GEOLOGICAL SURVEY OF CANADA

OPEN FILE 4022

Capsule Geology of the Vancouver Area and Teacher's Field-Trip Guide

J.A. Roddick

2001

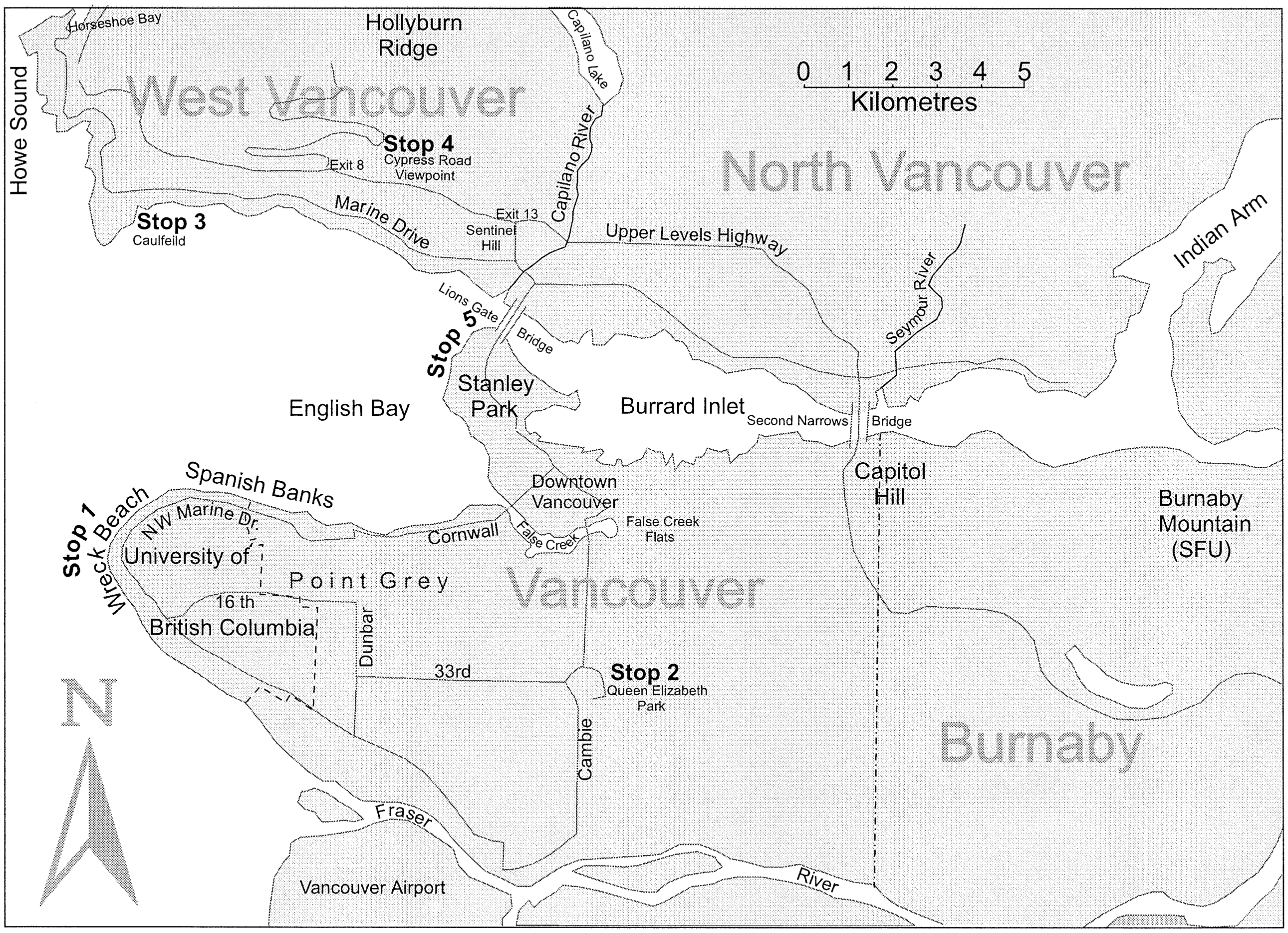
For more on granites go to the Roddick granite webpage at : http:// 132.156.108.210/personalpages/roddick

Natural Resources Canada Ressources naturelles Canada

Canada

GEOLOGICAL SURVEY OF CANADA
OPEN FILE 4022

Capsule Geology of the Vancouver Area and Teacher's Field-Trip Guide



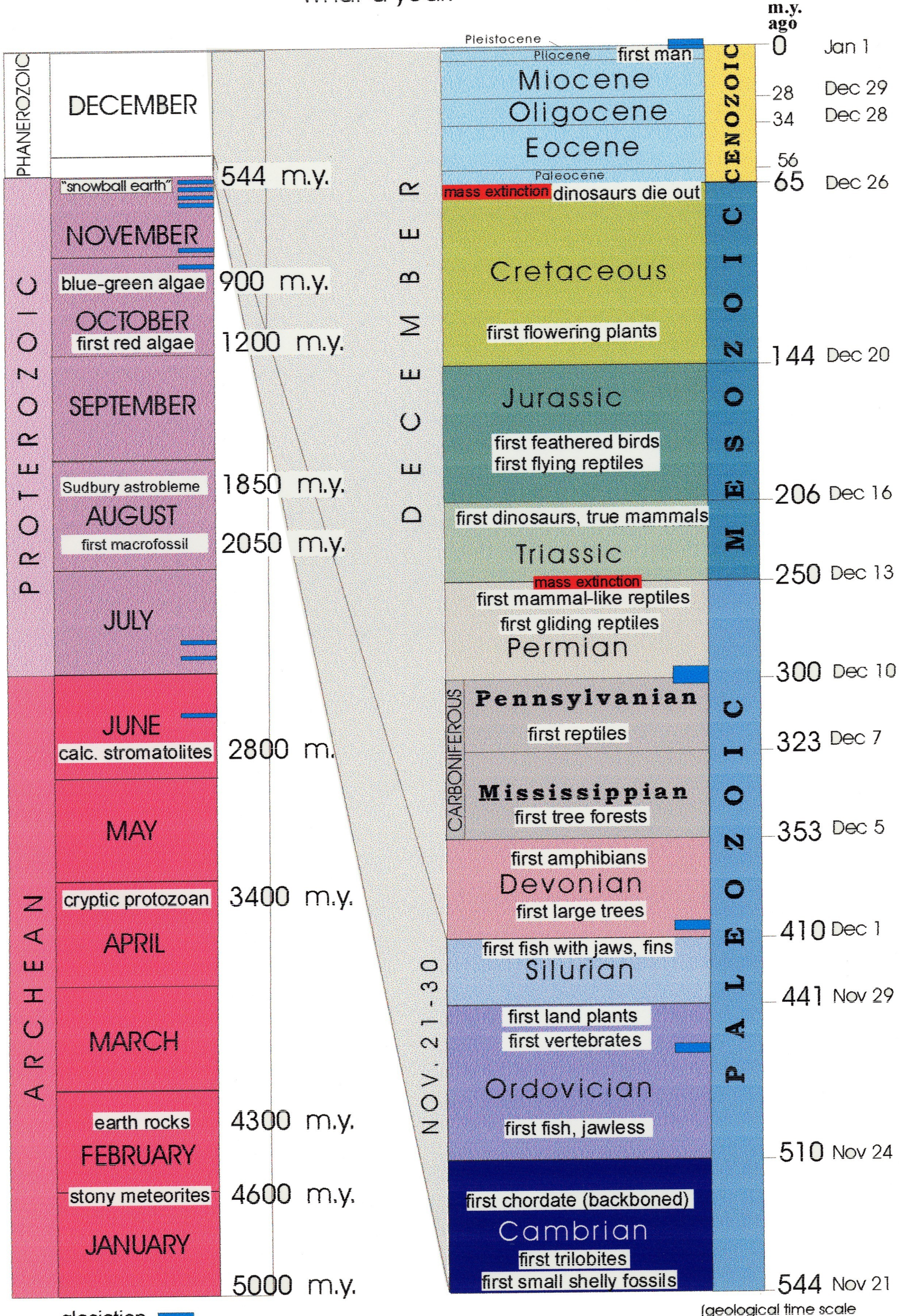
J.A. Roddick
2001

Natural Resources Canada | Ressources Naturelles Canada

Canada

GEOLOGIC TIME IN A YEAR

What a year!



J.A. Roddick, 2001
GSC, Vancouver

Figure 1. Geological Time Scale

(geological time scale based on Okulitch, 1995)

Capsule geology of the Vancouver area

J.A. Roddick

INTRODUCTION

This Open File contains a brief description of the Vancouver geological setting; descriptions of five field trip stops; a road log; a summary of glaciation theory; and a summary of the granite controversy.

The Vancouver area is composed of two distinct geographical areas: the Coast Mountains along the North Shore and the Fraser Lowland encompassing much of the city and extending south across the Canada-USA border. The region is flanked on the south and east by the Olympic and Cascade mountains and to the west by the Strait of Georgia and the Insular Ranges on Vancouver Island.

Coast Mountains

The most striking geologic feature in the Vancouver area is the backdrop of rugged mountains rising up from the North Shore. These mountains form the southern end of the Coast Mountains (Fig. 2) which extend 1700 km north from Vancouver to the Yukon. They are underlain mainly by the Coast Plutonic Complex (CPC) which consists mostly of granitoid (more than 80% near Vancouver) and metamorphic rocks. Mount Baker (Fig. 2), the only volcano of the Cascade chain visible from Vancouver, is south of the United States border, and is built on rocks similar to the CPC by magma which originated far below.

The Lithoprobe project (Varsek et al., 1993) provided reflection seismic data along a transverse profile north of this region that indicated the Moho to be about 35 km deep along the northeast side of Georgia Strait, and gradually deepening to the east.

Glaciation

The Vancouver region was strongly affected by glaciation. When the ice age began depends on what is meant by ice age. Definitions vary so widely that the term has little significance beyond its local context. Vancouver has been ice free for almost 13,000 years. Antarctica has not been free of ice for, at least, the last 11 million years (Miocene). In fact, a few believe the earth has not been entirely ice-free since the Cretaceous. Small alpine glaciers have probably existed in the highest mountains throughout much of earth's history.

As summarized by Clague (1994), the first incursion of ice during the Fraser Glaciation, the local name for the most recent ice age, took place about 25 ka BP. Glaciers retreated briefly from the Vancouver area about 19 ka BP, and re-advanced about a thousand years later. At the height of Fraser Glaciation, about 15,000 years ago, the site of Vancouver was under about 1800 m of ice. Glaciers from the Coast Mountains and the Cascade Mountains (Fig. 2) formed large ice sheets which covered the Fraser Lowland and extended out into the Strait of Georgia. At its maximum extent, ice extended a few miles south of Puget Sound in Washington State. The Puget lobe began to retreat about 14 ka BP and the Fraser Lowland is thought to have been free of ice about 11 ka BP.

The entire region was isostatically depressed during Fraser Glaciation, in places more than 250 m. Sea level was lower during each major ice advance, and the glaciers were able to cut below the present day sea level in the coastal portions of many valleys. With deglaciation the region rose, unevenly, but greater than the coeval eustatic rise in sea level. Valleys were partly submerged, resulting in the creation of fjords in the region. Those in the Vancouver area, such as Indian Arm and Howe Sound, are among the most southerly in the Northern Hemisphere. Incidentally, as first observed by J.D. Dana in 1830s, the most northerly fjords south of the equator, in Chile, appear at a similar latitude (Jackson and Clague, 1991). During the last 2,000 years relative sea level has varied no more than one metre (Clague, 1994).

The unevenness of isostatic depression and rebound on the coasts of British Columbia and southeast Alaska makes it difficult to place the level of compensation as deep as the asthenosphere. The mid- to lower-crust, having a lesser viscosity than either the upper crust or upper mantle (Mooney and Meissner, 1992; Bailey, 1998) would, in this writer's opinion, be a more probable zone of isostatic compensation.

Before being blocked by Fraser River sediments, Pitt Lake was an arm of the sea, as Indian Arm is today. It is an unusual lake in that a delta is forming at the outlet of the lake. At high tide, flow of the Pitt River is reversed and Fraser River sediments are carried into the lake. (See page 19 for a general summary of glaciation theory).

Fraser Lowland

The Fraser Lowland (Fig. 2) forms the southwest corner of the Pacific Coast mainland of Canada and the adjoining northwest corner of the United States, an area of approximately 3,500 km². Most of this description is only slightly modified from Armstrong, (1990). It is bordered by the Coast Mountains to the north and the Cascade Mountains to the east and southeast.

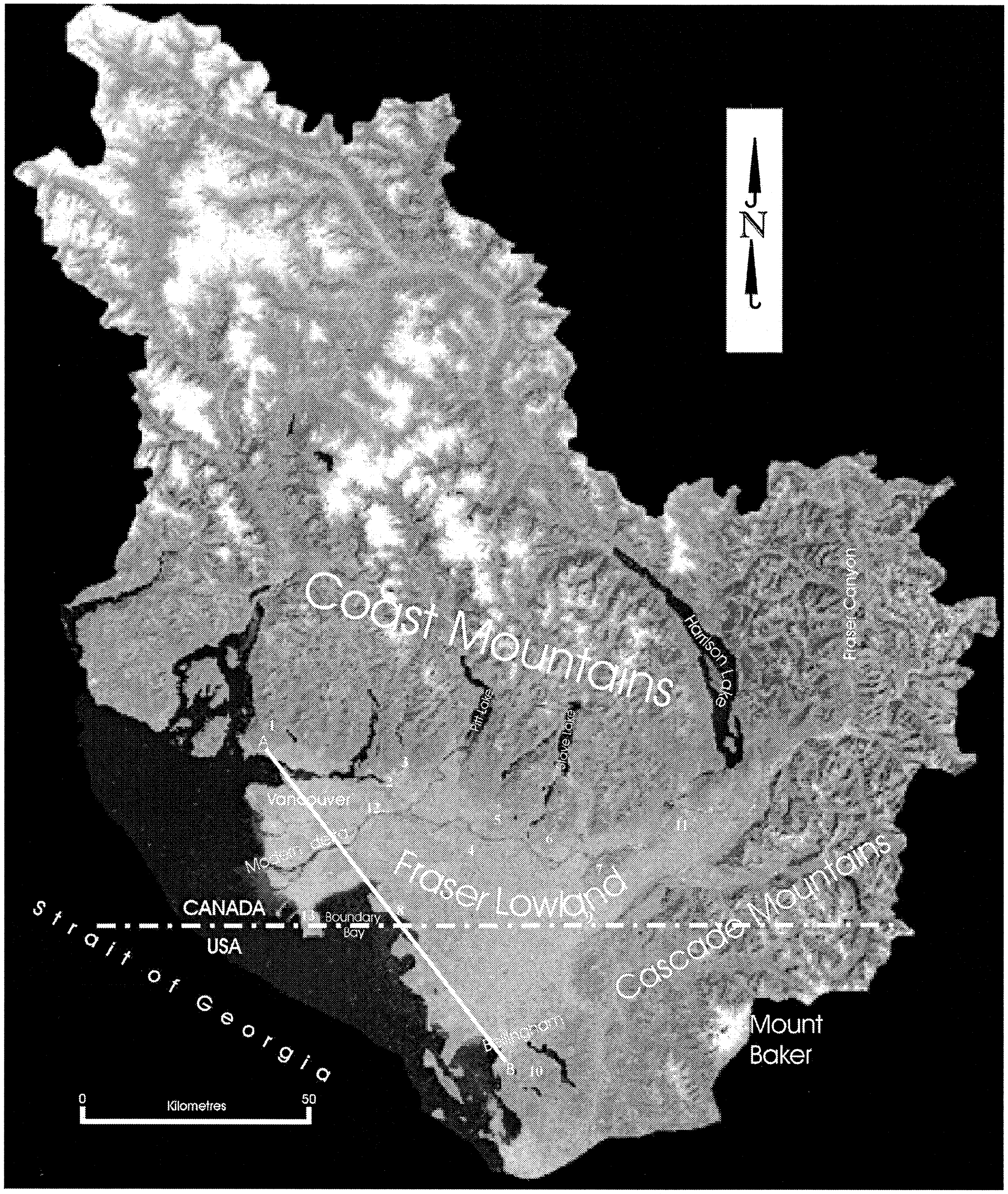


FIGURE 2. Fraser Lowland and confining mountainous region. 1- Mount Strachan; 2- Port Moody; 3- Coquitlam River; 4- Fort Langley; 5- Grant Hill; 6- Silverdale Hill; 7- Sumas Mountain; 8- White Rock; 9- Huntingdon; 10- Chuckanut Hills; 11- Chilliwack; 12- New Westminster; 13- Tsawwassen; A-B, line of cross-section shown in Figure 3. (From Turner, et al., 1996)

The Fraser Lowland consists of gently rolling and flat-topped uplands, separated by wide, flat-bottomed valleys. Several larger hills and low isolated mountains rise above it. Prominent landmarks, such as Burnaby Mountain (cover), Grant Hill, Silverdale Hill and farther east, Sumas Mountain (Fig. 2), expose the oldest rocks. These are mainly freshwater sandstone and conglomerate that were deposited in a depression now called the Georgia Basin starting about 80 m.y. ago. They form the Late Cretaceous Lower Nanaimo Group, and the overlying Late Paleocene to Early Oligocene Kitsilano Member of the Huntingdon Formation.

These strata are consistently tilted about 10 degrees to the south, reflecting a concurrent rise of the Coast Mountains and sinking of Georgia Basin (mostly since mid-Miocene time). Along the northern edge of the Fraser Lowland they form only a thin veneer over the granitic rocks of the Coast Mountains. However, they thicken to the south and reach a maximum thickness of 4400 m near the Canada-USA border (Fig. 3).

The Late Cretaceous and Tertiary sedimentary rocks are covered by up to 300 m of unconsolidated sediments, deposited during glaciation by both water and ice together with that deposited in water between glacial events. They are best seen in the sea cliffs around UBC (Stop 1 at Wreck Beach), at White Rock and along the narrow Coquitlam River valley near Port Moody (cover). They are potentially unstable and can be easily washed away by major floods along rivers and streams that cut through them. Exposed headlands composed of this material, such as Point Grey, are vulnerable to erosion by storm-generated waves and would be quickly eroded if there were a significant rise in sea level.

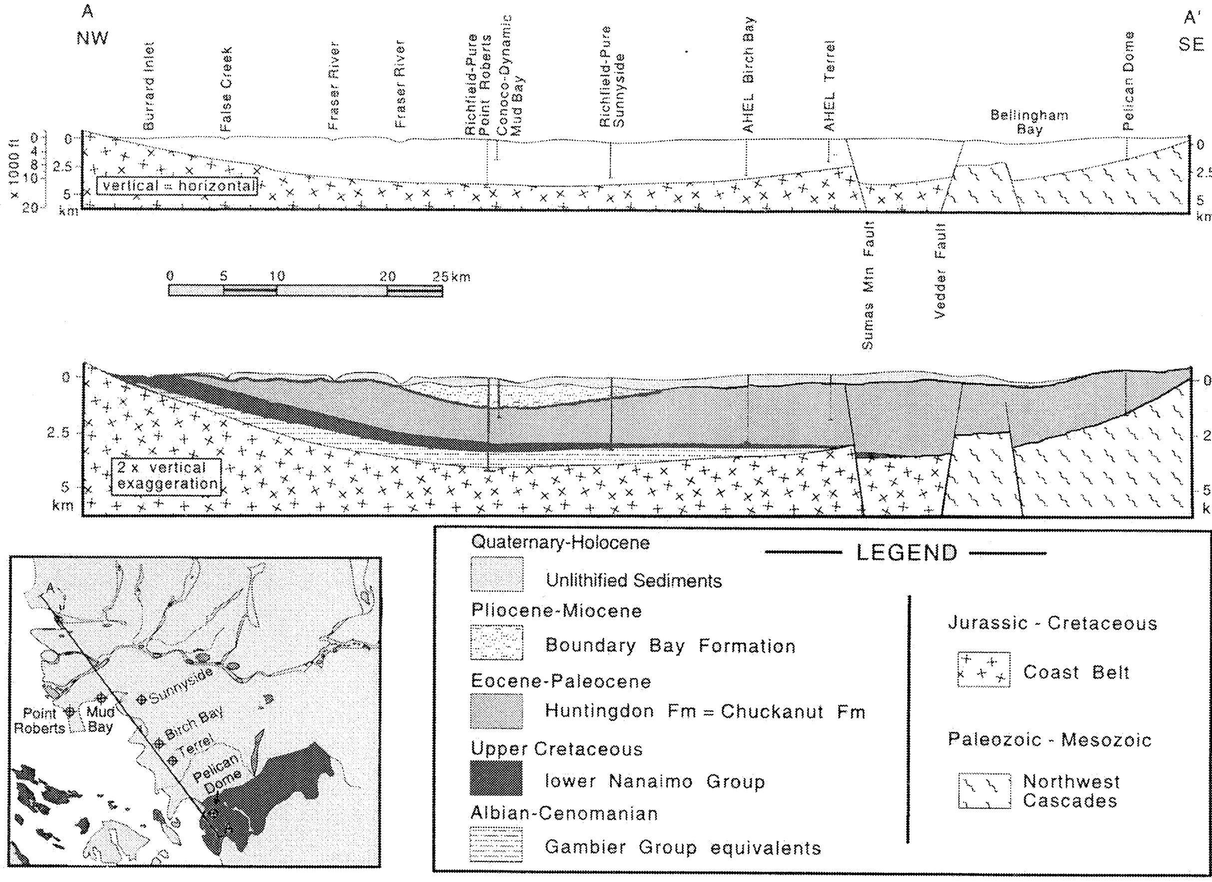


FIGURE 3. Stratigraphic cross-section from the North Shore to Chuckanut Mountain. Locations of exploration wells which provided subsurface data near the line of section are shown. (From Mustard and Rouse, 1994.)

