relatively stable pattern that prevailed until the next glaciation.

CORDILLERAN ICE SHEET

J.J. Clague

Character and extent

Several times during the Pleistocene, large parts of the Canadian Cordillera were covered by an interconnected mass of valley and piedmont glaciers and mountain ice sheets, collectively known as the Cordilleran Ice Sheet (Fig. 1.12; Flint, 1971). At its maximum, this ice sheet and its satellite glaciers buried almost all of British Columbia, southern Yukon Territory, and southern Alaska, and extended south into the northwestern United States. To a considerable degree, the ice sheet was confined between the high bordering ranges of the Cordillera, although large areas on the east flank of the Rockies and west of the Coast Mountains were also covered by ice. Glaciers in several mountain ranges were more or less independent of the ice sheet, even at the climax of glaciation (Fig. 1.12).

The Cordilleran Ice Sheet attained its maximum size in British Columbia where it was up to 900 km wide and at the climax of glaciation extended to 2000-3000 m elevation over much of the Stikine and Interior plateaus (Wilson et al., 1958). When fully formed, the ice sheet probably had the shape of an elongate inverted dish, with gentle slopes in the broad interior region and steeper slopes at the periphery. At such times, it closely resembled the present-day Greenland Ice Sheet.

In western British Columbia, ice streamed down fiords and valleys in the coastal mountains and covered large areas of the continental shelf; in places ice lobes extended to the shelf edge where they calved into deep water (Prest, 1969, 1970; Luternauer and Murray, 1983). Ice from the southern Coast Mountains and Vancouver Island coalesced to produce a great piedmont lobe that flowed far into Puget Lowland in Washington (Armstrong et al., 1965; Waitt and Thorson, 1983). Glaciers streaming down valleys in south-central and southeastern British Columbia likewise terminated as large lobes on plateaus and in broad valleys in eastern Washington, Idaho, and Montana.

The Cordilleran Ice Sheet in southern Alaska consisted of ice fields and large valley and piedmont glaciers that flowed from montane source areas across the continental shelf to the south and into the broad low country drained by Yukon River to the north (Hamilton and Thorson, 1983). Ice was thicker and more extensive on the seaward side of the mountains than on the interior side, reflecting the predominant Pacific source of snowfall (Flint, 1971). Most of interior Alaska and central and northern Yukon Territory were dry throughout the Quaternary and consequently remained unglaciated. The Cordilleran Ice Sheet in southern Yukon Territory was fed principally from the Selwyn and Cassiar mountains; some ice also was supplied from the St. Elias Mountains (Hughes et al., 1969).

On the east, ice flowed out of the Rocky Mountains and locally coalesced with the Keewatin portion of the Laurentide Ice Sheet. Some valley glaciers in the eastern Mackenzie Mountains also came into contact with Laurentide ice. Coalescence probably occurred only at the climaxes of major glaciations; at other times, an ice-free zone existed between Cordilleran and Laurentide glaciers on the westernmost Interior Plains (see Relationship of Cordilleran and Laurentide glaciers).

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1989: Cordilleran Ice Sheet; in Chapter 1 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

Growth

At the end of each major Pleistocene nonglacial period, glaciers were restricted to the high mountains of the Canadian Cordillera. As climate deteriorated during the early part of each glaciation, small mountain ice fields grew and alpine glaciers advanced (alpine phase of glaciation; Kerr, 1934; Davis and Mathews, 1944; Flint, 1971; Fig. 1.13). With continued cooling and perhaps increased precipitation, glaciers expanded and coalesced to form a more extensive cover of ice in mountain areas (intense alpine phase). During long sustained cold periods, these glaciers advanced across plateaus and lowlands and eventually grew into large confluent masses of ice that covered most of British Columbia and adjacent regions (mountain ice sheet phase). Throughout this period, the major mountain systems remained the principal source areas of glaciers, and ice flow was controlled by topography. During the final phase of glaciation, which probably was infrequently achieved, ice thickened to such an extent that one or more domes with surface flow radially away from their centres became established over the interior of British Columbia (continental ice sheet phase; Dawson, 1881, 1891; Kerr, 1934; Mathews, 1955; Wilson et al., 1958; Fulton, 1967; Flint, 1971).

In this general sequence of glacier growth, the transition from the third to the fourth phase was accompanied by a local reversal of glacier flow in the Coast Mountains as the ice divide (i.e., the axis of outflow) shifted from the mountain crest eastward to a position over the British Columbia interior (Kerr, 1934; Flint, 1971). A comparable westward shift and reversal of flow also may have occurred locally in the Rocky Mountains. These flow reversals resulted from the buildup of ice in the interior to levels higher than the main accumulation areas in the flanking mountains.

Drumlins, flutings, and other streamlined forms indicate that there were at least three ice divides over the Interior System at the climax of the last glaciation or shortly thereafter (Fig. 1.12). This suggests that the Cordilleran Ice Sheet did not constitute a single monolithic ice dome as has often been suggested. These ice divides, however, probably were subordinate to the main divide along the axis of the Coast Mountains which may have persisted throughout the last glaciation. If this is true, the continental ice sheet phase of glaciation was not completely achieved in the Canadian Cordillera during Late Wisconsinan time.

The successional four-phase model outlined above provides a useful framework for conceptualizing the growth of the Cordilleran Ice Sheet; however, the actual history of glacier growth in the Cordillera during the Pleistocene is more complicated, because ice did not build up in a uniform fashion. Rather, periods of growth were interrupted, at least in some areas, by intervals during which glaciers stabilized or receded. These complex glacier fluctuations were controlled mainly by global climatic changes and, to a lesser extent, by secondary local and regional factors induced by glaciation but only indirectly related to climate (e.g., eustatic sea level lowering, ocean cooling, and changes in local atmospheric circulation due to glacier growth).

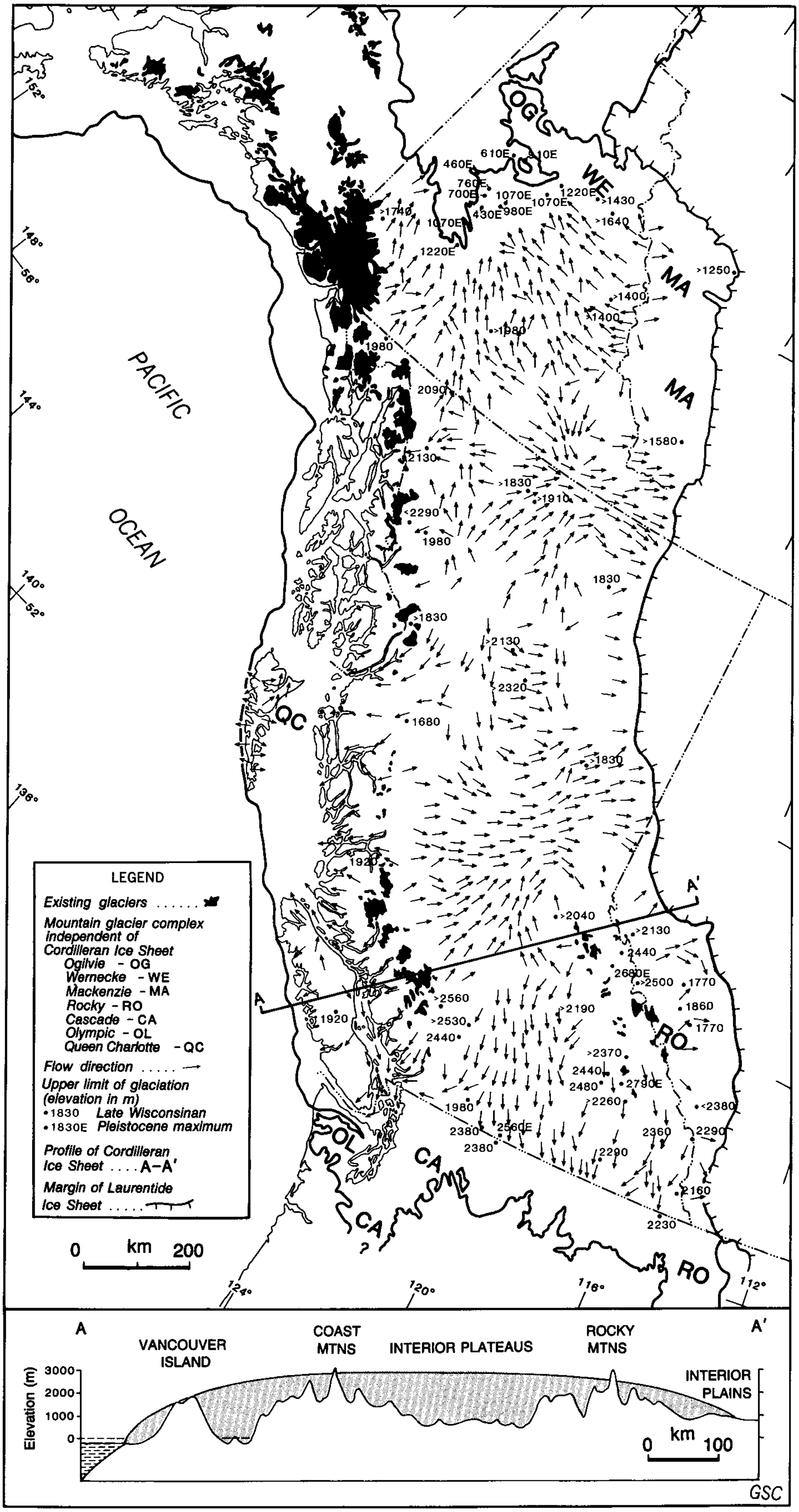

Figure 1.12. Maximum extent of Pleistocene glaciation in the Canadian Cordillera and adjacent areas, and ice flow pattern during the Late Wisconsinan. The glacier complex shown in this figure includes the Cordilleran Ice Sheet and independent and nearly independent glacier systems in some peripheral mountain ranges. Nunataks were small in ice sheet areas at the climax of glaciation. In contrast, there were large ice-free areas in some peripheral glaciated mountain ranges. Extent of glaciation, in part, from Crandell (1965), Lemke et al. (1965), Richmond et al. (1965), Prest et al. (1968), Hamilton and Thorson (1983), and Porter et al. (1983); ice flow pattern from Prest et al. (1968); data on upper limit of glaciation from Wilson et al. (1958).

Decay

Each glacial cycle terminated with rapid climatic amelioration. Deglaciation was characterized by complex frontal retreat in peripheral glaciated areas and by downwasting accompanied by widespread stagnation throughout much of the interior (Fig. 1.13; Rice, 1936; Fulton, 1967; Tipper, 1971a, b; Clague, 1981).

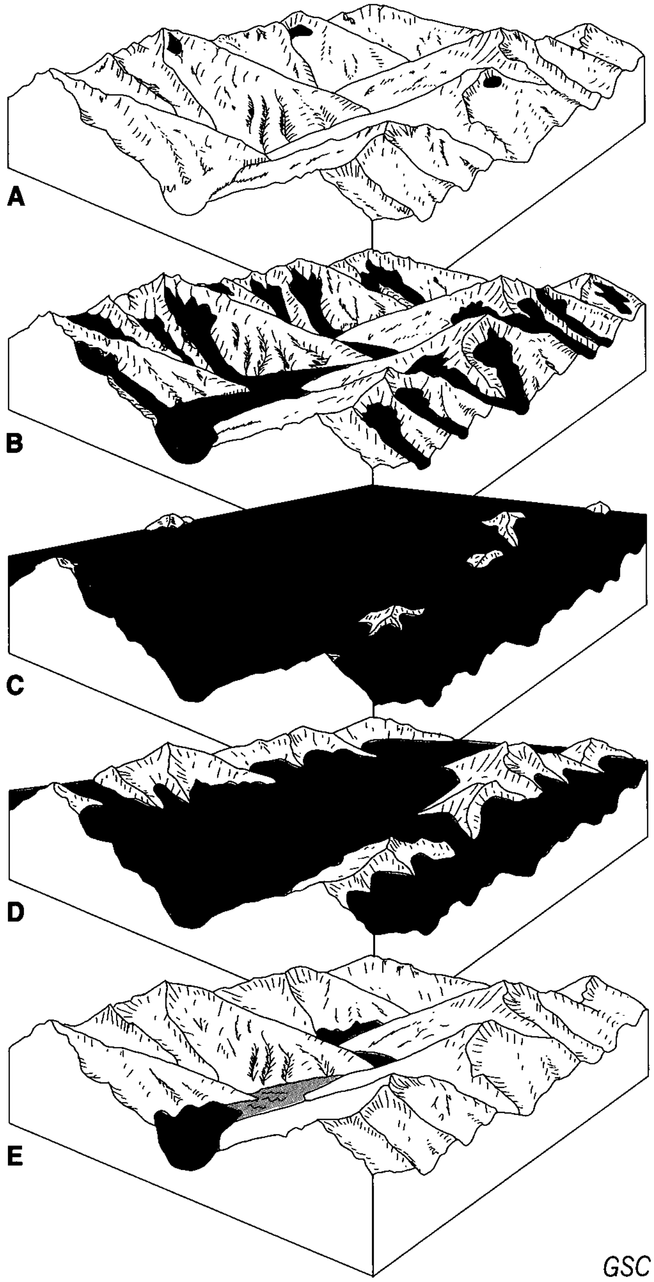

Figure 1.13. Growth and decay of the Cordilleran Ice Sheet. A. Mountain area at the beginning of a glaciation. B. Development of a network of valley glaciers. C. Coalescence of valley and piedmont lobes to form an ice sheet. D. Decay of ice sheet by downwasting; upland areas are deglaciated before adjacent valleys. E. Residual dead ice masses confined to valleys.

At the close of each glaciation, the western periphery of the Cordilleran Ice Sheet became unstable, probably due to a eustatic rise in sea level. The British Columbia continental shelf was rapidly deglaciated as glaciers calved back to fiord heads and shallow protected coastal embayments. Frontal retreat also occurred elsewhere along the periphery of the ice sheet, for example in northeastern Washington and southern Yukon Territory.

A different pattern of deglaciation has been documented for areas of low and moderate relief nearer the centre of the ice sheet. On the basis of observations of late Pleistocene landforms and sediments in south-central British Columbia, Fulton (1967) suggested that deglaciation in such areas occurred mainly by downwasting and stagnation and proceeded through four stages: (1) active ice phase — regional flow continued but diminished as ice thinned; (2) transitional upland phase — highest uplands appeared through the ice sheet, but regional flow continued in major valleys; (3) stagnant ice phase — ice was confined to valleys but was still thick enough to flow; and (4) dead ice phase — valley tongues thinned to the point where plasticity was lost. This model has been used with minor modifications to describe deglaciation of other parts of the British Columbia interior (Tipper, 1971a, b; Heginbottom, 1972; Ryder, 1976, 1981b; Howes, 1977).

The first areas to become ice-free at the end of each glaciation were those near the periphery of the ice sheet, for example the British Columbia continental shelf. Active glaciers probably persisted longest in some mountain valleys; however, these glaciers may have coexisted with large remnant dead ice masses on the plateaus of the Cordilleran interior. In general, retreat in the interior proceeded from both southern and northern peripheral areas towards the centre of the ice sheet. In detail, however, the pattern of retreat was complex, with uplands in each region becoming ice-free before adjacent valleys and other low-lying areas.

Decay of the Cordilleran Ice Sheet at the close of each glacial cycle was interrupted repeatedly by glacier still-stands and readvances. Most of these fluctuations affected relatively small areas and were not synchronous from one region to another; thus, they may have resulted from local factors rather than global climatic change.

RELATIONSHIP OF CORDILLERAN AND LAURENTIDE GLACIERS

J.J. Clague

Rocky Mountain glaciers on occasion advanced onto the Interior Plains and coalesced with the Laurentide Ice Sheet. In the Athabasca, Peace, and Liard river areas and perhaps elsewhere, these glaciers were augmented by streams of ice flowing from the interior of British Columbia and southern Yukon Territory. The Laurentide Ice Sheet spread into the region from the north and east, at times reaching into the foothills and mountains at the eastern edge of the Cordillera. During lesser glaciations, Rocky Mountain glaciers failed to reach the Interior Plains and the Laurentide

Clague, J.J.
1989: Relationship of Cordilleran and Laurentide glaciers; in Chapter 1 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

Ice Sheet terminated short of the mountain front. At such times, montane outwash trains ended as deltas in lakes impounded by Laurentide ice.

Some glaciers in the Mackenzie Mountains also reached the eastern edge of the Cordillera on one or more occasions during the Pleistocene. Within this region, there are small areas that were affected by both montane and Laurentide ice masses; locally these ice masses coalesced.

Evidence on the relationship of Cordilleran and Laurentide glaciers comes from studies of till and associated stratified drift along the eastern margin of the Canadian Cordillera. Deposits of the Laurentide Ice Sheet and of Cordilleran glaciers are very different in lithology and thus are easily distinguished. The former are rich in plutonic and metamorphic detritus derived from the Canadian Shield, whereas the latter consist dominantly of detritus of sedimentary rocks eroded from the Rocky and Mackenzie mountains. Using lithological criteria as well as stratigraphic and geomorphic evidence, scientists have recognized several advances of the Laurentide Ice Sheet and broadly correlative episodes of mountain glaciation in the eastern Cordillera (see Regional Quaternary stratigraphy and history). However, there probably were other glaciations that have not been recognized because of their antiquity and because their deposits have been eroded or covered with younger sediments.

In general, older known glaciations in the eastern Cordillera were more extensive than younger ones. Differences in the extent of glaciation in this region are reflected in the degree of contact between Cordilleran and Laurentide glaciers and the width of the ice-free zone between them. The pattern of glaciation might be likened to "the periodical closing and opening of a gigantic zipper along the Albertan mountain front, with the zipper closing a lesser distance during each successive glaciation. During each glaciation the zipper would close from north to south as the Cordilleran and Laurentide glaciers spread farther east and southwest respectively. Then, when they began to wane, the zipper would part from south to north, opening up an ever-widening corridor between them" (Stalker and Harrison, 1977, p. 887). Although various scientists differ in opinion as to the amount of contact between Cordilleran and Laurentide glaciers during each glaciation, they all visualize the ice-free corridor as narrowing northward in western Alberta. During the most extensive glaciation, there may have been fairly complete contact from Waterton Lakes to beyond the British Columbia-Yukon border. In contrast, at the climax of the last glaciation, closure did not extend as far south, and many intervening high areas in the eastern Rocky Mountains were left unglaciated.

In some areas, Cordilleran and Laurentide advances were out-of-phase. For example, during the pre-Wisconsinan "Great Glaciation" of Waterton Lakes National Park, Cordilleran glaciers advanced and retreated before the Laurentide Ice Sheet invaded the area (Stalker and Harrison, 1977). The main evidence for this is the occurrence of Laurentide Labuma Till over Cordilleran Albertan Till deposited during the same glaciation. The two tills do not interfinger, nor do they exhibit a transitional zone of mixed lithology as might be expected if the two ice masses had coalesced. Locally, the two tills are separated by Cordilleran outwash or by glacial lacustrine sediments deposited in lakes dammed by advancing Laurentide ice. Altogether, the evidence points to the fact that Cordilleran glaciers had largely receded from the region before the arrival of the Laurentide Ice Sheet. Similar arguments have been made for nonsynchronous advances in Oldman River basin directly north of Waterton Lakes (Alley, 1973). In addition, mountain and Laurentide glaciers may have been out-of-phase in the Mackenzie Mountains and on the adjacent Interior Plains (Ford, 1976).

In some cases, however, Cordilleran and Laurentide advances were broadly synchronous, or at least the Laurentide Ice Sheet began to retreat along its western margin at about the same time as mountain glaciers and lobes of the Cordilleran Ice Sheet. General synchroneity is suggested by the fact that Cordilleran and Laurentide ice coalesced in several areas during the last glaciation. Till of mixed Cordilleran and Keewatin provenance and associated quartzose conglomerate erratics derived from the Rocky Mountains near Jasper (Foothills Erratics Train; Stalker, 1956; Mountjoy, 1958) define the zone of coalescence of Cordilleran ice flowing from the west and Laurentide ice flowing from the north and east at the climax of this glaciation. The till of mixed provenance extends southeastward at the edge of the Rocky Mountains from Athabasca River on the north to near Oldman River on the south. There is no evidence within this region for overriding of the mixed provenance till or the Foothills Erratics Train by either Cordilleran or Laurentide ice; rather, the two ice masses apparently retreated at about the same time. It thus is clear that the Cordilleran and Laurentide advances during the last glaciation were broadly synchronous, although there undoubtedly were some differences in timing.

There is no satisfactory explanation for the apparent nonsynchroneity of Cordilleran and Laurentide advances in some areas and for synchroneity in others. However, it seems reasonable that mountain glaciers, with their shorter travel distance, would reach positions at or beyond the mountain front before the more massive Laurentide Ice Sheet entered the region. The early retreat of mountain glaciers in some areas may have resulted from a reduction in precipitation in the eastern Cordillera due to growth of the Cordilleran Ice Sheet to the west. Ice covering the British Columbia interior may have depleted or diverted moist air masses that previously had flowed across the Rocky Mountains, making the air reaching that area rather dry. This, in turn, may have caused some local glaciers in the Rocky Mountains to retreat at a time when both the Cordilleran and Laurentide ice sheets were growing.

QUATERNARY SEA LEVELS

J.J. Clague

Changes in sea level along the Pacific coast of Canada during Quaternary time are attributable mainly to diastrophism, eustasy, and isostasy (Mathews et al., 1970; Clague, 1975a, 1981, 1983; Fladmark, 1975; Andrews and Retherford, 1978; Clague et al., 1982b; Riddihough, 1982). The net effect on past sea levels of these three factors can be determined by studying former shorelines and associated

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deposits, but it is difficult to assess the individual effect of each. This arises because eustasy and isostasy are interdependent and because there is disagreement about the nature of past eustatic sea level changes (e.g., Mörner, 1976; Clark et al., 1978). Notwithstanding these problems, it is possible to summarize late Quaternary sea level fluctuations on the British Columbia coast and to comment on their likely causes.

Sea levels during the last glaciation

The position of the sea relative to the land at the beginning of the last glaciation (i.e., Fraser Glaciation in British Columbia) is uncertain because there is no consensus concerning eustatic sea levels at this time and because subsequent vertical tectonic movements are poorly known. It seems most likely, however, that British Columbia shorelines at the beginning of the Fraser Glaciation were somewhat lower than, although perhaps close to, the present.

Loading of British Columbia by glacier ice during the Fraser Glaciation caused major isostatic adjustments in the crust and mantle of the region (Clague, 1983). Gradual growth of glaciers led to progressive isostatic depression of the land surface. At first, this depression was localized beneath major mountain ranges that served as loci of glacier growth. Lateral movement of material in the asthenosphere away from these areas probably produced outward migrating forebulges. Initially, relative sea levels may have fallen as these forebulges passed through coastal areas, as oceans cooled and contracted, and as water was transferred from oceans to expanding Late Wisconsinan ice sheets. Eustatic sea level lowering, in turn, may have caused hydro-isostatic uplift of the continental shelf, resulting in a further fall of the sea relative to the land. As glaciers advanced from mountains onto plateaus and into lowlands, however, isostatically depressed areas grew in size, and the coastal region began to subside. Eventually, glacial isostatic depression became dominant in most areas, and the sea rose above its present level relative to the land.

At the climax of the Fraser Glaciation, the entire glaciated Cordillera was isostatically depressed, with areas near the centre of the ice sheet displaced downward more than areas near the periphery. Although the magnitude of isostatic depression at this time is unknown, limits are provided by levels of shorelines that formed at the end of the Pleistocene as glaciers withdrew from the British Columbia coast. Relict shorelines occur on the mainland coast up to about 200 m above present sea level. Taking into account a eustatic sea level lowering of perhaps 50-100 m at the time the highest shorelines formed, local glacial isostatic depression of more than 250 m is indicated. In fact, this is a minimum value for isostatic depression in this area because the Cordilleran Ice Sheet had decreased in size before the highest shorelines formed, and consequently isostatic rebound probably had commenced earlier.

The highest shorelines on the coasts of Vancouver Island and mainland British Columbia date to the time of glacier retreat at the close of the Pleistocene. The elevation of the marine limit in these areas varies in relation to the distance from main centres of ice accumulation and, to a lesser extent, to the timing of retreat. In general, the marine limit is highest (ca. 200 m) on the mainland coast and declines towards the west and southwest (Fig. 1.14; Mathews et al., 1970; Clague, 1975a, 1981). It is less than 50 m in elevation on the west coast of Vancouver Island near the margin of the former ice sheet (Clague, 1981; Howes, 1981b). In contrast, relative sea level on the Queen Charlotte Islands and in parts of Dixon Entrance, Hecate Strait, and Queen Charlotte Sound was lower at the end of the last glaciation than at present (Fladmark, 1975; Clague, 1981; Clague et al., 1982a, b; Warner et al., 1982). This indicates that glacial isostatic depression of these areas was minor.

Rapid deglaciation at the close of the Pleistocene triggered isostatic movements in the Canadian Cordillera that were opposite in direction to those that occurred during ice sheet growth. Isostatic uplift of Vancouver Island and the mainland was greater than the coeval eustatic rise, thus the sea fell from the marine limit in these areas as deglaciation progressed (Fig. 1.15). The fall in sea level was extremely rapid. For example, at Courtenay on eastern Vancouver Island, the highest marine delta, 150 m above the present shore, formed about 12 500 ± 450 BP (I(GSC)-9,Table 1.2). In contrast, deltas 60-90 m below marine limit about 65 km to the southeast have yielded radiocarbon dates ranging from 12 400 ± 200 BP to 12 000 ± 450 BP (GSC-1, I(GSC)-1). Similarly, at Kitimat on the northern mainland coast, the sea fell 85 m with respect to the land between 10 100 ± 160 BP and 9300 ± 90 BP (GSC-2492, GSC-2425). The error terms associated with these and other radiocarbon dates are sufficiently large and the dates themselves so closely spaced in time that it is not possible to assign precise figures to rates of uplift. Nevertheless, most recorded uplift at each site occurred within a time interval of less than 2 ka.

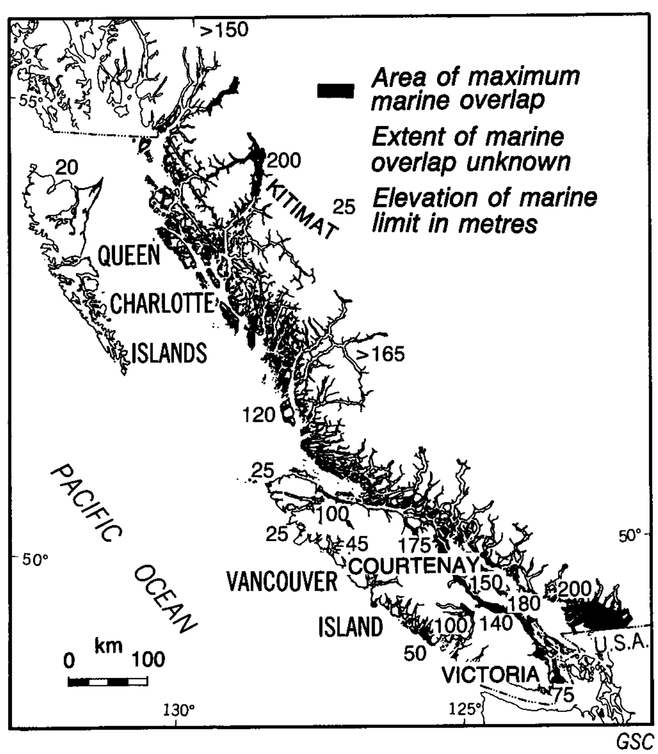


Figure 1.14. Maximum late Quaternary marine overlap in British Columbia; extent of overlap inland from the heads of most mainland fiords is unknown (adapted from Clague, 1981, Fig. 5).

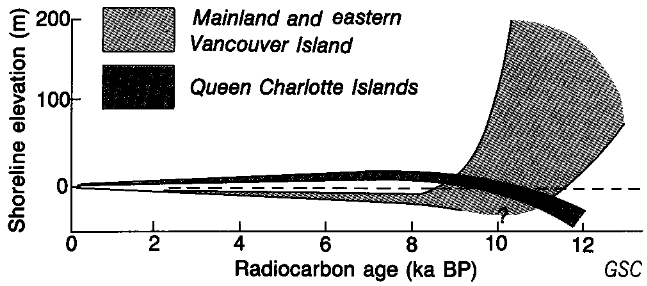


Figure 1.15. Generalized patterns of sea level change on the British Columbia coast since the end of the last glaciation. Each envelope depicts a range of sea level positions, reflecting regional and local differences in ice loads and time of deglaciation as well as uncertainties in locating former shorelines.

Isostatic uplift occurred at different times along the British Columbia coast due to diachronous retreat of the Cordilleran Ice Sheet. In general, regions that were deglaciated first rebounded earlier than those that were deglaciated at a late date. For example, the sea had fallen to its present level at Victoria on southern Vancouver Island by about 11.5 ka, about 1.5 ka after initial deglaciation of the area. In contrast, the sea was at its upper limit at Kitimat about 11 ka when the area was first deglaciated, and most isostatic uplift there occurred between 10.5 ka and 9 ka. Such data indicate that there were regional differences in the timing of the isostatic response to deglaciation and show that the crust deformed in a complex nonuniform manner.

Even within a region, the sea probably did not fall uniformly relative to the land at the end of the Fraser Glaciation. There are at least three reasons for this: (1) the rate and direction of eustatic change during this period varied; (2) the rate of isostatic uplift may have varied due to local stillstands and readvances of glaciers; and (3) some vertical displacements may have occurred instantaneously and sporadically along faults during earthquakes. Evidence for nonuniform emergence has been found at several sites on the British Columbia coast. At Kitimat, for example, shorelines apparently were relatively stable for hundreds of years after initial deglaciation of the area; about 10-10.5 ka, however, the sea began to fall rapidly relative to the land (Clague, 1984, 1985). In some areas, emergence probably was interrupted by short periods of submergence. Mathews et al. (1970) and Armstrong (1981), for example, proposed that emergence of Fraser Lowland at the close of the Fraser Glaciation was interrupted by a strong short-lived submergence that culminated about 11.5 ka.