The oldest Quaternary sediments were deposited in a shallow arm of the sea which probably covered much of the present day Fraser Lowland. The hills of older rocks, such as, Burnaby Mountain (cover), formed islands similar to the Gulf Islands we know today. During the major glaciations, and the intervening nonglacial intervals, the areas between the islands were filled by debris carved from the surrounding mountain valleys by the ice.

Fraser River and Delta

The lower part of Fraser River, draining an area about the size of Oregon State, occupies a late glacial to postglacial valley which is 5 km wide and 225 m deep. The river branches into three main arms as it flows across its modern delta (cover), which extends from New Westminster into the Strait of Georgia, a distance of nearly 30 km (Fig. 2). On the coast, the delta is 20 km wide and covers the area between the Point Grey Peninsula (Fig. 1) and Tsawwassen (Fig. 2). The delta also extends 6 to 9 km underwater into the Strait of Georgia as an area of shallow water and banks. Beyond the delta front slope, submarine flows carry deltaic material practically across the strait. The delta could eventually form a land bridge connecting the mainland with the Gulf Islands.

Because of its combination of rich silt and sand, the delta is some of the best agricultural land in the Fraser Lowland. Although the water-saturated sand under static conditions is normally suitable for the foundations of large buildings, it is prone to liquefaction during earthquakes. Submarine slides at the delta front can also be triggered by earthquakes. The slides, however, occur periodically, as a matter of course, following rapid sedimentation at the delta front.

In addition to the Fraser River, two other flat-bottomed valleys cross the Fraser Lowland. The Nicomekl and Serpentine rivers begin near Fort Langley and flow into Boundary Bay (Fig. 2). The valley in which they flow is about 30 km long and 5 km wide and is bordered by low lying hills. Farther east, the Sumas valley, which extends from Chilliwack to the Canada-USA border, is about 25 km long and averages 5 km in width.

Cascade Mountains

Although the Cascade Mountains (Fig. 2) do not form part of the immediate Vancouver area, they bound the east and southeast edge of the Fraser Lowland and are prominent on the Vancouver skyline.

The Cascade Mountains are separated from the Coast Mountains by major faults in the Fraser River Canyon (Fig. 2), and extend from Washington State northward to Lytton, at the north end of the Fraser Canyon. They are made up of sedimentary and volcanic rocks of Mesozoic and Paleozoic age, many of which have been highly deformed and metamorphosed. These rocks are cut by Mesozoic and Tertiary plutonic rocks.

The most prominent features of the Cascade Mountains in Washington are major volcanoes, one of which, Mt. Baker, is visible from Vancouver. All of the Cascade volcanoes are young and most are potentially active.

Mt. Baker is the northernmost volcano in the conterminous U.S. It is a 10,800-foot stratovolcano whose main cone was constructed during the Pleistocene, but before the Fraser glaciation (about 25,000 years ago). After the main ice sheet departed, eruptions took place at 6750 and 6525 BC, and 1100 AD. Possibly 9 eruptions took place in the 1800s. The 1843 eruption resulted in a major fish kill in Baker River, forest fires and widespread volcanic ash. Steaming from Sherman Crater on the south flank was common until about 1950, and still occurs from time to time.

Burrard Inlet

Burrard Inlet is part of the Indian Arm fjord. It is tempting to speculate that Burrard Inlet was once a branch of the Fraser River. However, there is no geologic evidence to support this and it appears that the present course of the Fraser River was established after the final retreat of the last ice sheets about 11,000 years ago. Burrard Inlet is, in fact, the valley of Indian River now drowned up to the north end of Indian Arm. Seymour River, Lynn Creek and Capilano River were once tributaries of Indian River. Capilano River (Cover) presently deposits sediment at the Inlet's entrance and until the Capilano dam was built in 1951, frequent dredging was needed to remove this sediment so that the channel could be kept open for ship traffic. Periodic dredging is still required.

Water Supply For Vancouver

Unlike most of the rest of the Fraser Lowland which depends on groundwater, the City of Vancouver relies on rivers, namely, the Capilano and Seymour Rivers flowing into Burrard Inlet, and the Coquitlam River flowing into the Fraser River. Their watersheds are closed to the public. The high annual precipitation in the Coast Mountains, normally in excess of 2500 mm, ensures a good runoff, which is stored for use in reservoirs created by dams on all three rivers. Several pipelines under Burrard Inlet and the Fraser River carry the water to the city.

The rivers flow in U-shaped valleys developed in granitic and metamorphic rocks of the Coast Plutonic Complex, which in most places is covered by up to several metres of soil, glacial deposits, and talus. This mantle maintains a 60- to 80-year old second-growth conifer forest. Because these rocks are not easily dissolved or otherwise weathered, the water at the source is of outstanding quality, the hardness rarely exceeding 10 parts per million (ppm), with total dissolved solids at 23 ppm. Chlorination is the only treatment required. On occasions, mainly during the period from October to April, the water may become dirty looking as a result of landslides, which increase the silt content of the water in the reservoirs. However, it remains perfectly safe to drink.

FIELD TRIP STOPS

A detailed road log appears on page 17 at the end of the general descriptions of the five stops.

Stop 1. Point Grey sea cliffs

A short distance west of the exit from the parking lot at the Museum of Anthropology is the head of the trail that leads down to the beach. The sea cliffs here at the west end of the University of British Columbia expose 62 m of sand, overlain by thin till, glacio-marine mud with pebbles, and minor beach gravel (Fig. 4). Most of the geological information about these strata that follows is taken from Clague (1977). A small, poor outcrop of till can be seen at the head of the steep part of the trail. The scenario here is that the Quadra sand was deposited in front of the advancing ice of the Fraser Glaciation, and this till at the top of the section announces that the ice had arrived. The time was about 22,000 years ago, and it concluded what is called the Olympia nonglacial interval.

The Quadra Sand, below the till, is found in many places around the margins of the Georgia Basin. The unit here consists of well-sorted sand and minor gravel and silt. Silt is restricted to the lower 18 m, and forms horizontal beds and laminae within the sand. Plant remains are common in this lower part of the unit and have yielded four radiocarbon dates ranging from 24,400 +/- 900 to 26,100 +/- 320 years BP (Fig. 4).

At the shore turn right (north) and follow the beach for about a half kilometre to another trail that ends at NW Marine Dr. a short distance below the junction with Chancellor Blvd.

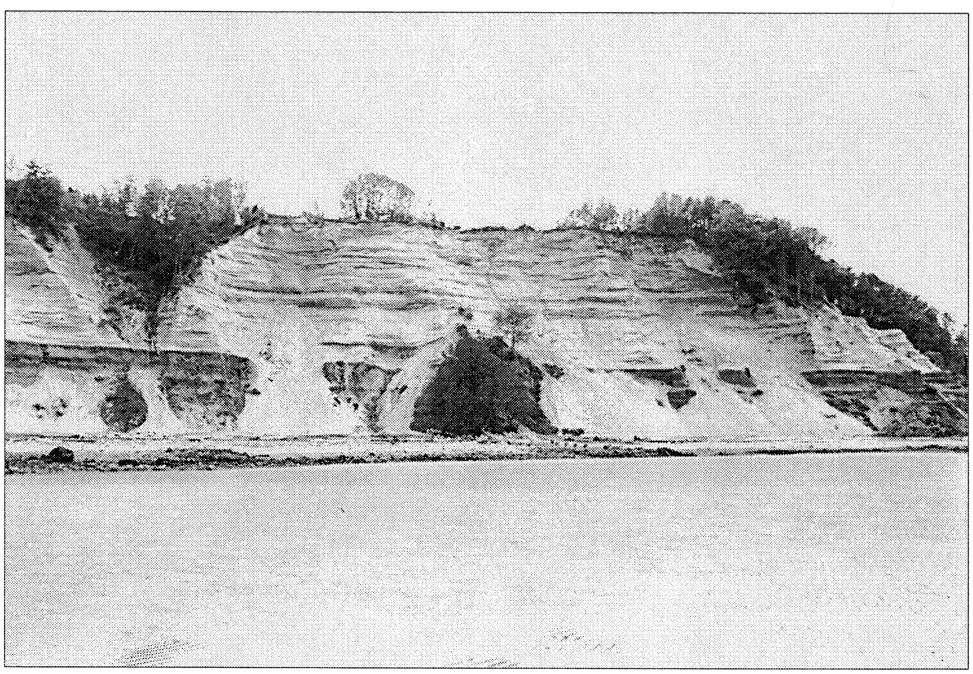


FIGURE 4. Quadra sand capped by till at west end of Point Grey. The lower darker part of the unit contains silt layers with plant remains ranging from 26.1 ka BP at the bottom to 24.4 ka BP at the top. Photo by J.J. Clague, 1976.

The sand is horizontally laminated and cross-stratified, the result of deposition in the channels of braided rivers. Detailed analysis of the axial orientation of crossbeds in troughs in the Quadra Sand was undertaken here and elsewhere in the Strait of Georgia region to determine flow directions during deposition of the unit. The study showed that the Quadra Sand was deposited by streams flowing of Georgia region to determine flow directions during deposition of the unit. The study showed that the Quadra Sand was deposited by streams flowing mainly from the west, northwest, and north, and that the flow direction shifted repeatedly during deposition. Flow indicators from the lower, silt-rich portion of the unit are more southerly directed than those from the upper part.

The lower part of the unit at Point Grey also differs mineralogically from the upper part. Sediment below 18 m elevation contains abundant volcanic rock fragments (including glass), and hypersthene is the dominant heavy mineral. The provenance of this volcanic detritus is the Mount Garibaldi area about 70 km north of Vancouver. In contrast, sand above 18 m elevation consists almost entirely of feldspar (mainly plagioclase), quartz, and minor lithic fragments derived from granitic rocks of the Coast Mountains; the main heavy minerals are hornblende, biotite, chlorite, magnetite, and hematite. These observations indicate that the lower part of Quadra Sand at this site was deposited by streams flowing south-southwest down Howe Sound from the Mount Garibaldi area, and that about 24,500 years ago (the age of the uppermost silt bed) the flow changed towards the east and southeast, bringing granitic detritus down the axis of the Georgia Depression.

Mathewes (1979) examined plant microfossils from stringers of peaty silt 16 to 18 m above sea level at this site and reported that the non-arboreal pollen (especially *Cyperaceae* and *Gramineae*) are dominant, and support the geologic evidence that the Quadra Sand was deposited in a braided-river environment under a cooler macroclimate than exists at present. High concentrations of pollen in the peat stringers, however, are incompatible with tundra conditions. Rather, the climate 24,500 years ago seems to have been comparable to that of higher forests, or possibly subalpine areas of the contemporary coast mountains.

The sand is younger south of the U.S. border, and was not overridden by the Puget Lobe of the Cordilleran ice sheet in the Seattle area for another 6000 years (about 16 ka BP). The ice there began retreating about 14,000 years ago, but Point Grey was not free of ice until 12,900 years ago, and Langley (about 50 km east of Vancouver), more than a thousand years later.

was responsible for the volcanic rocks here at Little Mountain, as well as the dykes at Stop 5 in Stanley Park and the flow (or sill) on Sentinel Hill in West Vancouver.

Little Mountain is a prominence because the augite-olivine basalt is more resistant to weathering than the surrounding sandstone. It forms an elliptical outcrop about 1800 feet along the northwest-trending axis and 1200 feet across the short axis. The best exposures are in an old quarry (Fig. 5) which was operated in the early part of the century, to provide road material for the Gastown area. The quarry has since been converted into an impressive garden. The adjacent Bloedel Conservatory, a geodesic dome, houses more-tropical plants.

When the quarry was less covered by vegetation, a GSC geologist, Johnston (1923), examined it and noted that the columns converged in places as if cooled mainly from a restricted or an uneven surface. In most places the basalt exhibits fairly well developed columns which range from about 20 cm to a metre across; some are about 10 m high. Their attitude generally suggests a dip to the south, but they are not consistent. No conclusive evidence has been found to establish whether the body is a flow or a sill. Johnston thought that the finely vesicular character and converging columns suggested a sill rather than a flow. The upper surface, however, had been rounded by glaciation and a more vesicular and scoriaceous top could have easily been removed by coeval erosional processes or the later glaciation. Wooton, in his B.Sc. thesis at UBC in 1959, recorded

FIGURE 5. Oligocene (34 m.y.) columnar basalt in Queen Elizabeth Park

Stop 2. Little Mountain, Queen Elizabeth Park

Most volcanic rocks in the Vancouver area are remnants of Cascade volcanism. The oldest examples, with an age of about 50 m.y., are not seen at the surface but were found by drilling beneath the city in the northern part of False Creek flats (cover). About 180 m below sea level are thin layers of basalt within Eocene sandstone and shale beds. They are notably vesicular, indicating surface lava flows.

The next period of Cascade volcanism took place between 31 and 34 m.y. ago. That particular event that rare vesicles are as large as 2 cm across and have been filled with amygdaloidal, radiating natrolite. He also mentioned a basalt dyke (feeder to the flow or sill?), by a house on the west side of the park, had picked up pebbles and much sand, evidently derived from the underlying sandstone and conglomerate. He did not state an opinion as to whether the main basaltic body was a sill or a dyke.

A radiogenic age determination by R.L. Armstrong (U.B.C.) in 1980 yielded a cooling age for the basalt of 34.3 ± 1.2 m.y. which places the extrusion in the Early Oligocene. Although lithologically similar, this rock appears to be slightly older than the dyke at Prospect Point (Stop 5) in Stanley Park.

To the north from the viewpoint, downtown Vancouver can be seen, as well as Burrard Inlet and the North Shore mountains. In recent years this view has been increasingly obscured by trees.

Lions Gate Bridge

We cannot stop here, but we cross it twice and it has an interesting history, as well as being the most contentious structure in the Vancouver area (Fig. 6). It is both a symbol of Vancouver and a vital transportation link, although clearly inadequate for the latter role. The following description is part of a newspaper article for the general public by this writer, which was published a couple of years ago.

"One of the largest and most interesting highway structures in Canada is now nearing completion in Vancouver, B.C." So began an article by P.L. Prately in the October 1938 issue of the *Canadian Engineer* in reference to the graceful bridge spanning the entrance to Vancouver harbour. Now in its 63rd year, the bridge has long outlived the *Canadian Engineer* (last issue 1939), and nearly all of the men who built it.

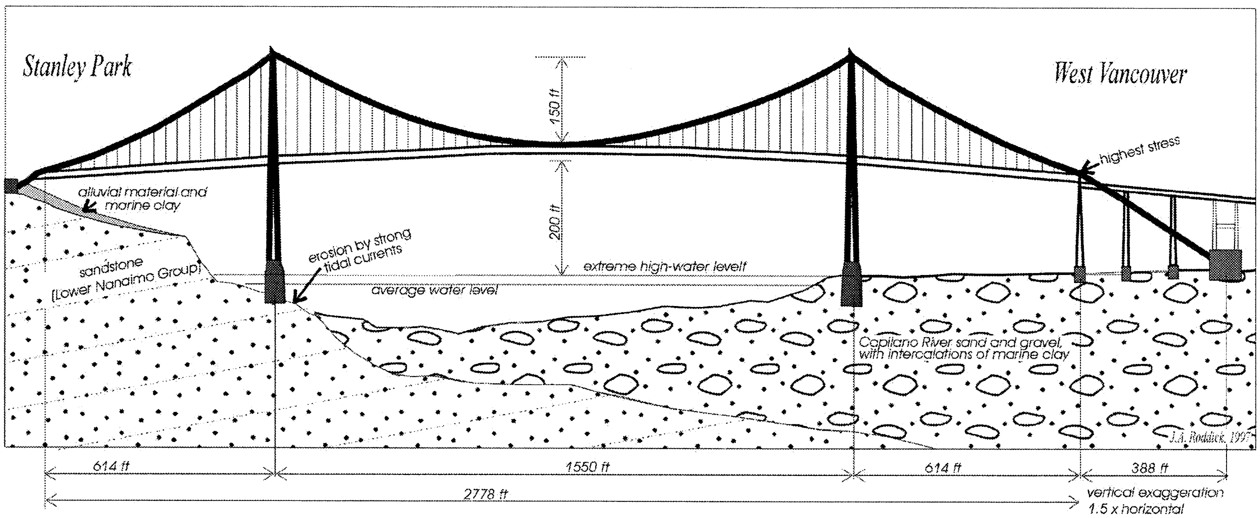


FIGURE 6. Lion's Gate Bridge: suspension span and part of the north viaduct.

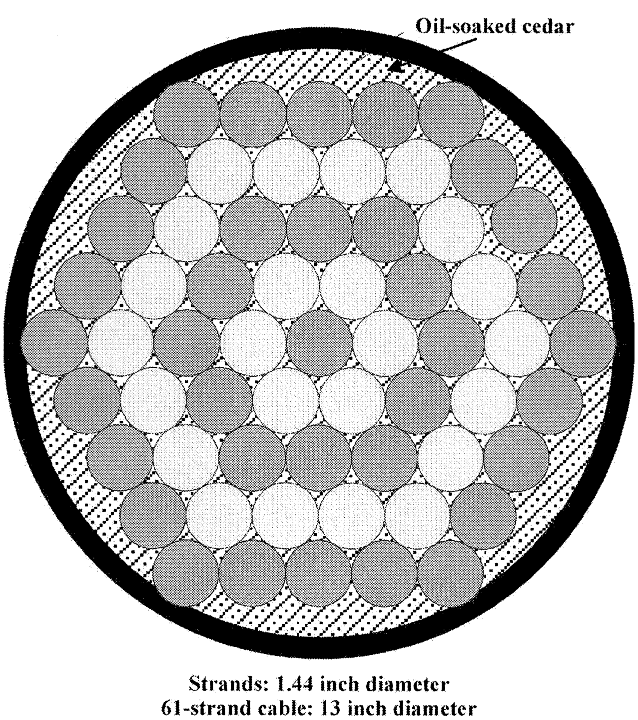


FIGURE 7. Cross-section of main cable, Lions Gate Bridge

The first proposal to construct a bridge from Prospect Point to the then remote North Shore was made in 1909 by the Burrard Wire Cable Bridge Company. Because of its impact on Stanley Park, it was rejected outright. In 1926 a proposal from the Guinness company of Britain got the same reception. In 1933, however, the Parks Board approved a second proposal by Guinness. The plan had not been improved much, but the economic climate had switched from the optimism of the Roaring Twenties to the desperation of the Great Depression. It was now no big deal that it meant slashing in two the already famous Stanley Park; in fact, only one member of the board voted against the bridge; and, in truth, Park usage did not decline after the bridge rose. To the contrary, like a famous New York restaurant, it has "become so crowded that no one goes there anymore."