The record of sea level changes during deglaciation on the Queen Charlotte Islands is different from that outlined above. Shorelines there were lower than at present from at least 15 ka until 9.5-10 ka; this contrasts sharply with the situation on the mainland coast directly to the east (Fig. 1.15; Clague, 1981; Clague et al., 1982a, b; Mathewes and Clague, 1982). This period of low sea levels was followed by a marine transgression that culminated about 7.5-8.5 ka with shorelines in some areas about 15 m higher than at present.

The opposing character of late Quaternary sea level changes on the Queen Charlotte Islands and the adjacent mainland is best explained in terms of differing ice loads and forebulge migration during deglaciation. Clague (1983) proposed that shorelines on the Queen Charlotte Islands were low at the close of the last glaciation because ice loads there were insufficient to depress the crust below lowered eustatic water levels prevailing at that time. He further suggested that mantle material flowed away from peripheral glaciated areas as the Cordilleran Ice Sheet decayed. As a result, the asthenosphere was depleted beneath the Queen Charlotte Islands, the crust subsided, and lowland areas were transgressed by the sea. In contrast, eastward forebulge migration beneath the mainland coast caused uplift and a marine regression there.

Sea levels during postglacial time

The coastal lowlands of eastern Vancouver Island and mainland British Columbia continued to emerge during the early Holocene, although generally more slowly than during deglaciation (Fig. 1.15; Mathews et al., 1970). This indicates that glacial isostatic uplift must have persisted into the early Holocene, because seas were still rising eustatically at that time due to final disintegration of the Laurentide and Fennoscandian ice sheets (Curray, 1965; Hopkins, 1967; Denton and Hughes, 1981). Emergence culminated during early or middle Holocene time when shorelines in some areas were considerably lower than they are today. The best evidence for low shorelines at this time comes from southern Vancouver Island and Fraser Lowland where Holocene terrestrial peats occur below present sea level (Mathews et al., 1970; Clague and Luternauer, 1982, 1983; Clague et al., 1982b); supporting evidence has been found on eastern Vancouver Island (Mathews et al., 1970) and on the central mainland coast (Andrews and Retherford, 1978). In these areas, the sea apparently rose during middle Holocene time, inundating some coastal archeological sites and terrestrial vegetation. Shorelines have deviated no more than a few metres during the last 5 ka, thus late Holocene sea level fluctuations have been relatively minor in comparison to those accompanying and immediately following deglaciation (Clague et al., 1982b).

In contrast to eastern Vancouver Island and the mainland, the Queen Charlotte Islands experienced sea levels higher than at present throughout the Holocene (Fig. 1.15; Fladmark, 1975; Clague, 1981; Clague et al., 1982a, b; Hebda and Mathewes, 1986). At the northeast end of this island chain, the sea has fallen about 15 m relative to the land since 7.5-8.5 ka. Features that formed during this regression and are now elevated above sea level include wave-cut scarps, wave-cut benches overlain by littoral gravel and sand, bars, spits, and beach and dune ridges parallel to the modern shoreline (Fig. 1.16). Sea level changes on westernmost Vancouver Island during the last 4 ka perhaps were similar to those on the Queen Charlotte Islands during the same period (Clague et al., 1982b).

In conclusion, significant isostatic uplift continued at gradually decreasing rates in most coastal areas until the early Holocene. The subsequent marine transgression on eastern Vancouver Island and the British Columbia mainland probably resulted from the combined effects of the global eustatic rise in sea level and forebulge collapse. In contrast, the recent regression on the Queen Charlotte Islands and western Vancouver Island is due largely to tectonic uplift, although some late Holocene sea level change in these areas may be eustatic in nature or even a residual isostatic response to deglaciation at the close of the Pleistocene (Riddihough, 1982; Clague et al., 1982b).

Table 1.2. Radiocarbon dates

Age (years BP)	Laboratory dating no.[1]	Locality	Location lat.	Location long.	Material	Reference	Comment
5 260 ± 200	Y-140bis	Mount Garibaldi	49°52'	122°59'	wood	Stuiver et al., 1960	treeline higher than at present
7 390 ± 250	GX-4039	Dunn Peak	51°27'	119°55'	charcoal	Duford and Osborn, 1978	minimum date for Dunn Peak advance
7 985 ± 125	I-9162	Dunn Peak	51°25'	119°57'	wood	Alley, 1980	minimum date for Harper Creek advance
8 460 ± 120[2]	GSC-2605	Old Crow Flats	68°13'	140°00'	wood	Lowdon and Blake, 1979	postdates proglacial lake
9 070 ± 130	GSC-3173	Bridge Glacier	40°48.3'	123°25.4'	wood	Blake, 1983	treeline higher than at present
9 300 ± 90	GSC-2425	Kitimat River	54°06.0'	128°36.8'	wood	Lowdon and Blake, 1979	sea level higher than at present
9 350 ± 80[3]	GSC-3120	Graham Island	53°41.7'	131°52.8'	marine shells	Blake, 1982	sea level higher than at present
9 670 ± 140[4]	I-5677	Rocky Mountain House	52°28.3'	114°32.0'	bone	Boydell, 1972	minimum date for deglaciation
9 780 ± 80	Y-1483	Kaskawulsh Glacier	60°49'	138°35'	grass	Denton and Stuiver, 1966	Kaskawulsh Glacier less extensive
9 960 ± 170	GSC-1548	Dawson Creek	55°58.8'	120°14.6'	freshwater shells	Lowdon and Blake, 1973	minimum date for deglaciation
10 100 ± 160	GSC-2492	Hirsch Creek	54°03.8'	128°35.6'	marine shells	Lowdon and Blake, 1979	sea level higher than at present
10 200 ± 280	GSC-3065	Calgary	51°02.9'	114°04.4'	bone	Blake, 1986	minimum date for deglaciation
10 250 ± 165[4]	I-5675	Rocky Mountain House	52°28.3'	114°32.0'	freshwater shells	Boydell, 1972	minimum date for deglaciation
10 400 ± 110[4]	GSC-2965	Kananaskis valley	50°52'	115°10'	wood	Lowdon and Blake, 1980	minimum date for deglaciation
10 400 ± 170[4]	GSC-1654	Dawson Creek	55°59.0'	120°15.7'	freshwater shells	Lowdon and Blake, 1973	minimum date for deglaciation
10 850 ± 320[2]	I-4224	Old Crow Flats	67°51'	139°48'	freshwater shells	Harington, 1977	Old Crow River near present level
11 100 ± 90[3]	GSC-3337	Graham Island	53°41.7'	131°52.8'	peat	Warner et al., 1982	sea level less than 8 m
11 370 ± 170	GSC-613	Cochrane	51°10.7'	114°27.5'	bone	Lowdon et al., 1967	minimum date for deglaciation
11 900 ± 120[4]	GSC-3885	Pocahontas	53°13'	117°55'	freshwater shells	Blake, 1986	minimum date for deglaciation
12 000 ± 450	I(GSC)-1	Parksville	ca. 49°17'	124°16'	wood	Walton et al., 1961	sea level higher than at present
12 400 ± 200	GSC-1	Parksville	ca. 49°17'	124°16'	wood	Dyck and Fyles, 1962	sea level higher than at present
12 460 ± 440[2,5]	I-3574	Old Crow Flats	67°50'	139°50'	bone	Harington, 1977	minimum date for proglacial lake
12 500 ± 200[5]	Y-1386	Kluane Lake	61°03'	138°21'	plant detritus	Denton and Stuiver, 1966	minimum date for deglaciation
12 500 ± 450	I(GSC)-9	Courtenay	49°38.7'	125°00.3'	marine shells	Walton et al., 1961	sea level higher than at present
13 660 ± 180[5]	GSC-495	Macauley Ridge	62°17'	140°42.5'	organic silt	Lowdon et al., 1967	minimum date for deglaciation
13 740 ± 190[5]	GSC-515	East Blackstone R	64°38'	138°24'	organic silt	Lowdon and Blake, 1968	minimum date for deglaciation
13 870 ± 180	GSC-296	Chapman Lake	64°51.5'	138°19'	organic silt	Dyck et al., 1966	minimum date for intermediate glaciation
16 000 ± 420	GSC-2690	Doll Creek	66°02'	135°42'	organic silt	Lowdon and Blake, 1981	minimum date for Hungry Creek Glaciation
16 000 ± 570[3]	GSC-3340	Graham Island	53°41.7'	131°52.8'	plant detritus	Clague et al., 1982a	minimum date for deglaciation
18 300 ± 380[6]	GSC-2668	Chalmer's bog	50°39.5'	114°33.5'	moss	Lowdon and Blake, 1979	ice-free conditions in Rocky Mountains
18 400 ± 1090[6]	GSC-2670	Chalmer's bog	50°39.5'	114°33.5'	moss	Lowdon and Blake, 1979	ice-free conditions in Rocky Mountains
19 100 ± 240	GSC-913	Bessette Creek	50°17.9'	118°51.8'	plant detritus	Lowdon and Blake, 1970	maximum date for last ice advance
20 800 ± 200	GSC-3946	Bluefish Flats	67°23.1'	140°21.7'	plant detritus	Blake, 1987	maximum date for proglacial lake
21 500 ± 240[7]	GSC-2536	Coquitlam valley	49°18.7'	122°46.8'	wood	Lowdon and Blake, 1978	Coquitlam Drift
21 700 ± 130[7]	GSC-2416	Coquitlam valley	49°18.8'	122°46.6'	wood	Lowdon and Blake, 1978	Quadra Sand (above Coquitlam Drift)
22 700 ± 1000[4]	GaK-2336	Eagle Cave	49°37'	114°38'	bone	Kigoshi et al., 1973	maximum date for last glaciation
23 840 ± 300	GSC-518	Mill Bay	48°37'	123°31'	wood	Lowdon et al., 1967	Cowichan Head Formation
23 900 ± 1140[5,8]	GSC-2811	Tom Creek	60°13.7'	129°00.4'	twigs	Lowdon and Blake, 1981	Tom Creek Silt
23 920 ± 400	GSC-59	Sidney Island	48°38.7'	123°19.7'	wood	Dyck and Fyles, 1963	Quadra Sand
24 400 ± 900[9]	L-502	Spanish Banks	49°17'	123°13'	wood	Olson and Broecker, 1961	Quadra Sand
25 170 ± 630[5]	NMC-1232	Cadzow Bluff	67°33.5'	138°53.5'	mammoth tusk	Morlan, 1986	maximum date for proglacial lake
25 840 ± 320[10]	GSC-715	Meadow Creek	50°15.1'	116°59.0'	wood	Lowdon and Blake, 1968	Bessette Sediments
25 940 ± 380[4,11]	GSC-573	Finlay River	56°18'	124°21'	plant detritus	Lowdon et al., 1971	Olympia Nonglacial Interval sediments
26 100 ± 320[9]	GSC-1635	Point Grey	49°15.9'	123°15.8'	wood	Lowdon and Blake, 1973	Quadra Sand
26 800 + 1200/-1000[11]	GX-2032	Sand Creek	49°21.4'	115°17.1'	wood	Clague, 1973	Olympia Nonglacial Interval sediments

Table 1.2 (cont.)

Age (years BP)	Laboratory dating no.[1]	Locality	Location		Material	Reference	Comment
			lat.	long.			
27 400 ± 580[4]	GSC-2034	Taylor	56°09'	120°42'	tooth	Lowdon and Blake, 1979	maximum date for last ice advance
27 400 ± 850	I-4878	Watino	55°43'	117°38'		Westgate et al., 1972	maximum date for last ice advance
27 500 ± 400	GSC-3530	Yakoun valley	53°31.2'	132°11.9'	peat	Blake, 1984	maximum date for last ice advance
28 800 ± 740	GSC-95	Willemar Bluff	49°40.2'	124°53.8'	wood	Dyck and Fyles, 1963	Quadra Sand
29 100 ± 560	GSC-3792	Jasper	52°53'	118°06'	wood	Blake, 1986	maximum date for Marlboro (?) advance
29 600 ± 460[5,12]	GSC-769	Silver Creek	61°00'	138°19'	plant detritus	Lowdon and Blake, 1970	maximum date for Kluane Glaciation
30 100 ± 600[12]	Y-1385	Silver Creek	61°00'	138°19'	plant detritus	Denton and Stuiver, 1967	maximum date for Kluane Glaciation
31 300 ± 640[2]	GSC-1191	Old Crow Flats	68°03'	139°49'	plant detritus	Lowdon and Blake, 1979	maximum date for proglacial lake
31 400 ± 660[2]	GSC-2739	Old Crow Flats	67°50.0'	139°51.8'	peat	Lowdon and Blake, 1979	maximum date for proglacial lake
32 710 ± 800[10]	GSC-493	Meadow Creek	50°15.1'	116°59.0'	wood	Lowdon and Blake, 1968	Bessette Sediments
33 400 ± 800[12]	Y-1488	Silver Creek	61°00'	138°19'	plant detritus	Denton and Stuiver, 1967	maximum date for Kluane Glaciation
33 700 ± 300[10]	GSC-542	Meadow Creek	50°15.1'	116°59.0'	wood	Lowdon and Blake, 1968	Bessette Sediments
36 900 ± 300	GSC-2422	Hungry Creek	64°34.5'	135°30'	wood	Hughes et al., 1981	maximum date for Hungry Creek Glaciation
37 700 + 1500/-1300[5,12]	Y-1356	Silver Creek	61°00'	138°19'	plant detritus	Denton and Stuiver, 1967	minimum date for Icefield Glaciation
40 900 ± 2000[11]	GSC-2591	Gold River	49°50.7'	126°07'	wood	Lowdon and Blake, 1981	minimum age of Muchalat River Drift
41 100 ± 1650[2]	GSC-2574	Old Crow Flats	67°51.5'	139°49.6'	wood	Blake, 1984	predates proglacial lake
41 500 ± 520[10]	GSC-1017	Meadow Creek	50°15.1'	116°59.0'	peat	Lowdon and Blake, 1970	Bessette Sediments
41 800 ± 600[10]	GSC-716	Meadow Creek	50°15.1'	116°59.0'	wood	Lowdon and Blake, 1968	Bessette Sediments
41 900 ± 600[10]	GSC-733	Meadow Creek	50°15.1'	116°59.0'	wood	Lowdon and Blake, 1968	Bessette Sediments
42 300 ± 650[10]	GSC-1015	Meadow Creek	50°15.1'	116°59.0'	wood, moss	Lowdon and Blake, 1970	Bessette Sediments
42 300 ± 700[10]	GSC-720	Meadow Creek	50°15.1'	116°59.0'	peat	Lowdon and Blake, 1968	Bessette Sediments
43 000 ± 600[10]	GSC-740-2	Meadow Creek	50°15.1'	116°59.0'	wood	Lowdon et al., 1971	Bessette Sediments
43 600 ± 700[10]	GSC-1017-2	Meadow Creek	50°15.1'	116°59.0'	peat	Lowdon et al., 1971	Bessette Sediments
43 800 ± 800[10,11]	GSC-740	Meadow Creek	50°15.1'	116°59.0'	wood	Lowdon and Blake, 1968	Bessette Sediments
48 000 ± 1300[5]	GSC-732	White River	62°00'	140°34'	wood	Lowdon and Blake, 1970	minimum date for Mirror Ck. Glaciation
58 800 + 2900/-2100	QL-195	East Delta	49°09.3'	122°55.6'	wood	Clague, 1977a	Cowichan Head Formation
>30 000[8]	GSC-2949	Tom Creek	60°13.7'	129°00.4'	twigs	Lowdon and Blake, 1981	Tom Creek Silt
>33 000	GSC-3208	Graham Island	53°42.5'	131°52.4'	peat	Blake, 1982	minimum date for penultimate glaciation
>38 000[11]	I-9772	Port McNeill	50°34.5'	127°02.5'	wood	Howes, 1983	older drift
>42 900	GSC-524	Stewart River	63°30.2'	137°16'	wood	Lowdon and Blake, 1968	minimum date for Reid Glaciation
>46 400[12]	Y-1355	Silver Creek	61°00'	138°19'	plant detritus	Denton and Stuiver, 1967	Shakwak Drift
>46 580[5]	GSC-331	Mayo	63°36'	135°56'	wood	Dyck et al., 1966	minimum date for Reid Glaciation
>49 000[12]	Y-1486	Silver Creek	61°00'	138°19'	peat	Denton and Stuiver, 1967	Icefield Drift
>49 000[7]	GSC-2094-2	Coquitlam valley	49°18.8'	122°46.8'	wood	Lowdon et al., 1977	Cowichan Head Formation?
>51 000	GSC-94-2	Cowichan Head	48°34'	123°22'	wood	Fulton and Halstead, 1972	Dashwood Drift
>52 000	GSC-3151-2	Graham Island	53°39.7'	131°54.3'	peat	Blake, 1982	minimum date for penultimate glaciation
>53 900[5]	GSC-527	Hunker Creek	63°58'	138°57'	wood	Lowdon and Blake, 1968	minimum date for intermediate glaciation
>62 000[11]	QL-194	Mary Hill	49°13.8'	122°46.5'	wood	Armstrong and Hicock, 1976	Semiahmoo Drift

1 Laboratories: GaK — Gakushuin University; GSC — Geological Survey of Canada; GX - Geochron Laboratories; I — Teledyne Isotopes; L — Lamont; NMC - National Museums of Canada (AMS age determination for National Museums of Canada by Atomic Energy of Canada); QL - Quaternary Isotope Laboratory; Y - Yale
2 See Figure 1.32.
3 See Figure 1.31.
4 See Figure 1.30.
5 See Figure 1.26.
6 Contamination by dead carbon suspected.
7 See Figure 1.18.
8 See Figure 1.27.
9 See Figure 1.20.
10 See Figure 1.19.
11 See Figure 1.17.
12 See Figure 1.28.

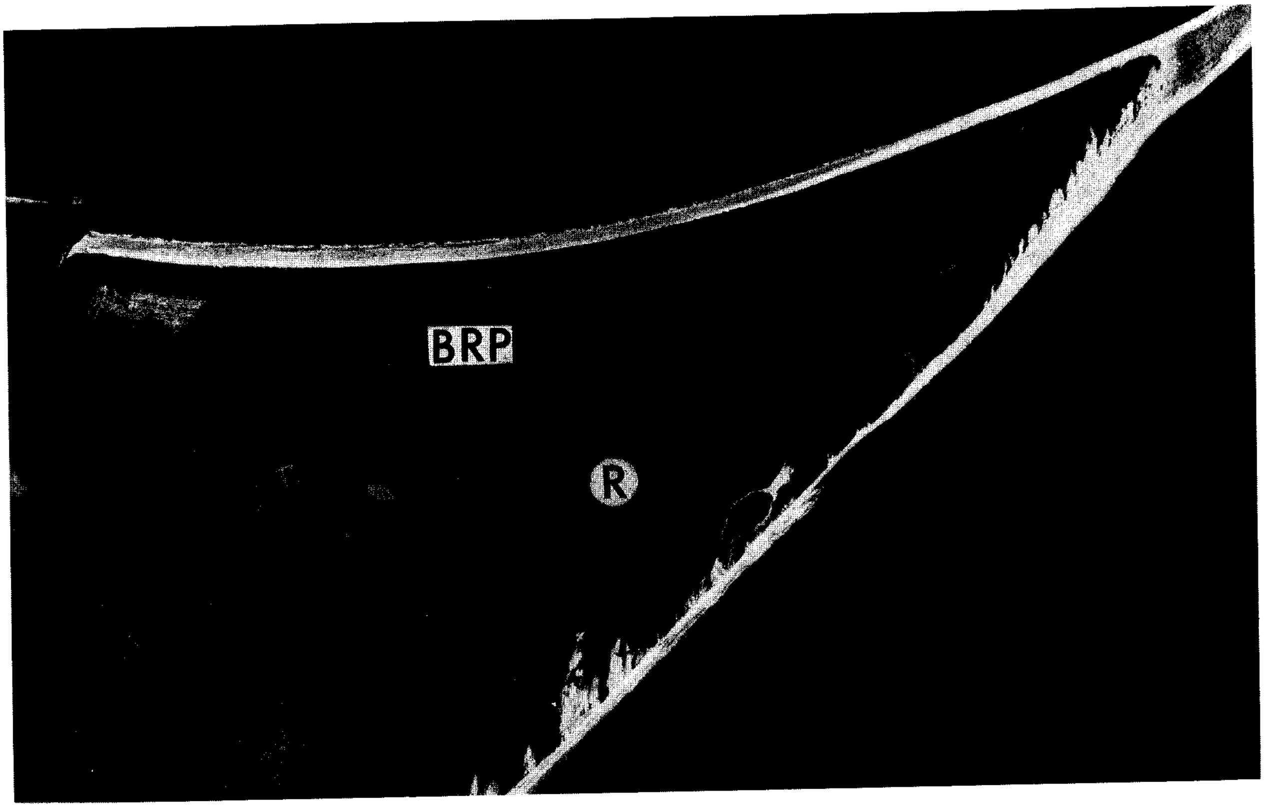


Figure 1.16. Elevated beach ridge plain (BRP) and older recurves (R) at the northeast corner of Graham Island, British Columbia. These features formed during a middle and late Holocene marine regression. Province of British Columbia BC5630-198.

QUATERNARY STRATIGRAPHY AND HISTORY

Area of Cordilleran Ice Sheet

Introduction

J.J. Clague

Most of British Columbia and southern and central Yukon Territory were repeatedly enveloped by the Cordilleran Ice Sheet during the Pleistocene. Surface sediments and landforms in these areas are mainly products of Late Wisconsinan glaciation (locally termed Fraser, McConnell, Macauley, and Kluane glaciations) and of postglacial time; Middle Wisconsinan and older sediments, in general, are covered by these younger materials. Our understanding of the Quaternary history of the area affected by the Cordilleran Ice Sheet thus is based on studies of relatively young surface sediments and associated landforms and of older subsurface deposits. At the northern limit of ice sheet glaciation in Yukon Territory, however, pre-Late Wisconsinan glacial and proglacial sediments occur at the surface beyond the McConnell and Macauley limits. Here, surface studies similar to those conducted on younger materials farther south have provided information on events predating the last glaciation.

Clague, J.J.
1989: Introduction (Quaternary stratigraphy and history, Cordilleran Ice Sheet); in Chapter 1 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

British Columbia

J.M. Ryder and J.J. Clague

Thick Quaternary sediments representing several glacial and nonglacial intervals are present in some valleys and lowlands in British Columbia and also occur on the continental shelf. These sediments are most common at sites where ice flow was transverse to valley axes, where bedrock projections sheltered downstream areas from scour, and where ice flow was sluggish, as for example in some lowlands near the margin of the Cordilleran Ice Sheet and in some valleys near its centre.

The stratigraphic relationships and internal characteristics of Quaternary sedimentary units in British Columbia are complex. Valleys and lowlands experienced

Ryder, J.M. and Clague, J.J.
1989: British Columbia (Quaternary stratigraphy and history, Cordilleran Ice Sheet); in Chapter 1 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

several cycles of infilling and partial or complete reexcavation, thus younger sediments commonly are inset into older deposits or are draped over an irregular older landscape. Younger sediments typically have been reworked from older Quaternary materials, thus units of different age within a given area may be physically similar. On the other hand, individual units may display abrupt facies changes due to the local variability of depositional environments within rugged terrain. Consequently, physical characteristics generally cannot be used to correlate discontinuous stratigraphic units. The firm correlation of major glacial and nonglacial units from valley to valley and from region to region has depended largely on radiocarbon dating and the presence of dated tephra layers.

Most Quaternary stratigraphic studies in British Columbia have been carried out in the southern part of the province. Chronologies that have been established and local names for the principal stratigraphic units are summarized in Figure 1.17 (see also Fulton et al., 1984). The longest record comes from Fraser Lowland near Vancouver, where three major glaciations and three nonglacial intervals have been recognized (Armstrong, 1981). Two major glaciations and three nonglacial intervals have been described from several areas, for example eastern Vancouver Island (Fyles, 1963) and south-central British Columbia (Fulton and Smith, 1978). Late Wisconsinan drift and postglacial sediments are common everywhere in British Columbia.

Old glaciations

There was glaciation in British Columbia in the Middle and Early Pleistocene and probably during the late Tertiary as well. Anahim Peak volcano on Chilcotin Plateau may have erupted into or against glacier ice in late Miocene or Pliocene time. On Mount Edziza in northwestern British Columbia, till and glacial fluvial sediments containing clasts foreign to the mountain are interbedded with lava flows and pyroclastic deposits dating back to perhaps 3-4 Ma (Souther et al., 1984). Till, proglacial sediments, and a glacial pavement at Dog Creek on Cariboo Plateau date to about 1.2 Ma (Mathews and Rouse, 1986). Tuyas and other ice contact volcanic landforms in the Clearwater River area of east-central British Columbia are at least 0.3 Ma old (Hickson and Souther, 1984). Finally, basal lavas about 0.4 Ma old in the Grand Canyon of Stikine River (P.B. Read, Geotex Consultants Ltd., personal communication, 1983) overlie both glacially scoured rock surfaces and glacial fluvial gravel containing clasts derived from the Coast Mountains to the west (Kerr, 1948).

Exposures of Early and Middle Pleistocene glacial deposits are rare outside of volcanic areas, in part because most of these materials occur below present base level. On southern Vancouver Island and in Fraser Lowland, however, there are exposures of drift beneath Sangamonian or older nonglacial deposits (Armstrong, 1975, 1981; Hicock and Armstrong, 1983). In Fraser Lowland, these old glacial deposits are termed Westlynn Drift and consist of a complex of till and glacial marine, glacial fluvial, and (?) glacial lacustrine sediments similar to materials in younger drift sheets. Deposits which correlate with, or are older than, Westlynn Drift may be present in other parts of British Columbia. Armstrong and Leaming (1968), for example, noted the presence of three or four till sheets in central British Columbia, and Ryder (1976) recognized two tills beneath Fraser Glaciation drift along Fraser River north of Lillooet.

Old nonglacial intervals

Outside of volcanic areas, the oldest known Pleistocene nonglacial deposits in British Columbia occur in Fraser Lowland (Armstrong, 1975, 1981; Hicock and Armstrong, 1983). These deposits, which underlie Westlynn Drift and

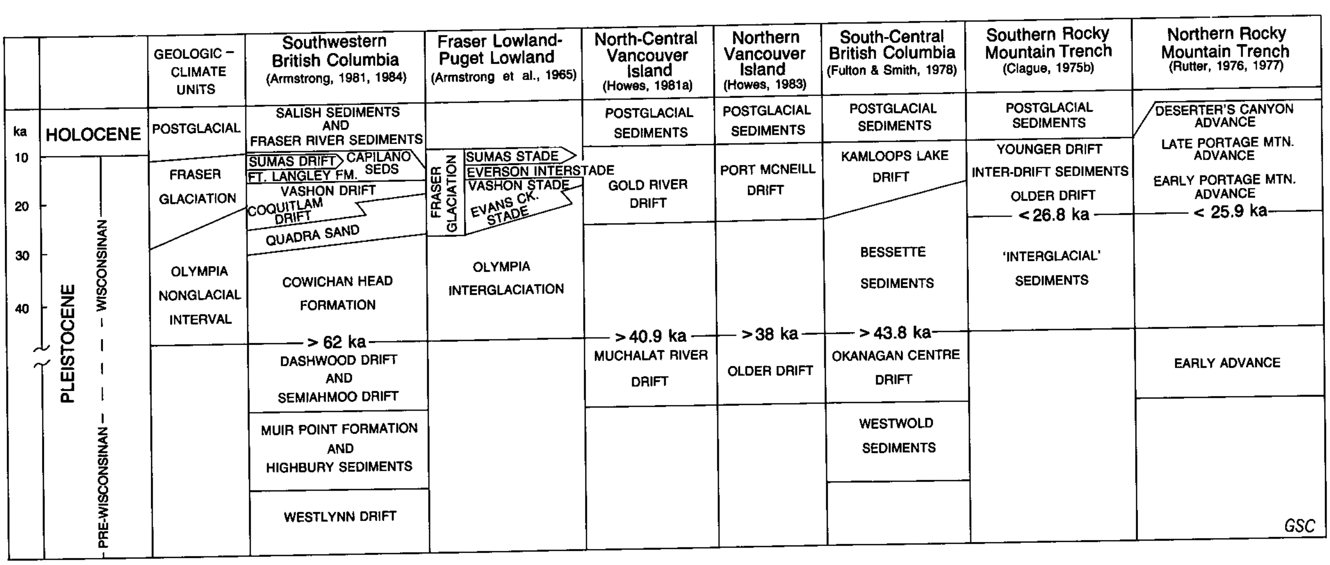


Figure 1.17. Subdivisions of Quaternary events and deposits in British Columbia. Different ages have been assigned to Northern Rocky Mountain Trench events by Bobrowsky et al. (1987).

are pre-Sangamonian in age, consist of more than 60 m of sand and silt encountered in drill holes. No outcrops are known.

Possible Sangamonian deposits have been described for only a few localities. In Fraser Lowland, Highbury Sediments consist of fluvial, deltaic, marine, and organic sediments, mainly sand and silt. Highbury deltaic and marine sediments grade upward into fluvial sediments, recording seaward progradation of floodplains, a process repeated during subsequent nonglacial intervals. The Muir Point Formation on southern Vancouver Island, which is Sangamonian or older in age and may correlate with Highbury Sediments, comprises gravel, sand, silt, peat, and diamicton, presumably deposited as alluvium and colluvium on a coastal floodplain (Hicock and Armstrong, 1983; Alley and Hicock, 1986). At the type section, the Muir Point Formation is more than 30 m thick and is unconformably overlain by Middle Wisconsinan nonglacial sediments and Late Wisconsinan drift.

Strongly oxidized, fluvial gravel and sand containing fragments of wood and blocks of peat and marl occur beneath the oldest till in Northern Rocky Mountain Trench (Rutter, 1976, 1977). These sediments, which commonly are more than 15 m thick, are beyond the range of radiocarbon dating and may correlate with Highbury Sediments and/or the Muir Point Formation.

Two well defined stratigraphic units which underlie till of the penultimate glaciation in southern British Columbia may record the transition from the Sangamonian Stage to the Wisconsinan Stage; alternatively, they may be pre-Sangamonian in age. Mapleguard Sediments, exposed at the base of some sea cliffs on eastern Vancouver Island, consist mainly of bedded sand, silt, and minor gravel of fluvial or perhaps deltaic or marine origin (Fyles, 1963; Hicock, 1980). The relationship of Mapleguard Sediments to the Muir Point Formation is unknown, although Hicock and Armstrong (1983) proposed that the former are younger than, and thus stratigraphically separate from, the latter. Mapleguard Sediments may be strictly nonglacial in origin, although they more likely are outwash deposits laid down during the early part of the penultimate glaciation. Westwold Sediments, which probably correlate with Mapleguard Sediments, are exposed at two sites in south-central British Columbia (Fulton and Smith, 1978). At the type section, Westwold Sediments consist of 16.5 m of cross-bedded gravelly sand capped by 1.8 m of sand, silt, clay, and marl. Features resembling ice-wedge pseudomorphs are present near the top of the gravelly sand unit, suggesting that permafrost may have been present at one time during the final stages of deposition of Westwold Sediments. However, the thin, fine grained upper unit lacks such features and contains molluscan shells, plant impressions, and fragments of bison bones, fish, beetles, and rodents, suggesting subsequent climatic warming.

Penultimate glaciation

Drift of the penultimate glaciation has been identified in most parts of British Columbia where detailed stratigraphic work has been carried out, although it generally is confined to major valleys and coastal lowlands (Fig. 1.18).

In the interior of British Columbia and on Vancouver Island, this drift consists of a single till bounded by stratified sediments. Okanagan Centre Drift in south-central British Columbia, for example, comprises a till, an underlying unit of glacial lacustrine silt and glacial fluvial gravel, and an overlying unit of glacial lacustrine silt and minor glacial fluvial and beach gravel (Fulton and Smith, 1978). A similar succession has been found at other sites in the southern interior (Ryder, 1976, 1981b) and in Northern Rocky Mountain Trench (Rutter, 1976, 1977). On north-central Vancouver Island, poorly exposed Muchalat River Drift consists of till and overlying glacial lacustrine silt (Howes, 1981a). Along the coast, the same till is overlain by fossiliferous glacial marine mud (Howes, 1983). On eastern Vancouver Island south of Howes' study area, Dashwood Drift comprises 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.

In Fraser Lowland, drift of the penultimate glaciation (Semiahmoo Drift) apparently is more complex than in other areas. Semiahmoo Drift consists of two or more tills interlayered with 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 probably results from the fact that tidewater glaciers fluctuated in this area during the decay phase of the penultimate glaciation. Semiahmoo and Dashwood drifts are similar in character and complexity to Late Wisconsinan drift in the same area, thus the pattern of glaciation in the Strait of Georgia region during the last two glacial periods probably was similar.

Olympia Nonglacial Interval

Sediments assigned to the Olympia Nonglacial Interval (or Olympia Interglaciation) are common in southern British Columbia and also occur locally in the central and northern parts of the province. These sediments were deposited during a lengthy nonglacial period that preceded the Late Wisconsinan Fraser Glaciation.

The Cowichan Head Formation underlies lowlands adjacent to the Strait of Georgia (Armstrong and Clague, 1977). The marine lower member, which consists mainly of sand and mud, conformably overlies Dashwood and Semiahmoo glacial marine sediments deposited during the transition from glacial to nonglacial conditions at the end of the penultimate glaciation. The upper member comprises interbedded organic-rich gravel, sand, and silt of fluvial and estuarine origin. The marine member is less widely distributed on land than the terrestrial upper member, but may occur extensively beneath the seafloor of the Strait of Georgia. Where the marine member is absent, upper Cowichan Head sediments overlie an irregular erosion surface developed on older drift. The Cowichan Head Formation is less than 10 m thick in most exposures and is much thinner than bounding drift units.

Deposition of the Cowichan Head Formation probably commenced while the sea was falling relative to the land at the end of the Semiahmoo Glaciation. At many coastal sites, shallow marine environments were gradually replaced by estuarine and fluvial environments as streams prograded seaward or extended across the emergent sea floor. Upper Cowichan Head sediments were deposited in channel, overbank, and swamp settings on coastal floodplains; they possibly also accumulated along prograding delta shorelines in bars, dunes, lagoons, and marshes (Fyles, 1963; Armstrong and Clague, 1977). Nonglacial sedimentation ceased with the deposition of outwash (Quadra Sand) from advancing glaciers at the onset of the Fraser Glaciation.

Bessette Sediments of south-central and southeastern British Columbia are time-equivalents of the Cowichan Head Formation (Fulton, 1968; Fulton and Smith, 1978). They consist chiefly of fluvial, deltaic, lacustrine, and colluvial materials deposited in a physiographic setting similar to that prevailing today in the southern interior. At the type section on Bessette Creek near Lumby, Bessette Sediments comprise 22 m of interbedded fluvial gravel, sand, and silt containing plant remains and two tephra layers. These sediments are sharply overlain by laminated silt deposited when the regional drainage was impounded by glaciers during the Fraser Glaciation. Another section of Bessette Sediments at Meadow Creek in southeastern British Columbia illustrates the wide range of Olympia-age materials in the southern interior (Fig. 1.19). Here, Bessette Sediments overlie a sloping surface developed on Okanagan Centre Drift; they consist of (1) thin colluvial and eolian sediments containing organic matter and well developed soil horizons, and (2) fluvial gravel, sand, and silt containing at least one tephra layer, plant fragments, and peat beds. The fluvial succession, which extends through a stratigraphic interval of about 15 m, intertongues with and overlies the colluvial and eolian sediments which are 1-2 m thick. At some sites in south-central British Columbia, Bessette Sediments include thick (up to about 100 m) sand and silt of probable deltaic origin.

Depositional analogues for Bessette Sediments exist under present-day conditions in the valleys of southern British Columbia (Fulton, 1975a; Fulton and Smith, 1978):

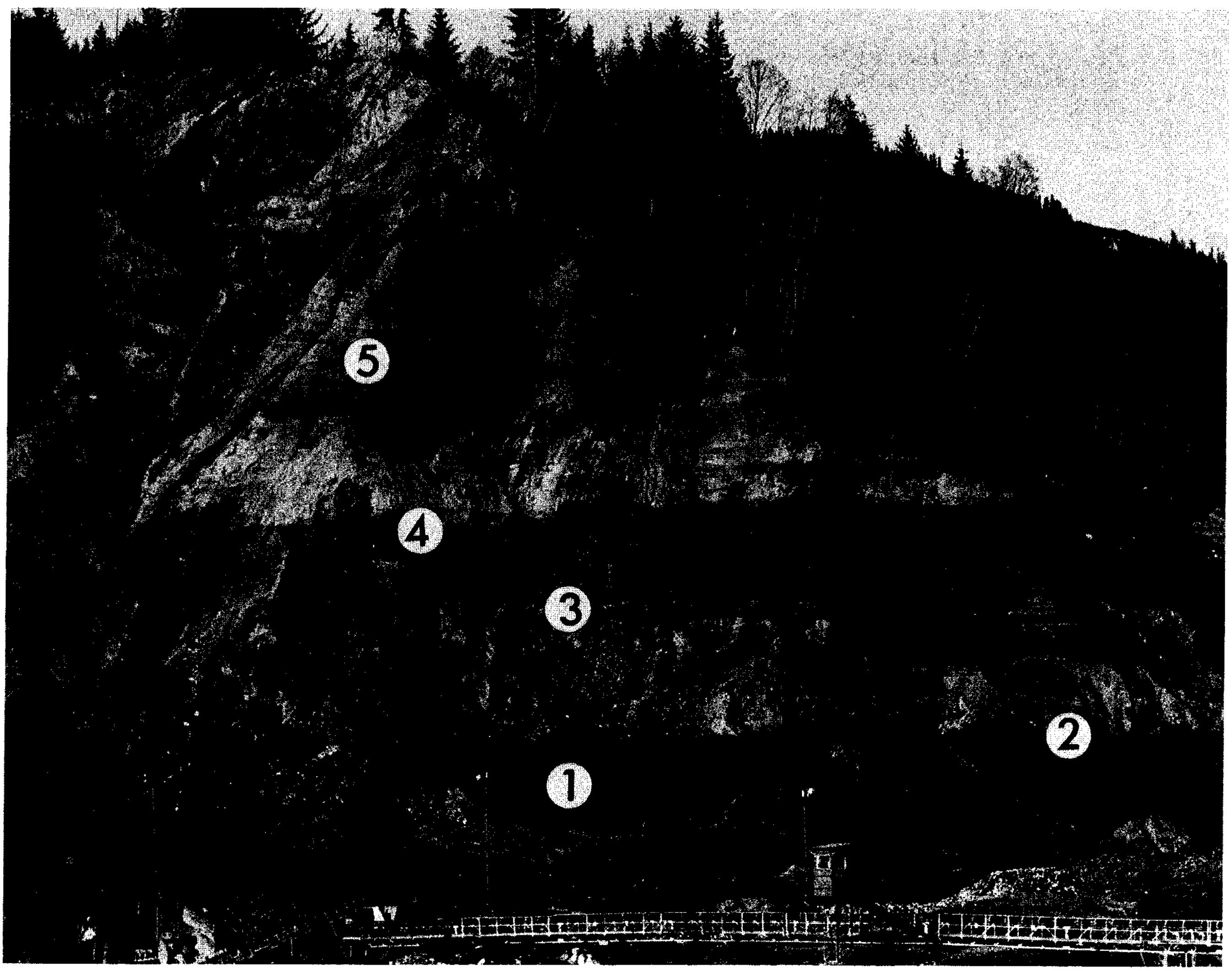


Figure 1.18. Exposure in Coquitlam River valley, southwestern British Columbia. Glacial fluvial gravel of the penultimate glaciation (1) is overlain successively by thin sand and silt of the Cowichan Head Formation (2), upward-coarsening glacial fluvial or fluvial sediments (3), Fraser Glaciation till (4), and Fraser Glaciation gravelly and sandy subaqueous (?) outwash (5). Wood from units 2, 4, and 5 yielded radiocarbon dates of >49 000 BP, 21 500 ± 240 BP, and 21 700 ± 130 BP, respectively (GSC-2094-2, GSC-2536, GSC-2416; Table 1.2).

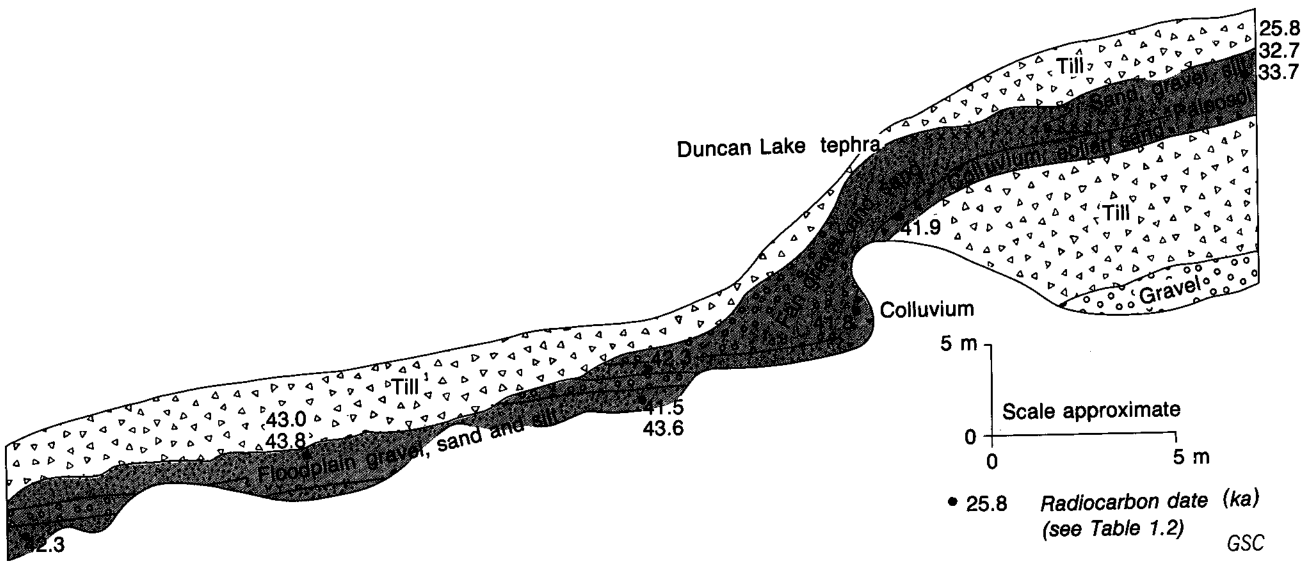


Figure 1.19. Exposure near Meadow Creek, southeastern British Columbia, showing varied Middle Wisconsinan nonglacial sediments (Bessette Sediments, shaded) overlain and underlain by till (adapted from Fulton, 1975b, Fig. I-1).

gravel and sand with cut-and-fill structures are present in the channels of most streams; finer sediments occur on many floodplains, on delta foreslopes, and elsewhere on the floors of lakes; and soil and colluvium cover valley walls. However, the present-day situation is not completely analogous to that of the Olympia Nonglacial Interval because Holocene fluvial deposits comparable in thickness to those of Olympia age are rare in interior valleys. During the Holocene, most streams in this region degraded their valleys, and they are more or less at equilibrium today. In contrast, Bessette Sediments in part record significant aggradation by major streams in southern British Columbia during the Olympia Nonglacial Interval. Fulton (1975a) proposed that aggradation occurred at the end of this period, probably in response to climatic change heralding the Fraser Glaciation. At that time, base level rose, and fluvial and lacustrine sediments accumulated up to 180 m above present valley floors.

This late aggradational phase was preceded by a period during which streams either were in equilibrium or were incising their valleys. This period may include most of the Olympia Nonglacial Interval, for in many areas where drift of both the penultimate glaciation and the Fraser Glaciation is present, the Olympia interval is represented only by an unconformity. This is the case, for example, on northern Vancouver Island (Howes, 1981a, 1983) and in Thompson River valley near Ashcroft (Ryder, 1976). Again, comparison with present-day conditions suggests that erosion was probably the dominant process in many valleys during the Olympia Nonglacial Interval.

Olympia-age sediments also have been reported at a few localities in central and northern British Columbia, but have not been studied as thoroughly as those in the southern part of the province. Lacustrine sediments at Babine Lake (Harington et al., 1974), intertill sediments in Northern Rocky Mountain Trench (Rutter, 1976, 1977) and near Atlin (Miller, 1976), and fluvial gravel and sand along Peace River and its tributaries near Fort St. John (Mathews, 1978) probably were deposited during the Olympia Nonglacial Interval. The fluvial sediments near Fort St. John may record the transition from the Olympia interval to the Fraser Glaciation because they are conformably overlain by glacial lacustrine silt and clay laid down in lakes dammed by the advancing Laurentide Ice Sheet. Aggradation of these sediments followed a period of erosion during which Peace River and its tributaries cut trenches comparable in size to present-day valleys in the region. This pattern of erosion followed by fluvial aggradation is identical to that reconstructed for many valleys in southern British Columbia during the Olympia Nonglacial Interval (Clague, 1986, 1987).

Fraser Glaciation

The Fraser Glaciation is the last period of ice sheet glaciation in British Columbia. In general, Fraser Glaciation drift consists of till and underlying and overlying glacial fluvial and glacial lacustrine sediments; in addition, glacial marine sediments are common in some coastal areas. At many localities, there is only a single Fraser Glaciation till, but at others two or more tills are present.

Early Fraser Glaciation stratigraphy and chronology are best documented in southwestern British Columbia where a prominent unit of advance outwash, Quadra Sand, has been described in detail (Fig. 1.20; Clague, 1976, 1977a; Armstrong and Clague, 1977). Quadra Sand is widely distributed in the Strait of Georgia region below an elevation of about 100 m and also occurs at higher elevations in some valleys extending into the bordering mountains. It overlies the Cowichan Head Formation and Semiahmoo and Dashwood drifts and underlies younger Fraser Glaciation deposits, including till. The unit, which locally is more than 50 m thick, consists mainly of well sorted, horizontally and cross-stratified sand. It includes small amounts of silt and clay in discrete beds and locally coarsens to gravel (e.g., Saanichton Gravel of Halstead, 1968). Paleocurrent indicators and provenance studies show that Quadra Sand was derived largely from the southern Coast Mountains and was deposited as outwash aprons in front of, and perhaps along the margins of, glaciers moving southward into the Strait of Georgia region and Puget Lowland during Late Wisconsinan time. Deposition occurred in channels of braided streams, on adjacent floodplains, and at the fronts of deltas that were prograding into the sea. Quadra Sand was progressively overridden by glaciers and partly eroded, giving rise to the present patchy distribution of the unit. Radiocarbon dates show that Quadra Sand is markedly diachronous and overlaps the Cowichan Head Formation in age (Clague, 1976, 1977a).

Thick subtill gravel and sand in interior valleys, mentioned in the preceding section, may correlate with Quadra Sand. These sediments are horizontally bedded, well sorted, and typically occur far above present base level. Although some scientists have grouped these materials with Olympia nonglacial units, they probably are outwash deposited in response to the expansion of glaciers during the early part of the Fraser Glaciation. The gravel and sand commonly grade up into, or are sharply overlain by, silt and clay deposited in lakes impounded by advancing glaciers or drift (Clague, 1987).

Thick and extensive glacial lacustrine sediments underlie Fraser Glaciation till in Stikine River basin. These sediments accumulated in a lake that formed when glaciers in the Coast Mountains advanced across and blocked Stikine River near its mouth. At its maximum, this lake extended at least 200-300 km upriver from the ice dam. Glaciers today reach almost to river level in this area, so the lake must have developed very early in the Fraser Glaciation and probably persisted for thousands of years. Similar lakes developed in the basins of other rivers that flow through the Coast Mountains, for example Skeena River basin (Clague, 1984).

In areas near the centre of the former Cordilleran Ice Sheet, Fraser Glaciation drift generally includes one till unit deposited during a single glacial event. This till typically is overlain and underlain by stratified sediments of glacial fluvial or glacial lacustrine origin. For example, Kamloops Lake Drift in south-central British Columbia is divisible into three units: (1) a lower stratified unit deposited during glacier expansion, mainly in proglacial lakes; (2) a middle till unit; and (3) an upper stratified unit deposited during deglaciation on floodplains and other subaerial surfaces and in glacier-dammed lakes (Fulton, 1975a; Fulton and Smith, 1978). Fraser Glaciation drift exhibiting this same tripartite stratigraphy is present elsewhere in the British Columbia interior (Ryder, 1976, 1981b; Clague, 1984) and on north-central Vancouver Island (Howes, 1981a).

In many areas, especially near the margins of the former ice sheet, Fraser Glaciation drift is more complex

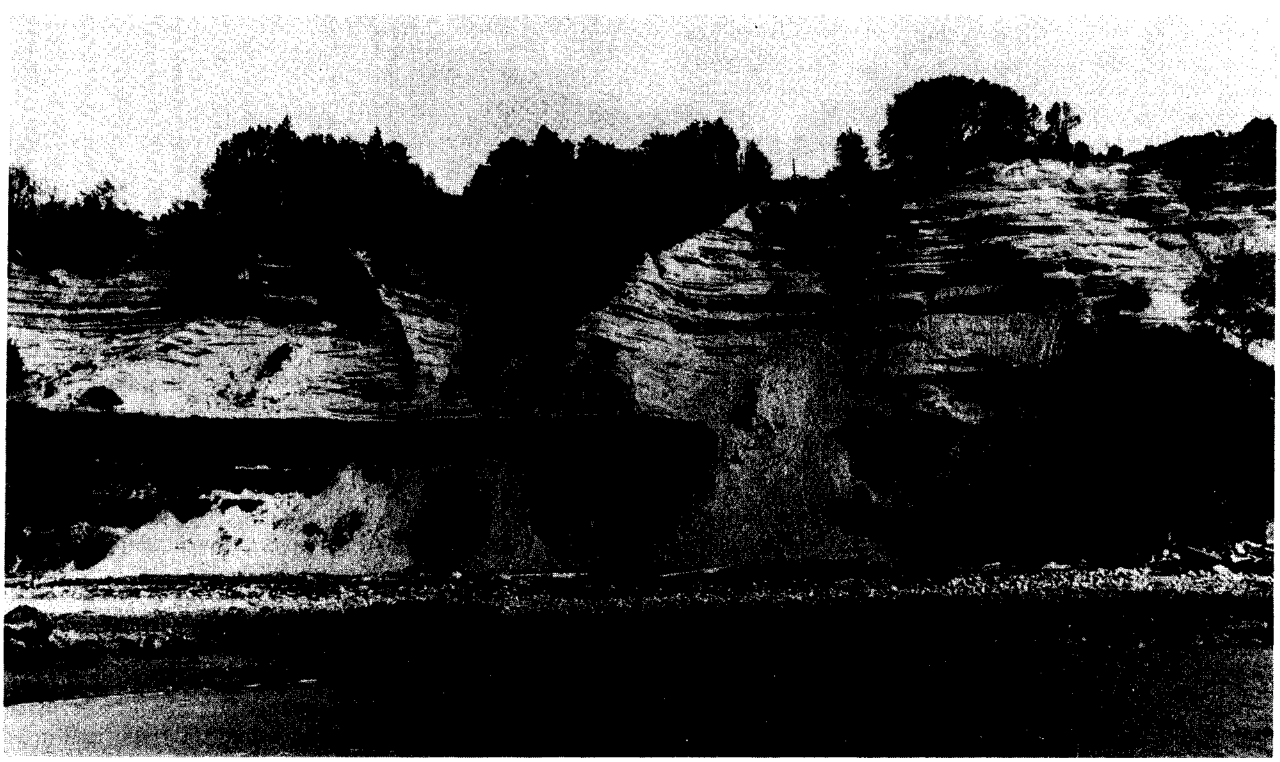


Figure 1.20. Quadra Sand, Vancouver, British Columbia. The lower darker part of the unit contains interbeds of silt and has yielded radiocarbon dates ranging from 26 100 ± 320 BP at the bottom to 24 400 ± 900 BP at the top (GSC-1635, L-502; Table 1.2).

and includes two or more till units. Such is the case, for example, in Fraser Lowland. There, Coquitlam Drift (Hicock and Armstrong, 1981) was deposited early during the Fraser Glaciation by valley and piedmont glaciers that expanded into lowland areas from the Coast Mountains. These glaciers receded from at least the western part of Fraser Lowland before 18.7 ka when forests became reestablished in the area. Glaciers subsequently readvanced and deposited Vashon Drift at the climax of the Fraser Glaciation (Armstrong et al., 1965; Hicock and Armstrong, 1985). During the Vashon Stade, the piedmont glacier in the Strait of Georgia advanced south into Puget Lowland and west into Juan de Fuca Strait; at its maximum, the Puget lobe reached 250 km south of the International Boundary (Waitt and Thorson, 1983). As this lobe retreated, a variety of sediments were deposited in Fraser Lowland in glacial marine environments. These sediments include till, subaqueous outwash, glacial marine mud and gravelly mud, and deltaic and beach gravel and sand (Fort Langley Formation and Capilano Sediments of Armstrong, 1981). Sumas Drift (Armstrong, 1981) is the youngest unit in the Fraser Glaciation drift sequence in this area and consists of till and outwash deposited during a local readvance at the end of the Pleistocene.

Another area with evidence for more than one ice advance during the Fraser Glaciation is Southern Rocky Mountain Trench (Clague, 1975b). During an early advance, some time after 27 ka, till and glacial fluvial gravel were deposited on the floor of the trench. This "older drift" is overlain by "inter-drift sediments" which consist mainly of glacial lacustrine and (?) lacustrine sand and silt deposited in one or more lakes that formed on the trench floor during an interval of glacier recession. Inter-drift sediments, in turn, are overlain by "younger drift" comprising two tills and associated glacial fluvial and glacial lacustrine sediments. Younger drift is the product of two glacier advances separated by a weak, short-lived interstade during which the floor of Southern Rocky Mountain Trench was only partially deglaciated.

Multiple Fraser Glaciation advances also have been reported for Northern Rocky Mountain Trench adjacent to the eastern margin of the Cordilleran Ice Sheet. Rutter (1976, 1977) recognized three advances in this area during Late Wisconsinan time: Early Portage Mountain (oldest and most extensive), Late Portage Mountain, and Deserter's Canyon. These advances were separated by periods of retreat during which glacial fluvial and glacial lacustrine sediments were deposited on the floor of the trench. This interpretation, however, has recently been questioned by Bobrowsky et al. (1987) who suggested that the Early Portage Mountain advance is Early Wisconsinan in age and that there was no Deserter's Canyon advance. According to them, the only Late Wisconsinan event in this area is the Late Portage Mountain advance.

Minor readvances of glaciers at the end of the Fraser Glaciation, such as the Sumas, have been recognized in many parts of British Columbia. For example, Alley and Chatwin (1979) presented evidence for resurgence of ice in Juan de Fuca Strait and on adjacent southern Vancouver Island during deglaciation of that area. Armstrong (1966) found stratigraphic evidence for a local readvance in the Coast Mountains near Terrace at the end of the Fraser Glaciation (see also Clague, 1985). Most of these advances occurred at different times and probably were controlled more by local factors than by global or regional climatic change (Clague, 1981).

The distribution of different types of surface sediments and the relationships of these materials to subglacial, ice marginal, and proglacial landforms enable a reconstruction to be made of the sequence and pattern of late Fraser Glaciation events in British Columbia. Such an approach was used by Fulton (1967), for example, to document deglaciation in south-central British Columbia. He showed that deglaciation took place largely by downwasting accompanied by stagnation and frontal retreat. Uplands appeared through the ice cover first, dividing the ice sheet into a series of valley tongues that retreated in response to local conditions.

As deglaciation progressed, glacier- and drift-dammed lakes formed in valleys and lowlands, trapping large quantities of fine sediment (Fig. 1.21). The most extensive lakes were in central and south-central British Columbia (Mathews, 1944; Armstrong and Tipper, 1948; Fulton, 1965, 1969; Tipper, 1971a). Some of these have been documented in considerable detail by studying shorelines, outlet channels, glacial lacustrine deposits, and ice marginal features (Fig. 1.22). Smaller lakes were present in many other areas during deglaciation, for example in parts of Rocky Mountain Trench and adjacent valleys (Clague, 1975b; Rutter, 1976, 1977; Ryder, 1981a), in Bulkley River valley (Clague, 1984), and on northern Vancouver Island (Howes, 1981a, 1983).

Postglacial

During and immediately following deglaciation, streams established courses across drift-covered terrain, including the floors of former glacial lakes and isostatically emergent coastal lowlands. Lakes remained in some basins dammed by drift and in basins previously occupied by stagnant ice.

Deglaciation initiated a period of redistribution of glacial materials by fluvial and mass wasting processes that continued into the Holocene (i.e., "paraglacial sedimentation"; Church and Ryder, 1972; Jackson et al., 1982). Freshly deglaciated drift was removed from slopes by landslides and running water and transported to lower elevations. As a result, large fluvial and debris-flow fans were constructed, and streams aggraded their valleys. Much of the fine detritus eroded from drift was carried through the fluvial system and contributed to the rapid growth of deltas in lakes and the sea (Clague et al., 1983).

Paraglacial sedimentation commenced as soon as local areas became ice-free. Processes were initially rapid due to plentiful meltwater and the lack of vegetation; they continued at decreasing rates as slopes stabilized and the supply of available drift declined. Mazama tephra commonly occurs within the uppermost 2-3 m of debris-flow fans in the southern Canadian Cordillera, indicating that most deposition took place before 6.8 ka (Ryder, 1971).

Eolian sedimentation likewise occurred mainly during latest Pleistocene and early Holocene time (Fulton, 1975a). During deglaciation, silt and clay were blown from unvegetated surfaces, especially floodplains, and deposited in protected areas as loess. At the same time, wherever there was an abundant supply of sand, dunes were constructed. There was a marked reduction in loess deposition and dune formation when bare surfaces became stabilized by vegetation.

Aggradation due to paraglacial effects was followed by degradation as the supply of sediment to streams decreased. Streams entrenched paraglacial fills and older deposits, thus producing terraces and dissected fans. As a consequence of this degradation, many streams in British Columbia today flow in trenches cut into broad sediment-filled valleys. For example, Fraser and Thompson rivers locally occupy trenches incised up to 300 m below late glacial valley floors (Fig. 1.5; Ryder, 1971).

This period of degradation ended in the middle to late Holocene with streams near their present levels. Since then, most fluvial systems have been, more or less, in dynamic equilibrium; consequently, Holocene alluvium beneath floodplains generally is thin. Streams issuing from present-day glaciers, however, have deposited thick bodies of outwash in some mountain valleys during late Holocene time.

The main sedimentation sites in British Columbia during the Holocene have been fans, deltas, and offshore basins (both lacustrine and marine). A variety of sediments have accumulated at other sites during this period, but most of the deposits are small.

Chronology and correlation

In southwestern British Columbia, all units below the Cowichan Head Formation (e.g., Highbury Sediments, Muir Point Formation, and Semiahmoo and Dashwood drifts) have consistently yielded "infinite" radiocarbon dates (i.e., dates beyond the limit of the radiocarbon dating method) (Fig. 1.17). On the basis of relative stratigraphic position and palynological data, Hicock and Armstrong (1983) suggested that the Muir Point Formation was deposited during the Sangamonian Stage and that it might correlate with the Whidbey Formation in nearby Puget Lowland, which has yielded amino acid dates of about 100 ka (Easterbrook and Rutter, 1982). However, there still seems to be some uncertainty about the age(s) of the Muir Point Formation and Whidbey Formation; Alley and Hicock (1986) recently stated that the former could be either Sangamonian or pre-Sangamonian. Correlation of the Muir Point Formation and Highbury Sediments is based largely on lithological similarity and relative stratigraphic position. If this correlation is valid, Westlynn Drift must be Illinoian (stage 6 of the marine isotope record) or older.