A month before the opening of the Golden Gate Bridge to the south in May 1937, construction began on the substructure of the Lions Gate Bridge. Although the bridge looks symmetrical the main piers differ. The south pier (at Prospect Point) consists of two cylindrical caissons 48 feet (14.6 m) in diameter seated about 30 feet (9.1 m) below low water on the Late Cretaceous (80 m.y.) Lower Nanaimo Group sandstone. It is well exposed along the seawall between Prospect Point and Siwash Rock, which will be visited on the last stop of this excursion. The same strata near Nanaimo on Vancouver Island contained economic coal beds, but near Stanley Park only rare thin beds are present. Nevertheless, those coal seams gave rise to the names Coal Peninsula (now the West End and Stanley Park), Coal Basin (now Lost Lagoon) and Coal Harbour.

The geologic situation is much different for the north pier where bedrock is deeply covered by deltaic deposits of the Capilano River and intercalated marine clay. It was not feasible to seat the north pier on bedrock there, but they did dredge down to 65 feet below low-water level and seated a single large caisson (117 by 48 feet (35.7 by 14.6 m)), presumably on gravel. Newspaper reports and photographs of the construction, however, refer only to blue clay with scattered cobbles (typical of the marine clay of the region). Depending on the proportions of clay to gravel, the foundation may be either good enough or, perhaps, dog meat for the next Richter 8-plus shakeup.

Meanwhile the Capilano River, dammed since 1951, continues depositing sand and gravel into the First Narrows, and to crowd the tidal currents against the foundation of the south pier. To slow erosion there a protective barrier of steel and riprap has been installed.

In February 1938, after 10 months of work on the substructure (about half of the total construction time), the first steel went up. The engineer-in-charge and his group placed their small change, 35 cents, below the foot steel, where it was to remain until the bridge is dismantled and constitute 'seed' money for the replacement. Part of crew consisted of experienced workers from the more massive San Francisco project. The foreman of the steelworkers, in fact, had driven the first rivet of the Golden Gate Bridge in 1934.

In late April the two towers reached their full height of 364 feet (111 m) above high water, and at high slack in the early morning of May 2nd, 1938, the catwalk cables were dragged by tugboat and barge across the narrows. The cables were immediately hoisted to the tops of the towers. The waterway was closed only for two short periods during the entire construction of the Lions Gate Bridge.

The main cables are 13 inches (33 cm) in diameter (Fig. 7), each being. made up of 61 1.44-inch (3.66-cm) strands. From large spools, each strand was hauled along the catwalks and over the towers. The strands were then gathered in layers of 5,6,7,8,9,8,7,6,5 to form a hexagon (5 cables per edge), which was then filled out to a circular section using oil-soaked, cedar fillers and finally wrapped with a continuous serving of soft galvanized steel wire. Recent testing shows that this treatment successfully prevented any rust-disintegration of the cable strands. Curiously, for less than 100 cables, only five clusters can make a hexagon, 7, 19, 37, 61 and 91. Both the Lions Gate and the Golden Gate use the 61-cluster, but the latter used much thicker strands (3-inch (7.6-cm)).

The cable anchors consist of large concrete blocks, which stay the structure mainly by their weight rather than by ground attachment. Even the south anchors at Prospect Point are not in the bedrock but in pits excavated in alluvial, glacial, and marine clay deposits. Collapse of one of those pits caused the only fatality during construction of the bridge.

When the bridge opened on November 12th, 1938, tolls were set at 25 cents for car and driver, five cents for passengers, pedestrians, bicycle and rider, and 25 cents for a horse-drawn vehicle. The tolls remained until 1963.

Although the Lions Gate Bridge suffers the natural defects of old age, they are confined mainly to the roadway. The cables and towers are presently okay, in spite of the rusted rivet heads found on the south pier. Some of the rivets have been replaced by strong bolts. As might be expected neither tower is now vertical, because of cable stretch, but there has been no change since 1970.

The susceptibility of the bridge to a large earthquake is a complex matter. The bridge has two main parts, the suspension span, and a long rigid viaduct on the North Shore. Because the natural resonance of the suspension has such a long wavelength, a major quake would scarcely be felt in the midspan segment, unless the cable anchors were dislodged, which is quite possible.

The rigid viaduct at the north end of the bridge, however, is at a much greater risk. It makes up 45% of the length of the Lions Gate Bridge and is the most susceptible to seismic damage — in fact major collapse is possible, although not likely. It would probably survive the long-expected Richter-8 quake. Although the Capilano River deltaic deposits would certainly amplify the shaking, liquefaction is unlikely unless thick lenses of sand, silt and marine clay are present within the gravel. For added insurance seismic upgrading in 2000 involved, among other measures, driving many piles around the bases of the viaduct supports.

Even major displacement on one of the several west-trending faults, that were identified in the tunnel that carries water across the First Narrows from Cleveland Lake, would not cause the bridge to col-

TABLE 1. Lions Gate and Golden Gate bridges

Feature	Lions Gate	Golden Gate
Total length, including approaches	5820 ft = 1774 m	8981 ft = 2737 m
Length of suspended span plus side spans	2778 ft = 847 m	6450 ft = 1966 m
Length of main span	1550 ft = 472 m	4200 ft = 1280 m
Width of roadway between curbs	29 ft = 9 m	62 ft = 19 m
Number of lanes	3	6
Width of each sidewalk	4 ft = 1.2 m	10 ft = 3 m
Clearance above mean high water	209 ft = 64 m	220 ft = 67 m
Deepest foundation beneath mean low water	65 ft = 20 m	110 ft = 34 m
Height of tower above mean sea level	364 ft = 111 m	746 ft = 227 m
Diameter of main cable including wrapping	13 in (33 cm)	36 in (91 cm)
Vehicle crossings; annual	about 25 million	41,267,000
Time of construction	April 1937 - Nov. 1938	Jan 1933 - May 1937
Number of strands in each cable	61	61
Current toll	none	$3 southbound only
Annual toll revenue	0	$58,306,000
Monthly toll revenue	0	$4,859,000
Daily toll revenue	0	$160,000
Total toll revenue since opening	?	$834,810,000
Total cost of bridge and approaches	about $5 million	about $24 million
Total fatalities during construction	1	11

lapse, although the resulting distortion would be bad financial news.

When it comes to length, the Lions Gate Bridge is in the minor leagues. Its 472-m main span does not approach that of the Golden Gate Bridge (1280 m), which was the longest span in the world from its opening in 1937 until New York City's Verrazano Narrows Bridge, with a suspended span of 1298 m, was completed in 1964. The Verrazano held the record as the longest single span bridge until 1981, when the Humber Bridge in England was opened for traffic. It held a couple of records, one for its main suspended span of 1410 m, and one for being the greatest financial disaster in bridge history. The Humber opened with a debt of £151 million, but because the population growth expected for South Humberside never materialized, the tolls collected failed to cover even the interest charges. By 1992 the debt had grown to £439 million. In 1996 a special act of Parliament gave relief by writing off part of the debt.

Since then, the main-span record passed first in 1997 to the Great Belt East Bridge (1624 m) in Denmark, and then in 1998 to the Akashi-Kaikyo Bridge (1990 m) in Japan, joining the main island of Honshu to Shikoku Island. The Akashi-Kaikyo Bridge is of considerable engineering interest, not only because of its length, but also because during construction the distance between the towers increased. It happened on January 17th 1995 when the site was hit by the Great Hanshin Earthquake (Richter 7.2) which was centred in Akashi Narrows directly beneath the bridge. The towers, however, survived without structural damage and it opened on schedule in 1998.

Stop 3. Caulfeild

Most of the coastline in West Vancouver consists of rounded, wave-washed granite outcrops, but these have deteriorated over the years owing to big-city grunge and encroaching vegetation. Here at Caulfeild Park (the unconventional spelling is correct; it was derived from the name of the original settler) the small scenic headland is underlain by a large inclusion of amphibolitic gneiss (exposed for about 900 m) immersed in and cut by granodiorite (varying to quartz diorite), pegmatite and aplite (Fig. 8). In its general complexity it is typical of many areas on the west side of the Coast Plutonic Complex, but in its particular combination of diverse features this area is possibly unique. As one of the more interesting sites of geological controversy in the Vancouver area, it has since been visited by countless U.B.C. students and many practicing geologists. It was first studied in considerable detail in 1930 by Phemister (1945), a visiting professor at U.B.C. from Aberdeen, Scotland. His report was delayed until 1945 by his other duties and the war.

west. The gneiss is cut by dykes of aplite, pegmatite, granodiorite and andesite (intermediate volcanic rock). Discontinuous layers of feldspathization and,

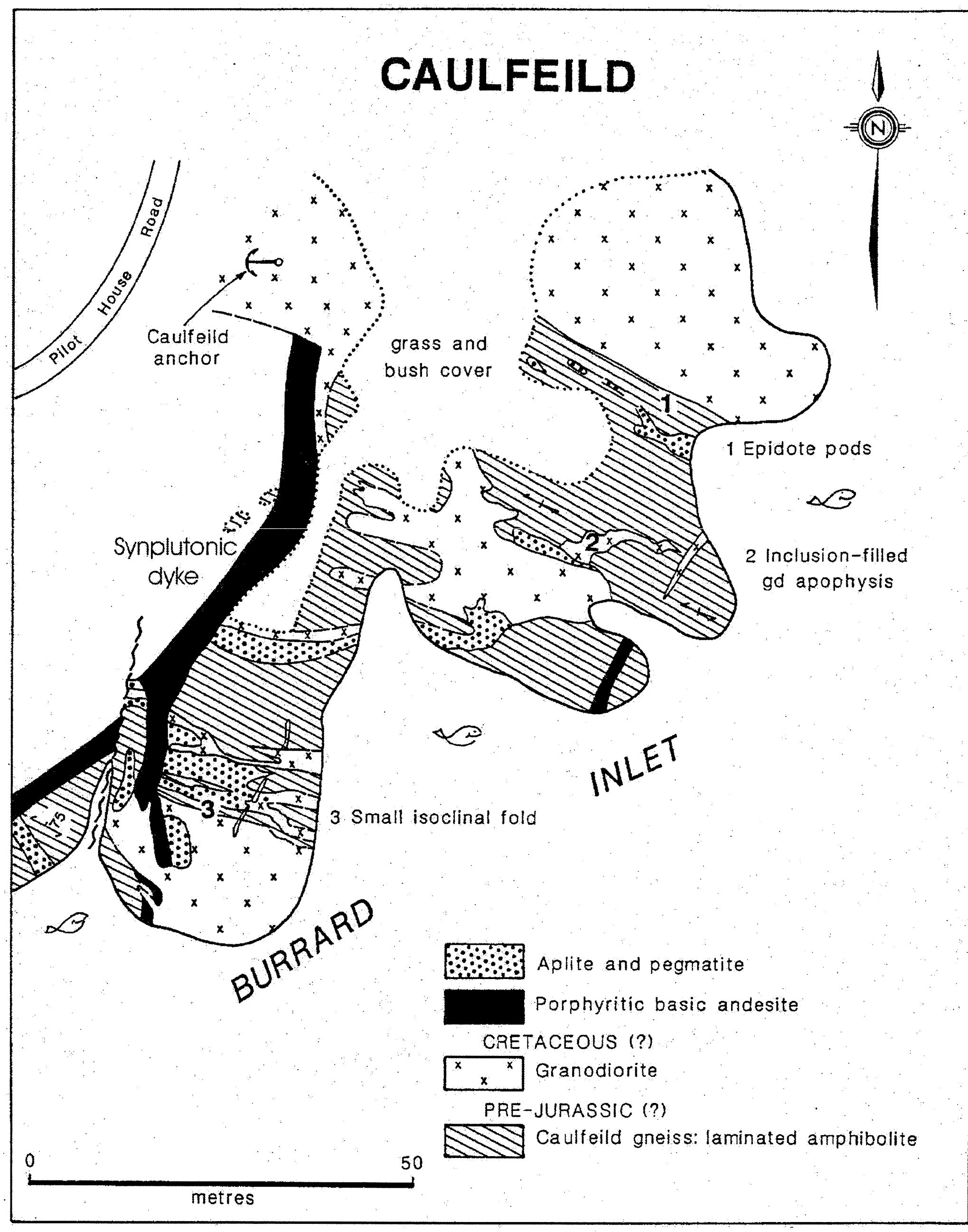


FIGURE 8. Geology of Caulfeild headland. After McTaggart, Greenwood and Read, 1972.

The gneiss is layered (near vertically) and streaky. The dark layers (amphibolite) and consist mainly of plagioclase feldspar and hornblende, whereas the broader light-coloured granitoid-looking areas consist mainly of a slightly more sodic plagioclase, quartz and minor K-feldspar. Phemister concluded that the gneiss was derived from mafic flows, tuffs and breccias, possibly correlative with those on Bowen Island to the west.

The Caulfeild gneiss is lithologically similar to and structurally parallel with that on Mount Strachan to the northeast and at Horseshoe Bay to the north-

in places, irregularly-shaped pegmatite masses can be seen in the gneiss. The inclusions in the granodiorite are irregularly distributed and locally abundant. They show all stages of conversion from amphibolite and gneiss to granodiorite.

The origin of granite remains, in my opinion, a major unsettled problem in geology. It is too complex a problem to deal with here, except in the broadest terms. One aspect is shown by the synplutonic dyke on the west side of the Caulfeild outcrop. Phemister thought that it was actually prebatholithic. He noted that although the dyke is bordered by granodiorite, it is also, in places, cut by the granodiorite. He further pointed out that here and there remnants of Caulfeild gneiss seem to be adhering to it. He postulated that the dyke had originally intruded the gneiss, but most of that had later been removed by delicate magmatic stoping during intrusion of the granodiorite. Magmatic stoping is a popular concept (Daly, 1914) by which magma advances by detaching and engulfing pieces of country rock, which subsequently sink downward. Phemister

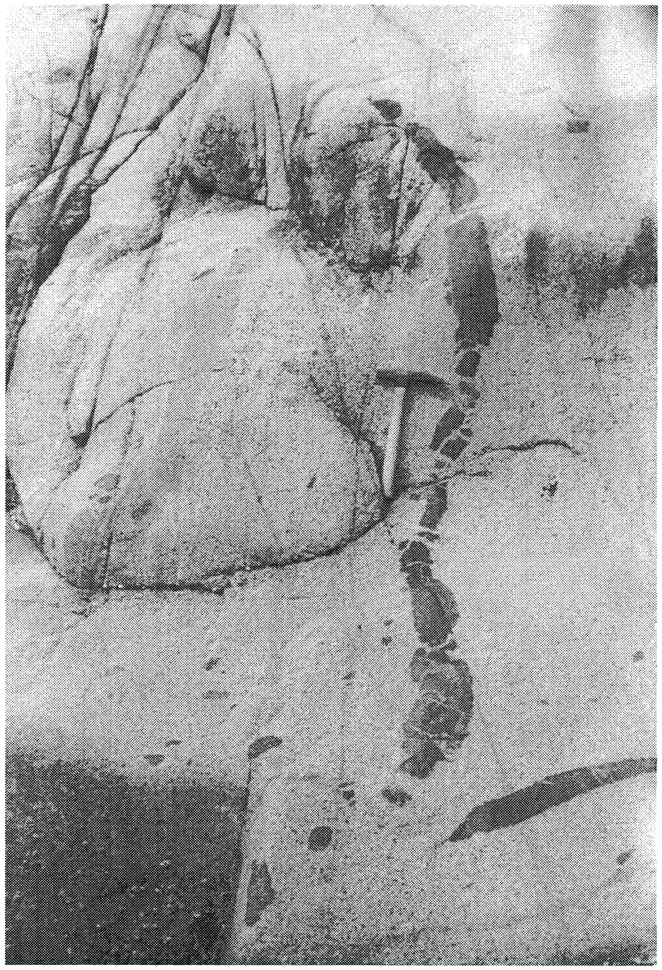


FIGURE 9. Synplutonic dyke near Gorge Harbour, Cortes Island. The fainter dyke on the left has been heavily feldspathized, and is apparently the older of the two.

favoured a magmatic origin for the granodiorite based mainly on dilation-offset shown by granodiorite dykes, rotated slabby xenoliths, and inclusions of different lithologies in close proximity to one another. He maintained that had the granodiorite formed by granitization, such features could not exist. H.H. Read disagreed (Phemister, 1945, Discussion).

Phemister was unable to attend the meeting of the Geological Society of London in 1945 where his Caulfeild paper was scheduled. Read, the most well known of British granite geologists of his time and a good friend of Phemister, agreed to present Phemister's paper although he disagreed with it. Read was a great and very humorous lecturer, and he presented Phemister's paper on these rocks, probably better than Phemister, himself, could have done.

Most geologists were and still are magmatists in that they consider granites to be mere products of slow-cooling magmas which, had they reached the surface, would have produced common volcanic rock. Read, on the other hand, was 'an old soak' as he often said, that is, a granitizer, who maintained that granites formed at below melting temperatures by metamorphism and metasomatism (that is, by passive mineralogical and textural evolution).

"Read stated that he had endeavoured to present a fair and adequate summary of Professor Phemister's long and detailed paper and now he wished to direct attention to certain features of it which he believed were fundamentally wrong. ....Professor Read had received the impression that time and again the author had been on the brink of loosing a thundering broadside against the magmatists, but, remembering his revered instructors just in time, he had withheld his hand; instead of the thunder, there was and uncomfortable silence." (Phemister, 1945, Discussion).

Read maintained that Phemister's 'pre-batholithic' dyke at Caulfeild could have been more easily preserved if the original country rock (i.e., the gneiss) had been removed by selective replacement.

The magmatic view is overwhelmingly predominant in the geological community of today, but after forty years of working on granites in the Coast Mountains and elsewhere around the Circum-Pacific region, this writer remains on the opposing side.

As a graduate student (Cal Tech and University of Washington) in the early 1950s, the writer investigated many dykes similar to that at Caulfeild in the surrounding region and concluded that neither process could free the dykes from the country rocks. The dykes had formed from magma that obviously had filled fractures in granitic rock. Magma cannot sustain fractures; the granitic rock at that time, therefore, could not have been magmatic. Yet, the dykes had later clearly been intruded, torn apart, and feldspathized by that same granitic rock. The evidence indicated that they had intruded the granitic mass as well as its inclusions and pendants when the plutonic rock had sufficient strength to sustain fractures, yet reacted to long applied stress by very slow plastic creep which was more than sufficient to rupture bodies such as dykes and country rocks (Figs. 9, 37).

Also, the granitic rock was either in a stage of active recrystallization or became so, and then could cause the same type of crystals to form in the dykes as were forming in the granitic rock. In the subsequent GSC memoir (Roddick, 1965), the writer introduced the term 'synplutonic' for these dykes. Synplutonic dykes are an intriguing feature not only of the Coast Plutonic Complex, but also in many plutonic terranes around the world. Actually the synplutonic dyke at Caulfeild lacks many features of the better examples elsewhere in the Coast Mountains (Figs. 9, 36-39). Photomicrographs of some crystals in the Caulfeild granodiorite, as seen in thin section, are shown in Figures 27, 30, 33, and 35. For a capsulation of the granite controversy, see page 23.

Stop 4. Lookout on Cypress Bowl Road

At this stop we are directly across English Bay from U.B.C. at the end of Point Grey (Cover). When weather permits, this stop affords a good view of the Lower Mainland. Vancouver Island appears on the horizon to the south and west, separated from the mainland by Georgia Strait and the Gulf Islands. To the southeast is Mount Baker, a 10,800-foot (3300-m) andesitic stratovolcano, and farther south the Chuckanut Hills (Fig. 2) which mark the southern margin of the Fraser Lowland.

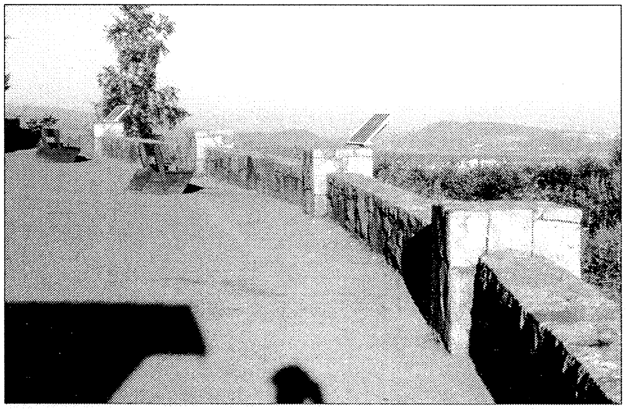


FIGURE 10. Stop 4; wall at Cypress Bowl Road Lookout, constructed of granodiorite blocks quarried from higher on the road.