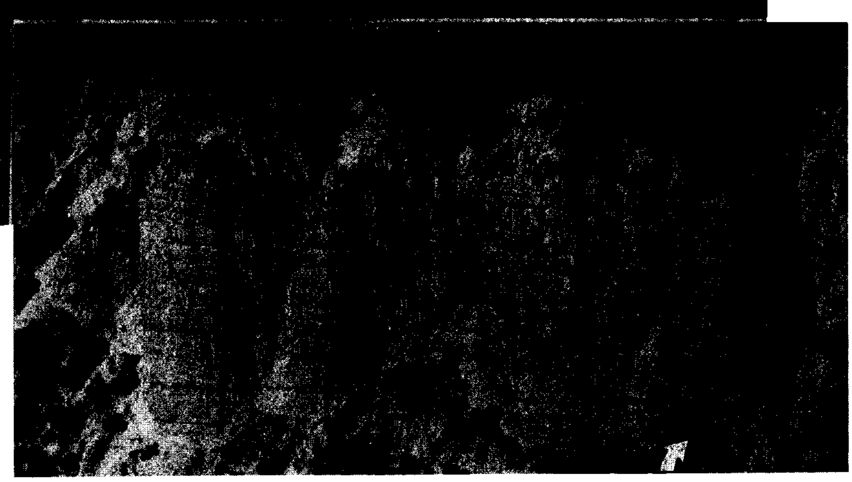
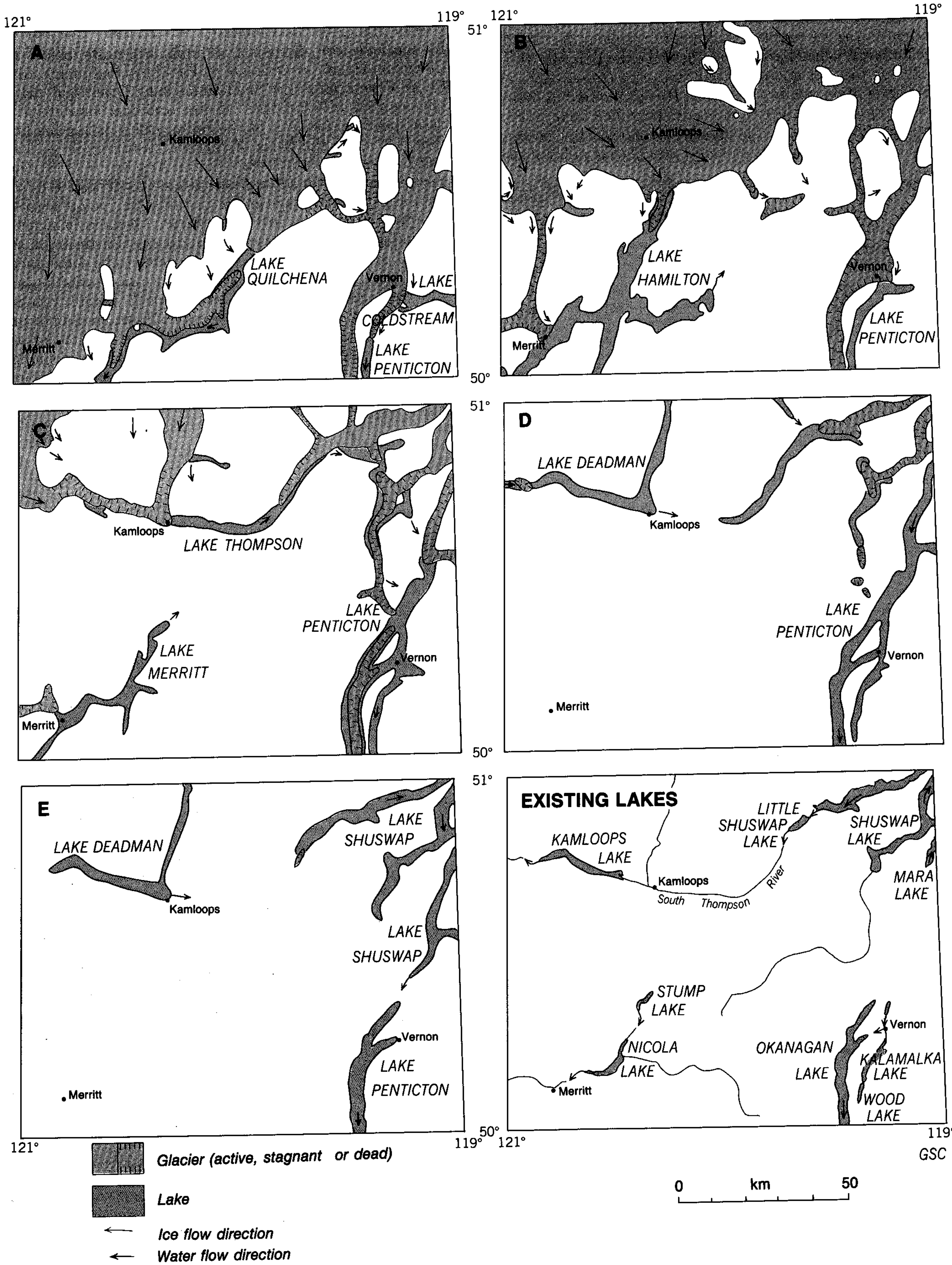


Figure 1.21. Glacial lacustrine sediments in Thompson River valley near Kamloops, British Columbia. These sediments were deposited over a period of 100-200 years at the close of the Fraser Glaciation (Fulton, 1965). The layers in the lower photograph are varves (person indicated by arrow provides scale). 204640-B, 204640


121°
119°
51°
50°
A
Kamloops
LAKE QUILCHENA
Vernon
LAKE COLDSTREAM
LAKE PENTICTON
Merritt
B
Kamloops
LAKE HAMILTON
Vernon
LAKE PENTICTON
Merritt
C
Kamloops
LAKE THOMPSON
LAKE PENTICTON
Vernon
LAKE MERRITT
Merritt
D
LAKE DEADMAN
Kamloops
LAKE PENTICTON
Vernon
Merritt
E
LAKE DEADMAN
Kamloops
LAKE SHUSWAP
LAKE SHUSWAP
Vernon
LAKE PENTICTON
Merritt
EXISTING LAKES
KAMLOOPS LAKE
Kamloops
South Thompson River
LITTLE SHUSWAP LAKE
SHUSWAP LAKE
MARA LAKE
STUMP LAKE
NICOLA LAKE
Merritt
OKANAGAN LAKE
Vernon
KALAMALKA LAKE
WOOD LAKE
GSC
Glacier (active, stagnant or dead)
Lake
Ice flow direction
Water flow direction
0 km 50

The absolute age of Westwold Sediments is unknown, although it too exceeds the limit of radiocarbon dating. On the basis of relative stratigraphic position, Fulton and Smith (1978) correlated this unit with Highbury and Mapleguard sediments and suggested that it most likely is of Sangamonian and perhaps Early Wisconsinan age.

Most workers have assigned the penultimate glaciation in British Columbia to the Early Wisconsinan Substage. However, dating control is poor, and it is possible that this glaciation is Illinoian. Semiahmoo and Dashwood drifts, correlated by Hicock and Armstrong (1983), have yielded radiocarbon dates of >62 000 BP and >51 000 BP, respectively (QL-194, GSC-94-2; Table 1.2). Dashwood Drift may correlate with Possession Drift in Puget Lowland, which has yielded wood and shell amino acid dates of 50-80 ka (Easterbrook and Rutter, 1981). Okanagan Centre Drift, which almost certainly correlates with Semiahmoo and Dashwood drifts, is known to be older than 43 800 ± 800 BP (GSC-740).

A large number of radiocarbon dates have been obtained from the Cowichan Head Formation, Bessette Sediments, and correlative deposits. These dates indicate that the Olympia Nonglacial Interval commenced more than 59 ka and persisted until the beginning of the Fraser Glaciation, 25-29 ka. Finite dates from the Cowichan Head Formation range from 58 800 +2900/−2100 BP to 23 840 ± 300 BP (QL-195, GSC-518; see also Clague, 1980). Bessette Sediments have yielded dates from 43 800 ± 800 BP to 19 100 ± 240 BP (GSC-740, GSC-913; Clague, 1980). Independent evidence of the duration of the Olympia Nonglacial Interval is provided by $^{230}Th/^{234}U$ dates on speleothems from a cave on central Vancouver Island (Gascoyne et al., 1981). The main period of speleothem growth in this cave, from 67 ka to 28 ka, corresponds to at least part of the Olympia Nonglacial Interval. It is possible, however, that the Olympia interval began well before 67 ka and may, in fact, include part or all of the Early Wisconsinan. Stratigraphic evidence and radiocarbon dates indicate that glaciers probably were confined to mountain areas throughout the Olympia interval (Fulton et al., 1976; Clague, 1978).

Climatic deterioration marking the end of the Olympia Nonglacial Interval may have begun as early as 29 ka on the Pacific coast, based on radiocarbon dates from Quadra Sand in the Strait of Georgia area (Clague, 1976, 1977a, 1980; Alley, 1979). Glacier growth was slow during the early stages of the Fraser Glaciation, and many thousands of years elapsed before ice reached lowland areas outside of the major mountain systems (Fig. 1.23). This is indicated, in part, by the long period of time during which Quadra Sand was deposited and by the presence of relatively young nonglacial sediments beneath till in the vicinity of some mountain ranges in southern British Columbia. Radiocarbon dates on that part of Quadra Sand predating Coquitlam Drift range from 28 800 ± 740 BP to 23 920 ± 400 BP (GSC-95, GSC-59; Clague, 1980). Dates on other subtill stratified sediments in southern British Columbia indicate that most plateaus and coastal lowlands remained ice-free until after 21 ka and that some areas were not overridden until after 17 ka (Clague et al., 1980). Chronological control on the growth of the Cordilleran Ice Sheet in northern British Columbia is sparse. However, glaciers in the Omineca Mountains and northern Rocky Mountains did not extend into bordering lowlands until some time after 25 940 ± 380 BP (GSC-573).

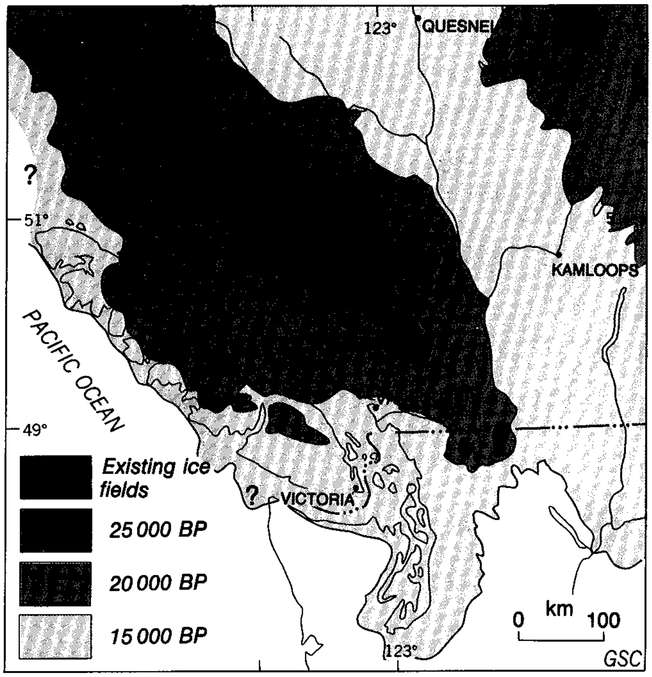


Figure 1.23. Growth of the Cordilleran Ice Sheet in southern British Columbia and northern Washington during the Fraser Glaciation (from Clague, 1981, Fig. 3). Approximate glacier margins at 25 ka, 20 ka, and 15 ka are depicted. Unglaciated areas within the confines of the ice sheet are not shown.

The Coquitlam and Vashon stades are fairly well dated. Coquitlam Drift has yielded a radiocarbon date of 21 500 ± 240 BP (GSC-2536). The interval of glacier recession separating the Coquitlam and Vashon stades lasted from 19-20 ka to 18 ka, based on radiocarbon dates from organic beds in western Fraser Lowland. The Puget lobe of the Cordilleran Ice Sheet achieved its maximum extent about 14-14.5 ka, and other lobes along the southern periphery of the ice sheet also may have reached their Late Wisconsinan limits at about the same time (Waitt and Thorson, 1983). The Puget lobe advanced 200-250 km between 17 ka and 14.5 ka, thus ice sheet growth during this period was relatively rapid.

Decay of the Cordilleran Ice Sheet at the end of the Fraser Glaciation was equally rapid. Radiocarbon dates from glacial marine sediments in the Strait of Georgia region and on northern Vancouver Island indicate that deglaciation was in progress in these areas about 13-13.5 ka

Figure 1.22. Glacial lake evolution and ice retreat in part of south-central British Columbia (adapted from Fulton, 1969, Fig. 3). Five stages in the development of glacial lakes in this region are shown, A being the oldest and E the youngest. Existing lakes are shown in the lower right panel. Although the age of each stage is unknown, deglaciation of this area probably began about 11-12 ka, and the modern drainage pattern was established before 9 ka.

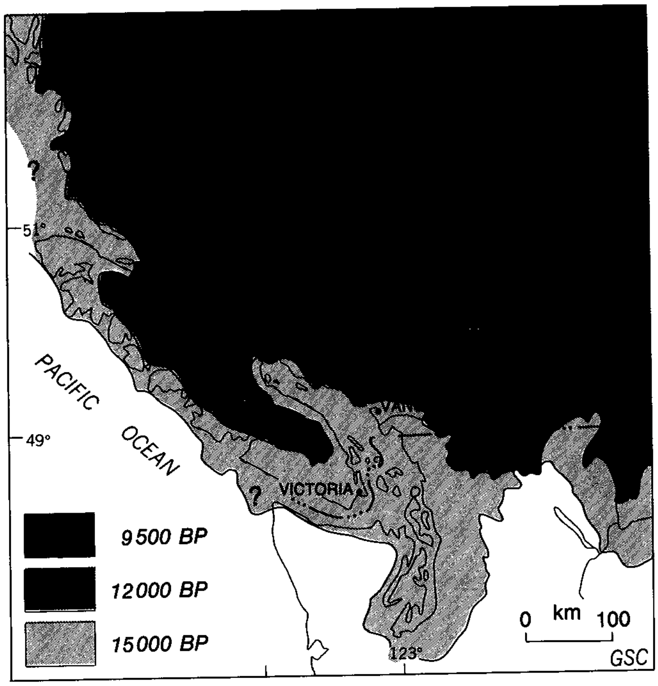


Figure 1.24. Decay of the Cordilleran Ice Sheet in southern British Columbia and northern Washington at the close of the Fraser Glaciation (from Clague, 1983, Fig. 5). Approximate glacier margins at 15 ka, 12.5 ka, and 9.5 ka are depicted. Unglaciated areas within the confines of the ice sheet are not shown.

(Fig. 1.24; Clague, 1980). Fraser Lowland and fiords bordering the Strait of Georgia were completely ice-free about 11 ka at the end of the Sumas Stade (Armstrong, 1981; Saunders et al., 1987). Some upland areas in the southern interior may have become deglaciated as early as 13-14 ka; low plateaus and intermontane valleys remained covered until about 10.5-11 ka (Fulton, 1971; Clague, 1981). The Terrace-Kitimat area in the northern Coast Mountains still supported glaciers as late as 10 ka (Clague, 1984, 1985), but by 9.5 ka glaciers throughout British Columbia were no more extensive than they are today (Clague, 1981).

Yukon Territory

O.L. Hughes, N.W. Rutter, and J.J. Clague

Our understanding of the Quaternary history of that part of Yukon Territory covered by the Cordilleran Ice Sheet has come mainly from studies of glacial sediments and landforms, thus the main emphasis in this section is on the glacial record.

Hughes, O.L., Rutter, N.W., and Clague, J.J.
1989: Yukon Territory (Quaternary stratigraphy and history, Cordilleran Ice Sheet); in Chapter 1 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

Old glaciations

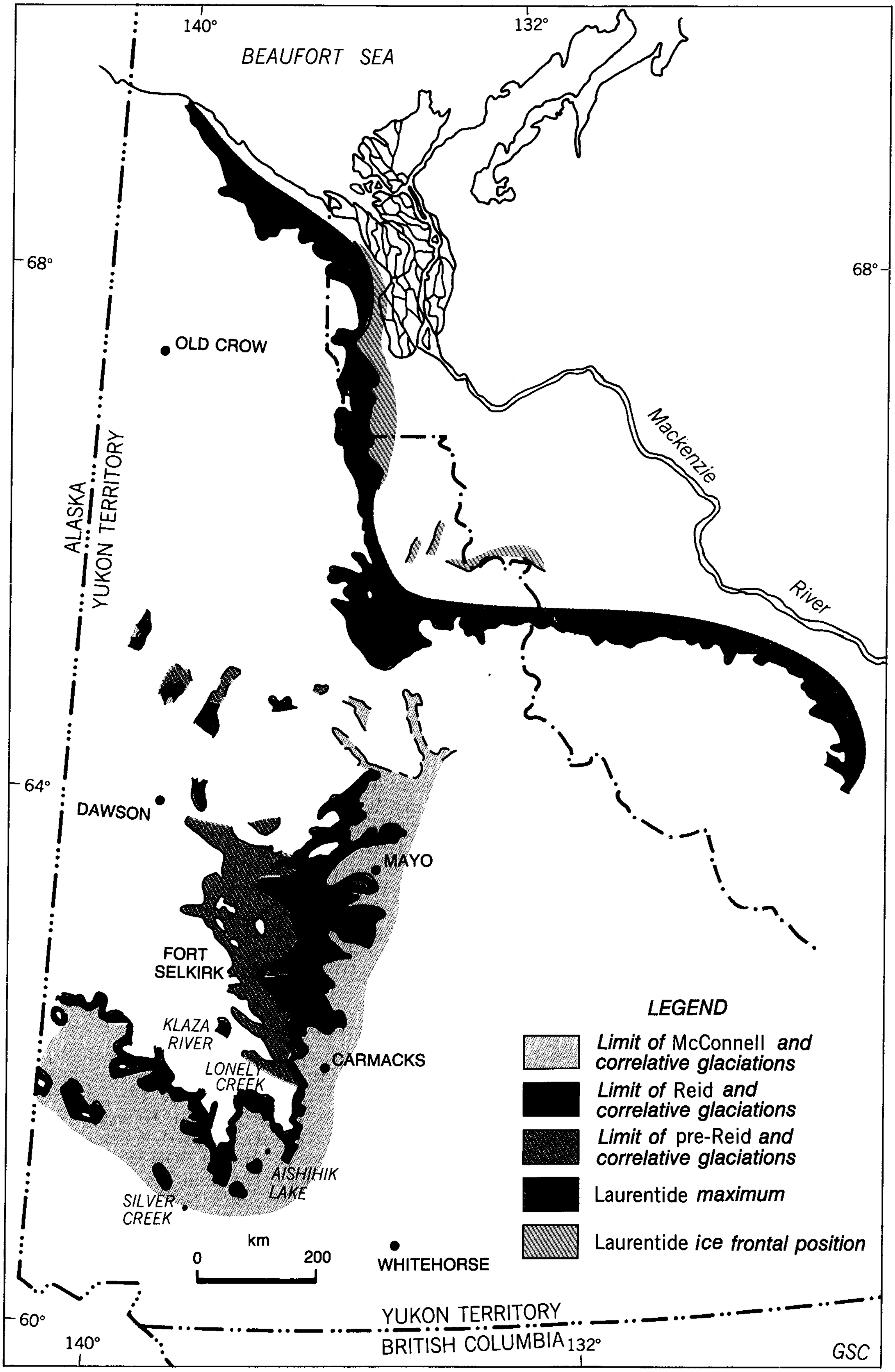
Bostock (1966) inferred four advances of the Cordilleran Ice Sheet in southern and central Yukon Territory, with each successive advance being less extensive than its predecessor: Nansen (oldest), Klaza, Reid, and McConnell (Fig. 1.25, 1.26). Few readily recognizable landforms remain from the Nansen and Klaza glaciations, hence their limits are poorly known. In contrast, ice marginal features of the Reid Glaciation are moderately well preserved and those of the McConnell Glaciation very well preserved, permitting air-photo interpretation of their limits across much of central Yukon Territory (Hughes et al., 1969). Reid and McConnell deposits also may be distinguished from each other and from older glacial deposits by conspicuous differences in soil development (Hughes et al., 1972; Foscolos et al., 1977; Rutter et al., 1978; Tarnocai et al., 1985). Soils on pre-Reid drift at well drained sites are Luvisols with thick Bt horizons; soils on Reid Drift are moderately developed Luvisols, or Brunisols with thick Bm horizons; and soils on McConnell Drift are Brunisols with thin Bm horizons (see Canada Soil Survey Committee, 1978, for definition of soil terms).

The Nansen Glaciation was named for occurrences of weathered till and weathered gravel of presumed glacial origin in the upper reaches of Nansen Creek and other streams west of Carmacks (Bostock, 1966). A later glaciation, the Klaza, was inferred by Bostock on the basis of evidence for northward diversion of the headwaters of south-flowing Lonely Creek into the Klaza River system. Although it cannot be proven that this diversion was not accomplished during the more extensive Nansen Glaciation, there is evidence elsewhere for two pre-Reid glaciations. For example, at Fort Selkirk outside the Reid limit, a sequence of till, gravel, sand, and silt deposited during an earlier (Nansen?) glaciation is overlain by about 100 m of basaltic lava flows. Glacial striae on the surface of the flows are the product of a later glaciation, presumably the Klaza.

Deposits of two pre-Reid glaciations also are exposed in sections in Liard Lowland, southeastern Yukon Territory (Fig. 1.27; Klassen, 1978, 1987). The older glaciation is represented by a till (Till A) which overlies Tertiary sediments and volcanics (Liard Formation) and is overlain by thick sand and silt. Later ice sheet glaciation of Liard Lowland is indicated by a second till (Till B) which underlies dense lacustrine clay containing plant detritus and gastropod shells. Basalt flows occur directly above Till A and between Till B and the dense lacustrine clay (Fig. 1.27).

In southwestern Yukon Territory, which was affected by successive advances of glaciers from the St. Elias Mountains, there are no surface deposits comparable either in soil or landform development to Nansen or Klaza drifts. However, the lowest of the three drift units exposed at Silver Creek near Kluane Lake probably predates the Reid Glaciation (Fig. 1.28; Denton and Stuiver, 1967). This unit, termed Shakwak Drift, consists of till and outwash gravel that have been oxidized throughout their exposed thickness

Figure 1.25. Maximum extent of glaciers in Yukon Territory and western District of Mackenzie during various Pleistocene glaciations (adapted from Hughes et al., 1983, Fig. 2).


140°
132°
BEAUFORT SEA
68°
68°
OLD CROW
ALASKA
YUKON TERRITORY
Mackenzie
River
64°
DAWSON
MAYO
FORT SELKIRK
KLAZA RIVER
LONELY CREEK
CARMACKS
AISHIHIK LAKE
SILVER CREEK
km
0
200
WHITEHORSE
LEGEND
Limit of McConnell and correlative glaciations
Limit of Reid and correlative glaciations
Limit of pre-Reid and correlative glaciations
Laurentide maximum
Laurentide ice frontal position
60°
140°
YUKON TERRITORY
BRITISH COLUMBIA
132°
GSC

	Yukon Plateaus (Bostock, 1966; Hughes et al., 1969)	Snag-Klutlan area (Rampton, 1971a)	Silver Creek (Denton and Stuiver, 1967)	Liard Lowland (Klassen, 1978, 1987)	Southern Ogilvie Mtns (Vernon and Hughes, 1966)	Old Crow Flats (Morlan, 1980)
HOLOCENE	POSTGLACIAL	NEOGLACIATION	NEOGLACIATION SLIMS NONGLACIAL INTERVAL	POSTGLACIAL		
		13.7 ka	12.5 ka		13.7 ka	12.5 ka
PLEISTOCENE — WISCONSINAN	McCONNELL GLACIATION	MACAULEY GLACIATION	KLUANE GLACIATION	TILL D	LAST GLACIATION	GLACIAL-LACUSTRINE CLAY
			< 29.6 ka	< 23.9 ka		< 25.2 ka
	REID SOIL SHEEP CK TEPHRA (ca. 80 ka)	OLD CROW TEPHRA (80-130 ka)	BOUTELLIER NONGLACIAL INTERVAL	INTERTILL UNIT C-D		ALLUVIUM "DISCONFORMITY A" OLD CROW TEPHRA (80-130 ka)
	> 46.6 ka	> 48 ka	> 37.7 ka		> 53.9 ka	
	REID GLACIATION	MIRROR CREEK GLACIATION	ICEFIELD GLACIATION	TILL C	INTERMEDIATE GLACIATION	
	PRE-REID SOIL		SILVER NONGLACIAL INTERVAL	INTERTILL UNIT B-C?		ALLUVIUM
				> 0.23 Ma		
PLEISTOCENE — PRE-WISCONSINAN	KLAZA GLACIATION	?	SHAKWAK GLACIATION?	TILL B?	OLD GLACIATION?	
	< 1.08 Ma					
	FORT SELKIRK TEPHRA (0.84-0.94 Ma)			INTERTILL UNIT A-B?	MOSQUITO GULCH TEPHRA (1.22 Ma)	LITTLE TIMBER TEPHRA (>1.2 Ma)
				> 0.76 Ma		
	NANSEN GLACIATION		SHAKWAK GLACIATION?	TILL A?	OLD GLACIATION	LACUSTRINE CLAY

GSC

Figure 1.26. Subdivisions of Quaternary events and deposits in Yukon Territory.

(24 m). At Silver Creek, Shakwak Drift is overlain by two younger drift units (Icefield and Kluane drifts), each of which comprises lower and upper outwash gravels and one or more intervening tills.

Reid Glaciation

The Reid Glaciation was named by Bostock (1966) for prominent moraines south and west of Reid Lakes in central Yukon Territory. These mark the limit of an advance of the Cordilleran Ice Sheet in Stewart River valley that was less extensive than the older Nansen and Klaza advances. Moraines and other ice marginal features of Reid age have been modified by mass wasting processes, but in general they are readily recognizable on airphotos.

Reid ice marginal features can be traced with some breaks across the Yukon Plateaus into the St. Elias area of southwestern Yukon Territory. Here, they appear to be continuous with features marking the limit of the Mirror Creek Glaciation (Hughes et al., 1969, 1983; Rampton, 1971a).

Units of possible Reid age exposed in sections in southern Yukon Territory include Till C in Liard Lowland and Icefield Drift at Silver Creek (Fig. 1.27, 1.28). Till C is underlain by lacustrine clay and is overlain successively by Intertill unit C-D and Till D, the youngest till in Liard Lowland. Intertill unit C-D comprises a lower clay layer, a middle gravel, and an upper fossiliferous silt (Tom Creek Silt) which contains an interstadial pollen assemblage. Icefield Drift is overlain by Kluane Drift, the youngest Pleistocene glacial deposit in southwestern Yukon Territory.

McConnell Glaciation

Bostock (1966) applied the name McConnell to a prominent morainal loop that crosses Stewart River about 18 km southwest of Mayo. The moraine and associated features can be traced almost continuously across the Yukon Plateaus to delineate the highly digitate northwestern margin of the Cordilleran Ice Sheet at the climax of the last glaciation (Fig. 1.29). The McConnell limit is continuous with the Macauley glacial limit in the St. Elias area of southwestern Yukon Territory (Hughes et al., 1969, 1983; Rampton, 1971a). Other deposits of McConnell age include Till D and associated glacial fluvial sediments in Liard Lowland and Kluane Drift in the Kluane Lake area (Fig. 1.27, 1.28).

Chronology and correlation

Evidence bearing on the age of pre-Reid glaciations has been obtained at sections near Fort Selkirk and in Liard Lowland. Till at the base of the Fort Selkirk section, thought by Bostock (1966) to be Nansen in age or possibly older, is overlain by stratified sediments containing Fort Selkirk tephra and by lava flows with a striated upper surface. The tephra has yielded glass fission-track ages of 0.84 ± 0.13 Ma and 0.86 ± 0.18 Ma and a zircon fission-track age of 0.94 ± 0.40 Ma (Naeser et al., 1982). A sample of basalt near the base of the flows gave a whole-rock K-Ar age of 1.08 ± 0.05 Ma (Naeser et al., 1982), which does not differ significantly from the fission-track ages at the 2σ level. The K-Ar age is compatible with the reversed magnetic polarity of the dated lava flow. It follows that the till at the base of the section is older than about 1 Ma and that the advance

(Klaza?) that left striae on the surface of the lava flows is younger than 1 Ma.

In Liard Lowland, basalt flows below Till C but above the Liard Formation range in age from 0.765 ± 0.049 Ma to 0.232 ± 0.021 Ma (Fig. 1.27). The flow that yielded the 0.765 Ma date is thought to overlie Till A and underlie Till B. If so, Till B is Middle Pleistocene and Till A Middle or Early Pleistocene in age. Till C, which presumably was deposited during the Reid Glaciation, is younger than about 0.232 Ma.

Radiometric age determinations and continuity of surface deposits indicate that the Reid and Mirror Creek glaciations are equivalent. The Cordilleran Ice Sheet advanced to the Reid limit near the type locality and began to retreat >42 900 BP (GSC-524, Table 1.2) and probably before 80 ka. The former date was obtained on wood collected from a layer of Sheep Creek tephra overlying Reid Drift; the tephra is thought to be about 80 ka old (Hamilton and Bischoff, 1984). A radiocarbon date of >46 580 BP (GSC-331) on wood collected from beneath McConnell till is also a minimum for the Reid Glaciation. Radiocarbon dates from organic material above Mirror Creek Drift and below Macauley Drift in southwestern Yukon Territory are all infinite, except for one of 48 000 ± 1300 BP (GSC-732) for which contamination by modern rootlets is considered possible (Rampton, 1971a). At one site near Kluane Lake, Mirror Creek Drift is overlain by Old Crow tephra, which may be older than 100 ka (Schweger and Matthews, 1985; Westgate et al., 1985; Wintle and Westgate, 1986; Berger, 1987). It thus is clear that the Mirror Creek and Reid glaciations are no younger than Early Wisconsinan; many workers favour an Illinoian age for these events.

The Silver Creek section of Denton and Stuiver (1967) and a section in Liard Lowland studied by Klassen (1978, 1987) have yielded finite radiocarbon dates from sediments beneath the surface drift. At Silver Creek, Icefield till is overlain by gravel containing organic-rich silt beds dating from 37 700 +1500/−1300 BP to 29 600 ± 460 BP (Y-1356, GSC-769; Fig. 1.28). Twig fragments from the lower and upper parts of Tom Creek Silt in Liard Lowland have yielded radiocarbon dates of >30 000 BP and 23 900 ± 1140 BP, respectively (GSC-2949, GSC-2811). These dates suggest that Intertill unit C-D (which includes Tom Creek Silt)

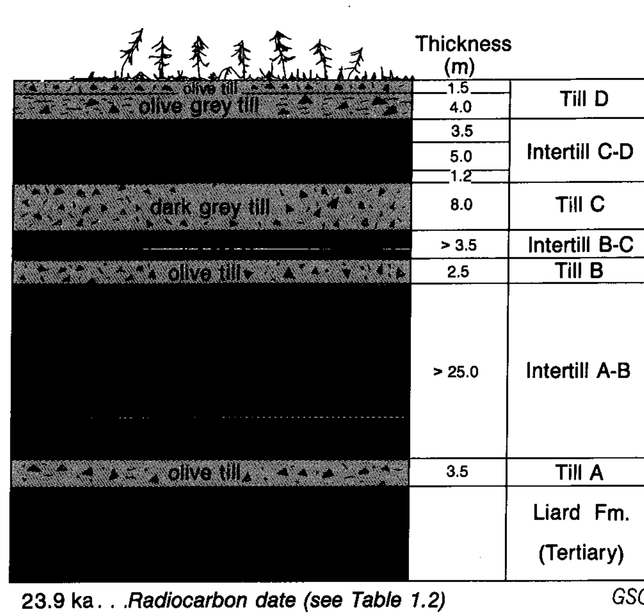


Figure 1.27. Composite Tertiary-Quaternary stratigraphic section, Liard Lowland, Yukon Territory (adapted from Klassen, 1978, Fig. 2, and 1987, Fig. 14).

is Middle Wisconsinan in age and that southern Yukon Territory was deglaciated during the Boutellier Nonglacial Interval.

The date of 29 600 ± 460 BP (GSC-769) at Silver Creek is a maximum for the initial advance of ice out of the St. Elias Mountains during the Kluane Glaciation. The date of 23 900 ± 1140 BP (GSC-2811) from Liard Lowland likewise is a maximum for the last glacier incursion into that area. Deglaciation began on the east side of the St. Elias Mountains before 13 660 ± 180 BP (GSC-495), and Kaskawulsh Glacier, one of the large valley glaciers in the St. Elias Mountains, was less extensive than at present by 9780 ± 80 BP (Y-1483).

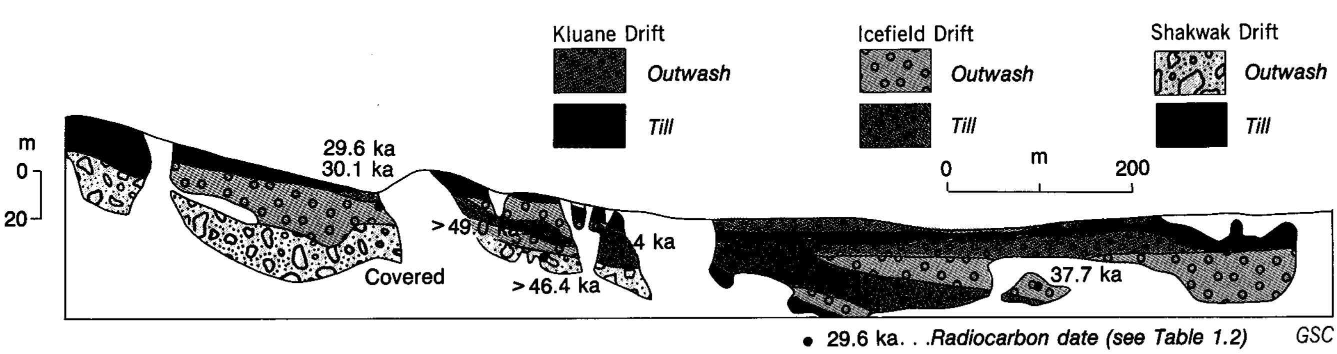


Figure 1.28. Exposure along Silver Creek, Yukon Territory, showing deposits of three glaciations (adapted from Denton and Stuiver, 1967, Plate 4A).

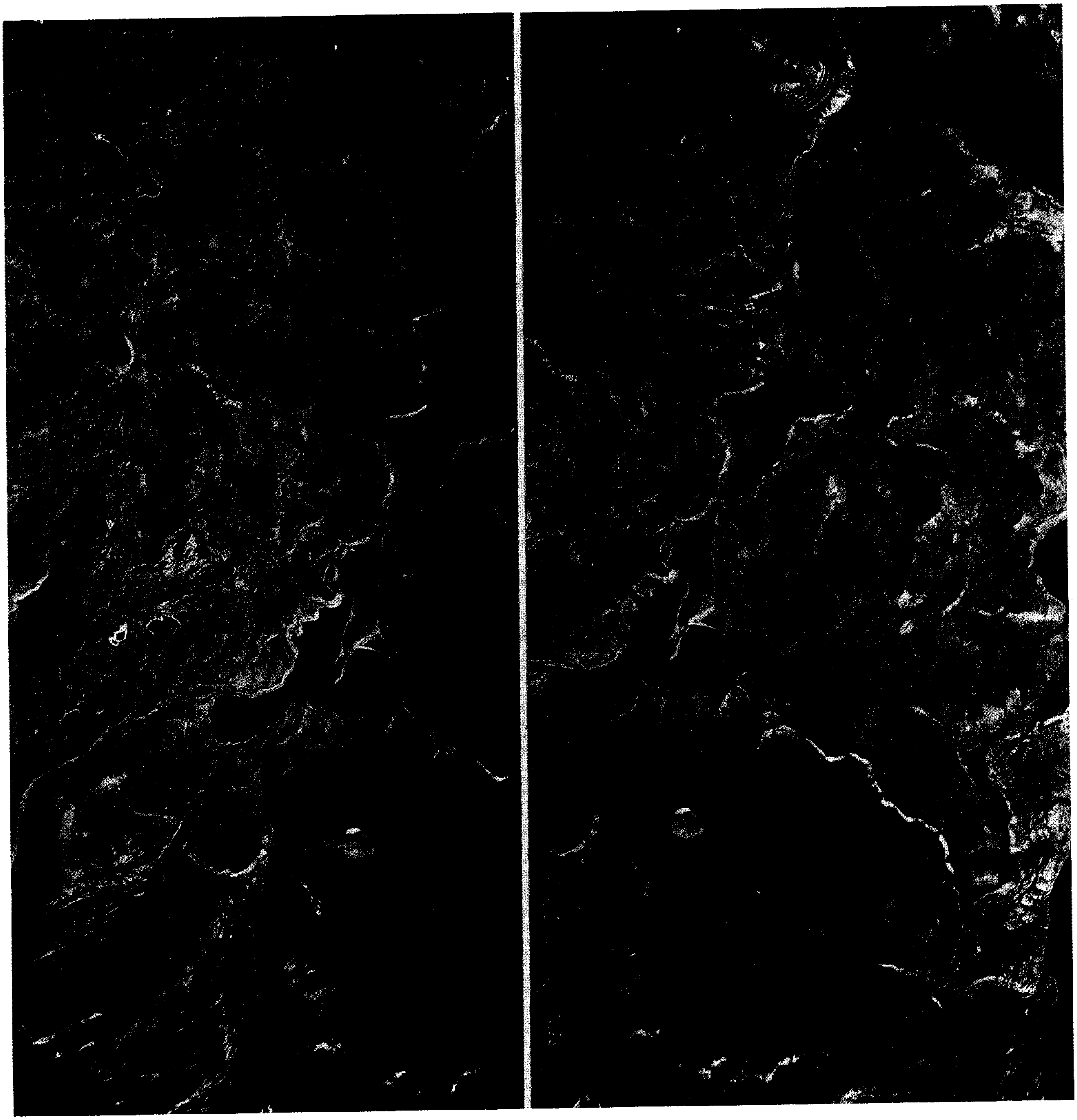


Figure 1.29. Stereogram showing Late Wisconsinan glacial limit (dotted lines) on the upland east of Aishihik Lake, southwestern Yukon Territory. The upland formed a reentrant between two north-flowing lobes of the Cordilleran Ice Sheet; the area between the dotted lines was not glaciated during the Late Wisconsinan. The delta (d) and beaches (b) formed in a glacial lake impounded at the ice margin. NAPL A15739-76, -77

Glaciated fringe

L.E. Jackson, Jr., N.W. Rutter, O.L. Hughes, and J.J. Clague

Introduction

Parts of the Canadian Cordillera were incompletely glaciated during the Pleistocene or escaped glaciation altogether. This section summarizes the Quaternary history of those areas affected mainly by glaciers that were independent or nearly independent of the Cordilleran Ice Sheet. These areas are located at the fringe of the former ice sheet and include parts of the Rocky Mountains and Mackenzie Mountains on the east, the Richardson, Southern Ogilvie, and Wernecke mountains on the north, and the Queen Charlotte Ranges on the west.[1] During Pleistocene glaciations, major valleys in these ranges were occupied by tongues of ice, and summit areas supported ice caps. In addition, the Laurentide Ice Sheet on a few occasions overrode the eastern foothills of the Rocky, Mackenzie, and Richardson mountains.

The Quaternary history of the glaciated fringe has been inferred mainly from stratigraphic and geomorphic evidence. Most deposits that have been studied are glacial in origin, thus Quaternary historical reconstructions for this region emphasize glacial events. The paucity of relevant absolute age determinations, however, has made it extremely difficult to date these events, and the lack of continuity of individual drift sheets from one drainage basin to the next and from one mountain range to another has made regional correlations inferential and tentative at best. Even in a single basin, stratigraphic relationships may not be straightforward, and consequently different conclusions as to the sequence of glacial events have been reached by different workers. As a result of these problems, there is considerable disagreement and confusion regarding the number and age of glaciations in the eastern Cordillera, even in the most intensively studied areas. The Quaternary stratigraphy and glacial history of the Cordilleran glaciated fringe are summarized on the following pages and in the accompanying correlation chart (Fig. 1.30). Undoubtedly, some of the interpretations and correlations that are made here will have to be modified as additional data become available.

Old glaciations

Drift deposited by mountain glaciers during various pre-Wisconsinan glaciations is found in several areas on the eastern flank of the Rocky Mountains and in the Southern Ogilvie Mountains.

Stalker and Harrison (1977), working in Waterton Lakes National Park, recognized four major glaciations, at least one of which is pre-Wisconsinan in age. During the oldest, or Great, glaciation, Rocky Mountain ice extended to high elevations in the foothills and far out onto the Interior Plains, where it deposited Albertan Till. Laurentide ice dropped erratics derived from the Canadian Shield high on the mountain front and spread Labuma Till over Albertan Till in lower areas. Mountain glaciers and the Laurentide Ice Sheet were out-of-phase during the Great Glaciation, the former having withdrawn before the latter reached its climax position. Whether the two ever coalesced in the southern Rocky Mountains remains uncertain, but there probably was at least local contact.

A study of paleosols by Karlstrom (1981, 1987) complements the work of Stalker and Harrison (1977) in the Waterton Lakes area. Karlstrom found five diamicton layers, each with a capping paleosol, on Mokowan Butte, a plateau-like hill rising over 400 m above surrounding terrain at the eastern boundary of Waterton Lakes National Park. According to Karlstrom, the diamictons probably are tills deposited during five separate glaciations, the youngest of which presumably correlates with the Great Glaciation.

Farther north, in Oldman and Crowsnest river valleys, Alley (1973) and Alley and Harris (1974) found evidence for a four-fold sequence of glaciation similar to that proposed by Stalker and Harrison (1977). The oldest advance, termed Glacial Episode 1, is probably equivalent to the Great Glaciation in Waterton Lakes National Park. The Laurentide Ice Sheet impounded lakes during the advance and recessional stages of Glacial Episode 1 and subsequent glaciations. Sediments deposited in these lakes locally separate tills of Rocky Mountain and Keewatin provenance and thus provide stratigraphic evidence that mountain glaciers and the Laurentide Ice Sheet were out-of-phase in the Oldman River area.

Jackson (1980), working in Highwood River basin north of Oldman River, identified an old glaciation equivalent to Alley's (1973) Glacial Episode 1. Evidence consists of erratics and scattered patches of till above 1400 m elevation in Porcupine Hills at the mountain front. This glaciation involved nonsynchronous advances of Rocky Mountain and Laurentide glaciers.

Farther north, Ford (1973, 1976) and Mathews (1978) documented pre-Wisconsinan advances of the Laurentide Ice Sheet to the western edge of the Interior Plains. Mathews (1978), working in the Peace River area, found Canadian Shield pebbles in an aggradational gravel unit deposited before the penultimate incursion of ice into the area. This gravel was laid down by streams flowing from the Rocky Mountains, thus Mathews inferred that some time prior to the penultimate glaciation, Laurentide ice extended west of the location of the shield pebbles (i.e., west of the British Columbia-Alberta boundary). Ford (1973, 1976) identified shield erratics in the foothills of the Mackenzie Mountains near South Nahanni River and assigned them to the oldest and most extensive of three major glaciations in the region (First Canyon Glaciation). This glaciation and the oldest glaciation in the Peace River area thus record major expansions of the Laurentide Ice Sheet. Presumably, mountain glaciers and the Cordilleran Ice Sheet grew at the same time, although sedimentary and other evidence for this has not been found.

Valley glaciers elsewhere in the Mackenzie Mountains reached Mackenzie Lowland and flowed onto Peel Plateau

[1] Glaciers in the St. Elias Mountains and Vancouver Island Ranges are not discussed here because they were confluent with, and thus part of, the Cordilleran Ice Sheet at the climaxes of most glaciations.

Jackson, L.E., Jr., Rutter, N.W., Hughes, O.L., and Clague, J.J.
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on at least one occasion. Extensive interfluve areas in the eastern ranges of the Mackenzie Mountains, however, were never covered by these glaciers, nor by the Laurentide Ice Sheet which abutted against the mountain front.

During one or more "old" glaciations, glaciers flowed from the Southern Ogilvie Mountains into Taiga Valley on the north and Tintina Trench on the south (Vernon and Hughes, 1966). These glaciers deposited more than 200 m of glacial lacustrine sediments, outwash gravel, and till in Tintina Trench east of Dawson (Flat Creek beds of McConnell, 1905). West of the trench, outwash gravel derived from the Southern Ogilvie Mountains (Klondike gravels of McConnell, 1907) lies on a high bedrock terrace along the lower reaches of Klondike River. Near Dawson, the Klondike gravels overlie and intertongue with the White Channel gravels (McConnell, 1907) of probable late Pliocene-early Pleistocene age.

Penultimate glaciation

Deposits of the penultimate glaciation are much more common and better preserved than those of older glaciations. In this section, the record of the penultimate glaciation is summarized, starting first in the Rocky Mountains, proceeding to the Mackenzie and Southern Ogilvie mountains, and ending on the Queen Charlotte Islands.

In Waterton Lakes National Park, the Great Glaciation was followed by the less extensive Waterton II advance, during which Rocky Mountain ice extended east onto the Interior Plains and the Laurentide Ice Sheet deposited till as far west as the mountain front (Stalker and Harrison, 1977). As was the case during the Great Glaciation, mountain glaciers receded before the Laurentide Ice Sheet achieved its maximum extent. With present information, it is not clear if Waterton II and Waterton III, the next major Rocky Mountain advance, were separated by a major nonglacial interval or were part of one large glaciation. The former interpretation is favoured here and is shown in Figure 1.30.

Maycroft Till, the oldest widespread montane till in Oldman, Crowsnest, and Highwood river valleys, was deposited during Glacial Episode 2 when Rocky Mountain ice extended east onto the Interior Plains (Alley, 1973; Jackson, 1980). This ice perhaps coalesced locally with Laurentide ice which deposited Maunsell Till. Glacial lacustrine sediments (Chain Lake Clays and Silts) accumulated in lakes at the margins of these ice masses.

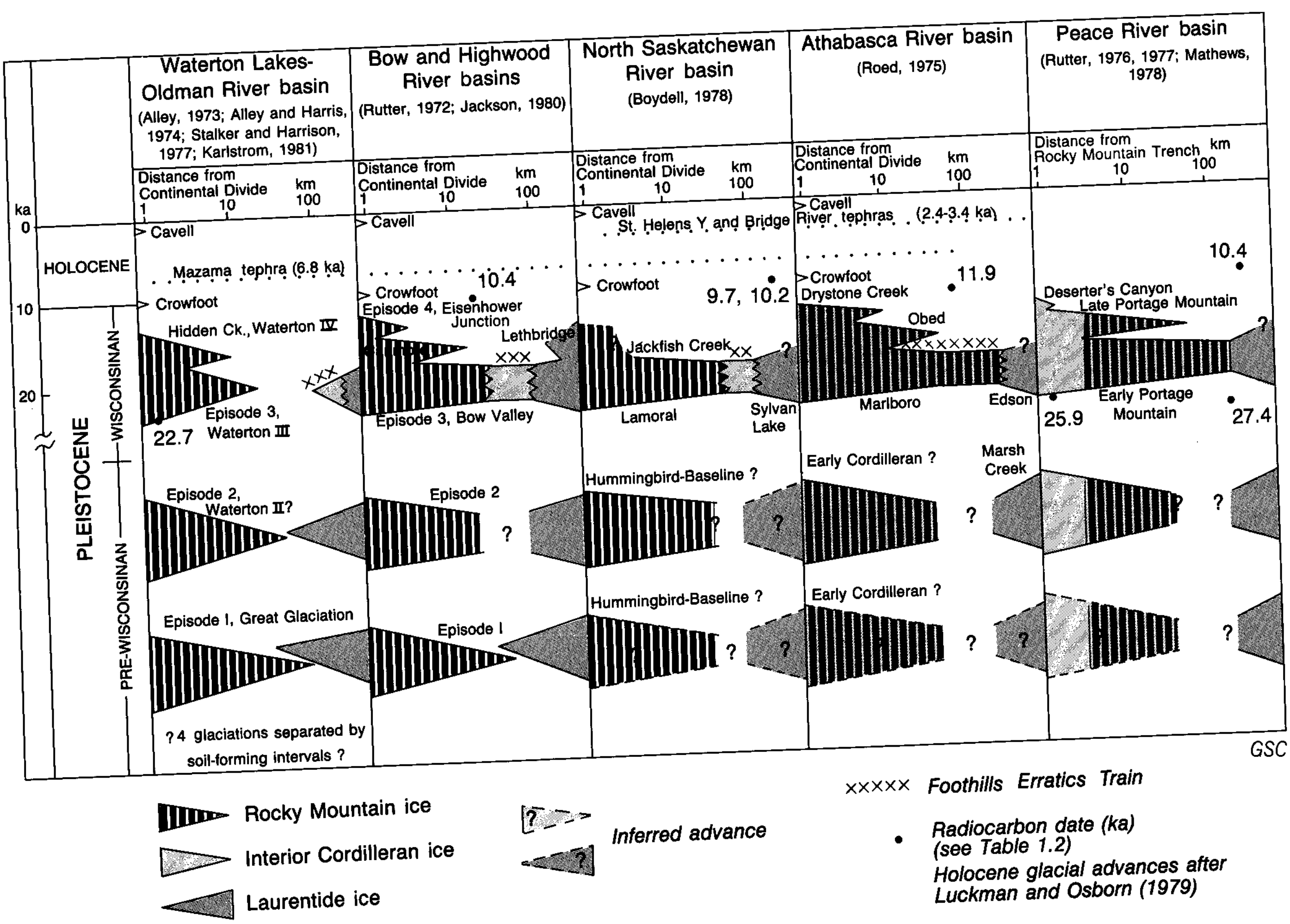

Figure 1.30. Quaternary glacial events in the Rocky Mountains and adjacent Interior Plains. The diagram shows the approximate extent of glaciers, the names of glaciations, and proposed inter-basin correlations. Some age assignments differ from those of original authors.

Fragmentary deposits attributable either to the penultimate glaciation or to an earlier glaciation include outwash underlying the oldest till in Bow River valley (Rutter, 1972), the montane Baseline and Hummingbird tills in North Saskatchewan River basin (Boydell, 1978), and the montane Early Cordilleran and Laurentide Marsh Creek tills in Athabasca River basin (Roed, 1975).

The penultimate glaciation in the Peace River area is recorded by a stratified valley fill capped by till of Keewatin provenance (Mathews, 1978). The valley fill consists of fluvial or glacial fluvial gravel and overlying glacial lacustrine sand, silt, and clay deposited when Peace River was ponded by advancing Laurentide ice. This ice eventually entered the area from the east or north and flowed an unknown distance to the west. The extent of Cordilleran ice at this time is unknown.

Undated moraines comparable in preservation to those of the penultimate (Reid) glaciation in southern and central Yukon Territory are present in many valleys in the eastern Mackenzie Mountains. The glaciers that deposited these moraines were smaller than those of at least one earlier glaciation, but larger than those of the last glaciation. At several localities along the northern front of the Mackenzie Mountains, Laurentide ice marginal features truncate moraines deposited by mountain glaciers during the penultimate glaciation. Elsewhere, however, these moraines are situated upvalley of the later limit of the Laurentide Ice Sheet. The trunk glacier in South Nahanni River valley, for example, terminated about 100 km short of the mountain front at the climax of the penultimate (Flat River-Clausen) glaciation; during the same glaciation, Laurentide ice extended into the eastern foothills of the Mackenzie Mountains, damming a lake (glacial Lake Nahanni) in which thick silt and clay accumulated (Ford, 1976). These sediments overlie the montane Flat River Till, suggesting that the South Nahanni valley glacier may have retreated prior to the incursion of Laurentide ice into the area.

Glaciers in the Southern Ogilvie Mountains during the penultimate ("intermediate") glaciation were similar to, although less extensive than, those of the "old" glaciation(s) (Vernon and Hughes, 1966). During the "intermediate" glaciation, ice in North Klondike River valley reached the east margin of Tintina Trench where it deposited drift on surfaces incised into deposits of an "old" glaciation. Apparently, these two glaciations were separated by a lengthy period of time during which Klondike River and its tributaries eroded their valleys to near present-day levels, about 200 m below "old" glaciation valley floors.

During the penultimate glaciation, the Queen Charlotte Islands supported valley and piedmont glaciers that flowed west and east from local mountain ice caps (Clague et al., 1982a). On eastern Graham Island, these glaciers probably came into contact with a westward-flowing lobe of the Cordilleran Ice Sheet. Deposits of this glaciation underlie a younger drift succession and are well exposed in coastal bluffs bordering northern Hecate Strait. These deposits comprise: (1) massive to weakly stratified, glacial marine mud containing foraminifera, molluscan shells, and scattered stones derived from the British Columbia mainland; (2) till of Queen Charlotte Islands provenance; and (3) stratified sand and gravel.

Last glaciation

Deposits and landforms of the last Pleistocene glaciation in the glaciated fringe are conspicuous and relatively easy to correlate, although in some areas their age is in dispute. In this section, we briefly review the record of the last glaciation, again proceeding in a counterclockwise direction around the Canadian Cordillera.

Sediments deposited at the maximum of the last glaciation in the southern and central Rocky Mountains have been recognized and correlated on the basis of their association with the Foothills Erratics Train, a band of distinctive erratics derived from the mountains near Jasper (Stalker, 1956; Mountjoy, 1958; Morgan, 1966, 1969). The Foothills Erratics Train and associated "mixed provenance" till were deposited by a south-flowing ice stream bordered on the west by montane ice and on the east by the Laurentide Ice Sheet. In the Highwood and Bow river areas, the montane Ernst and Bow Valley tills are in contact with the mixed provenance Erratics Train Till near the front of the Rocky Mountains. These tills were deposited when Rocky Mountain and Laurentide ice last coalesced in this region (Glacial Episode 3 of Jackson, 1980; Bow Valley advance of Rutter, 1972). Farther north in North Saskatchewan River basin, mountain glaciers coalesced with the Laurentide Ice Sheet during the Jackfish Creek-Sylvan Lake glacial episode (Boydell, 1978). There, Jackfish Creek Till of Rocky Mountain provenance grades eastward into mixed provenance Athabasca Till capped by the Foothills Erratics Train; Athabasca Till, in turn, is bordered on the east by Sylvan Lake Till of Keewatin provenance. Still farther north in Athabasca River basin, Marlboro Till, deposited in part by mountain glaciers and in part by ice streaming out of the British Columbia interior, grades laterally into mixed provenance till overlain by Foothills erratics. In this area, the mixed provenance till is bordered on the east by Edson Till, deposited by the Laurentide Ice Sheet. Thus, during the last glaciation, Rocky Mountain (and interior Cordilleran) ice coalesced with the Laurentide Ice Sheet along the eastern edge of the Canadian Cordillera from Highwood River on the south to at least Athabasca River on the north.

South of Highwood River, in Oldman River basin and in Waterton Lakes National Park, Rocky Mountain and Laurentide ice apparently did not coalesce at the climax of the last glaciation (Glacial Episode 3 of Alley, 1973; Waterton III of Stalker and Harrison, 1977). Mountain glaciers flowed east to the front of the Rockies and deposited Ernst Till, while at about the same time mixed provenance till, the Foothills Erratics Train, and Laurentide Buffalo Lake Till were being deposited on the adjacent Interior Plains. Lakes dammed by the Laurentide Ice Sheet covered parts of this region during Glacial Episode 3.

There were one or more readvances in the southern Rocky Mountains during the closing stages of the last glaciation. During the Waterton IV-Hidden Creek advance, glaciers in Waterton Lakes National Park and Oldman and Crowsnest valleys terminated well inside the mountain front and thus were much less extensive than at the last glacial maximum.[1] In Highwood, Bow, and Athabasca valleys,

[1] Some workers have questioned this interpretation, claiming that the Waterton IV and Hidden Creek advances are the climactic advances of the last glaciation and that the Foothills Erratics Train and associated mixed provenance till were deposited during the penultimate glaciation (Alley, 1973; Stalker and Harrison, 1977; Jackson, 1980; Rutter, 1980, 1984; Reeves, 1983).

glaciers reached the mountain front during the Canmore and Obed advances (probable correlatives of the Waterton IV — Hidden Creek event; Rutter, 1972; Roed, 1975); a subsequent advance (Eisenhower Junction, Drystone Creek) was much less extensive.

The Cordilleran and Laurentide ice sheets and Rocky Mountain glaciers coalesced in northeastern British Columbia and adjacent Alberta during the last glaciation. Drumlins and other streamlined forms in the Peace River area indicate that southwest-flowing Laurentide ice was deflected in a broad sweeping arc south and southeast by Cordilleran ice flowing east and northeast (Mathews, 1978). The surface till in this area is Cordilleran in provenance in the west and Keewatin in the east, and locally overlies an aggradational valley fill similar to that deposited during the early part of the penultimate glaciation (see preceding section).

The pattern of deglaciation in the northern Rocky Mountains and adjacent Interior Plains at the end of the last glaciation has been reconstructed by Mathews (1980) from geomorphic and stratigraphic evidence. Mathews was able to relate the retreat of the Cordilleran Ice Sheet to that of the Laurentide Ice Sheet by tracing the evolution of glacial Lake Peace which was ponded by Laurentide ice. He showed that the end moraine near Portage Mountain marking the terminus of the Late Portage Mountain advance (Rutter, 1976, 1977) was built out into a high-level lake, probably the Bessborough stage of Lake Peace. Cordilleran ice remained near the mountain front through the Bessborough stage and an unnamed stage that followed. During the next (Clayhurst) stage, Cordilleran ice retreated more than 100 km from the Portage Mountain area, while the Laurentide Ice Sheet continued to impound Lake Peace from positions east of the British Columbia-Alberta boundary. There was a minor readvance of glaciers in Halfway River valley on the east side of the Rocky Mountains during or shortly after the Clayhurst stage.

The last glaciation in the eastern Mackenzie and Southern Ogilvie mountains was restricted, with glaciers in many areas confined to valleys near the crests of the ranges. In South Nahanni National Park, for example, glaciers apparently reached no more than 30 km from cirques during the last (Hole-in-the-Wall) glaciation (Ford, 1976). Some valley glaciers in the eastern Mackenzie Mountains, however, extended into the foothills and coalesced with the Laurentide Ice Sheet at the climax of the last glaciation. These glaciers may have achieved their maximum extent shortly after Laurentide ice began to recede (A. Duk-Rodkin, Geological Survey of Canada, personal communication, 1987).

During the last glaciation, there were significant ice-free areas on the Queen Charlotte Islands, including intervalley ridges, some mountains and coastal lowlands, and possibly shallow offshore platforms (Warner et al., 1982). Glaciers flowed from the Queen Charlotte Ranges and terminated near the present shoreline; those on eastern Graham Island briefly coalesced with the Hecate lobe of the Cordilleran Ice Sheet. Drift of the last glaciation is particularly well exposed in sea cliffs bordering Argonaut Plain on northeastern Graham Island. In this area, the drift consists of: (1) thick, well sorted, crossbedded sand; and (2) overlying till and associated ice contact gravel of local provenance (Clague et al., 1982a). A unit of massive to weakly stratified, gravelly mud, which in places directly underlies the crossbedded sand, may be part of this drift sequence; alternatively, it may predate the last glaciation. Over much of Argonaut Plain, unit 1 sand extends to the surface, and unit 2 is absent. The sand, which typically is a few tens of metres thick, probably was deposited at the western edge of the Cordilleran Ice Sheet just before glaciers achieved their maximum extent in this area during the last glaciation. West of Argonaut Plain, thin local provenance till veneers the sand in an area of low hills and swales oriented in a northerly to northwesterly direction. The linear topography was produced when glaciers flowing east and northeast from the Queen Charlotte Ranges were deflected northward by the margin of the Cordilleran Ice Sheet.

On eastern Graham Island, drift of the last glaciation is conformably overlain by fossiliferous sediments that have yielded a detailed record of Late Pleistocene and Holocene paleoenvironments and sea level change (Clague et al., 1982a, b; Mathewes and Clague, 1982; Warner et al., 1982; Clague, 1983; Warner, 1984). These sediments are thickest in coastal areas below about 15 m elevation and consist of fluvial, shallow marine, littoral, and slopewash deposits, and surface and buried peats (Fig. 1.31).