The wall at this stop was constructed of blocks representative of the granodiorite and tonalite that underlie this slope of Hollyburn Ridge (Cover).

Stanley Park and Lions Gate bridge are in the foreground below. To the east, when the view is not obscured by bushes, the south-sloping, Late Cretaceous, exhumed, erosion surface can be seen (Fig. 11). This surface can be projected beneath Sentinel Hill in the near foreground (not visible in Figs. 10, 11). The southern slopes of Burnaby Mountain (Fig. 11) and Capital Hill are controlled by bedding in the conglomerate of the Eocene Kitsilano Member of the Huntingdon Formation, and is essentially parallel with the old erosion surface, which there, is perhaps a thousand metres below.

Stop 5. Prospect Point to Third Beach, Stanley Park

The path that leads down to the seawall begins between the lookout and Prospect Point Cafe. Make a couple of left turns and follow the switchback downward and beneath the bridge. Sand and gravel for use in the park is stored directly beneath the bridge. The path continues for some distance to the east before swinging back to the west. It merges with the sea wall about 100 metres east of the bridge. A new retaining wall has been built on the south side of the walkway. It consists of blocks of clean granodiorite, typical not of the main mass of the Coast Plutonic Complex, but of well-defined plutons within it. At low tide wave- and ice-cut, flattish outcrops of the Lower Nanaimo Group sandstone are visible. Follow the path westward past the south pier of the bridge to the white lighthouse.

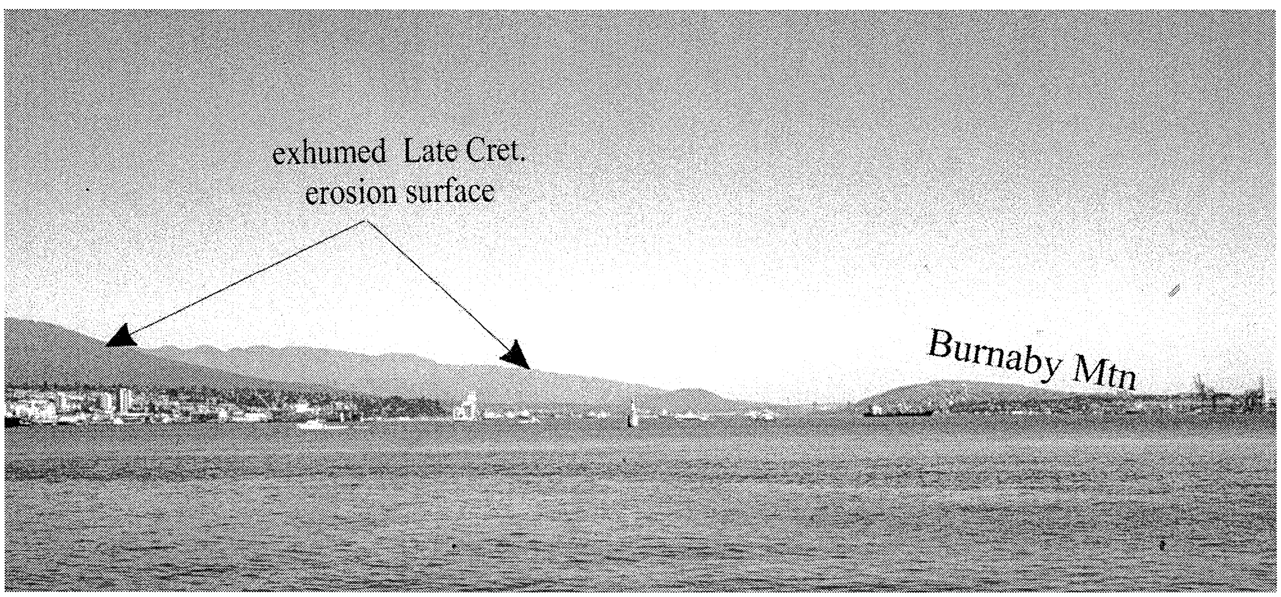


FIGURE 11. View east from Stanley Park shows the exhumed Late Cret. erosion surface developed on mainly granitic rocks of the Coast Plutonic Complex. Burnaby Mountain, on the right is formed by resistant conglomerate of the Kitsilano Member of the Huntingdon Formation. The surface is parallel with the underlying south-dipping erosion surface.

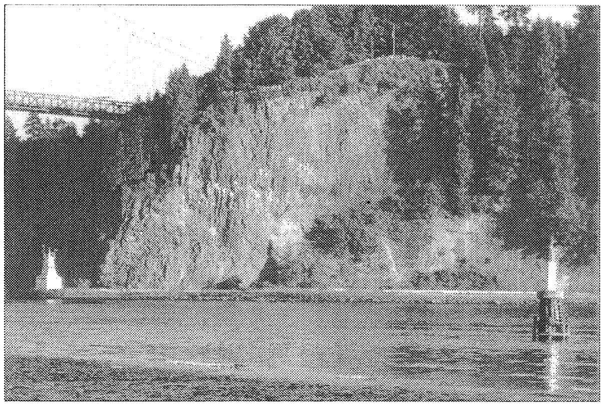


FIGURE 12. Cliff face formed by basalt dyke at Prospect Point.

This is Prospect Point. A large basaltic dyke rises from here vertically about 60 m from sea level (Fig. 12). The dyke is about 30 m wide at the base of the exposure, and extends southwest about 1200 m to just south of Siwash Rock (Fig. 13). The dyke consists of a slightly porphyritic, fine grained basalt. The phenocrysts are mainly augite, but some are olivine crystals partly altered to serpentine. The groundmass has a well-developed trachytic texture. Horizontal jointing at the base gives way to vertical columns in the upper parts, which display irregular vesicles partly filled with quartz (Johnston, 1923), possibly transitional to a flow. According to Wooton (1959), Victor Dolmage (a GSC geologist in the 1920s and a successful consultant later) told him that much volcanic rock was encountered in excavations made for the present road. The present surface outcrops by the road are, unfortunately, deeply weathered.

The upper part of the dyke and the surface extension to the southwest may be a flow or sill. If it is a flow it may be a remnant of what was once a fairly large lava field, extending perhaps from Sentinel Hill on the North Shore to Queen Elizabeth Park. This, however, is highly conjectural. The flow capping Sentinel Hill has not been dated, although Wooton concluded that it was identical in composition with the Prospect Point dyke. Although both are Oligocene the Prospect Point dyke is apparently a couple of million years younger than the basalt at Little Mountain. A whole rock K-Ar age of 31.5 + 1.1 Ma was determined by R.L. Armstrong of UBC. The Prospect Point dyke is so large it probably reached the surface.

West of the dyke the cliffs consist mainly of arkosic sandstone and fine conglomerate beds with concretionary layers and scattered pebbles, and dips gently (about 11 degrees) to the south. These rocks are about 80 m.y. old and strongly indurated. According to Rouse and others (1975) the pebbles include sandstone and siltstone from lower parts of the unit as well as granitic and metamorphic rocks. Granitic pebbles are rather rare in the exposures we will see here.

Assuming no complications by faulting, Rouse calculated that about 340 m of the strata were exposed in

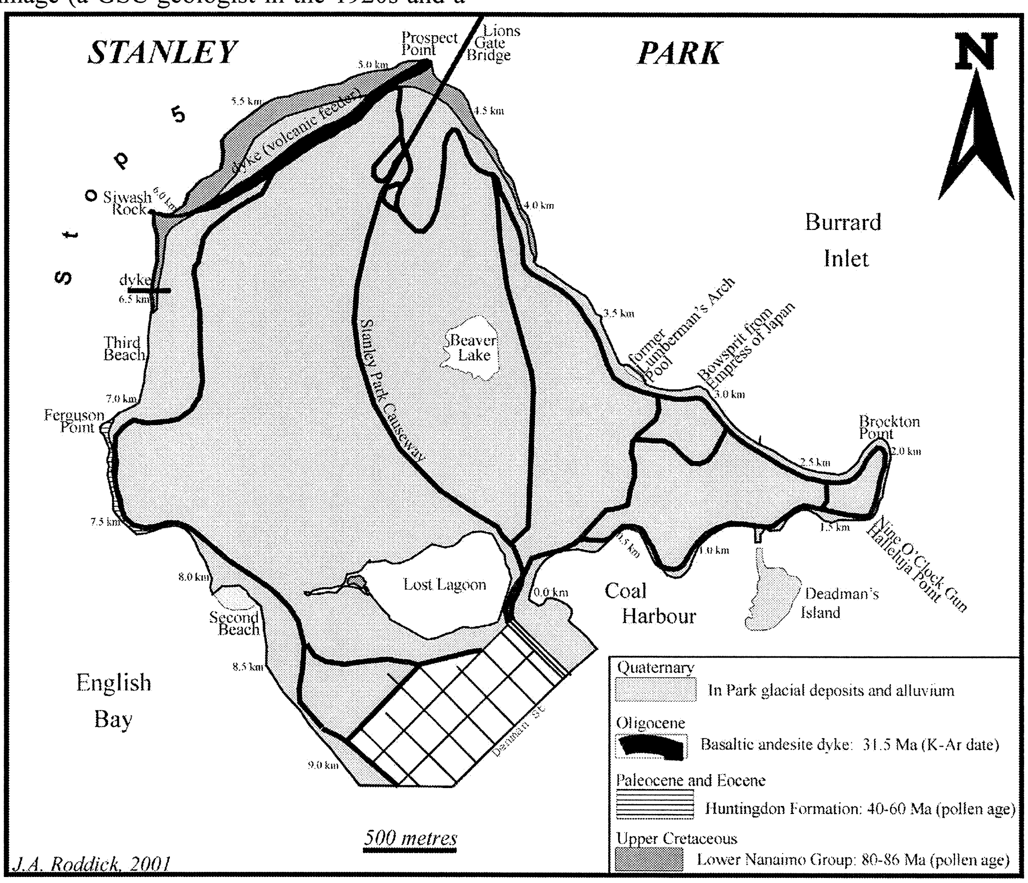


FIGURE 13. Stanley Park; the segment along the seawall from Prospect Point to Third Beach forms Stop 5.

the park, and that they lie about 500 m stratigraphically above the basal conglomerate exposed in the lower canyon of Capilano River on the North Shore. Complications, however, do exist. When the tunnel that brings water from Capilano Valley to Vancouver was driven, a number of steep faults in the Burrard Formation were encountered. They posed no serious engineering problems, and their offset and age were not established.

Across the First Narrows in Capilano River canyon, the basal conglomerate lies unconformably on a deeply weathered granitic surface that was part of a widespread peneplane that developed in the Late Cretaceous. An early study of pollen by Rouse (1962) from material from both the Capilano River canyon and Stanley Park identified flora of the Campanian Stage (84 to 71 Ma) of the Upper Cretaceous. The underlying plutonic rock in Capilano River canyon has not been dated there, but related plutonic rocks to the west have been dated at about 100 m.y. The peneplane development is, therefore, fairly well restricted to the early part of the Late Cretaceous. Rouse noted that the sediments above the erosion surface contain two distinct flora, one of Late Cretaceous and one of Eocene age.

The Cretaceous (and Paleocene) part qualifies as Lower Nanaimo Group; the Eocene part is considered to be the Kitsilano Member of the Huntingdon Formation, better defined south of the US border. Blunden (1971) estimated a total thickness of about 5000 feet (1500 m) for the early Tertiary sediments in the Vancouver area.

Between Siwash Rock and Third Beach (Fig. 13), if the tide is low, a 3 m-thick dyke can be seen trending westerly perpendicular to the shore. This dyke is strongly jointed and consists of a dense basalt, more basic than the basalt that forms most of Siwash Rock. It resembles Miocene dykes in the Coast Mountains but no isotopic dating has been done on it.

Siwash Rock appears to consist mainly of a narrower part of the large dyke at Prospect Point. The contact between the dyke near Siwash Rock and the Burrard sandstone shows only minor thermal alteration. The clay there, however, seems to be the result of increased alteration owing to the barrier effect of the dyke on groundwater.

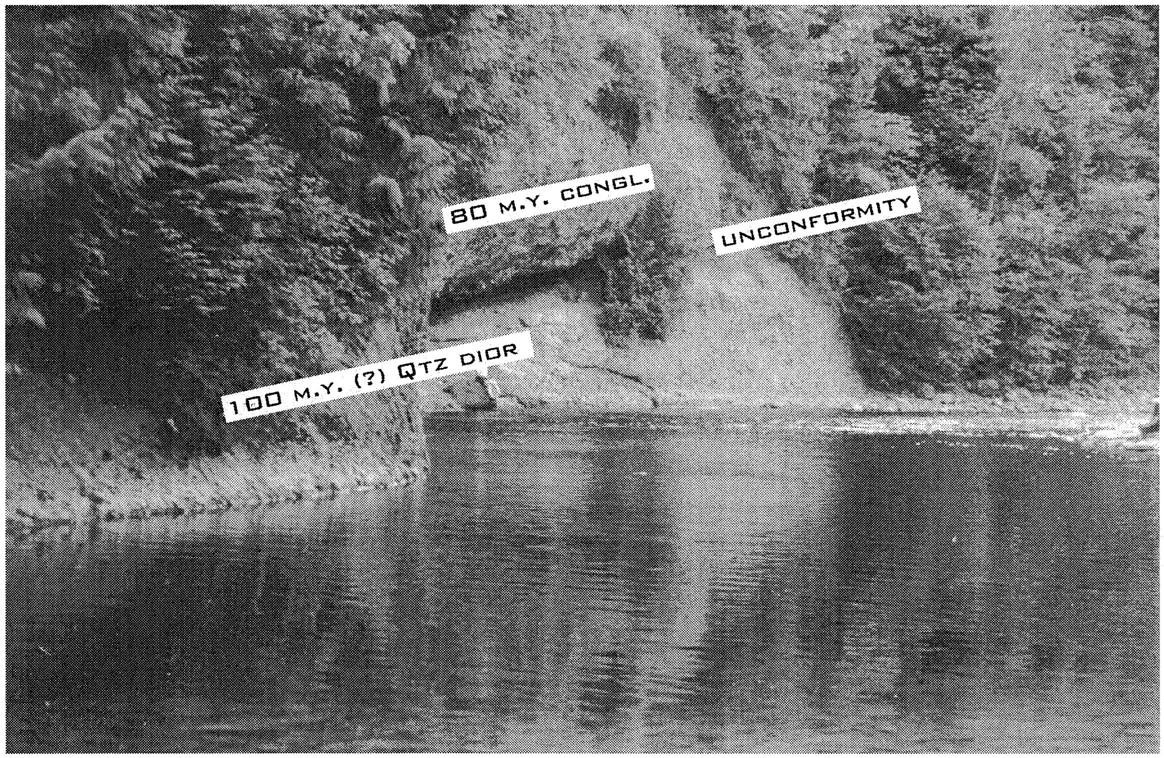


FIGURE 14. Base of Upper Cretaceous lower Nanaimo Group on weathered erosion surface of quartz diorite on west bank of Capilano River.

Capilano River Canyon section of the Lower Burrard Formation

Exposed on the west bank of the Capilano River (Fig. 14) is the base of the Late Cretaceous part of the Burrard Formation (Lions Gate Member of Rouse and others, 1975). Peter S. Mustard (then with the Geological Survey, now at Simon Fraser University) measured and described this section as follows (Figs. 15 and 16).

About 55 m of conglomerate and minor sandstone unconformably overlie undated quartz diorite of probable Early Cretaceous age. Pollen assemblages

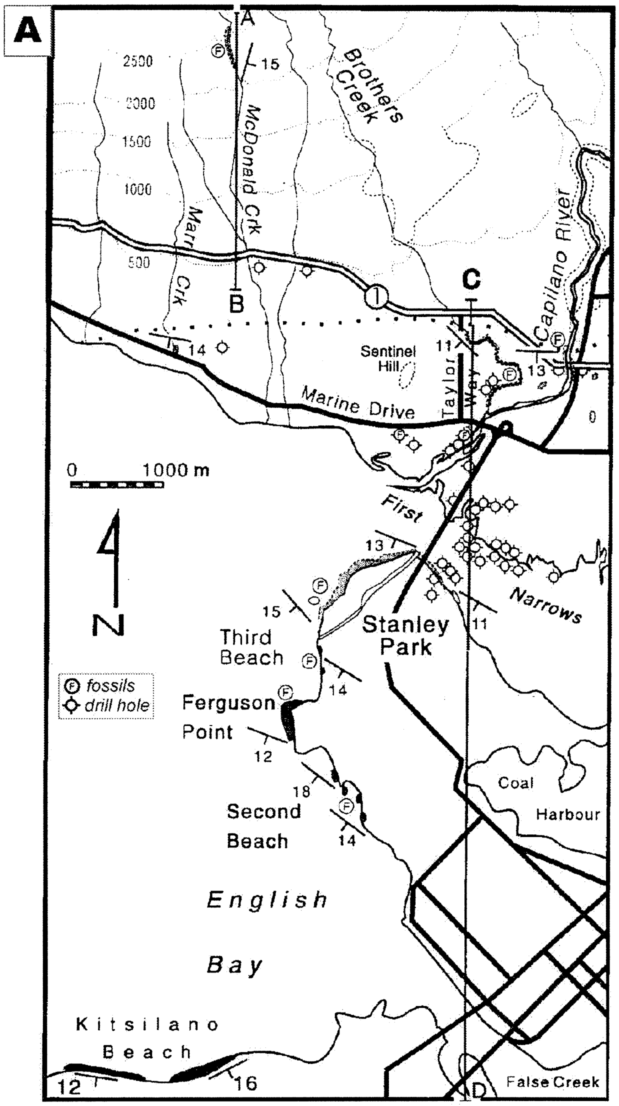


FIGURE 15A. Main outcrop areas of Upper Cretaceous strata on the North Shore and in Stanley Park.

from mudstones near here were interpreted as Campanian age (84 to 71 Ma), suggesting correlation with the Extension-Pender-Protection Formations of the Nanaimo Group on eastern Vancouver Island (Rouse et al., 1975). On the west bank of the river a few metres of the unconformity surface is exposed. The quartz diorite shows evidence of pre-Late Cretaceous exposure and weathering. It is directly overlain by poorly sorted conglomerates interpreted as alluvial fan debris flows. Higher in the section these change to interbedded arkosic sandstone and imbricated conglomerate (best seen on the east side of the canyon) which include silicified logs up to 20 cm in diameter and more than a metre long. Paleocurrents were southerly directed and the conglomerate cobbles are of local types. These are interpreted as lower fan, braided channel deposits.

Drill and tunnel construction logs suggest this exposure is overlain by about 300 to 350 m of mudstone and arkosic sandstone, all forming a fining- and thinning-upward megasequence. This succession is gradationally overlain by about 200 m of pebbly arkosic sandstone, well exposed along the seawall of Stanley Park. The overall succession is interpreted as an upward (and southward) transition from alluvial fan to fluvial floodplain environments, later buried by prograding fluvial sands.

Near Siwash Rock the sandstone cliffs have been
Near Siwash Rock the sandstone cliffs have been
conspicuously rounded by glaciation (Fig. 17). From Third Beach, at the end of this traverse, the view to the west shows glacial notches on the south slope of Black Mountain that appear to have been carved. by the overriding ice sheet (Fig. 18).

Road Log

0.0 km Set odometer at the first stop light met after taking the right hand exit from the Burrard St. Bridge (southbound). This point is the intersection of Cornwall Ave. and Cypress St. Proceed west on Cornwall Ave.

0.5 km Kitsilano Beach Park on the right. The pool, built in 1931, was hailed at the time as being the first tide-pool anywhere (maybe).

2.9 km Cornwall Ave. has changed to Point Grey Road (at Trafalgar St.). Although many of the houses we passed on our right are quite modest, they sit on the most expensive waterfront lots in the city. Turn left onto Alma Street.