Chronology and correlation

Old glaciations

There is very little chronological control on events predating the penultimate glaciation. However, because the penultimate glaciation is no younger than Early Wisconsinan, older glaciations must be pre-Sangamonian in age. Estimates of soil age based on degree of soil formation and paleomagnetic data suggest that the youngest of five diamictons on Mokowan Butte in southwestern Alberta is Middle Pleistocene and that the oldest is late Pliocene in age (Karlstrom, 1987). If these diamictons are tills, as argued by Karlstrom (1981, 1987), there were several episodes of late Tertiary and early Quaternary glaciation in the southern Canadian Cordillera. $^{230}Th/^{234}U$ age determinations on speleothems in caves in South Nahanni River valley suggest that the earliest incursion of the Laurentide Ice Sheet into that area (First Canyon Glaciation) occurred before 320 ka (Ford, 1976; Harmon et al., 1977). A fission track date of 1.22 ± 0.49 Ma on Mosquito Gulch tephra near Dawson provides a minimum age for deposition of the Klondike gravels and hence the "old" glaciation(s) of the Southern Ogilvie Mountains (Naeser et al., 1982; Hughes et al., 1983). This tephra occurs on a terrace cut into the White Channel gravels which, at least in part, are correlative with the Klondike gravels. This date and similar fission-track dates on Fort Selkirk tephra indicate at least comparable antiquity for the Nansen Glaciation of central Yukon Territory and the "old" glaciation(s) of the Southern Ogilvie Mountains. This is supported by the fact that pre-Reid drift, the Flat Creek beds, and Klondike gravels all support Luvisolic soils with thick Bt horizons (Tarnocai et al., 1985).

Because of the lack of adequate chronological control, the number of old glaciations is uncertain. There is some evidence for several in the Waterton Lakes area (Fig. 1.30), but elsewhere only one has been recognized. The correlation of old glaciations from region to region also is uncertain; those shown in Figure 1.30 are tentative.

Penultimate glaciation

Most workers consider the penultimate glaciation in the fringe areas of the Canadian Cordillera to be Early Wisconsinan in age. This event, however, is not well dated, and it could just as well be Illinoian. Radiocarbon dates indicate only that it is older than Middle Wisconsinan.

$^{230}Th/^{234}U$ dates on cave speleothems indicate that the penultimate (Clausen-Flat River) glaciation in South Nahanni River valley occurred sometime after 190 ka (Ford, 1976; Harmon et al., 1977). This glaciation may correspond to the Reid advance of the Cordilleran Ice Sheet in southern Yukon Territory. In terms of degree of preservation of glacial morphology and soil development, drift of the "intermediate" glaciation of the Southern Ogilvie Mountains is similar to Reid deposits (Tarnocai et al., 1985). Organic silt on the floor of Hunker Creek valley near Dawson has yielded a radiocarbon date of >53 900 BP (GSC-527). This is a minimum age for incision of Klondike River and its tributaries preceding the "intermediate" glaciation. The silt may be reworked loess deposited during or after this glaciation; if so, the advance is older than 53.9 ka. The oldest radiocarbon date from sediments overlying drift of the "intermediate" glaciation in the Southern Ogilvie Mountains is 13 870 ± 180 BP (GSC-296).

Radiocarbon dates ranging from >52 000 BP to >33 000 BP (GSC-3151-2, GSC-3208; Clague et al., 1982a) have been obtained on sediments underlying drift of the last glaciation on eastern Graham Island. The penultimate glaciation on the Queen Charlotte Islands thus is older than 52 ka.

Figure 1.31. Late Quaternary sediments exposed in a sea cliff on eastern Graham Island, British Columbia: (1) till; (2) outwash gravel; (3) fluvial and ponded water sand and silt; (4) peat; (5) marine and littoral sand and mud; and (6) peat. Plant detritus at the base of unit 3 yielded a radiocarbon date of 16 000 ± 570 BP (GSC-3340); peat at the base of unit 4 dated 11 100 ± 90 BP (GSC-3337); and shells in unit 5 dated 9350 ± 80 BP (GSC-3120, Table 1.2). Courtesy of R.L. Long.

Last glaciation

The Foothills Erratics Train and associated mixed provenance till provide a basis for correlating deposits of the last glaciation in the eastern Rocky Mountains from Waterton Lakes to Athabasca River. Drift units associated with the Foothills Erratics Train and thus assignable to the last glaciation include Ernst, Bow Valley, Lamoral, Jackfish Creek, and Marlboro tills of Rocky Mountain provenance, and Buffalo Lake, Sylvan Lake, and Edson tills of Keewatin provenance.

Boydell (1978) proposed that Sylvan Lake Till and, by inference, the mixed provenance Athabasca Till and Foothills Erratics Train in the North Saskatchewan River area are Late Wisconsinan in age. He cited as evidence radiocarbon dates of 10 250 ± 165 BP and 9670 ± 140 BP (I-5675, I-5677) obtained on gastropods and bison bone from diamicton-capped lake sediments in an area of Sylvan Lake dead-ice moraine. Boydell attributed the burial of the lake sediments by diamicton to melting and collapse of underlying ice and to flowage of supraglacial drift into ponds.

In contrast, some workers favour an Early Wisconsinan or older age for the Foothills Erratics Train and associated deposits. Jackson (1980), for example, assigned an Early Wisconsinan age to the Erratics Train on the basis of radiocarbon dates of 18 400 ± 1090 BP and 18 300 ± 380 BP (GSC-2670, GSC-2668) from a peat bog in a meltwater channel incised into Bow Valley Till (time equivalent of Erratics Train Till). Recent work, however, has shown that these dates probably are several thousand years too old (MacDonald et al., 1987). This suggests that Bow Valley Till may be Late Wisconsinan in age.

The last glaciation in Athabasca River valley probably occurred sometime after 29 100 ± 560 BP (GSC-3792; Levson and Rutter, 1986), which is a date on wood in gravel below a single till near Jasper. A date of 22 700 ± 1000 BP (GaK-2336; Kigoshi et al., 1973) on bone from a cave in Crowsnest Pass is a maximum for the last glaciation in that area.

Radiocarbon dates on ungulate remains from a terraced postglacial gravel fill along Bow River range from 11 370 ± 170 BP to 10 200 ± 280 BP (GSC-613, GSC-3065; Wilson, 1981; Jackson et al., 1982). On the basis of these and older dates from Elk River valley to the south (Harrison, 1976; Ferguson and Osborn, 1981), it appears that the Canmore advance ended before 12 ka and perhaps before 13 ka.

Several radiocarbon dates help fix the time of the last glaciation in the Peace River region. A tooth from gravel thought by Mathews (1978) to have been deposited before the last incursion of glaciers into the Fort St. John area gave a date of 27 400 ± 580 BP (GSC-2034). This gravel seems to correlate with similar sediments occupying a former valley of Smoky River at Watino, Alberta, 200 km east of Fort St. John (Westgate et al., 1972; Mathews, 1978). At Watino, fluvial gravel is overlain by lacustrine sediments that have yielded radiocarbon dates as young as 27 400 ± 850 BP (I-4878). These dates suggest that the surface tills in the Fort St. John area, shown by Rutter (1976, 1977) and Mathews (1978, 1980) to have been deposited by coalescing Cordilleran and Laurentide ice, are Late Wisconsinan in age. Deglaciation of this area was in progress long before 10 ka. Freshwater shells from silts near Dawson Creek that were deposited either in a late stage of glacial Lake Peace or in local ponds that postdate the lake have yielded radiocarbon dates of 10 400 ± 170 BP and 9960 ± 170 BP (GSC-1654, GSC-1548). Other dates from bog and lake bottoms, proglacial lacustrine sediments, and younger fossiliferous silt in the Swan Hills area about 240 km southeast of Fort St. John indicate that much of north-central Alberta was deglaciated before 11.5 ka, with some areas ice-free perhaps as early as 13.5 ka (St-Onge, 1972).

The beginning of the "last" glaciation in the Southern Ogilvie Mountains is unknown. Glaciers in these mountains, however, probably were in retreat before 13 740 ± 190 BP (GSC-515).

Peat beneath outwash gravel and till in Yakoun River valley on the Queen Charlotte Islands has yielded a radiocarbon date of 27 500 ± 400 BP (GSC-3530). The last advance of glaciers in the Queen Charlotte Ranges thus occurred after 28 ka. Recent work on sediments exposed in coastal bluffs on northern and eastern Graham Island indicate that Late Wisconsinan glaciers achieved their maximum extent in this area after 21 ka. Glaciers retreated from their climax positions on eastern Graham Island before 15 ka (Clague et al., 1982a; Warner et al., 1982).

Unglaciated areas

O.L. Hughes, N.W. Rutter, J.V. Matthews, Jr., and J.J. Clague

Unglaciated central and northern Yukon Territory includes diverse terrains that range from flat lowlands through rolling plateaus to locally rugged mountains. This area is unified only by its lack of glacial deposits.

Cryoplanation (altiplanation) terraces and tors are common features in this region (Hughes et al., 1972). Cryoplanation terraces exhibit all the characteristics described by Demek (1969) for European, Asian, and other North American examples and likely are products of parallel scarp retreat in a periglacial environment. They are absent in areas of readily recognizable glacial features and therefore are pre-Reid in age. Some tors may have formed by cryoplanation of larger rock masses. However, the common occurrence of these features independently of cryoplanation terraces, their presence on western Klondike Plateau at a lower elevation than the terraces, and their occurrence in some glaciated areas where there are no terraces suggest a different origin for many tors. One possible formative mechanism is downwasting by solifluction; the tors would be left where the rock is more resistant to weathering and erosion.

Throughout much of the unglaciated Yukon, the only surficial deposits are weathered rock and colluvium on slopes, organic silt and peat in depressions, and fluvial sediments along major streams (Hughes, 1972). Thick and extensive unconsolidated sediments have been identified in only three lowland areas within Porcupine River basin: Old Crow, Bluefish, and Bell flats. The sediments in Old Crow and Bluefish flats, and probably those in Bell Flats, extend

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1989: Unglaciated areas (Quaternary stratigraphy and history, Canadian Cordillera); in Chapter 1 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

below bedrock thresholds, therefore these lowlands cannot be entirely erosional in origin; Tertiary or Pleistocene warping or faulting must have contributed to their formation (Hughes, 1972).

Quaternary deposits in Old Crow, Bluefish, and Bell flats are broadly similar in character and are mainly of fluvial, lacustrine, and glacial lacustrine origin. The flats once were drained by streams that flowed east through the Richardson Mountains at McDougall Pass. These streams deposited a variety of fluvial, lacustrine, and deltaic sediments in a nonglacial setting. At least once during the Pleistocene, however, the Laurentide Ice Sheet advanced against the Richardson Mountains and blocked east-flowing streams to form a vast lake that inundated low-lying areas to the west (Hughes, 1972; Thorson and Dixon, 1983). This lake discharged through a canyon at the Alaska-Yukon boundary (The Ramparts of the Porcupine). Waters escaping from the lake gradually eroded the Ramparts; by the time McDougall Pass last became ice-free, the outlet was lower than the pass and westward drainage was permanently established. Drainage of the glacial lake was followed by incision, local deposition of fluvial and shallow water lacustrine sediments, and widespread peat deposition.

Quaternary deposits in Old Crow, Bluefish, and Bell flats have been extensively studied, in part because bones thought to be human-modified have been found at some exposures (Morlan, 1980, 1986; Jopling et al., 1981). A composite section of the deposits in Old Crow Flats, shown in Figure 1.32, illustrates the complex history of sedimentation and erosion in this area during the Quaternary.[1] The lowest unit (1a) commonly extends to a few metres above river level and consists of massive clay, formerly thought to be glacial lacustrine in origin (Hughes, 1972; Matthews, 1975b; Jopling et al., 1981), but now suspected of having accumulated in a lake impounded by tectonic warping. Unit 1b consists of clay reworked from unit 1a. At one locality, it contains a layer of volcanic ash, termed the Little Timber tephra. Small channels at the contact between units 1a and 1b contain concentrations of wood, plant macrofossils, insects, mammal and fish bones, and molluscs (Clarke and Harington, 1978; Cumbaa et al., 1981; Bobrowsky, 1982; Morlan and Matthews, 1983). Unit 1b is overlain by 20-30 m of well bedded sand, silt, and minor gravel (unit 2) deposited intermittently by streams over an extended period of time. Large channel cut-and-fill structures occur in unit 2, as do ice-wedge pseudomorphs, cryoturbation structures, paleosols, peat layers, wood, molluscs, and bone. Unconformities demarcated by concentrations of bone, truncated cryoturbation structures, and peat are also present, especially in the upper part of the succession. The most conspicuous of these unconformities (= boundary between units 2a and 2b in Fig. 1.32, "Disconformity A." of Morlan, 1980) merges with a paleosol and peat that formed on a coniferous forest floor (Morlan and Matthews, 1983). Some of the bones found on this surface may be artifacts (see Morlan, 1986, for a critique). Two tephras occur in unit 2: Surprise Creek tephra near the base and Old Crow tephra (Westgate et al., 1983, 1985; Schweger and Matthews, 1985) in the upper part, 1-2 m below "Disconformity A". Unit 2 is sharply overlain

[1] The lowest sediments in this section, however, may be pre-Pleistocene in age. A section of similar sediments exposed along Porcupine River in adjacent Bluefish Flats is described by Schweger (1989).

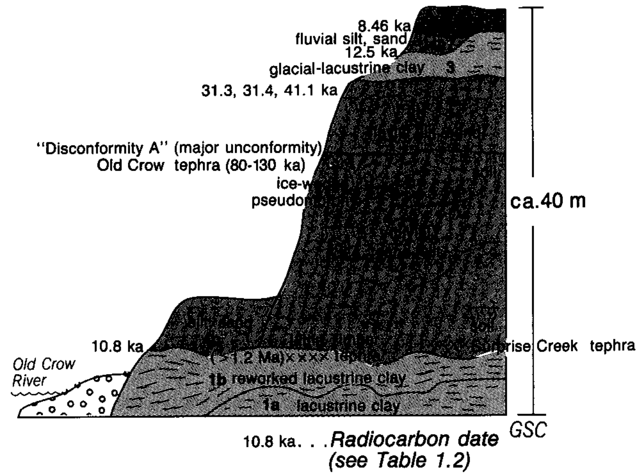

Figure 1.32. Composite stratigraphic section of Old Crow River bluffs, northern Yukon Territory (adapted from Morlan, 1980, Fig. 2.2).

by unfossiliferous glacial lacustrine clay (unit 3) up to 7 m thick. This clay, commonly with a capping of peat (unit 4b), forms the upper surface of much of Old Crow Flats. In places, however, the peat is underlain by fluvial sand and silt (unit 4a) which fill channels cut into the glacial lacustrine sediments.

Chronological control on the deposits in Old Crow Flats is provided by radiocarbon dates on wood, peat, and bone, fission-track and thermoluminescence dates on tephras, uranium-series dates on bone, and paleomagnetic stratigraphy. Part of unit 1b is reversely magnetized (J.A. Westgate, University of Toronto, personal communication, 1986) and may have been deposited during the Matuyama Reversal Epoch, in which case it is older than 790 ka but younger than 2.5 Ma. This is supported by fission-track dates of >1.2 Ma on Little Timber tephra and uranium-series dates of >460 ka on bone from the lower part of unit 2a. Estimates of the age of Old Crow tephra range from about 130 ka to 80 ka (Schweger and Matthews, 1985; Wintle and Westgate, 1986; Berger, 1987; J.A. Westgate, University of Toronto, personal communication, 1987). Finite radiocarbon dates ranging from 41 100 ± 1650 BP to 31 300 ± 640 BP (GSC-2754, GSC-1191; Table 1.2) have been obtained on plant material near the top of unit 2b, 4-5 m above Old Crow tephra. Organic detritus at about the same stratigraphic level in an exposure in Bluefish Flats yielded a radiocarbon date of 20 800 ± 200 BP (GSC-3946); mammoth tusk from this level in another exposure yielded an AMS (accelerator) radiocarbon date of 25 170 ± 630 BP (NMC-1232). Finally, bone from a channel fill incised into unit 3 has been dated at 12 460 ± 440 BP (I-3574).

Unit 2 thus spans a long period of time, perhaps much of the Middle and Late Pleistocene. During this period, intervals of fluvial deposition alternated with intervals of soil formation and permafrost degradation (Schweger and Matthews, 1985).

Glacial lacustrine sediments of unit 3 were deposited during the Late Wisconsinan Hungry Creek Glaciation when the Laurentide Ice Sheet blocked a major east-flowing river at McDougall Pass. At the climax of this glaciation, some time after 36 900±300 BP (GSC-2422) but before 16 000±420 BP (GSC-2690), Laurentide ice filled Bonnet Plume Depression and diverted Peel River northward via Eagle River into the Porcupine drainage (Hughes et al., 1981; Vincent, 1989). The ice sheet also backed up against the eastern Mackenzie and Richardson mountains, reaching an elevation of about 1500 m near Keele River and 885 m just south of Yukon Coastal Lowland (Hughes, 1987). McDougall Pass (314 m) thus was blocked and a lake impounded to the west. A Late Wisconsinan age for the Hungry Creek Glaciation, however, conflicts with evidence from the Interior Plains to the north and east which suggests that part of this area was ice-free during this period (Hughes et al., 1981; Rampton, 1982; see also Vincent, 1989).

The lake in which unit 3 was deposited had drained by 12 460±440 BP (I-3574). By 10 850±320 BP (I-4224), Old Crow River had cut 40 m through the Pleistocene fill to near its present level and was integrated with the Alaskan Porcupine River system.

Thorson and Dixon (1983) studied the alluvial history of Porcupine River in Alaska and suggested that there were at least three glacial lacustrine inundations of Old Crow Flats and Bluefish Flats during Wisconsinan time, two of which occurred after about 31 ka. In contrast, Morlan (1980) and Hughes et al. (1981) suggested, on the basis of stratigraphic and geochronological studies in Old Crow Flats and Bonnet Plume Depression, that there was only one lake episode during the Late Wisconsinan, represented by unit 3. However, because of the high ice content of unit 3, there are no sections where good exposures of the entire glacial lacustrine sequence may be seen; it thus may represent two or more ponding events.

PALEOECOLOGY AND PALEOCLIMATOLOGY

J.J. Clague and G.M. MacDonald

Paleoecological and paleoclimatic information for the Canadian Cordillera has come from studies of fossil pollen, plant macrofossils, paleosols, speleothems, and vertebrate and invertebrate faunal remains (Fig. 1.33). Most of this information is from the southern part of the region, but even here an understanding of Quaternary paleoecology remains fragmentary. This section provides a brief synopsis of Quaternary climates and vegetation in the Canadian Cordillera. Additional information is given in Chapter 7 of this volume.

Coastal British Columbia

Studies of plant microfossils from Middle Wisconsinan sediments in south-coastal British Columbia and adjacent Washington indicate that the climate of the Olympia Nonglacial Interval was variable, but generally cooler than the present-day climate (Hansen and Easterbrook, 1974; C.J. Heusser, 1977, 1983; Clague, 1978; Alley, 1979; Heusser et al., 1980; Hicock, 1980; Hebda et al., 1983). Some beds examined by these authors are dominated by trees that presently grow at low elevations; others contain assemblages indicative of subalpine forests or grass-herb meadows, implying a cooler climate. Slightly cooler and wetter conditions also prevailed on the Queen Charlotte Islands during part of the Olympia Nonglacial Interval (Warner et al., 1984).

Palynological investigations of early Fraser Glaciation sediments in the Strait of Georgia region indicate a gradual deterioration of climate after about 29 ka (Alley, 1979; Mathewes, 1979). By 21 ka, subalpine parkland and possibly alpine plant communities were dominant on the lowlands of eastern Vancouver Island, and large vertebrates, including *Mammuthus imperator* (mammoth) and *Symbos cavifrons* (muskox), were relatively common in the region (Harington, 1975; Hicock et al., 1982b).

Diverse floras existed in some British Columbia coastal areas near the climax of the Fraser Glaciation. About 18 ka, *Abies lasiocarpa-Picea* cf. *engelmannii* (subalpine fir-Engelmann spruce) forest and parkland grew under cold humid continental conditions at sea level near Vancouver (Hicock et al., 1982a). Mean annual temperature was depressed about 8°C, and treeline was 1200-1500 m lower than today. About 15 ka, the lowlands of the Queen Charlotte Islands supported a varied nonarboreal terrestrial and aquatic flora characteristic of alpine herbaceous and shrub meadows (Warner et al., 1982; Warner, 1984).

Vegetation became reestablished in coastal British Columbia soon after the area was deglaciated at the end of the Fraser Glaciation (Mathewes, 1973, 1985; Hebda, 1983; Warner, 1984). The earliest fossil pollen assemblages, dating from about 13-13.5 ka, are dominated by herbs and shrubs adapted to a cold, relatively dry climate. About 13 ka, forest became established and began to change in composition as plants migrated into the region from extraglacial areas and as climate ameliorated. There was rapid warming and drying between 10.5 ka and 10 ka in the Strait of Georgia region, marked by the appearance of *Pseudotsuga menziesii* (Douglas-fir) and *Tsuga heterophylla* (western hemlock). There was also climatic amelioration on the Queen Charlotte Islands at this time.

Pseudotsuga and *Alnus* (alder) dominated south-coastal forests during the early Holocene (Mathewes, 1985), and it is likely that the climate at that time was similar to or warmer than the present. Maximum dry and warm conditions during this "early Holocene xerothermic interval" (Mathewes and Heusser, 1981) occurred between 10 ka and 7 ka. The xerothemic (Hypsithermal) interval was followed by generally wetter and cooler conditions. By 4-5 ka, modern forests dominated by *Tsuga heterophylla* and *Thuja plicata* (western red cedar) had become established on the south coast (Mathewes, 1985). Clear evidence for the present cool wet climate on the Queen Charlotte Islands is not apparent in the fossil record until about 3 ka when *Thuja plicata* became dominant in lowland forests. Since then, there has been little change in vegetation along the British Columbia coast (Warner, 1984).

Interior British Columbia

Fossil animal and plant remains from Bessette Sediments suggest that the climate of south-central British Columbia during at least part of the Olympia Nonglacial Interval was similar to the present-day climate of the region (Fulton,

Clague, J.J. and MacDonald, G.M.
1989: Paleoecology and paleoclimatology (Canadian Cordillera); in Chapter 1 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

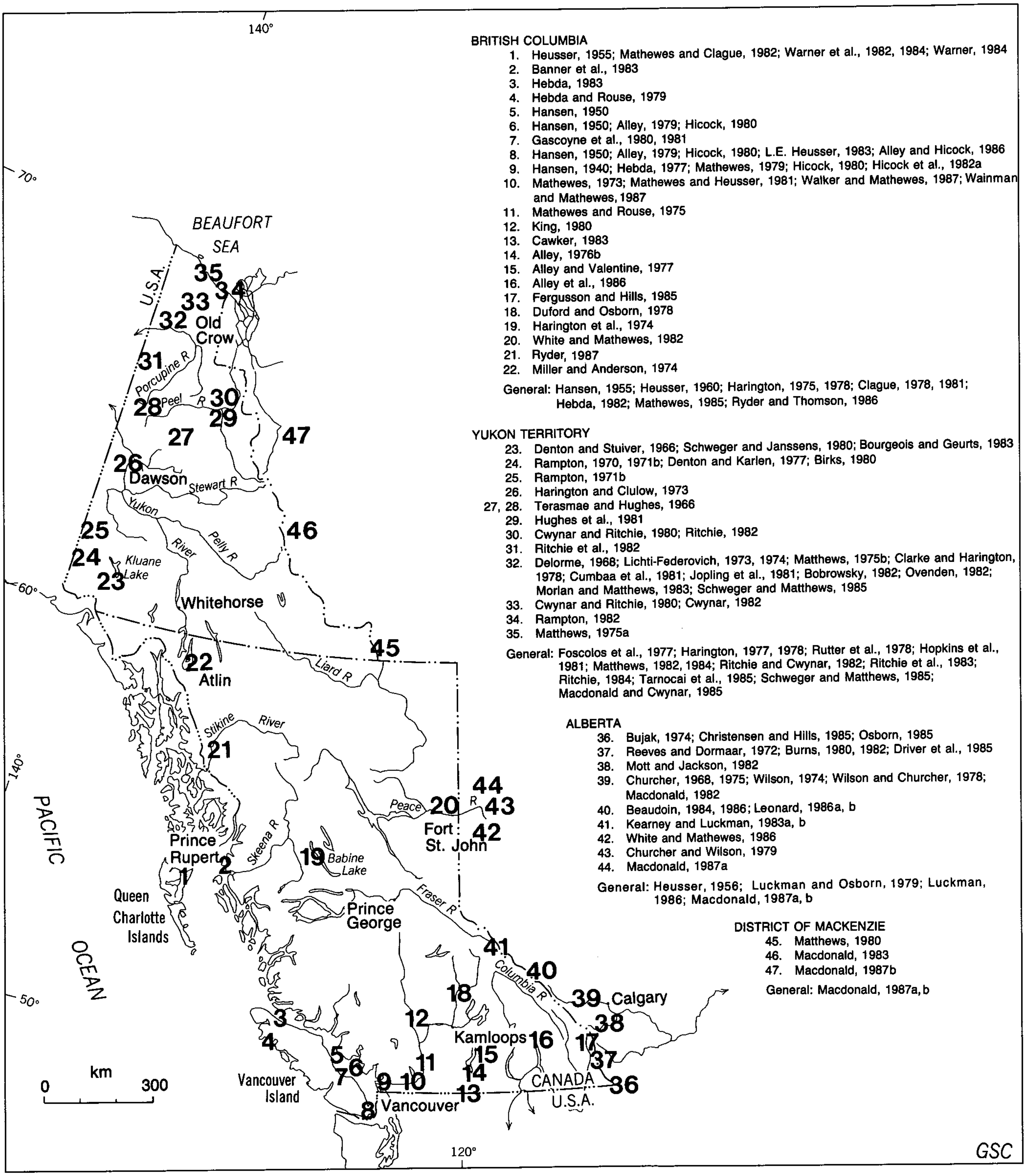

Figure 1.33. Selected Quaternary paleoecological and paleoclimatic studies in the Canadian Cordillera.

1975a; Alley et al., 1986). Major climatic deterioration accompanying the onset of the Fraser Glaciation began about 25 ka, but a full glacial climate was not attained until 19 ka (Alley and Valentine, 1977).

Early deglacial plant assemblages in the southern interior, as on the coast, were dominated by nonarboreal taxa, indicating an open landscape and generally cold and possibly dry climatic conditions (Hebda, 1982). A cold, relatively dry climate also prevailed in the Atlin area in northwestern British Columbia during deglaciation (Miller and Anderson, 1974); there, shrub tundra apparently grew at sites now covered by spruce forest.

Geological and botanical evidence pertaining to early and middle Holocene climatic conditions in the British Columbia interior is complex and in some cases contradictory. Pollen evidence generally supports the existence of a xerothermic interval during the Holocene, although the proposed dating varies considerably from locality to locality (Clague, 1981; Mathewes, 1985). For example, Hansen (1955) identified a warm dry period in south-central British Columbia between 7.5 ka and 3.5 ka, whereas Alley (1976b), working in the same region, concluded that the Hypsithermal interval extended from about 8.4 ka to 6.6 ka. King (1980) presented evidence for low lake levels and, by inference, xerothermic conditions near Lillooet from 8 ka until well after 7 ka. Miller and Anderson (1974) proposed that the period of maximum Holocene warmth at Atlin lasted from about 8 ka until 2.5 ka. Precipitation at Atlin during this period apparently equaled or exceeded present levels, in contrast to the situation in the southern part of the province.

Fossil tree remains found above present treeline in the mountains of southern British Columbia provide additional evidence for a Holocene xerothermic interval. Some of these remains are of early and middle Holocene age and have yielded radiocarbon dates ranging from 9070 ± 130 BP to 5260 ± 200 BP (GSC-3173,Y-140bis; Clague, 1980; Ryder and Thomson, 1986).

The xerothermic interval in southern British Columbia was followed by a generally cooler wetter period marked by sharp, but relatively minor fluctuations in climate. Hebda (1982) concluded that essentially modern climatic conditions were attained in the southern interior about 4.5 ka after a period of cooling and increased precipitation that began some time between 8 ka and 6 ka. Results of paleoecological studies by Alley (1976b), Hazell (1979), and King (1980) at various sites in southern British Columbia generally support this conclusion.

Yukon Territory

The fact that a large part of Yukon Territory was ice-free or glaciated only infrequently during the Quaternary has made this an important region for paleoecological research. Evidence for early and middle Quaternary environmental conditions has been obtained from sediment sections in river valleys, on plateaus, and along the coast (see Schweger, 1989, for pollen diagram of Twelvemile Bluff, Porcupine River). In contrast, evidence for late Quaternary environmental conditions comes mainly from sediment cores taken from extant lakes and bogs.

Fossil pollen and plant macrofossils from the Porcupine River area indicate that a forest dominated by *Picea* (spruce), *Pinus* (pine), and *Betula* (birch), but also containing *Larix* (tamarack) and *Corylus* (hazel), was present in northern Yukon Territory during the Pliocene and perhaps the Early Pleistocene (Lichti-Federovich, 1974; Pearce et al., 1982; Ritchie, 1984). The climate probably was warmer and moister than the present. The later part of the Quaternary in northern Yukon Territory appears to have been characterized by an alternation of tundra, represented in the fossil record by Cyperaceae (sedges), Gramineae (grasses), and various herbs, and forest, dominated by *Picea* (Lichti-Federovich, 1973, 1974; Matthews, 1975b; Schweger and Janssens, 1980; Hughes et al., 1981).

Additional climatic information comes from studies of paleosols developed on drift of the Cordilleran Ice Sheet in central Yukon Territory. Well developed Luvisolic soils on pre-Reid drift indicate a temperate and humid climate prior to the Klaza and/or Nansen glaciations (Foscolos et al., 1977; Rutter et al., 1978; Tarnocai et al., 1985). Moderately developed Luvisols and Brunisols on Reid Drift are evidence of a cool subhumid climate during the Boutellier Nonglacial Interval. Boreal forest conditions prevailed during part of the Boutellier interval, but by about 37.7 ka conditions were colder than at present (Denton and Stuiver, 1967; Schweger and Janssens, 1980). The paleosols contain periglacial and cryogenic features such as sand wedges, sand involutions, and oriented stones, which developed under cold dry conditions during the Reid and McConnell glaciations.