3.3 km At the traffic light, note guide signs to Locarno Beach and Spanish Banks, turn right (west) onto 4th Ave.

4.4 km At the next traffic light take the diversion to the right onto NW Marine Drive. This route leads

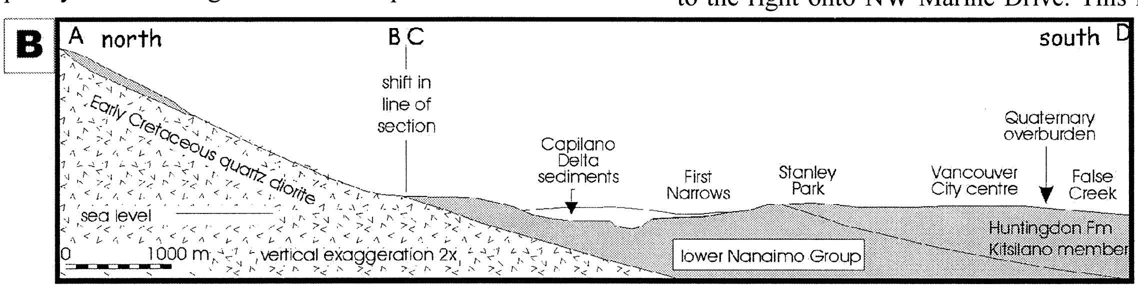


FIGURE 15B. Simplified cross-section along line A-B-C-D shown Fig. 15A (after Mustard and Rouse, 1994).

north to the shore and west along the lengthy beaches of Locarno and Spanish Banks. Good views to the north are available from any of the parking lots on the right. Clouds permitting the Tantalus Range on the west side of Howe Sound can be seen to the northwest. Closer are Bowen Island, the North Shore mountains, Stanley Park and the tall buildings of downtown Vancouver. Near shore the water is commonly brownish from Fraser River sediments which are swept around Point Grey by incoming tides.

6.5 km Spanish Banks East concession stand.

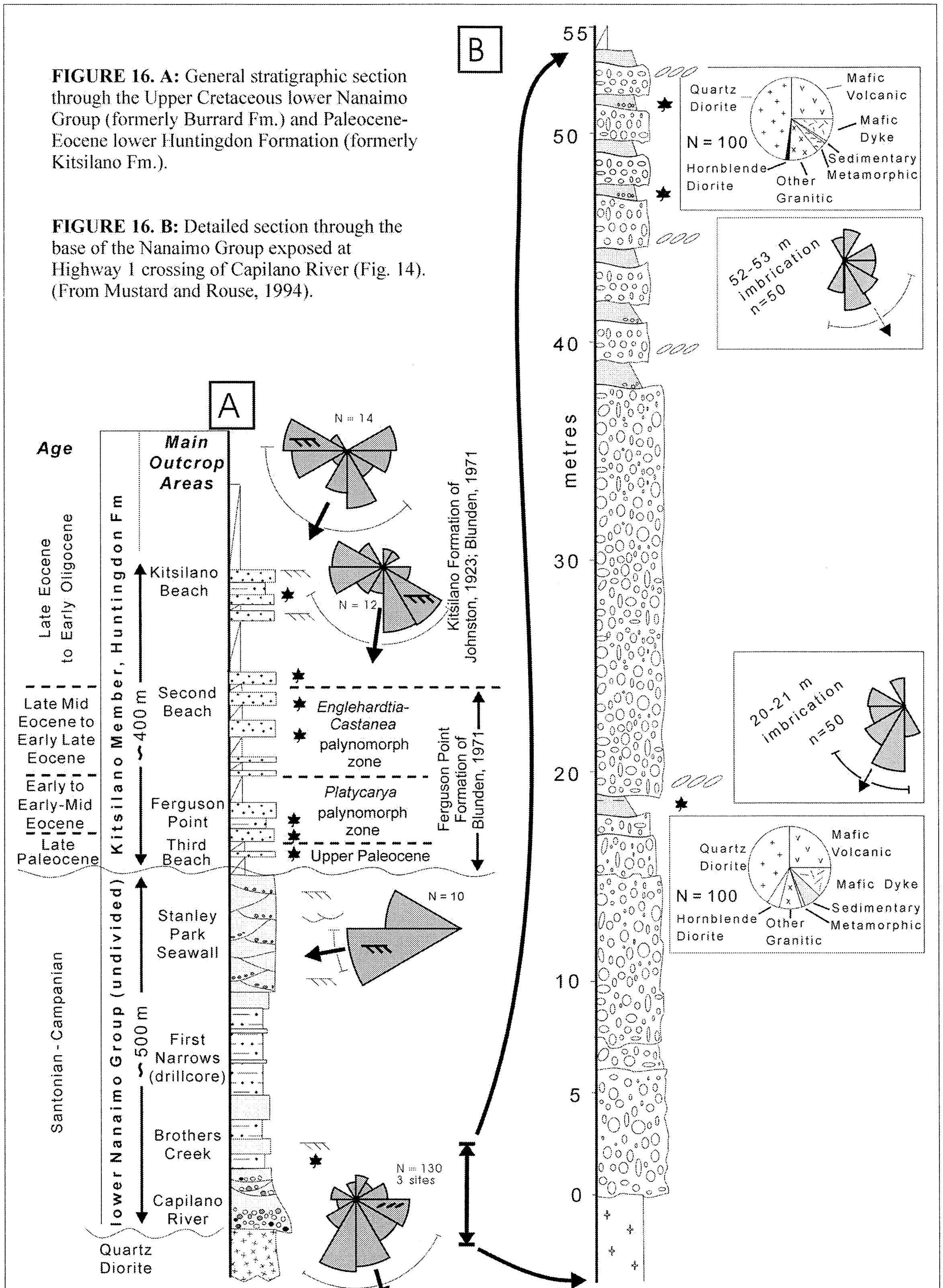


FIGURE 16. A: General stratigraphic section through the Upper Cretaceous lower Nanaimo Group (formerly Burrard Fm.) and Paleocene-Eocene lower Huntingdon Formation (formerly Kitsilano Fm.).

FIGURE 16. B: Detailed section through the base of the Nanaimo Group exposed at Highway 1 crossing of Capilano River (Fig. 14). (From Mustard and Rouse, 1994).

7.7 km The road leaves the beach and climbs the peninsula.

9.5 km Grassy areas and a few houses on the left mark the entrance into the University Endowment Lands, an enormous tract of land that is supposed to benefit the university. The stop sign at this point marks the edge of the campus. Turn right.

9.8 km Turn right again, into the parking lot (pay meter) of the Museum of Anthropology.

Stop 1. Point Grey sea cliffs

Leave the Museum of Anthropology parking lot and follow Marine Drive southward around the head of the peninsula.

11.6 km Turn left onto 16th Ave. and follow it east.

16.1 km At Dunbar Ave. turn right and travel south through the Dunbar commercial strip.

17.8 km At 33rd Ave. turn left. The route eastward on 33rd Ave. is on rocks which we won't see *en route* , but will at later stops. Vancouver is built on two low east-trending ridges separated by the estuary of False Creek. The ridges form two peninsulas, the northern between Burrard Inlet (the city's inner harbour) and the estuary of False Creek, and the southern between False Creek and the North Arm of Fraser River. We are traveling along the upper part of the southern ridge which is, within the city limits, the higher of the two. The crest is covered by a thin veneer of glacial deposits and alluvium which become thicker on the lower slopes.

22.7 km At 33rd Ave. and Cambie St. the hill in front is Little Mountain. Turn left and proceed northward.

23.2 km At 29th Ave turn right and proceed eastward.

24.2 km After passing Nat Bailey Stadium on the left, turn right onto the park access road and proceed up the hill, holding left at the intersection.

24.8 km **Stop 2. Little Mountain, Queen Elizabeth Park**

Leave the park and turn left, passing the stadium again; this time on the right. At Cambie St., turn right. From here our route will be down Cambie St. and across the bridge over False Creek.

29.8 km From the north end of the bridge near B.C. Place Stadium, proceed up Smithe St.

30.1 km Turn right at Homer (a one-way street). Pass Vancouver Public Library on the right.

30.4 km Turn left on Georgia St. and proceed westward to Stanley Park.

32.8 km Enter the causeway into Stanley Park. It is flanked on the left by Lost Lagoon, which was an extension of Coal Harbour (on the right) until 1916 when the narrow water connection was filled in for the causeway. Inside the Park, a choice must be made. The right lane continues onto the perimeter road around the Park. It is scenic and slower, but an exit to

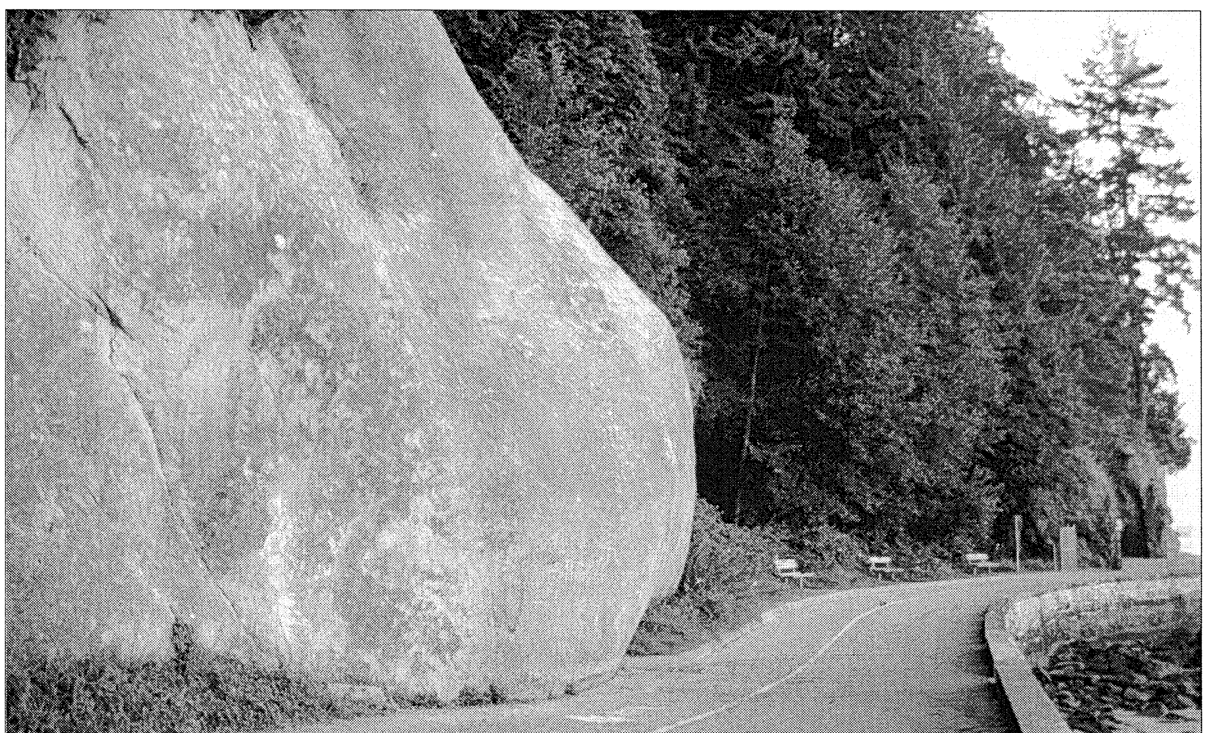


FIGURE 17. Glaciated outcrop of Lower Nanaimo Group sandstone along seawall in Stanley Park near Siwash Rock.

the causeway near the south end of Lions Gate Bridge is open except during the afternoon rush hour. The left two lanes pass into the 3-lane causeway that curves through the Park to the south end of Lions Gate Bridge (the distances assume this route).

35.0 km At the south end of Lions Gate Bridge.

36.9 km Near the north end of the bridge is the mouth of Capilano River which separates West Vancouver to the west from North Vancouver to the east. On the west bank of the Capilano River, about 2 km upstream just beyond the Upper Levels Highway bridge, and reachable only during low water, is the unconformity between Georgia Basin sediments (Nanaimo Group) and rocks of the Coast Plutonic Complex. The underlying rusty quartz diorite appears to have been weathered before Late Cretaceous time when the overlying conglomerate was deposited. The conglomerate is 15 m thick and contains clasts (pebble- to boulder-size) of granitic and metamorphic rock in a sandy matrix. It is one of the few places where the base of the Georgia Basin strata is exposed (Fig. 14). Unfortunately, access to this site is difficult and too time-consuming for this excursion.

37.7 km Now at the intersection of Marine Drive and Taylor Way, we have crossed over the Capilano River. Park Royal shopping centre, which flanks both sides of Marine Drive, opened in 1950 and is the oldest shopping mall in the country.

39.0 km Ambleside, the village's commercial area extends from 13th St. to 19th St.

41.2 km Dunderave commercial strip lies between 24th and 25th streets. Our route continues west along the mainly residential part of Marine Drive.

43.8 km At this point we are passing West Bay, a small beach out of sight to the south.

45.9 km We pass the Dept. of Fisheries and Oceans research centre at Sandy Cove. Their parking lot on the right at one time exposed a large glaciated outcrop of quartz diorite and a gneissic zone, cut by several andesitic dykes and healed synplutonic faults. It was a major stop for many field excursions in the 1960s and 1970s, including the Coast Mountains excursion of the 1972 International Geological Congress. (Roddick and Hutchison, 1972).

46.3 km The bridge across Cypress Creek.

46.4 km **Stop 3. Caulfeild.** At this point, Piccadilly St. emerges from the southwest to meet Marine Drive at a small angle and is marked by a small sign, *Anglican Church.* On the south side is a tree bearing the sign Caulfeild Trail, which comes up from the shore at this point. Parking is best found somewhere along the south side of Marine Drive. We will follow the trail along the shore to a small point known as Caulfeild Headland.

47.4 km The route from Caulfeild is east along Marine Drive. At this point turn left onto Keith Road (which is a north-trending fragment of the first east-west road along the north shore

47.8 km Keith Road passes beneath the B.C. Railway overpass.

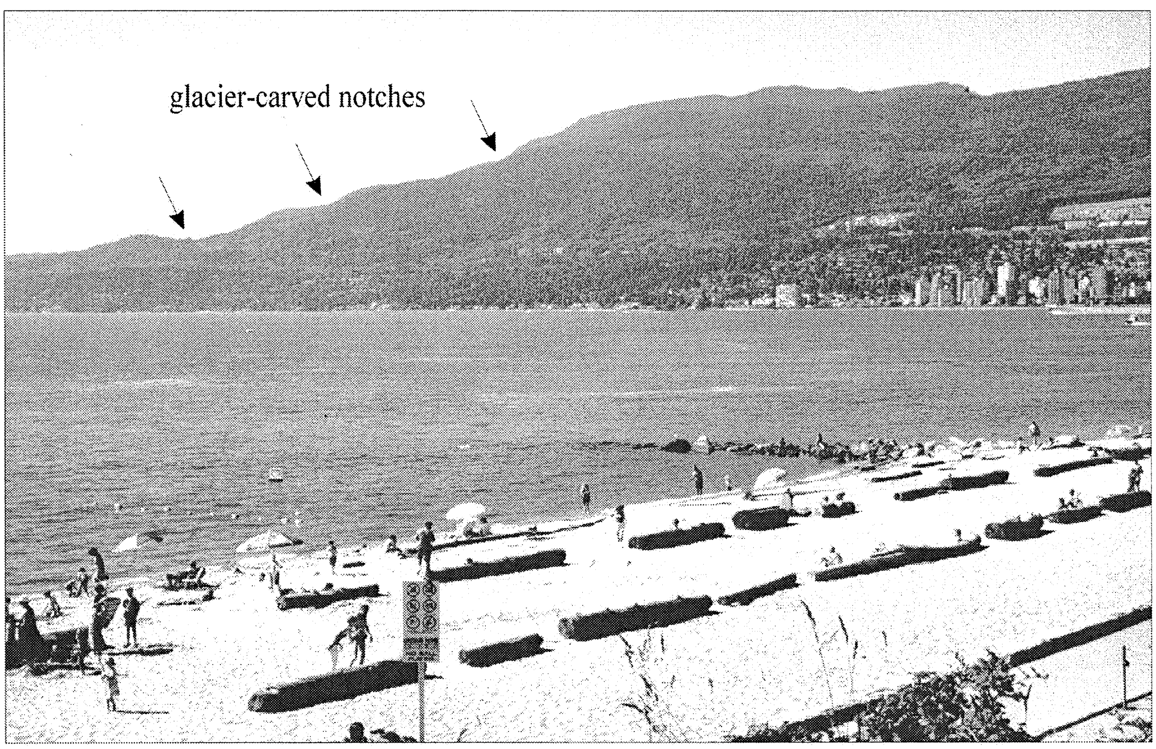


FIGURE 18. View to northwest from Third Beach in Stanley Park. South slope of Black Mountain shows notches probably carved by the ice sheet which covered the region 15,000 years ago.

48.0 km Turn right onto Willow Creek Road and continue up the hill, passing through a residential area with some substantial houses.

48.5 km Turn right onto Caulfeild Drive.

49.2 km At the stop sign in front of Caulfeild Shopping centre. Turn right and onto the access road to the Upper Levels Highway. Travel east on the highway.

50.4 km The highway passes over the Cypress Creek bridge.

53.2 km Leave the highway at Exit 8.

53.8 km Turn left onto Cypress Bowl Road, which after crossing the highway on an overpass trends to the west.

55.0 km On the right are some granodiorite outcrops of the Coast Plutonic Complex.

55.9 km West Vancouver Public Works yard on the right. The road swings around the west side of the yard and commences a long climb eastward up the side of Hollyburn Ridge.

56.8 km Along this section of road are many outcrops of the granodiorite with a glaciated surface that dips moderately to the south. The plutonic rock shows well-developed joint planes, subparallel with the present slope, believed to be the exhumed Late Cretaceous erosion surface modified by glaciation.

59.7 km Stop 4. Lookout on Cypress Bowl Road.

59.8 km Leave the parking lot and return to the highway.

66.0 km Join the highway and travel east.

69.3 km Sentinel Hill is directly ahead. It consists of Lower Nanaimo Group sandstone and conglomerate, capped by a flow of Oligocene (?) basalt.

70.1 km Leave the highway and turn south down Taylor Way.

71.6 km Taylor Way and Marine Drive, at Park Royal shopping centre. Turn left towards the Lions Gate Bridge. Cross the bridge to Stanley Park.

74.1 km A short distance past the park perimeter-road overpass, take the turnoff to the right to Park Drive and Prospect Point.

74.6 km Stop 5. Prospect Point to Third Beach, Stanley Park.

The excursion is on foot between here and Third Beach. The remaining log is for the driver(s). If they happen to be sounder in body than mind, they can drive to the Third Beach parking lot, run back along the seawall, and rejoin the group at Prospect Point.

75.7 km Hollow Tree on left has been a favorite group-photo site in the Park since the turn of the century.

76.5 km Just past the sign 'Third Beach Concession' turn right, and at the STOP sign right again, and proceed down the hill to the parking lot.

76.8 km End of road log.

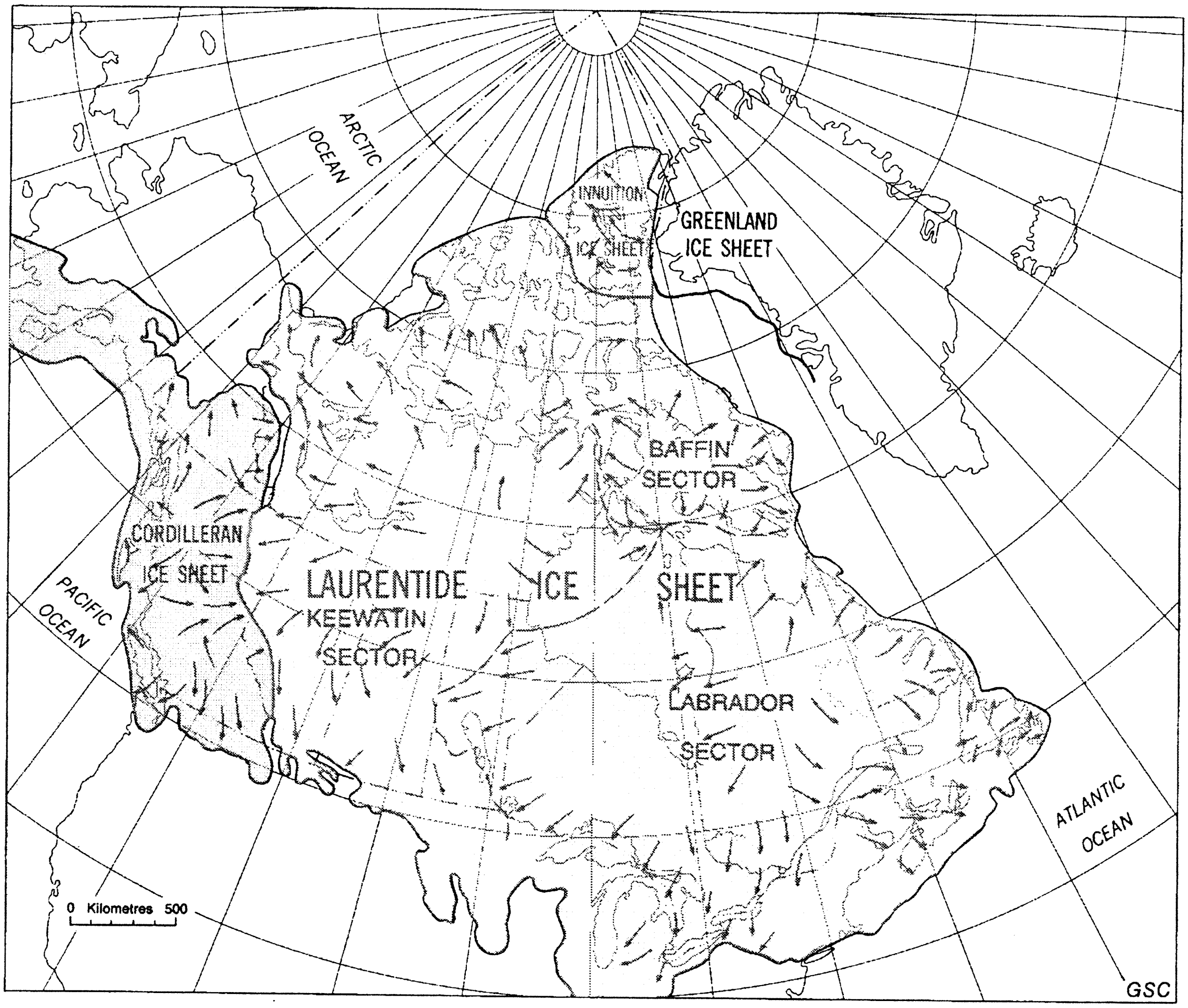


FIGURE 19. Approximate extent of main ice sheets during the last (Wisconsin) glaciation. Arrows indicate probable directions of ice flow at the glacial maximum. (Fulton, 1989; after Prest, 1984.)

CAPSULE OF GLACIATION THEORY

Ice ages have occurred more than a dozen times. It is perhaps surprising that glacial periods have not been more common than indicated by the geological record. The earth has an average temperature which is high enough to support liquid water, but with little freeboard. The difference in mean temperature between the present and the coldest of glacial times is only 5 to 10°C.

The oldest documented ice age took place about 2.6 billion years ago, near the end of the Archean. The period from 750 Ma to 570 Ma in the Neoproterozoic had at least four glaciations. Two are known to have extended to low latitudes. The 720 Ma glaciation seems to have lasted more than 10 m.y., and, according to Hoffman and others (1998), may have resulted in a "Snowball Earth". Such protracted ice cover is thought to have been initiated by a decrease in atmospheric CO_2 caused by a strong carbon drawdown, attributed to a major bloom of metazoans (large multicelled organisms) in the oceans, and subsequent exceptionally high rates of organic carbon burial. Extension of the ice cover would have been driven mainly by the increase in the earth's reflectivity (albedo), and ended when atmospheric carbon dioxide from volcanic outgassing increased sufficiently to overcome the effect of ice albedo. Ice albedo runaway has not reoccurred during the Phanerozoic, perhaps because of increased solar luminosity, or for more complex reasons. There were, nevertheless, at least three glaciations in the Paleozoic, but none recorded in the Mesozoic.

In the Tertiary, ice began to accumulate in Antarctica more than 20 million years ago, although only much later in the northern hemisphere. Ice sheets in North America began to grow about 2.4 million years ago. The individual ice ages since have been characterized by slow buildup and rapid deglaciation.

Scientists were slow to recognize even the most recent glaciation. It was not until the mid-1860s, that most geologists accepted the general tenets of Louis Agassiz's glacial theory, which he had proposed in a private publication in 1840. They then knew that not only had an ice age occurred, but it had occurred repeatedly. Yet, the reason for general ice advance and retreat remained a mystery, and for the most part, still is, but many theories have been formulated. An excellent and entertaining summary of these theories (up to 1976) was compiled by Imbrie and Imbrie (1979).

The most important, and still unresolved, problem is, "What initiates an ice age?" A number of possibilities have been suggested.

Decrease in solar radiation. This would certainly cause a drop in global temperature and an ice age, but except for these effects themselves, no independent evidence for a decrease in solar radiation during ice ages has been found, nor is there any evidence of an increase since the last ice retreat began.

Dust clouds in space. If the solar system moves through an interstellar dust cloud, the effects would be uncertain. Opinion is divided as to whether the filtering effect would offset the fueling effect of dust particles falling into the sun, which presumably would make it burn more brightly.

Carbon dioxide in the atmosphere. Carbon dioxide is relatively transparent to short-wave radiation from the sun, but more opaque to the longer-wave radiation reflected back into space. If the carbon dioxide concentration dropped, the earth would cool. As previously mentioned, accelerated carbon drawdown by increased rates of organic carbon burial in the oceans is a possibility, but why this would happen is not clear.

Volcanic eruptions. Large volcanic eruptions can eject large quantities of sulfur-rich gases into the stratosphere. Within a month they are oxidized to H_2SO_4 which can persist for several years. It can cause cooling on the order of a half a degree Celsius, by increasing the heat absorption of the atmosphere and by providing condensation nuclei for clouds. Yet, most dust ejected by eruptions settles very quickly and has only short term local effects on the climate. The eruption of Krakatoa in 1883, however, caused redder than normal sunsets for two years. Conceivably, a closely spaced sequence of major eruptions that happened to coincide with other factors (astronomic, etc.) could drop global temperatures sufficiently to bring on glaciation. Ice cores from Greenland and Antarctica provide excellent records of periodic volcanic eruptions in both hemispheres for the last 100,000 years but the climatic connection remains unclear. Volcanism might actually increase global temperatures by increasing atmospheric CO_2. Complications are further increased by evidence that crustal stress caused by loading and unloading of ice from continents, and loading melt water in the ocean basins may in itself promote volcanism (G.A. Zielinski, 1999).

Cyclical variations in Arctic precipitation. The essence of this, the Ewing-Donn theory, is that at the beginning of an ice age the Arctic Ocean is free of ice, allowing more evaporation and an increase in precipitation. Both land ice and ocean ice then grow. When ice again covers the Arctic Ocean, precipitation decreases, and the ice sheets retreat. At present, precipitation is very low in the Arctic; in fact, it is technically a desert. The Arctic ice is thinning, and fairly rapidly. Will it disappear and again cause high precipitation? It is possible, and from that point of view the theory seems credible. The main snag is that at the beginning of an ice age the Arctic Ocean sediments should contain fossils of organisms characteristic of sunlit waters. Yet, studies of these sediments indicate that at no time in the past several millions of years has the Arctic Ocean been ice free.

The stochastic theory. This is a statistical approach that maintains ice ages are simply the result of random changes in factors affecting global climate, and that such randomness is an inherent property of large complex systems. The theory is a delight to mathematicians, but just a white flag to geologists.

The astronomic theory of Adhémar. This theory was proposed in France in 1842 shortly after Agassiz got the glaciation theory rolling. Adhémar sought intellectual solace in the fact that in the northern hemisphere the number of daylight hours exceeds the hours of darkness by 168 hours (seven days), whereas the reverse is true for the southern hemisphere. He maintained that this is causing a gradual warming in the north, and a cooling in the south. By this theory the southern hemisphere is currently in an ice age. Due to the wobble in axis of rotation (a 22,000 year progression), the situation reverses every 11,000 years. Such a periodicity of ice ages seemed reasonable to some in the 1850s, until Baron Alexander von Humbolt, a German naturalist, showed that the average temperature of both hemispheres is controlled by the number of solar calories received, not by number of hours of daylight, and that both receive exactly the same number of calories in the course of a year. The southern hemisphere is actually colder, but that was learned much later, and is caused by the location of the Antarctic continent directly over the south pole.

The astronomic theory of Croll. In 1875 James Croll, then with the Geological Survey of Scotland, published *Climate and Time.* He expanded Adhémar's theory by adding the effects of changes in the eccentricity of the earth's orbit, caused by the gravitational effects of other planets, and to the wobble of the rotational axis. Croll maintained that when the distance between the sun and the earth on December 21st exceeds a critical value, winters in the northern hemisphere are cold enough to trigger an ice age; when it is less than a critical value an ice age occurs in the southern hemisphere. The critical values are exceeded about every 100,000 years when the earth's eccentricity is greatest (about 6%). He recognized two positive feedbacks. As ice expands increased reflectivity promotes further expansion. Also, as a polar region gets colder, the trade winds get stronger in that hemisphere and force more of the warm currents into the opposite hemisphere, further cooling the colder hemisphere.

Croll's theory would be substantiated if it could be shown that ice ages alternated between hemispheres, and matched the eccentricity record. They don't, but establishing that was beyond the capability of nineteenth-century geology. Nevertheless, most geologists, especially in North America, became skeptical of Croll's theory, mainly because it required that the last ice age terminated about 80,000 years ago. Contrary evidence came from the Niagara River, which flows over a glacial till that must have been deposited before the river occupied its present course. The length of the gorge and the estimated rate of retreat of the falls (about a metre per year) indicated that the ice had left only about 10,000 years ago.

Milankovitch modification of the Croll theory. In 1920 Milutin Milankovitch (1879-1958), a talented Serbo-Coatian mathematician, showed that Croll was partly correct. He calculated the effects of the 41,000-year oscillation of the inclination of the earth's axis (large at the poles, little at the equator), and the 22,000-year oscillation of the earth-sun distance (small at the poles and large at the equator), on the radiation received at various latitudes. Milankovitch was aided by Vladimir Köppen, a German climatologist, who pointed out that it was the decrease in radiation during the summer, not in the winter, that would cause glaciation. Temperatures in the polar regions even now are clearly low enough for ice to accumulate, but that is more than offset by summer melting. Milankovitch was convinced, and concentrated his calculations on this effect. Eventually, he produced his influential radiation curves which appeared in, *Climates of the Geological Past,* a book by Köppen and Wegener (1924). The curves fitted moderately well with what was then known about the timing of European glaciations during the past 650,000 years. The last three radiation minima indicated ice ages 25,000, 72,000, and 115,000 years ago, but the ages of glacial drift could not be accurately determined at that time.

Miscellaneous lines of evidence. Help with the timing problem arrived shortly after World War II, when Willard Libby, at the University of Chicago, developed the radiocarbon method of dating. An accurate means of determining the ages of glacial deposits now existed. Unfortunately, the relatively short half-life of radiogenic carbon 14 (only 5,730 years) limits its usefulness to material less than about 40,000 years old. In North America dating of the Wisconsin drift showed that the ice had reached its maximum 18,000 years ago (Fig. 19), and rapidly disappeared about 10,000 years ago. Older tills existed in the drift, but were beyond the range of the carbon 14 method. The 18,000-year date did, however, damage the Milankovitch theory because it was 7,000 years later than predicted. Even more damaging was the dating of 25,000-year-old peat from many locations in North America and Europe. This confirmed a warm interval precisely when radiation was at its minimum.

Additional evidence existed in sediments from the deep ocean basins. Sedimentary records on land are notoriously incomplete, but in the oceans deposition is continuous, and in the deep oceans the layers are only rarely disturbed (mainly by turbidity currents). They are also very thin, only two to three millimetres in the Atlantic and less, about one millimetre, in the Pacific. Short cores, therefore, provide good records for long periods of time. Unlike continental margin deposits, deep-sea oozes consist almost entirely of organic remains. In temperate and tropical seas, the material is mainly from foraminifera which are calcareous. In colder waters the remains are mostly from radiolaria, which are siliceous. Below about 4,000 metres, however, both types are dissolved, and the ooze is not fossiliferous. A combination of simple chemical analyses, and the relative abundance of certain fossil species, established a number of fluctuations in water temperature during the past million or so years, but firm time lines were still missing.

A different type of evidence came from terraces cut in coral reefs in the Barbados and in Australia. They could be dated by the thorium method which was good for ages up to 150,000 years, a considerable extension of the 40,000-year limit of the carbon 14 method. The terraces, when dated, marked episodes of high sea levels at 125,000, 105,000, and 82,000 years ago. Those numbers corresponded well with the radiation highs at 45° latitude determined by Milankovitch, providing that the greater emphasis is placed on the 22,000 precession cycle of axial tilt than on the eccentricity cycle. A longer record, however, was definitely required.

It came when accurate dating of paleomagnetic reversals became possible. Reversals in the earth's magnetic field had been first discovered in 1906, by a French geophysicist, Bernard Brunhes. Most geologists did not accept the idea, even after Motonori Matuyama showed in 1929 that multiple reversals occur in a series of lava flows in both Japan and Korea. He also found that at least one of these reversals had occurred during the Pleistocene. The prevalent disbelief was based on some lab work which showed that certain minerals could, under specific conditions of cooling, acquire a reverse magnetization. By 1963, the reversal theory was finally confirmed by the dating of reversed lava flows around the world by the potassium-argon method, and finding them to be in agreement. The pioneers were recognized by naming the current interval of normal polarity the Brunhes Epoch, and the preceding interval of reverse polarity, the Matuyama Epoch. Of particular interest was the reversal at 780,000 years, and a very short interval of normal polarity at 1.8 million years within a long period of reversal. The latter was important, in that it coincided with the age of the first occurrence of cold water species in the well-exposed sediments of southern Italy. By agreement, the 1.8 million-year age was designated as the beginning of the Pleistocene.

Further work established that during the Pleistocene climate cycles and, by inference, glacial periods had a pulse of about 100,000 years, and that the periods of cooling were long compared to the abrupt warming intervals. This did not sit well with astronomic theory, so ways to modify the theory were investigated.

Oxygen isotope analyses of planktonic shells do not reflect temperature of the seawater accurately, but they do indicate how much of the lighter isotope (^{16}O) has been extracted by evaporation from the ocean, and subsequently bound up in glacial ice, leaving the seawater richer in the heavier isotope (^{18}O). The oxygen-isotope-ratio is, therefore, a reflection of ice volume. Within the 700,000-year Brunhes Epoch of normal magnetization, 19 isotopic stages were found in a Pacific deep-sea core that penetrated the entire period. This made possible fairly accurate time estimates by interpolation, based on the assumption of a near-constant rate of deposition.

During glacial maxima about 5.5% of the world's surface water was locked up in ice, as opposed to only 1.7% at present. The larger Quaternary ice sheets were up to 4 km thick. They lowered the glacio-eustatic sea level about 150 metres, and depressed the crust. The accumulation of land ice, however, is only the second major cause of changes in sea level. The largest changes in sea level (about 500 m) are caused by fluctuations in the volume of oceanic ridges, but these take place very slowly (over tens of millions of years). The effects of glaciation and deglaciation are comparatively rapid. It has been suggested that when continents are depressed by ice, ocean floors rise, and conversely, isostatic uplift of continents after ice retreat is matched by subsidence of ocean floors. This complicates the interpretation of sea level changes, because the ice-loaded areas are much smaller than ocean basins. On the coast of British Columbia and southeast Alaska the scene is further complicated by tectonic uplift caused by the subduction of oceanic plates beneath North America. The net change on the coast since preglacial time is depression of the land compared with sea level. The fjords, however, exaggerate the degree of flooding. Those valleys have been greatly deepened by glacial erosion. The floor of Queen Charlotte Sound, where erosion was less concentrated, is perhaps a better indicator of the degree of submersion. In-place plant material there indicates about 100 m of submersion since the ice left, which is about equivalent to the calculated rise in sea level resulting from the return of water to the oceans from the last glacial maximum.

CAPSULE PLUTONISM

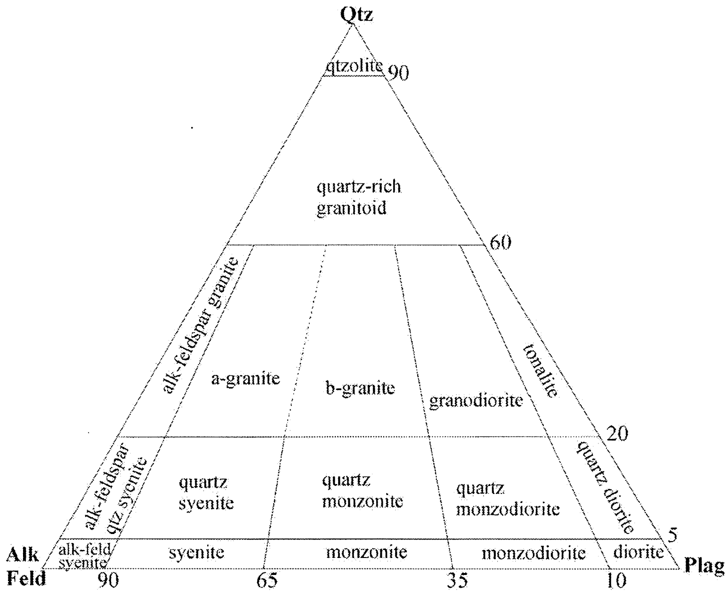


FIGURE 20. I.U.G.S. classification of plutonic rocks.
(after Streckeisen, 1976)

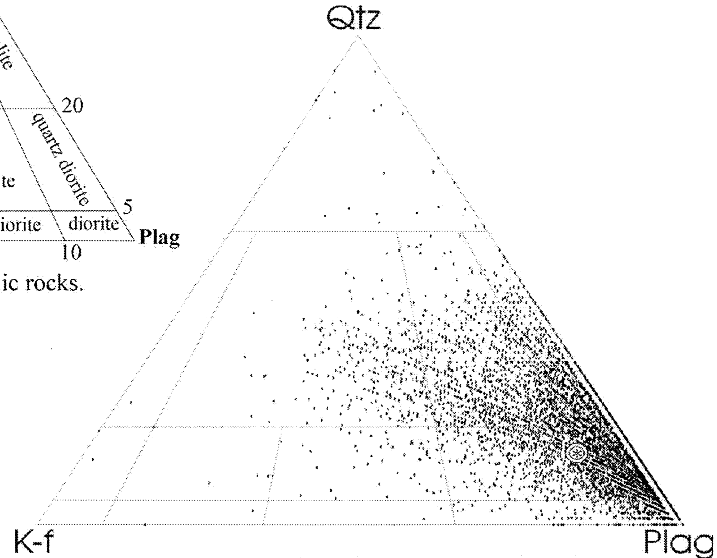


FIGURE 21. Coast Plutonic Complex. Modes of 8,983 specimens, determined by comparison with spot charts. Averages: total mafics-17%, k-feldspar-5%, quartz-14%, density-2.75, plot position of average-⊛

The granitic rocks which underlie the North Shore mountains form a small part of the Coast Plutonic Complex, which extends northward about 1700 km, into the Yukon. The belt is narrow (about 100 km) and dominated by intermediate and basic plutonic rock. It is one of the largest plutonic terranes in the world, and also the most basic (see Figures 20, 21).

The average plutonic rock, both in the region near Vancouver, and for the Coast Plutonic Complex (CPC) as a whole, is quartz diorite, which underlies about 40% of the region.

In the broader picture, about three quarters of the Coast Mountains are underlain by plutonic rock. The order of abundance of the different plutonic rock types, as determined by the GSC's long-lived Coast Mountains Project is as follows: quartz diorite, 38%; tonalite, 18%; diorite, 17%; granodiorite, 12%; quartz monzodiorite, 8%; gabbro, 3%; beta granite, 3%; monzodiorite, <1%; and quartz monzonite <1%.