Abundant Pleistocene vertebrate remains have been found at several sites in unglaciated northern Yukon Territory, for example near Dawson and Old Crow (Harington, 1970, 1978). More than 60 species of vertebrates have been recovered, including *Equus* (horse), *Bison* (bison), and *Mammuthus* (mammoth). Unfortunately, most of the collections that have been described are reworked and lack stratigraphic context (Morlan, 1980), thus their paleoecological usefulness is limited.

Radiocarbon-dated cores from lakes and bogs have provided a wealth of paleoenvironmental information for the Late Pleistocene and Holocene of Yukon Territory (e.g., Terasmae and Hughes, 1966; Rampton, 1971b; Birks, 1980; Cwynar, 1982; Ovenden, 1982; Bourgeois and Geurts, 1983; MacDonald and Cwynar, 1985). Late Quaternary pollen, plant macrofossils, insect fossils, and vertebrate remains also have been obtained from cave deposits (Ritchie et al., 1982) and from sediment sections along streams and the Yukon coast (Matthews, 1975a; Cumbaa et al., 1981; Rampton, 1982; Morlan and Matthews, 1983). These studies collectively provide a detailed record of environmental change that spans the last 30 ka.

During Late Wisconsinan time, Yukon Territory was occupied by tundra, characterized in the pollen and plant macrofossil record by Cyperaceae, Gramineae, *Artemisia* (sage), and various herbs. This vegetation was widely established in the Yukon by 30 ka (Rampton, 1971b; Cwynar, 1982). It has been suggested that the vegetation from 30 ka to 14 ka resembled a sparse, fell-field tundra typical of the modern vegetation of the Arctic islands (Cwynar and Ritchie, 1980; Cwynar, 1982; Ritchie and Cwynar, 1982; Ritchie, 1984). This interpretation runs counter to the view that a highly productive "steppe tundra" existed in the Yukon during this period (Guthrie, 1968, 1982; Matthews, 1976, 1982). Although there is a continuing dispute over this issue, it generally is agreed that a reasonably large mammal population was supported by the available plant cover and that the climate was very cold and arid (Guthrie, 1982; Ritchie, 1984).

The climate in northern Yukon Territory ameliorated slightly between about 23 ka and 18.5 ka. The evidence for this is an increase in *Picea, Alnus*, and *Betula* pollen at some sites and the presence of fossil beetle and ostracode species typical of modern shrub tundra and the forest-tundra boundary (Cwynar, 1982; Rampton, 1982).

This interval was followed by a herb and willow tundra phase which ended about 12 ka with the establishment of birch tundra over most of northern Yukon Territory (Matthews, 1975a; Ovenden, 1982; Ritchie, 1982; Ritchie et al., 1982). The change from herb-willow to birch tundra generally has been attributed to climatic amelioration, although the diachronous nature of birch expansion in northwestern Canada remains an enigma (Ritchie, 1984). *Populus* (poplar), *Myrica* (bog myrtle), and *Typha* (cat-tail) grew north of their present limits in this region at 10-11 ka (Cwynar, 1982; Rampton, 1982; Ritchie et al., 1983), and the climate was warmer than it is today.

Picea pollen increased significantly in the northern Yukon as early as 10 ka, possibly reflecting the invasion of spruce from the east (Hopkins et al., 1981) and south (Ritchie and MacDonald, 1986; MacDonald, 1987b). The dispersal of *Picea* and, later, *Alnus* apparently was diachronous: *Picea* did not reach some sites in south and central Yukon Territory until 9 ka or later, and *Alnus* did not achieve its present distribution and density in this region until after 6 ka (Rampton, 1971b; Hopkins et al., 1981). Since the establishment of *Picea* and *Alnus,* the vegetation appears to have changed little, except at climatically stressed sites near treeline and in parts of southern Yukon Territory where *Pinus* only recently (<0.5 ka) has reached its present northern limit (MacDonald and Cwynar, 1985).

There exists a rich record of middle and late Holocene climatic change in the St. Elias Mountains. Stumps above present treeline in this region have been dated at 5.35 ka, 5.25 ka, 3.0-3.6 ka, and 1.23-2.1 ka (Rampton, 1971b; Denton and Karlén, 1977); climate probably was warmer at these times than it is today. Times of cooler climate (2.9-2.1 ka, 1.23-1.05 ka, and 0.5-0.1 ka) are recorded by glacier advances beyond present limits. These and other data (e.g., Bourgeois and Geurts, 1983) indicate that there have been minor adjustments of vegetation at sensitive sites in Yukon Territory during the last half of the Holocene.

Eastern Cordillera

Pollen records from the Rocky and Richardson mountains indicate that the Late Wisconsinan climate of the eastern Canadian Cordillera was cold and probably arid. A sparse herbaceous tundra, dominated by Cyperaceae, Gramineae, *Salix, Artemisia,* and various herbs, existed in the foothills of the Rocky Mountains early during deglaciation (Mott and Jackson, 1982; MacDonald, 1982). The vegetation in this area changed to shrub tundra near the end of the Pleistocene. Herb tundra existed in the Richardson Mountains between 15 ka and 12 ka (Ritchie, 1982). Shortly after 12 ka, it was replaced by shrub tundra dominated by *Betula glandulosa* (resin birch).

A number of Holocene paleoecological records are available for the southern Rocky Mountains and adjacent Interior Plains, Peace River region, Mackenzie Mountains, and Richardson Mountains:

Southern Rocky Mountains and adjacent Interior Plains. Coniferous forests dominated by *Picea* and *Pinus* invaded lower mountain valleys in southwestern Alberta and southeastern British Columbia between about 12 ka and 10 ka, replacing nonarboreal vegetation and, in places, pioneering stands of *Populus* (poplar-aspen) (Harrison, 1976; Schweger et al., 1981; MacDonald, 1982; Mott and Jackson, 1982; Fergusson and Hills, 1985). *Pinus contorta* (lodgepole pine) was growing near present treeline by 9.7 ka (Kearney and Luckman, 1983b; Beaudoin, 1984). Since then, significant changes in vegetation in the southern Rocky Mountains probably have been restricted to sites at the upper and lower limits of forests.

Studies of pollen, plant macrofossils, and paleosols have shown that there were shifts in treeline in the southern Rocky Mountains during the Holocene, presumably in response to changes in temperature and precipitation (Reeves and Dormaar, 1972; Reeves, 1975; Kearney and Luckman, 1983a, b; Beaudoin, 1984, 1986). During most of early and middle Holocene time, treeline was higher than at present, and the climate thus warmer. A cool climate, similar to the present, was achieved after 5 ka. There were additional complex, but minor, timberline fluctuations during the late Holocene associated with neoglacial climatic oscillations.

Holocene malacological data from the southern Rocky Mountains and Interior Plains support these conclusions (Harris and Pip, 1973; MacDonald, 1982). They show that the regional molluscan fauna contained a high complement of south-ranging species from approximately 9 ka to 6 ka; after 6 ka, north-ranging taxa increased in abundance.

Fossil vertebrates in 10-11 ka terrace gravels flanking Bow River near the front of the Rocky Mountains represent a diverse ungulate fauna, including *Camelops* cf. *hesternus* (camel), *Bison bison antiquus* (bison), *Equus conversidens* (Mexican half-ass), *Mammuthus, Cervus canadensis* (wapiti), *Rangifer tarandus* (caribou), and *Ovis canadensis* (bighorn sheep) (Churcher, 1968, 1975; Wilson, 1974; Wilson and Churcher, 1978). This fauna may have lived in a parkland vegetational setting similar to that of present-day southern and central Alberta (Wilson and Churcher, 1978), although paleobotanical information to confirm or negate this hypothesis is lacking.

In summary, climate had ameliorated sufficiently by 10 ka for the establishment of regional vegetation in the southern Rocky Mountains similar to that of the present. The general warming trend seems to have continued to the middle Holocene, although there may have been brief periods of climatic deterioration. Warmer conditions terminated about 4-5 ka.

Peace River region. Holocene paleobotanical information has been obtained from sediment cores taken from lakes in the Clear Hills (MacDonald, 1987a), Saddle Hills (White and Mathewes, 1986), and near Fort St. John (White and Mathewes, 1982). The Clear Hills pollen record shows that the vegetation was dominated by *Artemisia,* Cyperaceae, Gramineae, *Salix,* and *Populus* from 10.7 ka to 9.9 ka. Spruce forest expanded into the Clear Hills between 9.9 ka and 9 ka, and *Pinus* appeared about 7.5 ka. *Pinus* and *Picea* may have arrived in the Saddle Hills as early as 11 ka. The Fort St. John pollen record indicates that there have been no major changes in the boreal forest of this region during the last 7.25 ka.

Early postglacial vertebrate fossils from the eastern Peace River region include *Mammuthus primigenius, Equus,* three species of *Bison,* a camelid, *Cervus canadensis,* and *Ovis,* a fauna indicative of an environmental setting similar to the modern parkland of southern and central Alberta (Churcher and Wilson, 1979).

Neither the pollen nor faunal evidence from this part of the eastern Cordillera provides a clear climatic signal. The transition from parkland to coniferous forest during the early Holocene may reflect differing rates of migration of the dominant plant species rather than climatic change. However, widespread development of small lakes and peatlands between 7 ka and 5 ka may reflect moister conditions during the middle and late Holocene (White and Mathewes, 1986; MacDonald, 1987a).

Mackenzie and Richardson mountains. Two Holocene pollen and plant macrofossil studies have been reported for the Mackenzie Mountains, one from the Boreal Forest zone in the southeast (Matthews, 1980), and another from a small bog above treeline near the crest of the range (MacDonald, 1983). In the former area, a spruce-dominated forest was established by 9.6 ka, and *Alnus* and *Pinus* appeared at 8.7 ka and 6.7 ka, respectively. In the latter area, shrub birch tundra prevailed from at least 8.6 ka until 7.7 ka, after which *Picea* became established. Treeline at this site was higher than at present, and the climate thus warmer, between about 7.7 ka and 5 ka.

Farther north in the Richardson Mountains, birch shrub tundra existed from 11 ka to 9 ka (Ritchie, 1982). At about 9 ka, *Picea* expanded into the region and within 1 ka was well established. *Alnus* appeared at about 7.5 ka and achieved its present abundance by 6.5 ka. *Betula neoalaskanis* (Alaskan paper birch) apparently attained its modern extent and abundance by 6 ka. While all of these changes may be the result of migrational lags, it is worth noting that pollen, plant macrofossil, and invertebrate fossil data from sites on the Mackenzie River delta and Yukon Coastal Lowland indicate that the climate in areas bordering the Richardson Mountains was significantly warmer than at present from about 11 ka until 5-6 ka (Ritchie et al., 1983).

HOLOCENE GLACIER FLUCTUATIONS

J.M. Ryder

Fluctuations of glaciers in the Canadian Cordillera during the Holocene (Table 1.3) have been recognized from detailed mapping of moraines and from stratigraphic investigations. These fluctuations have been dated by dendrochronology, lichenometry, radiocarbon ages, and stratigraphic relationships of drift to various tephras.

Moraines of latest Pleistocene or early Holocene age have been described from Shuswap Highland in south-central British Columbia (Alley, 1976a; Duford and Osborn, 1978), the southern Rocky Mountains (Luckman and Osborn, 1979; Osborn, 1985), and the southern Coast Mountains (Ricker, 1983). In Shuswap Highland, the Harper Creek and Dunn Peak moraines occur in cirques beyond late Holocene moraines and are older than 7985 ± 125 BP and 7390 ± 250 BP, respectively (I-9162, GX-4039; Table 1.2). In the Rocky Mountains, moraines and rock glaciers of the Crowfoot advance generally lie within 1 km of present ice margins and are older than 6.8 ka. Both the Dunn Peak and Crowfoot advances occurred after substantial or complete retreat of Late Wisconsinan glaciers and thus are younger than 12 ka. Davis and Osborn (1987) argued that these and comparable advances elsewhere in the North American Cordillera are probably latest Pleistocene rather than early Holocene in age.

Ryder, J.M.
1989: Holocene glacier fluctuations (Canadian Cordillera); in Chapter 1 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

Glaciers expanded during middle and late Holocene time after the Hypsithermal warm interval. This period of glacier expansion has been referred to as "Neoglaciation" by Porter and Denton (1967). Early neoglacial advances are recognized in the southern Coast Mountains ("Garibaldi phase" of Ryder and Thomson, 1986) and in nearby Washington (Miller, 1969; Beget, 1984). The Garibaldi phase is poorly dated, but probably occurred between 6 ka and 5 ka (Ryder and Thomson, 1986). An early neoglacial advance on Dome Peak in northern Washington took place shortly after 5 ka (Miller, 1969).

Glaciers grew in many parts of the Canadian Cordillera and elsewhere in western North America between about 3.3 ka and 1.9 ka (Porter and Denton, 1967; Denton and Karlén, 1973, 1977; Burke and Birkeland, 1983; Ryder and Thomson, 1986). Advances during this period have been documented for three glaciers in the Coast Mountains (Ryder and Thomson, 1986): (1) Tiedemann Glacier achieved its maximum Holocene extent at 2.3 ka; (2) an advance of Gilbert Glacier began before 2.2 ka and culminated shortly after 2.0 ka; (3) Frank Mackie Glacier had about the same extent at 2.7 ka as it does today. A major neoglacial advance of Bugaboo Glacier in the Columbia Mountains began before 2.4 ka (Osborn, 1986). The Battle Mountain advance in Shuswap Highland culminated between 3.4 ka and 2.4 ka (Alley, 1976a). During this same general period, glaciers also advanced in the St. Elias Mountains, with a culmination about 2.6-2.8 ka (Denton and Karlén, 1977). Glaciers in the southern Rocky Mountains began to advance from retracted middle Holocene positions before 3 ka; by 2.7 ka, they may have been as extensive as at present (Leonard, 1986b).

A short-lived advance occurred locally in the St. Elias Mountains about 1.23-1.05 ka (Denton and Karlén, 1977). This advance has not been documented elsewhere in the Canadian Cordillera and thus likely was a minor event. However, Alley (1976b) proposed on the basis of palynological evidence that the present relatively moist phase at Kelowna began about this time.

In many parts of the Cordillera, glaciers attained their maximum Holocene extent during the last several centuries, destroying or burying evidence of earlier advances. This interval of glacier growth has been referred to as the "Little Ice Age" and is recognized in mountain ranges throughout the world (for a summary and references, see Porter and Denton, 1967; Denton and Karlén, 1973). Little Ice Age advances produced the fresh, sparsely vegetated morainal ridges adjacent to most existing glaciers (Fig. 1.34). Dates from in situ stumps and roots exposed by recent glacier recession in the southern Coast Mountains show that the Little Ice Age began there before 0.9 ka and continued without notable interruption until 0.1-0.2 ka (Mathews, 1951a;

Table 1.3. Holocene glacier fluctuations in the Canadian Cordillera

	Coast Mountains	Shuswap Highland		Rocky Mountains	St. Elias Mountains
	Ryder and Thomson, 1986 Ryder, 1987	Alley, 1976a	Duford and Osborn, 1978	Luckman and Osborn, 1979 Luckman, 1986	Denton and Karlén, 1977
	Late neoglacial advance	Mammoth Creek advance	Raft Mountain advance	Cavell advance	Late Neoglacial advances
Late Neoglacial phase	maximum - 0.1-0.3 ka began - >0.9 ka	late phase - <0.4 ka early phase - 0.8-1.0 ka	"last few centuries"	maximum - <0.1-0.4 ka began - >0.6 ka	late phase - max. - 0.1-0.5 ka late phase - began - >0.5 ka early phase - 1.05-1.23 ka
	Tiedemann advance	Battle Mountain advance			Middle Neoglacial advance
Middle Neoglacial phase	ended - <1.9 ka maximum - 1.9-2.3 ka[1] began - $\geq$3.3 ka	>2.4 ka <3.4 ka			ended - 2.1 ka maximum - 2.6-2.8 ka began - 2.9 ka
Early Neoglacial phase	Garibaldi phase ca. 5-6 ka				
Early Holocene-Late Pleistocene phase		Dunn Peak advance[2]	Dunn Peak advance >7.4 ka <11 ka	Crowfoot advance >6.8 ka <12 ka	

1 2.3 ka for Tiedemann Glacier, 1.9 ka for Gilbert Glacier.
2 Alley (1976a) assigned an age of 4-5 ka to the Dunn Peak Advance, but Duford and Osborn (1978, 1980) later showed that this advance is older than 7.4 ka.

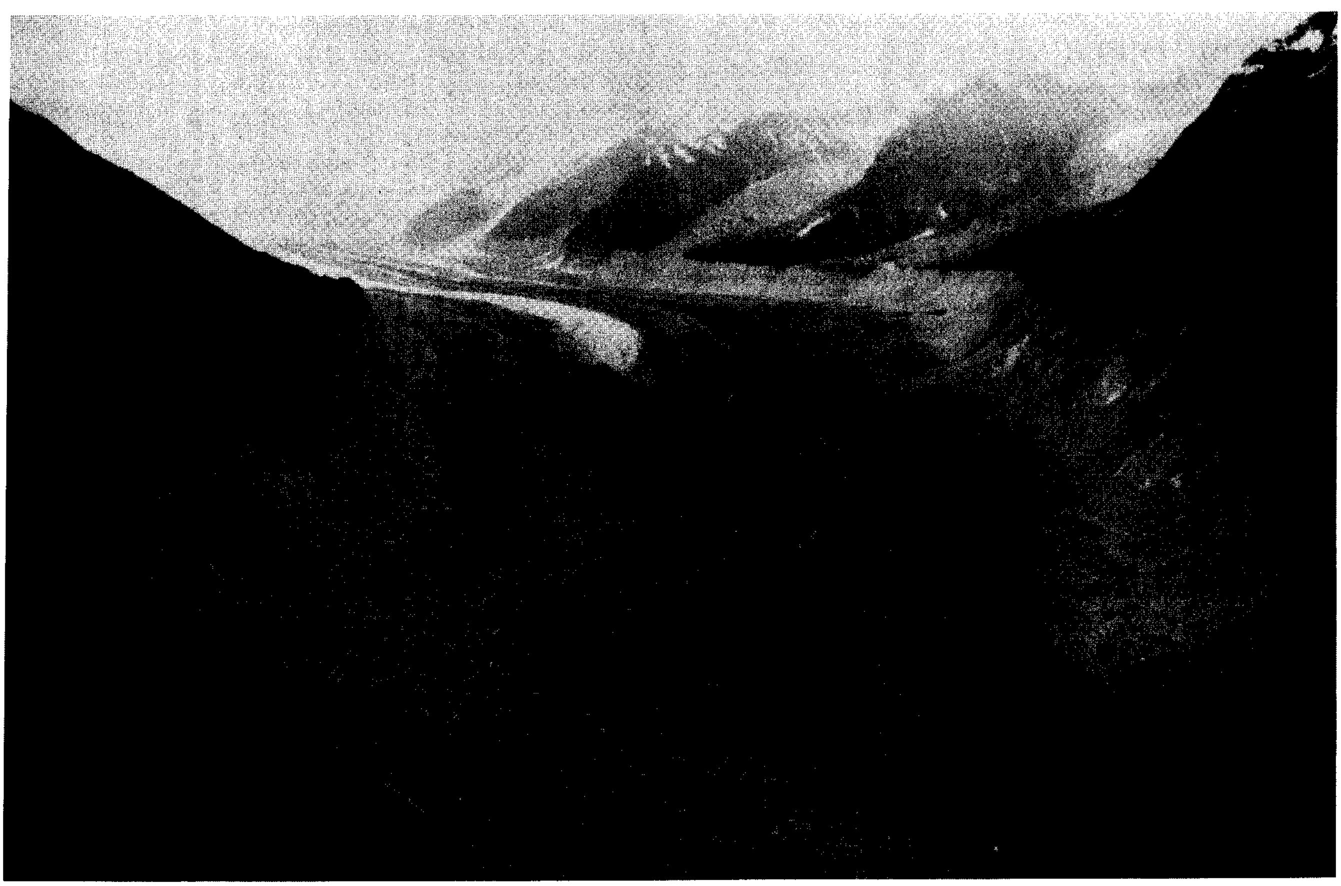

Figure 1.34. "Little Ice Age" moraines bordering Tiedemann Glacier, southern Coast Mountains, British Columbia.

Ryder and Thomson, 1986). Dates from similar materials in the northern Coast Mountains indicate that glaciers began to advance there before 0.6 ka (Ryder, 1987). The earliest Little Ice Age advances in the Rocky Mountains may have occurred before 0.6 ka (Luckman and Osborn, 1979; Leonard, 1986a; Luckman, 1986), and in the St. Elias Mountains before 0.5 ka (Denton and Stuiver, 1966; Rampton, 1970; Denton and Karlén, 1977).

Dates obtained by dendrochronology and lichenometry indicate that most glaciers in the Canadian Cordillera began to recede from their maximum Little Ice Age positions at various times during the Eighteenth, Nineteenth, and early Twentieth centuries (Mathews, 1951a; Denton and Karlén, 1977; Duford and Osborn, 1978; Luckman and Osborn, 1979; Osborn, 1985; Luckman, 1986; Ryder, 1987). Since then, glaciers generally have undergone sporadic retreat interrupted by stillstands and readvances. Today, almost all glaciers are much less extensive than they were at the climax of the Little Ice Age.

In the Coast and Rocky mountains, most glaciers reached their greatest Holocene extent during the Little Ice Age. In contrast, early Holocene advance(s) in Shuswap Highland were more extensive than subsequent advances. In the St. Elias Mountains, some glaciers climaxed about 2.6-2.8 ka, whereas others attained maximum Holocene positions during the Little Ice Age.

ECONOMIC IMPLICATIONS OF QUATERNARY GEOLOGY

J.J. Clague

Quaternary deposits and geological processes are fundamental to the very character of the Canadian Cordillera and have played a major role in economic development of the region. The significance of these deposits and processes to the economy of the region is briefly summarized in this section, with emphasis on urban and industrial development, resource exploitation, and natural hazards.

Foundations and excavations

Most towns and cities in the Canadian Cordillera are located in areas of low relief, low elevation, and thick Quaternary sediments. Excavation in such areas is relatively easy and inexpensive, facilitating construction of roads, sewers, and water lines. On the other hand, special care must be taken to provide protection against landslides and to ensure the stability of structures built on permafrost and loose, water-saturated or highly compressible materials. For example, compressible fine grained sediments of the Fraser River delta were preloaded at considerable expense prior to the expansion of the Vancouver International Airport in the 1960s (Meyerhof and Sebastyan, 1970). As another example, governments have restricted construction on unstable glacial lacustrine silt benches in Kamloops, Penticton, and some other interior communities to prevent or minimize property damage and loss of life due to landsliding.

Clague, J.J.
1989: Economic implications of Quaternary geology (Canadian Cordillera); in Chapter 1 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

Agriculture

Land suitable for farming in the Canadian Cordillera is restricted to valley bottoms, coastal and intermontane lowlands, and some plateaus, and represents less than 2% of the total land area of region (Barker, 1977). The most fertile and productive lands are those underlain by silt and clay of glacial lacustrine and glacial marine origin (old lake basins and coastal lowlands formerly submerged by the sea), and by sand and silt of fluvial origin (river terraces and present-day floodplains). Till is less suited for cultivation because it is gravelly and commonly is associated with irregular terrain; nevertheless, large areas of rolling ground moraine in the British Columbia interior are used as range land for livestock and also yield forage crops.

Forestry

About 60% of the Canadian Cordillera is forested, but commercial timber covers only one-third of British Columbia and one-tenth of Yukon Territory (Barker, 1977). Forest productivity is highest in the wet coastal belt of British Columbia, especially in lowland areas. In relatively dry areas of the interior, trees tend to be smaller and productivity lower. Productivity also is low where the water table is near the surface (e.g., bogs), in areas subject to sporadic flooding (e.g., floodplains), and throughout most of the northern Cordillera. Rugged mountains support little commercial timber because climatic conditions there are generally severe and because steep bedrock slopes mantled by thin bouldery drift and colluvium provide a poor substrate for trees and are subject to avalanching. Geological materials exert less of an influence on forest productivity outside major mountain ranges, although fine sediments with high moisture-holding capacities (e.g., clay-rich till, glacial lacustrine silt and clay) are more productive than coarse, well drained sediments (e.g., outwash gravel) in areas of summer drought (the opposite is true in areas of discontinuous permafrost). Substrate materials also may affect forest composition. Within a specific region, for example, the forest on rolling till surfaces may differ significantly from that on gravelly outwash terraces.

Geology and topography also have important economic implications for timber harvesting. In general, the costs of road construction and logging are much higher on steep rock slopes than on low-relief surfaces underlain by Quaternary sediments. Economic and environmental costs may be so high in some areas, for example on slopes prone to landsliding, that logging is not justified.

Mining

Glaciation has played a vital, although largely negative, role in mineral development in the Canadian Cordillera. Pleistocene glaciers have eroded the weathered supergene mantles of many orebodies, dispersing valuable secondary ores. Many primary orebodies thus exposed by glacier action have been buried by younger drift and are not amenable to discovery by traditional prospecting methods. Instead, they remain to be found by geophysical, geochemical, and aeromagnetic surveys, and by drilling. Some surface and subsurface orebodies may be found by tracing mineralized boulders along former glacier flow lines back to their sources. Others may be located by identifying and mapping areas of till containing elevated concentrations of base metals ("dispersal fans").

More than 50% of the gold and 4% of the silver produced in the Canadian Cordillera have come from placers in Quaternary and late Tertiary sediments (Fig. 1.35, 1.36; Debicki, 1983; see also Morison, 1989). Some of these placers predate one or more glaciations and either were not eroded by glaciers or were never overridden by them; others postdate the last glaciation. Most placers in the Cordillera, irrespective of age, are gravels of fluvial or glacial fluvial origin. The richest deposits occur in stream valleys and gulches in hilly terrain. Some are associated with and clearly derived from lodes (e.g., the placers of the Cariboo gold fields), but others have no obvious bedrock sources.

Three examples illustrate the range of gold placers in the Cordillera. The Klondike placers were discovered in 1896 and have yielded about 285 000 kg of gold, more than half of the placer production of the Canadian Cordillera

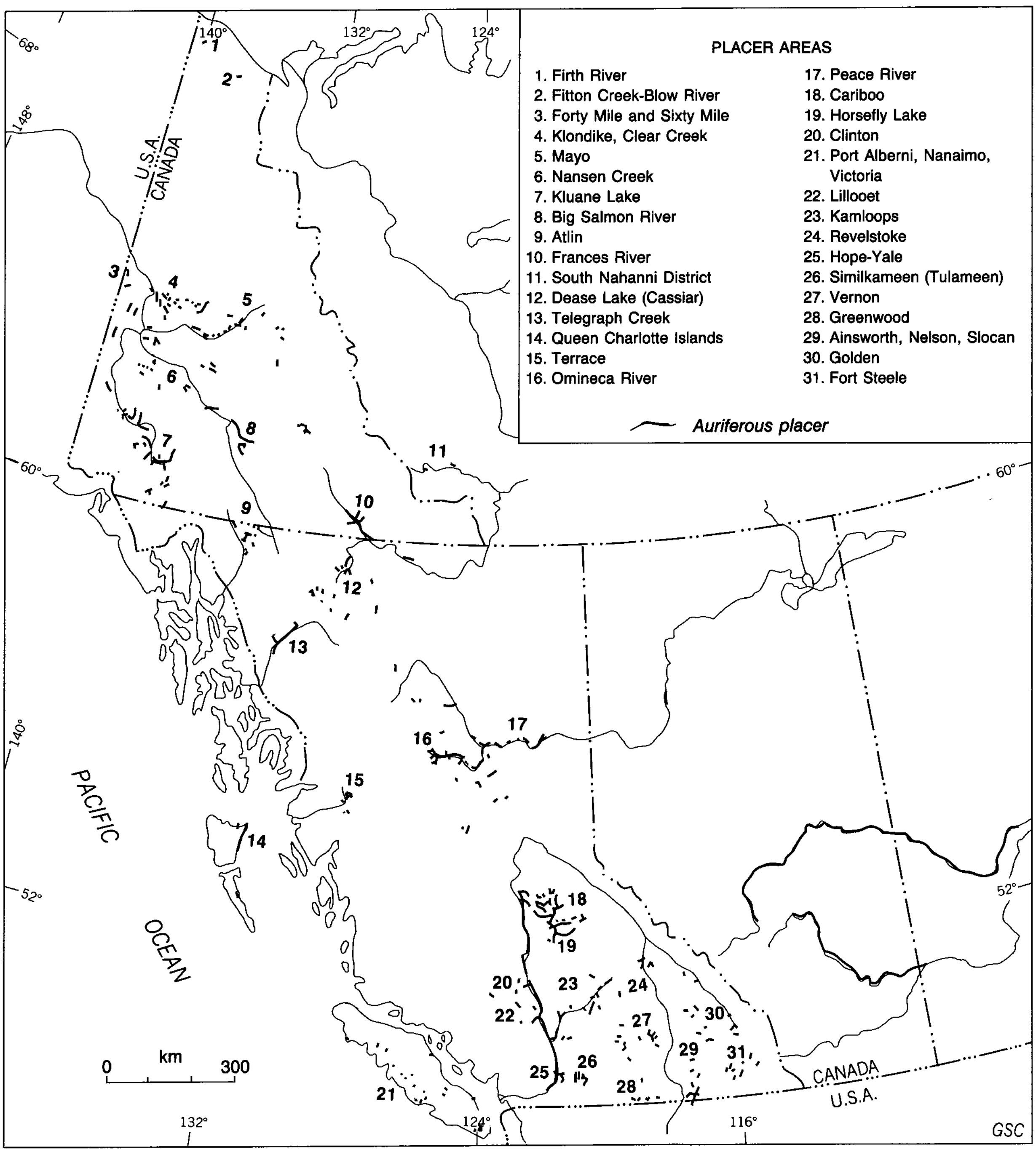

Figure 1.35. Placer gold deposits in the Canadian Cordillera (adapted from Boyle, 1979, Fig. 79).