Although a variety of common plutonic rocks are present in the area, they are composed of only five main plutonic minerals, and they have a definite genetic sequence, namely, plagioclase → hornblende → biotite → quartz → potassium feldspar. This sequence is independent of relative abundance or of plutonic rock type. An exception to the sequence is the albite component of perthite, an alkali feldspar consisting mainly of potassium feldspar but with blebs, strings, etc. of exsolved albite, the sodic end member of plagioclase(Figs. 28, 29).

A study of the main minerals in each of the different types of plutonic rock revealed a distinct evolution of these minerals from irregular shapes and numerous micro-inclusions (survivors from the protolith; Figs. 26, 27) in hornblende-rich diorites and quartz diorites to more regular shapes (that is, greater euhedralism) and almost no micro-inclusions in biotite granodiorite and granite. Most of the plutonic rocks exposed in the region appear to have evolved through the sequence amphibolite→ diorite→ quartz diorite→ tonalite→ granodiorite→ granite.

Zoning in the plagioclase evolves from simple in the hornblende diorites, to complex in the intermediate hornblende–biotite quartz diorites and tonalites, to simple again in the highly evolved, biotite granodiorites and granites.

Chemically, the average composition of plutonic rock in the Coast Plutonic Complex is nearly identical to that of typical continental crust in SiO_2, FeO, TiO_2, and P_2O_5, and similar in MgO and CaO (Table 2). The CPC, however, is markedly richer in Al_2O_3; in fact, it is equivalent to oceanic basalt in that oxide. The CPC is also somewhat richer in Na_2O, and distinctly poorer in K_2O than continental crust. The volcanic equivalent to the average plutonic rock in the CPC is a tholeiitic andesite.

Elements of the granite controversy

An integral part of the earth's evolution is the growth of granitic crust, considered herein as the 'grey hair' of moderate-size planets that orbit within the water window of their sun. It has not yet been proven, but development of a granitic crust seems to be a special form of planetary evolution, restricted to planets able to retain surface water and a hydrous atmosphere. Beyond the planetary requirements, the origin of granite is still a matter of dispute. The subject is too paradoxical, made up of too many components, and too many subprocesses for comfortable comprehension.

TABLE 2. Average chemical composition of the Coast Plutonic Complex, compared with average continental crust.

Oxide	CPC (%)	Cont. Crust (%)	% Diff
SiO_2	60.15	60.22	-0.1
Al_2O_3	16.98	15.18	12
Fe_2O_3	1.75	2.48	-29
FeO	3.72	3.77	-1
MgO	2.97	3.06	-3
CaO	5.85	5.51	6
Na_2O	4.37	2.97	47
K_2O	1.82	2.86	-36
TiO_2	0.72	0.73	-1
P_2O_5	0.24	0.24	0
MnO	0.12	0.14	-14
S	0.03	0.04	-25
H_2O	1.00	1.38	-28

CPC values from Roddick (1983); values for continental crust from Ronov and Yaroshevsky (1969).

In spite of relentless study over two centuries, the creation of granites (sensu lato), that is, plutonism, remains poorly understood, as does the mechanism of pluton emplacement. Nevertheless, most geologists consider plutonic rock to be former magma that attained its coarse grain size and texture by cooling slowly within the crust, and, where the magma reached the surface, a compositionally equivalent volcanic rock resulted. Opposing this welding of the relationship between volcanic and plutonic rocks, were the granitizers, who maintained that granites formed in the crust by metasomatic processes. They were always a minority, and became fewer, mainly because the processes they proposed were either clearly invalid, or at best unconvincing. Also, little supporting experimental work could be cited, owing to the extreme difficulty of replicating plutonic processes in the laboratory. On the other hand, many magmatic processes have been duplicated in the lab, but they seem more applicable to volcanic rocks than to granites. Common granitic textures have yet to be produced in the laboratory.

The dominant theme in the magmatic origin of granite is evolution by crystal differentiation in large magma chambers. Yet, whether magma chambers of pluton or batholithic size can exist anywhere in the crust, without violating the most basic tenets of physics and hydraulics, has never been established. The main problem is that magmas are the weakest of crustal materials. How a large volume of any such low-viscosity material can support the surrounding heterogeneous crust is not apparent. Nor is the mechanism intuitive by which such material can be brought up through many kilometres of crust in the shape and size of common plutons and batholiths.

Magma in the crust is necessarily contained in an unmelted, irregularly-shaped envelope of heterogeneous country rock of comparatively great physical strength. Because they are fluids, magmas act hydraulically (i.e., pressures are transmitted perpendicular to the chamber walls). Enormous stresses are concentrated against the walls of embayments, tending to break them open, and create fractures extending outwards from them. Stresses at points of wall or roof irregularity can neither be prevented, nor long resisted, and magma forced into the resultant fractures will either soon freeze or reach the surface as a volcanic flow. No opportunity for pluton creation exists in that scenario.

The difficulties inherent in large crustal magma chambers have led to an increasingly popular alternative mechanism for bringing magmatic granitic material up through the lower crust. In these models, magma (of the appropriate composition, attained, in this writer's opinion, miraculously) actually does succumb to hydraulic principles in the lower crust, and is forced into fractures to begin its passage upwards. When the upper crust is reached, the dykes are postulated to balloon, that is, to expand laterally, and form the common types of plutons. Such ballooning requires a very special structural setup, namely, some sort of barrier which can prevent further upward propagation of dyke fractures. No evidence of such barriers exists for most of the world's plutons, nor of internal layering which inevitably would result from this mechanism because of the normally fast freezing of dykes and sills in the upper crust (Rubin, 1995).

Unless one believes that the earth was born with stratified continental crust, any theory on the creation of granite should be consistent with its known stratification. Seismic records show that the crust is stratified, and that it is denser and more basic in composition near the Mohorovičić Discontinuity, or Moho. The Moho itself is marked by a sharp increase in seismic velocity, and is thought to separate the crust from the mantle. In reflection seismic images the Moho is considered to be the base of a commonly conspicuous zone of high reflectivity at or near the base of the crust (Fig. 22).

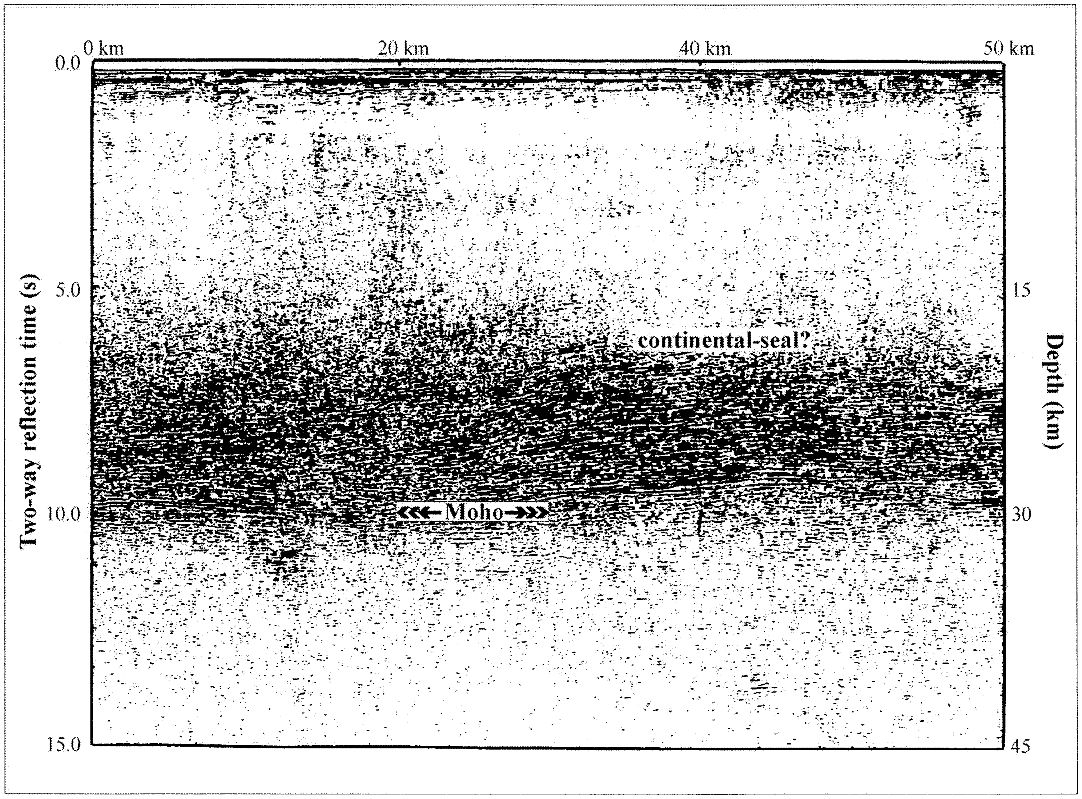


FIGURE 22. Highly reflective lower crust is unusually well-defined in this seismic profile offshore from southwest Britain. (from Warner, 1990).

Within the concept of cooling-differentiation in large magma chambers, crustal stratification is attributed to settling of heavier crystals, but within dykes, that process would be clearly ineffective.

A different approach to the problem is found in variations of the 'restite' theory, wherein the more sialic (quartzo-feldspathic) components are bled off by partial melting leaving the more mafic and unmelted components behind as 'restite'. The theory was developed mainly in Australia (in a number of papers by Bruce Chappell and Allan White), but is opposed by other magmatists, on chemical and other grounds. Whatever its merits regarding the stratification of the crust, the theory contributes little to the underlying problem of pluton and batholith creation.

The case for metasomatic plutonism

In the metasomatic hypothesis weathering is a necessary preconditioning process. By removing the more soluble components to the oceans and concentrating the more sialic components on the continents, weathering is an important and fundamental differentiation process. Equally important, it leads to the creation of hydrous minerals in surface and near-surface sedimentary and volcanic material. Subsequent burial and heating of that material introduces *distributed* water to the crust. This is an essential step, without which metasomatic plutonism cannot take place.

The importance of water in crustal processes can scarcely be overestimated. Above its critical temperature of 374° C and at a minimum pressure of 218 atmospheres (about 3200 lbs/in^2, attained at about 900 metres of depth), water and steam have the same density (0.324). Such hydrous fluid is the strongest general purpose solvent known. It can readily dissolve quartz (silica) and other rock-forming minerals. The critical-temperature isotherm is nearer the surface (10-20 km) in most Phanerozoic regions, than in the older shield regions (20-35 km) where thermal gradients are lower. In most continental crust the isotherm is well above the Moho, resulting in a chemically active lower crust, but in some places, such as Vancouver Island, which has a low thermal gradient, the isotherm lies below the Moho, within essentially dry mantle rocks.

In the lower crust some hydrous minerals break down, producing hydrous fluid. Little of it can be absorbed by high-grade metamorphic minerals, which are mainly anhydrous, and little can escape upwards because of the rapid-sealing properties of rock near the critical-temperature isotherm, which also marks the transition between brittle fracturing above and plastic deformation below. Initially the hydrous fluid is irregularly distributed, but plastic creep eventually forces much of the fluid into near-horizontal lamellae. These are strongly seismically reflective, and in places can be traced on seismic profiles for more than 10 kilometres (Warner, 1990).

Oxygen isotope studies indicate that the network of hydrous fluid may become integrated through very large crustal regions (Magaritz and Taylor, 1986). More importantly, the hot damp rock may persist almost indefinitely, and is susceptible to chronic, cancer-like metasomatism.

Chemical heterogeneity which is not in equilibrium with the prevailing temperature pattern cannot long persist if it is penetrated by hydrous fluid, which provides the medium for diffusion and metasomatism. In dry areas metasomatism can proceed only by solid diffusion, a comparatively weak and slow process, even by geological standards. The rock in such places changes little, even over long intervals.

Small isolated volumes of rock may remain dry. They become the mafic enclaves, commonly amphibolitic in composition, that are common in most plutonic terranes. This process of enclave creation in the lower crust is equivalent to spheroidal weathering at the surface, especially in tropical climates, where relatively fresh granite 'boulders' survive within a granitic matrix, so totally altered that it crumbles in the hand.

Although rarely forming more than one percent of any substantial volume of the crust, hydrous fluid is the signature component of active zones. Wherever it is concentrated, an environment is created for chemical transfer that lessen instabilities inherent in heterogeneous crust. The effects are profound because they are so chronic. The concatenated effects of pressure-solution and reprecipitation displace hydrous films, and over time cause them to sweep across large volumes of crust, much like streams move back and forth across deltas in much shorter geological intervals. Yet, actual hydrous flow within the films is probably not significant. Nor is it needed; only the presence of hydrous fluid as a medium for diffusion is required.

The reactions inexorably rearrange components of the lower crust. The temperature pattern drives metasomatic reactions always in the direction of stabilizing its normally heterogeneous chemical character. Given enough time, nothing can prevent the more basic components from migrating to hotter, generally deeper, areas, and the more sialic to cooler environs, usually upwards. The combination of processes, in other words, plutonism, is inevitable while the hydrous network exists. The slowness of the process is compensated for by the vastness of the time in which it operates. Plutonism is continuously active in most continental crust and not easily disabled, but it can be halted, locally and temporarily, by loss of hydrous

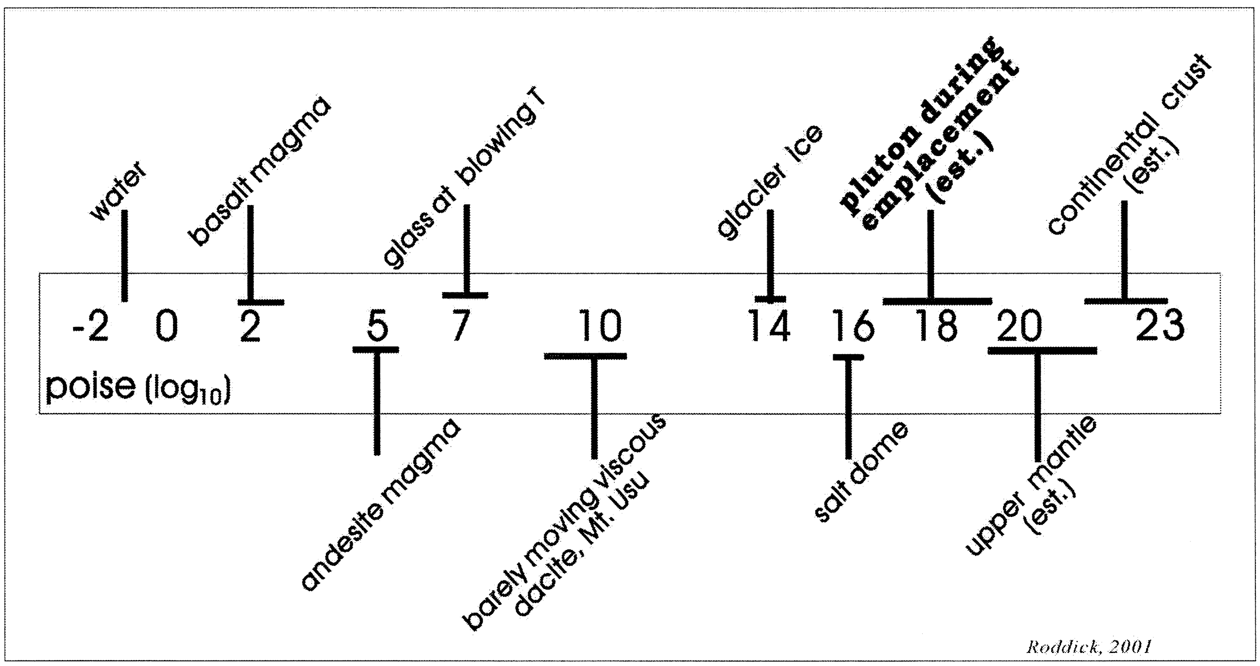


FIGURE 23. Viscosity of some common materials from water to continental crust.

fluid or by fusion (which immediately 'zaps' the sparse, but vital hydrous fluid).

The brush used in this scenario is necessarily broad, with emphasis on the underlying principles, rather than on individual responses to the innumerable factors of the local environment. Some details about the behavior of the major individual minerals can be found in Roddick (1965), and in Figs. 26-35.

The lower crust is where the 'action' is (read plutonism), simply because this zone is endowed with the critical properties. Most of the strata there are in the amphibolite grade of metamorphism (Table 3), and on compressional continental margins especially, amphibolite itself is common. (It is a metamorphic rock consisting mainly of hornblende and plagioclase.) Even with minimal metasomatism, simple recrystallization of amphibolite to coarser grain size will give large volumes of rock a dioritic appearance.

Emplacement

Although they overlap, plutonism and pluton emplacement are not synonymous. Substantial volumes of continental lower crust are made more granitoid in texture and composition by plutonism, but only parts are mobilized in the form of plutons.

The fundamental factor that makes mobilization possible is the rheological stratification of the crust. The lower crust, which has a viscosity of about $10^{18}\rho$, is confined between the relatively rigid upper crust and upper mantle, both of which have a viscosity of about $10^{23}\rho$ (Bailey, 1998). The lower viscosity of the enclosed material is a result of its content of hydrous films and its temperature, which is higher than the rocks above. The upper mantle is, of course, even hotter, but it lacks the hydrous fluid. Diapiric emplacement of plutons is mechanically the same as for salt intrusions. The driving force is not buoyancy, but crustal stress on material contained in a more rigid envelope. It depends primarily on the viscosity contrast between the two ("the pie crust and the filling"), and the magnitude of the stress applied to the envelope; the smaller the difference in viscosity, the greater is the stress required to force the core material out of its envelope.

That salt domes rise because salt is lighter than crustal rocks is still commonly taught in schools. Yet it has long been realized by those who work in salt tectonics (Balk, 1949), that buoyancy is not a significant factor driving salt intrusions (Fig. 24). This is particularly clear in the case of 'salt domes' in the Canadian Arctic, and elsewhere, that consist mainly of anhydrite, which is considerably denser than the strata they intrude (Heywood, 1957).

TABLE 3. Principal Metamorphic Grades

Metamorphic Grade	*Temperature (°C)*	*Characteristic minerals*
Greenschist facies	250 - 450	Chlorite, epidote, actinolite
Amphibolite facies	450 - 650	Hornblende, andesine (an intermediate plagioclase)
Granulite facies	Above 650	Pyroxene (augite, diopside)

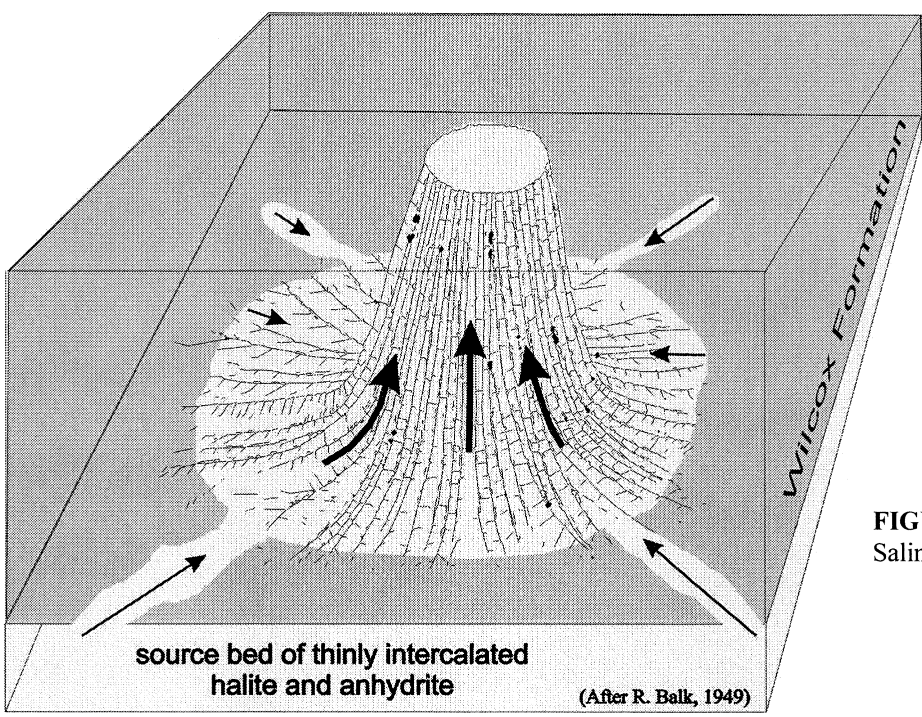


FIGURE 24. Grand Saline Salt Dome, Texas

For a pluton to form, the stress required must be great enough to exploit the viscosity difference between plutonic material, about $10^{18}\rho$, and the confining country rock, about $10^{23}\rho$. For salt the viscosity difference is considerably greater, $10^{16}\rho$ vs. $10^{23}\rho$, and the stress required to mobilize it is considerably less. (In the lower crust salt would be forced into dykes. Salt domes are therefore upper crustal phenomena.) The much greater stress needed to mobilize plutonic material is probably transmitted by converging tectonic plates (in our area, the North American and Juan de Fuca plates).