(Fig. 1.36; McConnell, 1905; Boyle, 1979; Morison, 1987). The gold occurs in several valleys dissecting an unglaciated upland near Dawson, Yukon Territory. These valleys are flat, wide, and locally terraced in their lower reaches, but they narrow into steep-sided gulches towards their heads and end abruptly in broad amphitheatres. The Klondike placers owe their existence to uplift and erosion of deeply weathered, gold-bearing schists in late Tertiary time. Auriferous White Channel gravels accumulated in broad valleys cut into the uplifted upland. Later, during the Pleistocene, streams cut down through the White Channel gravels into underlying bedrock, and a series of successively lower terraces were formed. Streams completed the incision of their valleys and deposited reworked White Channel gravels on the eroded valley floors some time before 50 ka (Naeser et al., 1982). Most Klondike gold has been recovered

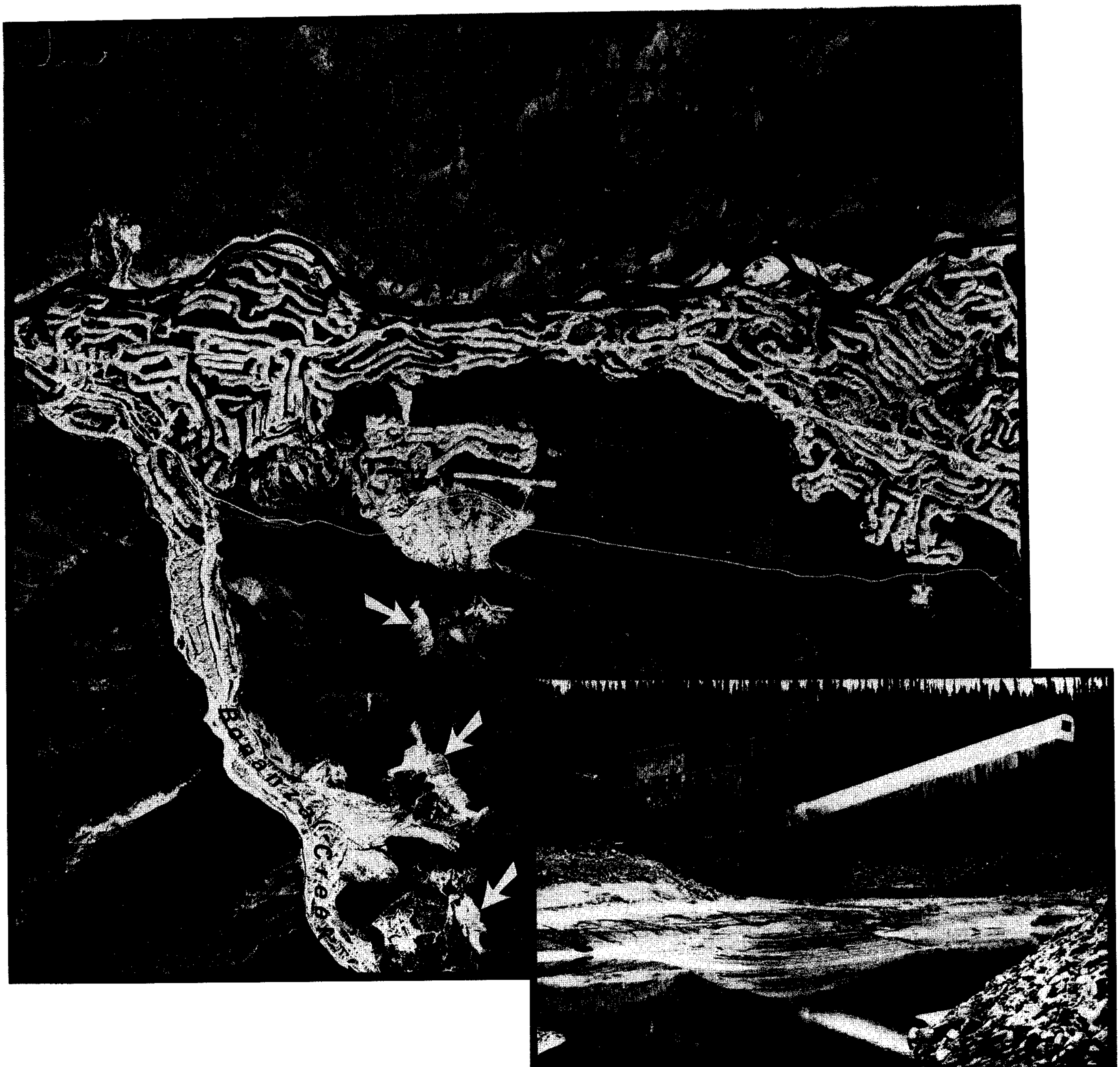

Figure 1.36. The Klondike gold district near Dawson, Yukon Territory. Large amounts of placer gold have been recovered from Plio-Pleistocene gravel which overlies high bedrock benches in this area; arrows indicate old hydraulic workings in these materials. Placer gold also is present in younger gravels underlying terraces and present-day valley floors. Valley floor deposits have been extensively mined by dredges (lower right), giving rise to the conspicuous, worm-like pattern of tailings along Klondike River and Bonanza Creek. NAPL A17155-99, 204655.

from basal White Channel deposits on benches along valley sides and from gravel underlying adjacent valley floors.

The Cariboo placers, discovered in 1860, have supplied over 70 000 kg of gold, more than any other placer district in British Columbia (Johnston and Uglow, 1926; Boyle, 1979). Several types of auriferous sediments are found in the district, and the gold has had a complex history following its release during the Tertiary from quartz veins and stringers in metamorphosed sedimentary rocks. First, there was deposition during the Tertiary of gold-bearing gravel and colluvium in valleys cut into the upland plateau. This was followed during the Pleistocene by repeated glaciation of the region. Some of the Tertiary placers were scoured away by Pleistocene glaciers and reworked by meltwater streams, but others were preserved beneath a thick mantle of drift. Gold placers also formed during Pleistocene nonglacial intervals when streams reworked older gold-bearing gravels. Finally, auriferous stream gravel was deposited in some valleys at the close of the Pleistocene and during the Holocene. These deposits underlie present-day floodplains and stream terraces, and contain variable although generally low amounts of gold.

The Atlin placers, discovered in 1898, occur in valleys cut into an upland that was repeatedly glaciated during the Pleistocene. Placer gold is found in stream gravels within a region containing base metal sulphide deposits and auriferous quartz veins (Boyle, 1979). There are two main types of auriferous gravels in the Atlin district: (1) highly decomposed yellow gravels of Tertiary and Pleistocene age; and (2) undecomposed gravel eroded from drift and old gold-bearing deposits during and following Late Wisconsinan deglaciation of the region.

Construction materials

Quaternary deposits are economically important as sources of construction aggregate. Gravel and sand are in great demand in all urban areas of the Canadian Cordillera and are used extensively in the construction of highways. The annual value of these materials in recent years has been of the order of 50-100 million dollars (British Columbia Ministry of Energy, Mines and Petroleum Resources, 1980). Production is concentrated near urban centres where most of the material is used in concrete, asphalt, and as fill. Supplies near Vancouver, Victoria, and a few other cities, however, are rapidly being depleted or covered with houses, necessitating lengthy transport at high cost. In some areas, transport costs are prohibitive, and crushed rock is substituted for gravel for some purposes.

A wide variety of Quaternary sediments are quarried for construction aggregate. Most of the gravel and sand produced in the Canadian Cordillera comes from outwash and ice contact deposits that were laid down during the advance and recessional phases of the last glaciation. These materials commonly occur above present base level and are terraced. They are thus relatively easy to mine, except where covered by thick till, silt, or clay. Large amounts of construction aggregate and fill also are obtained from Holocene fluvial deposits. In a few areas, Pleistocene beach deposits and even till and rubbly colluvium are mined for aggregate. Till, colluvium, and clay also are used where construction fill of low permeability is required, for example in earth-fill dams. Finally, gravel is dredged from the seafloor to meet the needs of some communities lacking suitable deposits on land (e.g., Prince Rupert).

Clay deposits of glacial marine origin were once exploited in the manufacture of brick, tile, and related products in British Columbia. This practice has been discontinued, despite large reserves of suitable material, due to competition from imported and substitute products.

Water resources

Abundant lakes and ponds which owe their origin mainly to Pleistocene glacial erosion and deposition play a major role in regulating runoff. Quaternary sediments also modulate the flow of streams by holding large amounts of water that would otherwise be lost through rapid runoff. In this manner, Quaternary sediments help maintain lake levels and enhance stream discharge during the dry summer season, both of which are important in hydroelectric power generation, irrigation, and the supply of water to communities. Glaciers and seasonal snowpack play a similar role in the seasonal storage of water in the high mountains of the Canadian Cordillera. In northern Yukon Territory where glaciers, lakes, and thick Quaternary deposits are generally lacking and where the subsoil is perennially frozen, drainage basins have only limited water storage capacity.

Thick Quaternary sediments in valleys and other lowland areas in the Canadian Cordillera contain large amounts of groundwater. At present, about 10% of the water used in the region is groundwater; the remainder is surface water from streams, natural lakes, and reservoirs (Barker, 1977). People in rural areas and in some communities rely almost exclusively on groundwater for domestic and farm needs. In some areas, however, groundwater use is limited by high concentrations of dissolved solids, by contamination from septic tanks, sanitary landfills, feed lots, fertilizers, and pesticides, and by the lack of adequate aquifers.

Although a variety of Quaternary materials contain groundwater, the best aquifers in the Cordillera are thick permeable glacial fluvial and fluvial deposits that occur in areas of high surface recharge such as valleys and lowlands adjacent to mountain slopes. Many of these deposits underlie one or more till sheets and thus are confined aquifers. In contrast, Holocene fluvial deposits directly underlie present-day floodplains and adjacent stream terraces and generally are unconfined.

Natural hazards

Landslides, floods, earthquakes, and volcanic eruptions have played a major role in shaping the landscape of the Canadian Cordillera. These and other natural processes also pose hazards to people and property in this region. Some of these hazards can be mitigated, but only after sufficient geological, geophysical, and historical data have been collected to appraise the probable location, magnitude, and frequency of future destructive events. Floods, earthquakes, and snow avalanches occur repeatedly, thus risk can be assessed by making observations over a long period of time. Areas susceptible to landslides and volcanic eruptions also can be identified, although predicting exactly when and where such events will occur is not yet possible. While in some cases the only option is to avoid a hazardous area, development may be possible after appropriate steps have been taken to reduce risk.

Landslides

A wide variety of potentially destructive mass movements occur in the Canadian Cordillera (Eisbacher, 1979; Evans, 1982; Cruden, 1985; Carson and Bovis, 1989; Evans and Gardner, 1989). Most are complex and are related to instabilities in steep bedrock slopes and to high-gradient mountain streams. They are controlled by unique combinations of topographic, geological, meteorological, and seismic factors and by human activity.

Landslides involving bedrock range from small falls and topples to very large (>10^6 m^3) slumps and slides.[1] Rockfalls are common on steep bedrock slopes in areas of intensely fractured or jointed rocks and are triggered by freeze-thaw activity, intense precipitation, and earthquakes (Peckover and Kerr, 1977; Piteau, 1977; Mathews, 1979). They occur frequently throughout the Cordillera, disrupt road and rail traffic, and occasionally take human lives. The small size of these landslides (typically 10-1000 m^3) belies the fact that they are among the most costly in the region. Aside from economic losses due to traffic delays, there is an enormous cost involved in scaling, blasting, and grouting threatening faces and in removing debris from roads and railways.

Rapid bedrock failures involving a sliding or flowing type of movement (rockslides and rock avalanches) are most common in areas of high relief where geological discontinuities (e.g., fractures, faults, bedding planes, cleavage, intrusive contacts) dip in the direction of the slope. They are triggered by excessive porewater pressure, seismic shaking, and human activity, among other things. Fundamental natural causes include erosion of slopes by streams and glaciers and the gradual destruction of cohesion along discontinuities by physical weathering and solution. At least 17 large rockslides and rock avalanches have occurred in historical time in the Canadian Cordillera, and 4 of these have claimed lives (Kerr, 1948; Mokievsky-Zubok, 1977; Cruden and Krahn, 1978; Mathews and McTaggart, 1978; Moore and Mathews, 1978; Clague and Souther, 1982; Cruden, 1982; Eisbacher, 1983; Evans, 1987; Evans et al., 1987). Although uncommon in comparison to small rockfalls and debris flows, rockslides and rock avalanches can be extremely destructive; they are responsible for about 40% (140) of the recorded landslide deaths in the Canadian Cordillera.

Rotational slides (slumps) of a range of sizes occur both in bedrock and Quaternary sediments, and are common in most parts of the Cordillera. Rotational sliding often takes place in association with other types of mass movements. For example, many landslides involving Quaternary sediments are retrogressive slumps at their heads but translational slides or flows at their distal ends. Also, some debris flows are initiated by the blockage of rain-swollen torrents by slumps.

Large deep-seated slope failures characterized by the slow downslope movement of internally broken rock masses along poorly defined rupture surfaces ("sagging slopes") are found in some areas of foliated metamorphic rocks, for example parts of the Columbia Mountains. An especially well documented example of such a failure is the Downie slide in Columbia River valley (Piteau et al., 1978). Although initiated thousands of years ago, this and many other large slope sags in the Cordillera are still active, and any permanent structures located on them ultimately will be damaged or destroyed. In addition, parts of some sags may detach from the main mass of creeping debris and move rapidly downslope; such a possibility should be considered when siting communities and major developments such as hydroelectric reservoirs.

Flows of unconsolidated sediment or weathered bedrock involve the displacement of a mass of material as a viscous fluid. Most such failures in the Canadian Cordillera are debris flows (Fig. 1.37) resulting from the failure of water-saturated, heterogeneous Quaternary sediments (e.g., till, glacial fluvial deposits, pyroclastics, colluvium). Debris flows are most common in mountain areas which receive abundant precipitation. They are triggered principally by intense fall, winter, and spring rainstorms, often accompanied by rapid snowmelt (Eisbacher and Clague, 1984; VanDine, 1985; Church and Miles, 1987); some, however, are initiated by the sudden draining of moraine- or glacier-dammed lakes (Jackson, 1979; Blown and Church, 1985; Clague et al., 1985). Most debris flows are funnelled along steep valleys and ravines and debouch onto fans or cones which often are sites of death and destruction (e.g., Nasmith and Mercer, 1979; Eisbacher, 1983; Jackson et al., 1985; VanDine, 1985). Debris flows are much smaller than most rockslides and rock avalanches, but occur more frequently and are responsible for over one-third (ca. 115) of all recorded landslide deaths in the Canadian Cordillera.

Flows and complex slides consisting mainly of silt and clay occur in areas of Pleistocene glacial lacustrine and glacial marine sediments (Evans, 1982). Failure commonly results from high pore water pressures attributable to both natural causes such as heavy rains and rapid snowmelt, and to human activity, for example irrigation. In drier parts of the Cordillera, special care must be taken with surface water in areas where glacial lacustrine silt and clay form bluffs. Several cities and towns (e.g., Vancouver, Victoria, Kamloops, Prince George, Kitimat, Whitehorse) are located partly on glacial lacustrine or glacial marine deposits, thus flows and related slides and lateral spreads pose a significant hazard to people and property. At least 25 people have died in landslides in these materials in the Twentieth Century, and property damage has been extensive.

Sediment flows also are common on slopes in permafrost areas. Failure occurs in zones of high porewater pressure at the base of the active layer during summer thaw (Hughes et al., 1973; Mackay and Mathews, 1973; Rutter et al., 1973; McRoberts and Morgenstern, 1974). These flows are found on slopes as gentle as about 7° and are especially common along river and coastal bluffs underlain by ice-rich Quaternary sediments.

Large slow-moving flows in weathered volcanic and sedimentary rocks of Cretaceous and Tertiary age are present in Fraser River basin in south-central British Columbia (Eisbacher, 1979; VanDine, 1980; Bovis, 1985). Bentonites and other clay-rich rocks and high porewater pressures are implicated in their movement. Many of these flows are presently active, but they move very slowly and consequently pose a hazard only to certain structures located on them (e.g., roads and rail lines).

Subaqueous mass movements similar to some terrestrial sediment flows are relatively common in British Columbia fiords, at several other sites along the coast, and in some lakes. Most occur at the fronts of active deltas and on steep slopes underlain by unconsolidated sediments (Prior

1 Landslide terminology after Varnes (1978).

and Bornhold, 1984, 1986; Prior et al., 1982, 1984, 1986). Several historical submarine slope failures have occurred in Howe Sound (Terzaghi, 1956; Prior et al., 1981) and Kitimat Arm (Luternauer and Swan, 1978; Swan and Luternauer, 1978; Prior et al., 1982, 1984); one in Kitimat Arm in 1975 generated a train of sea waves that damaged port facilities at Kitimat (see Tsunamis).

Snow avalanches

Avalanches of dry and wet snow, in some cases with dispersed sediment or freshly broken rock, are common in the mountains of western Canada during winter and spring. They occur both on open slopes and in ravines and gullies. Areas of frequent avalanche activity on forested slopes are easily identified because they lack trees; in contrast, avalanche areas above treeline may be more difficult to recognize.

Although they can occur almost any time on slopes that are moderately steep and snow-covered, avalanches are favoured by certain weather conditions (Fitzharris and Schaerer, 1980). They are especially common during thaws or periods of rain after heavy snowfalls, and when thick dry snow accumulates on old icy snow surfaces. In the latter case, the buried icy surface is a plane of weakness along which overlying snow may easily slide.

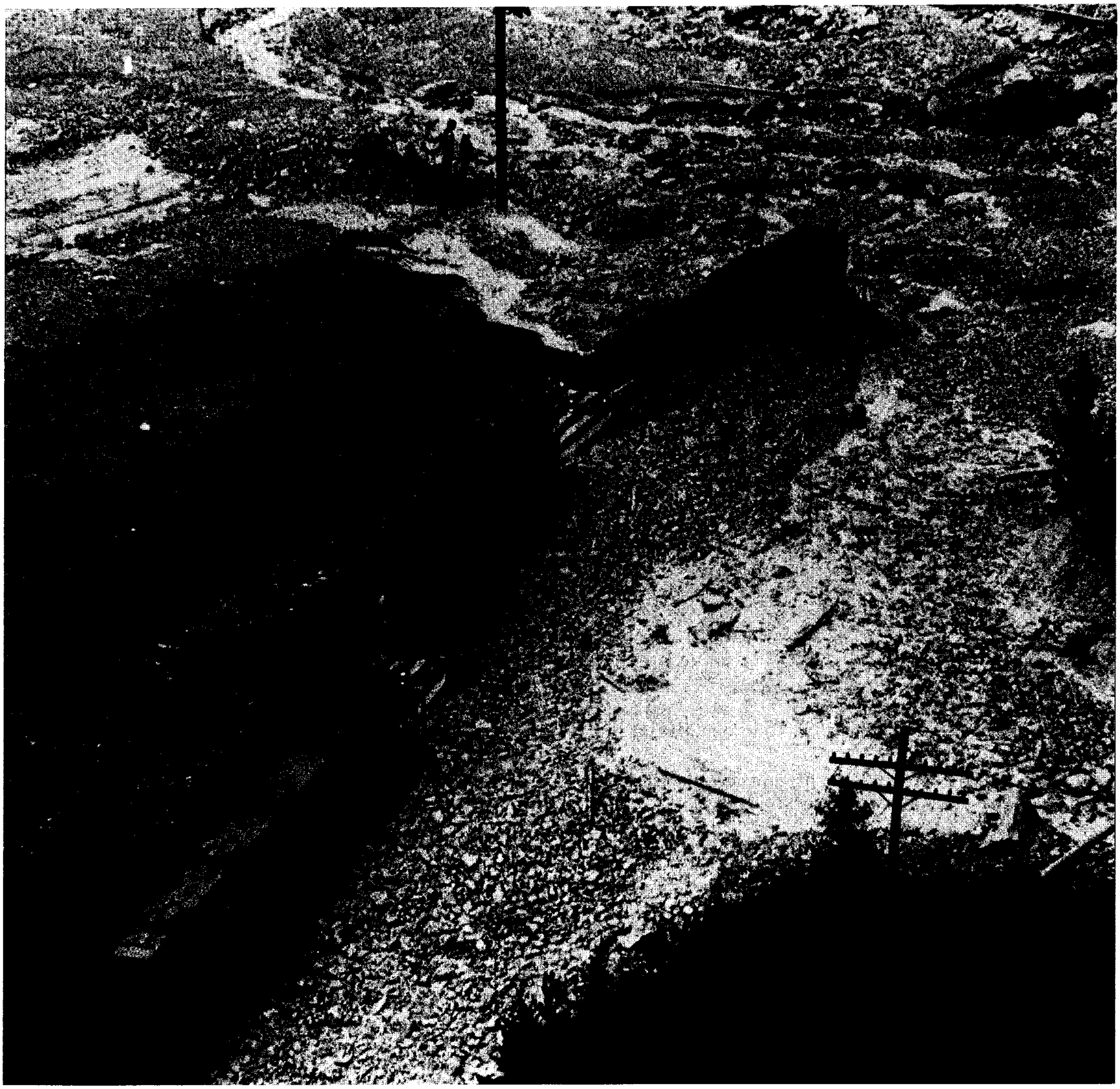

Figure 1.37. Locomotives derailed by debris-flow deposits in the southern Rocky Mountains, British Columbia, September 7, 1978. Photo by Calgary Herald, courtesy of Glenbow Archives, Calgary, Alberta.

In recent decades, avalanches have taken the lives of many people engaged in recreation activities in remote areas (Stethem and Schaerer, 1979, 1980). Avalanches also were responsible for two disastrous mining-related accidents (Chilkoot Pass, 1898; Granduc Mine, 1965), and regularly disrupt road and rail traffic in some mountain passes and valleys (Schaerer, 1962; Avalanche Task Force, 1974; Barker, 1977).

Floods

Although there has been little loss of life due to floods in the Canadian Cordillera, property damage has been extensive and the cost of flood-control measures high. The most destructive historical flood in this region occurred in southwestern British Columbia in May and June of 1948. At that time, Fraser River broke its dykes in Fraser Lowland and flooded large tracts of agricultural land. As a result, 16 000 people were evacuated and 2000 homes damaged; total losses amounted to more than $20 000 000 (Barker, 1977).

Flooding in the southern Cordillera occurs during extended periods of warm weather in the spring and during periods of heavy rainfall. Large rivers generally crest in late spring as the winter snowpack melts. In contrast, many medium-size rivers have maximum flows during rainy periods in the fall. Streams with small catchments may flood at any time of the year.

Although rainstorm and snowmelt floods also occur in the northern Cordillera, another type of flood perhaps is more common there. In winter, rivers in the north freeze over, often to a depth of more than 1 m. During spring thaw, these rivers swell and their ice crusts are broken into sheets that float downstream. Channel constrictions and river bars may obstruct the free passage of the ice blocks and thus create large jams that force the river over its banks. When the flow becomes powerful enough to break the ice jams, the backed-up waters flood downstream. Considerable damage may be caused in this manner both by the water and by the debris and ice carried along with it. Similar floods occasionally occur when aufeis impedes flow sufficiently to force a stream over its banks during spring thaw.

Some low-lying areas along the coast of British Columbia are flooded by the sea during severe storms, unusually high tides, or as a result of tsunamis. For example, exceptionally high spring tides raised the level of Fraser River during the 1948 flood, causing additional damage to parts of Fraser Lowland.

Finally, lakes impounded by present-day glaciers and by bulky neoglacial end moraines may drain catastrophically to produce severe downstream floods. There have been numerous historical jökulhlaups (glacier-outburst floods) in the Coast, Rocky, and St. Elias mountains (Marcus, 1960; Mathews, 1965, 1973; Gilbert, 1972; Jackson, 1979; Clague, 1982b; Clague and Rampton, 1982; Clarke, 1982; Clarke and Waldron, 1984; Blown and Church, 1985; Jones et al., 1985; Ryder, 1985; Gilbert and Desloges, 1987). Floods from moraine-dammed lakes have been recognized in the Selkirk Mountains and southern Coast Mountains (S.G. Evans, Geological Survey of Canada, personal communication, 1984; Blown and Church, 1985).

Erosion

During periods of high discharge, streams may erode their banks and, in some cases, occupy new channels through avulsion. Erosion may be particularly severe at meander bends and other inflection points and opposite the mouths of aggrading tributaries. Rapid lateral erosion by streams may create steep banks prone to landsliding. Braided streams, which are common in and around glaciated mountain ranges, are especially susceptible to channel migration and avulsion; these processes, in conjunction with normal flooding, render the floodplains of these streams hazardous sites for development.

Parts of the shoreline of British Columbia have receded rapidly (up to a few metres per year) in historical time, necessitating the abandonment or relocation of homes and roads. Erosion has been most severe along those parts of the coast bordered by thick Quaternary sediments, for example eastern Graham Island, eastern and southern Vancouver Island, and near Vancouver (Clague and Bornhold, 1980; Clague, 1982a).

Tsunamis

The most destructive historical tsunami in British Columbia resulted from the Alaska earthquake of March 27, 1964 (Wigen and White, 1964; Thomson, 1981). Waves up to 7 m high swept into the lower parts of Port Alberni, damaging 260 homes, 60 extensively. Hot Springs Cove and Zeballos also suffered wave damage. Seismic sea waves comparable to those generated by the 1964 earthquake are rare. Of the 176 tsunamis recorded in the Pacific Ocean between 1900 and 1970, 35 caused damage near their sources, but only 9 resulted in widespread destruction (Thomson, 1981). Even fewer produced significant runups on British Columbia shores.

Tsunamis generated by distant earthquakes do not affect all parts of the British Columbia coast equally. The 1964 tsunami, although highly destructive at Port Alberni, caused little damage at Tofino only 65 km away and had little effect on protected waterways such as the Strait of Georgia.

Not all tsunamis in British Columbia are the result of distant earthquakes; some are produced by local quakes and submarine landslides. For example, the Vancouver Island earthquake of June 23, 1946 ($M = 7.2$) produced a local tsunami that affected shores along the Strait of Georgia and nearby inlets (Murty, 1977). In April 1975, sea waves generated by a large submarine landslide near the head of Kitimat Arm slammed into shore installations at Kitimat, causing about $600 000 damage (Campbell and Skermer, 1975; Murty, 1979).

Earthquakes

Although the Canadian Cordillera is seismically active (Milne et al., 1978), damage from earthquakes in the region has been extremely low. The main reason for this is that most large quakes have occurred far offshore at lithospheric plate boundaries. Also, with the exception of the 1946 Vancouver Island earthquake, the few large earthquakes that have occurred on land have been sufficiently far from urban centres that damage has been minor. Nevertheless, some time in the future, a large earthquake probably will occur near a major population centre in western British Columbia. In general, earthquake risk decreases towards the east and is moderate to low in most of the interior of the Canadian Cordillera. High levels of seismicity, however, have been recorded in parts of Yukon Territory and District of Mackenzie, and earthquake risk in these remote, sparsely populated areas is considered to be relatively high (Basham et al., 1977; Horner, 1983).

Earthquake damage is produced both by direct ground motion and by secondary causes such as landslides and fires. The extent of damage is closely related to magnitude, epicentral proximity, and focal depth. In addition, earth materials respond differently to seismic shaking, thus geological factors are important determinants of earthquake damage. Structures located on Quaternary sediments generally are more prone to damage from shaking than structures on bedrock, although some dense compact sediments perform as well as rock in this respect. Saturated, fine grained sediments may liquefy during a major earthquake, giving rise to destructive flows and spreads. Delta-front sediments off the mouth of Fraser River presumably could fail in this way (Luternauer and Finn, 1983). Other materials prone to failure during earthquakes are artificial fills and fine glacial lacustrine and glacial marine deposits.

Volcanic eruptions

There have been many volcanic eruptions in the Canadian Cordillera during the last several thousand years, although none during this century. Quaternary volcanoes occur in clusters and narrow belts in the western Cordillera (Fig. 1.3; Souther, 1970, 1977), and it is likely that any future eruption will also take place in these areas.

Most Quaternary eruptive centres have been the locus of a single pulse of activity during which one or more small pyroclastic cones were built and a small volume of basalt erupted to form thin blocky flows. The past record indicates that an eruption of this type is likely to occur somewhere in the Canadian Cordillera in the next several centuries. Such an eruption probably would affect only a small area and would not be hazardous, unless it occurred in an inhabited region.

In contrast, a few volcanic centres in Alaska, British Columbia, Washington, and Oregon have erupted explosively on several occasions during the Holocene, with far-reaching effects (Table 1.1, Fig. 1.11). The most recent significant eruption of this sort occurred about 1.2 ka when a layer of tephra was deposited over a large area of southern Yukon Territory and western District of Mackenzie from a vent near the Alaska-Yukon boundary. Future eruptions of this sort would likely damage property and crops and disrupt transportation in the Canadian Cordillera. This would result mainly from fallout of ash and dust over large areas and from flooding and aggradation in stream valleys surrounding the volcano. Lahars, pyroclastic flows, and pyroclastic surges would produce additional damage near the volcano.

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Author's Addresses

J.J. Clague
Geological Survey of Canada
100 West Pender Street
Vancouver, British Coumbia
V6B 1R8

W.H. Mathews
Department of Geological Sciences
University of British Columbia
Vancouver, British Columbia
V6T 1W5

J.M. Ryder
Department of Geography
University of British Columbia
Vancouver, BritishColumbia
V6T 1W5

O.L. Hughes
Geological Survey of Canada
3303-33rd Street N.W.
Calgary, Alberta
T2L 2A7

N.W. Rutter
Department of Geology
University of Alberta
Edmonton, Alberta
T6G 2E3

L.E. Jackson, Jr.
Geological Survey of Canada
100 West Pender Street
Vancouver, British Columbia
V6B 1R8

J.V. Matthews, Jr.
Geological Survey of Canada
601 Booth Street
Ottawa, Ontario
K1A 0E8

G.M. MacDonald
Department of Geography
McMaster University
Hamilton, Ontario
L8S 4K1

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