Initiation of the process is probably analogous with that of salt domes. Slow plastic flowage in the lower crust driven by compressive stresses causes a ridge (probably a succession of ridges) to form in the lower crust roughly parallel to the line of subduction. As a ridge increases in height, local bulges form along it. Hydraulics now come into play. Plastic flow is concentrated within the bulges, and some of these break through the confining country rock and develop into diapiric plutons.

As a speculative scenario, we will focus on one of these juvenile plutons that is destined to become a normally- zoned pluton, that is, one with a more basic outer margin. The active pluton is somewhat like a tooth, having an inert exterior but a living core containing the original hot hydrous fluid, which is gradually concentrated in the nascent pluton's core by differential plastic flowage. There, as long as the hydrous fluid maintains its connection with the source region, metasomatic plutonism will continue, possibly at an accelerated pace because of fluid concentration, but at a somewhat lower ambiant temperature than the source region. Diorite-producing, hornblende-plagioclase metasomatism and recrystallization, which is thought to dominate the reactions in the source region gradually gives way to quartz metasomatism. The dioritic rock of the core zone evolves toward quartz diorite, and should the pluton remain in the quartz-in temperature range for an extended period, tonalite (see Figure 20) will form in the core. Because the outer zones of plutons, tend to lose their hydrous fluid faster than do the cores, they commonly lag metasomatically and remain more basic, in some cases, dioritic.

Fast-forward thousands or perhaps, hundreds of thousands of years. Horizontal inflow from the source layer continues to force the pluton head upward, in piston mode. Because of the horizontal plastic inflow to wards the base of the pluton, the Moho does not rise in response to pluton emplacement (Stewart, et al., 1986; Matthews, 1987; and Figure 21). Except for the pluton itself and its parent ridge, the action is, regionally, a crustal thinning process.

The pluton head eventually passes upward through the critical-temperature isotherm and into the upper crust. The hydrous fluid connection with the source layer still provides a medium for diffusion of incoming and outgoing components, and also a conduit for thermal energy. While the connection is maintained cooling will be retarded, and commonly irregular or oscillatory.

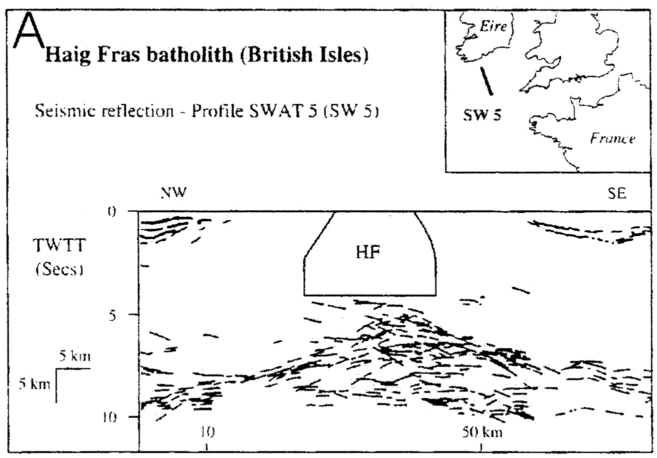


Figure 25A. Interpretation of a seismic reflection profile across the Haig Fras batholith by Matthews (1987). It also shows the HF batholith modelled to fit the observed gravity anomaly (Edwards, 1984). The reflections are shown as coming from beneath the batholith.

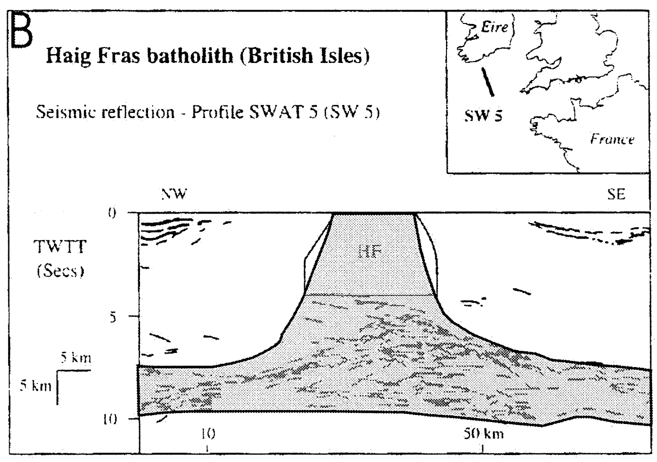


Figure 21 B. Same as Fig. 21A but extent of the Haig Fras Batholith has been reinterpreted, showing the reflections as originating in the lower part of the batholith and its source layer, in the manner of a salt dome.

Subduction forces keep the show moving, although less stress is required than for initiation of the pluton. When the pluton cools to the temperature range that favours crystallization of K-feldspar (at the expense of earlier minerals), tonalite evolves to granodiorite, some, perhaps, even to granite. Quartz diorite and diorite also will evolve leftward in the IUGS plutonic rock classification triangle (Fig. 20). (Even without pluton creation, the same general processes would apply on a much larger scale to the lower crust as it cools during erosional unloading.)

The invading pluton makes room for itself in the upper crust by a host of mechanisms, most commonly by roof uplift, but also by plastic deformation, folding and fracturing of the wall rocks, depending on their competence, structure and depth.

During the late stage of emplacement some plutons penetrate water-rich horizons of the upper crust. This water may penetrate the outer part of a pluton. Its effect can later be detected because it imposes the oxygen-isotope signature of the country rock on that part of the pluton, as it did for example in the Stuart Batholith in northern Washington (Paterson, et al., 1994). The addition of water from the country rock may also reduce the viscosity of the pluton margins, either by its mere presence, or by partial fusion if temperatures are sufficiently high, leading to the emanation of dykes. The phenomenon may also, lead to formation of small cupolas and miarolitic cavities in the roof area. The lowered viscosity of the margins also facilitates wedging of wall rock slabs, creating the 'stoped' appearance locally found in some plutons. Many plutons, however, show none of these features.

During emplacement, a large pluton forms an enormous stack of hot rock. It may retard the normal dissipation of heat flowing from the mantle sufficiently to cause fusion at the base of the pluton. If magma from there reaches the surface and is found to be chemically similar to nearby plutonic rock, it will be identified, erroneously, as the 'comagmatic' volcanic partner of the pluton. The extraction of magma from the base of a pluton also may cause it to sink back, and possibly form the classic pluton-cored caldera.

Synplutonic dykes (see Figures 8 and 9), common in many plutonic complexes, are accounted for in magmatic theory as basic intrusions into silicic magma. How this mechanism accounts for parallel walls is, at least, counter-intuitive. It is much easier to account for synplutonic dykes with metasomatic theory, because relative rigidity is normally maintained in both the source region and in the pluton itself. Suddenly applied high stress results in fractures which may penetrate deep into the mantle and admit basic magma. Whether they reach the surface or not, the dykes created will soon freeze, and when penetrated by hydrous fluid will be affected by the same metasomatism that thereafter modifies the plutonic rock. The dykes are far from immune, and may be partly or totally obliterated. They may also be torn apart by plastic creep, and form a trail of inclusions. Some geologists attribute most mafic enclaves in plutonic rocks to this process, although the generally random distribution of mafic enclaves does not support this view.

A frozen picture of the early stages of plutonism is captured in the walls of fiords which penetrate the Coast Plutonic Complex. Within the commonly thick sections of amphibolitic rock, are scattered and isolated granitic-textured patches, veinlets, layers, and other shapes. They emulate a kind of cancer, cumulating in places as large 'tumors'. To carry the analogy beyond what is justified, volcanism, in contrast, resembles ruptured aneurysms.

Acknowledgments

The paper was greatly improved by the constructive suggestions made by the reviewers, Lionel E. Jackson and Jennifer S. Getsinger.

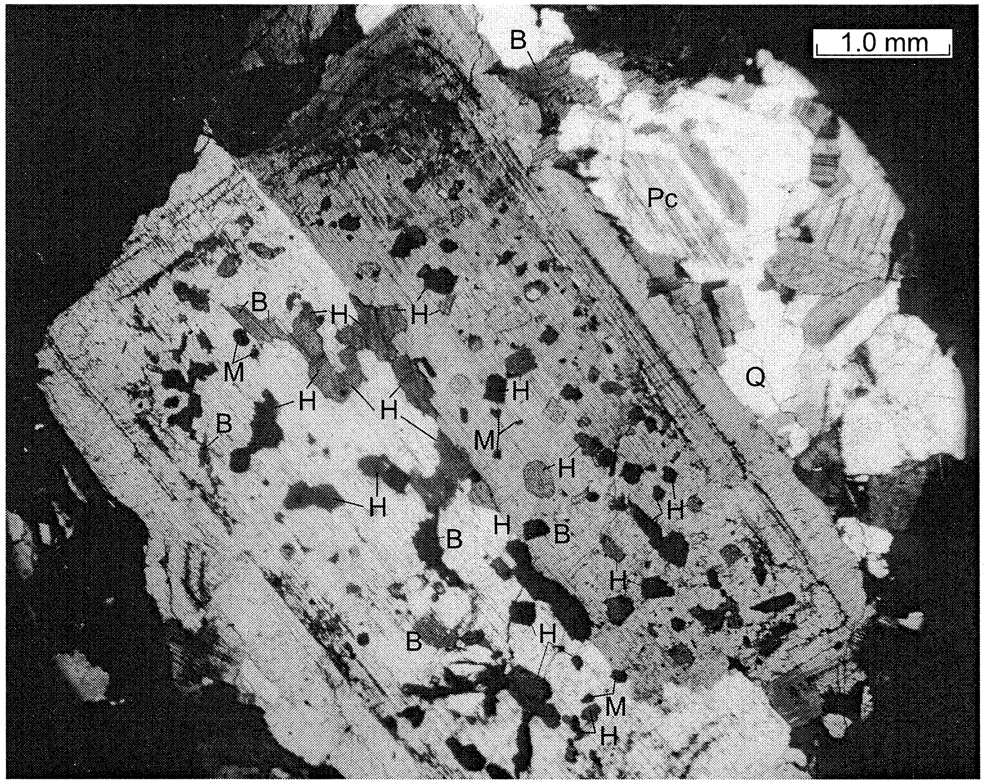


FIGURE 26. Large plagioclase crystal (a porphyroblast) containing granules which have survived from the amphibolitic protolith. Most of the included granules are hornblende (H). Also present are biotite (B) and magnetite (M) granules. In the upper right outside of the porphyroblast are crystals of biotite, plagioclase (Pc) and quartz (Q). From a quartz diorite on the ridge between Jacobs Lake and North Alouette River.

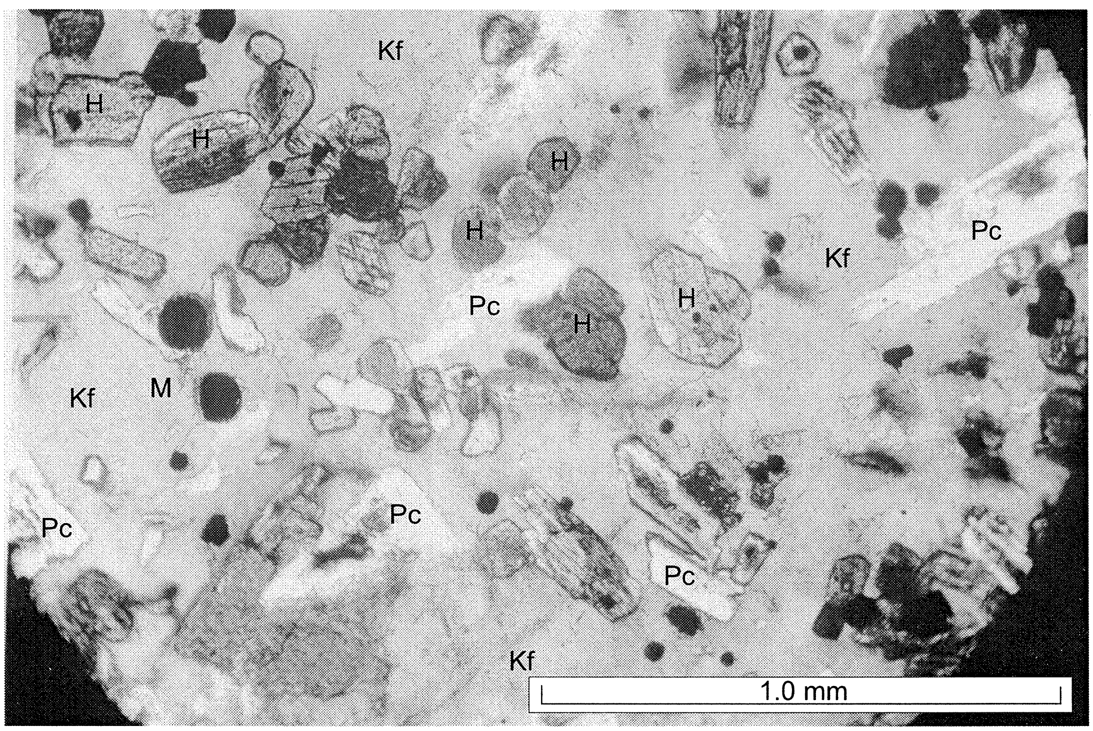


FIGURE 27. Part of a large K-feldspar (K-f) porphyroblast containing relic granules of hornblende (H) and plagioclase (Pc) from a pre-existing amphibolite layer in the Caulfeild gneiss. Some of the plagioclase crystals have been reduced to ghost-like forms;most have been made more euhedral by the slow corrosive effect of K-feldspar replacement. (In contrast, high energy replacement commonly leads to anhedralism and embayments.)
From granodiorite just east of the Caulfeild headland.

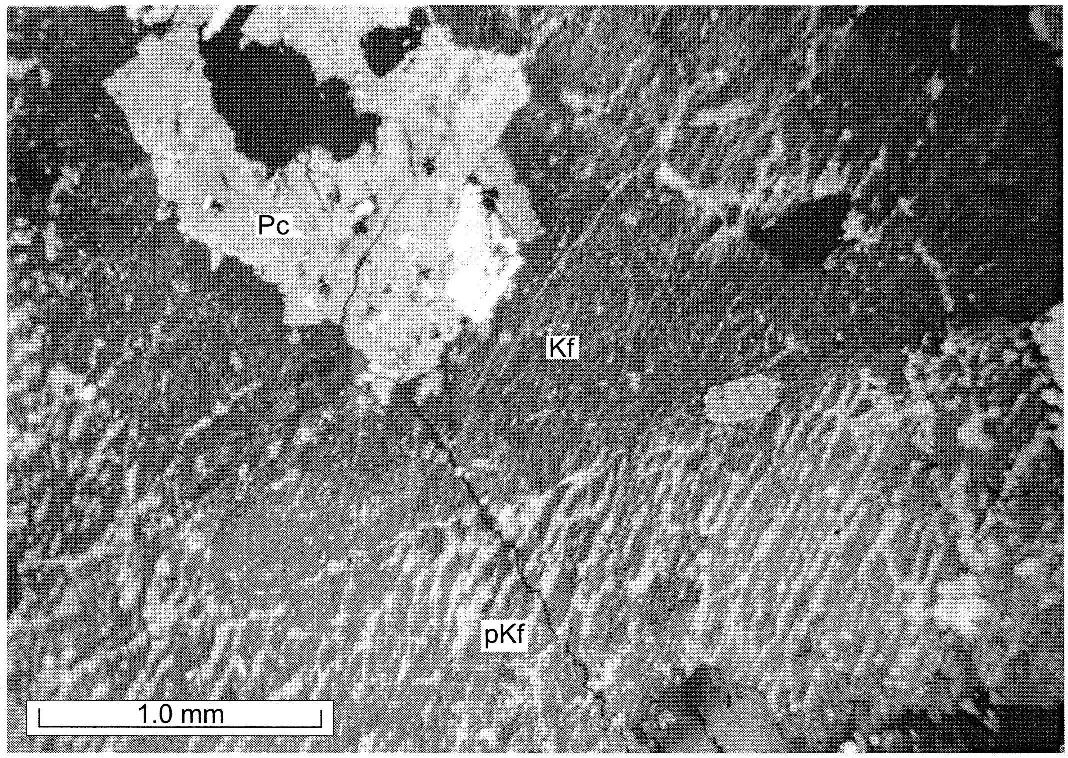


FIGURE 28. A large plagioclase crystal once occupied the entire field of this photomicrograph, but has been largely replaced by K-feldpar. Part of the calcic core of the Pc remains. The former extent of the core is marked by limit of the perthitic K-feldspar (pKf), which formed during replacement of the sodic outer zone of the plagioclase. The white strings and blebs in pKf consist of exsolved albite (sodic end-member of the plagioclase series). For more on this see Roddick, 1965. From a biotite granite west of Coquitlam Lake.

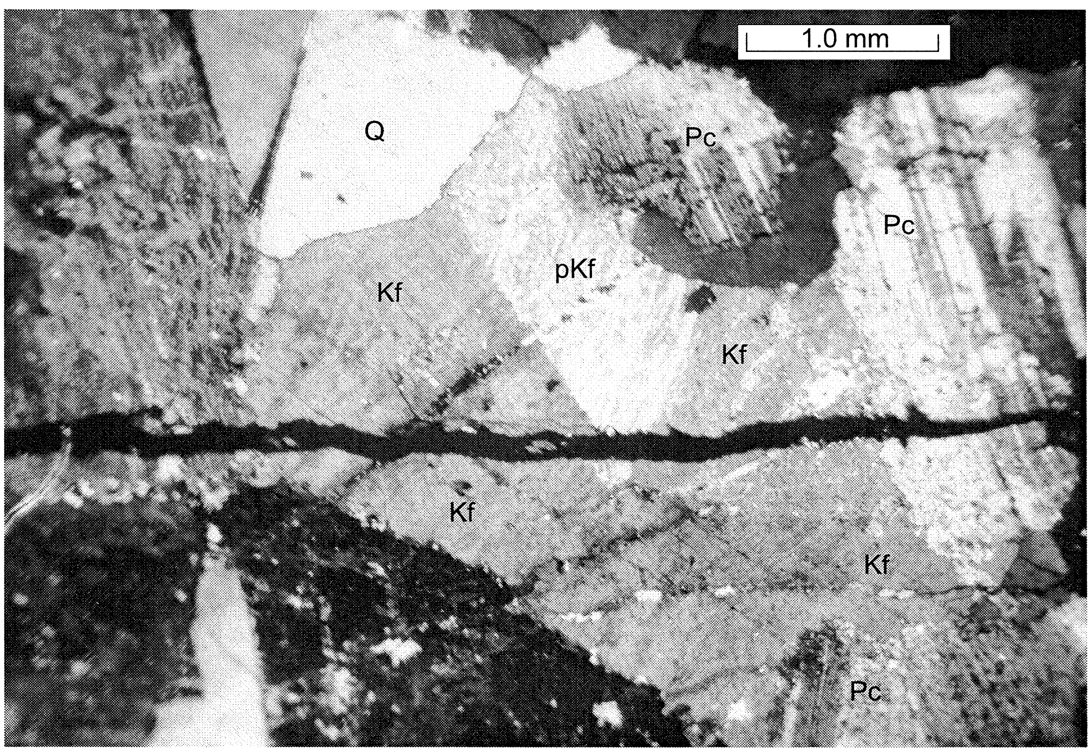


FIGURE 29. Highly perthitic area (pKf) in the K-feldpar (Kf) appears to be pseudomorphic after a very sodic plagioclase crystal. The relatively non-perthitic areas (Kf) were probably once occupied by calcic plagioclase and/or quartz. From a granite at Sunset Beach on Howe Sound.

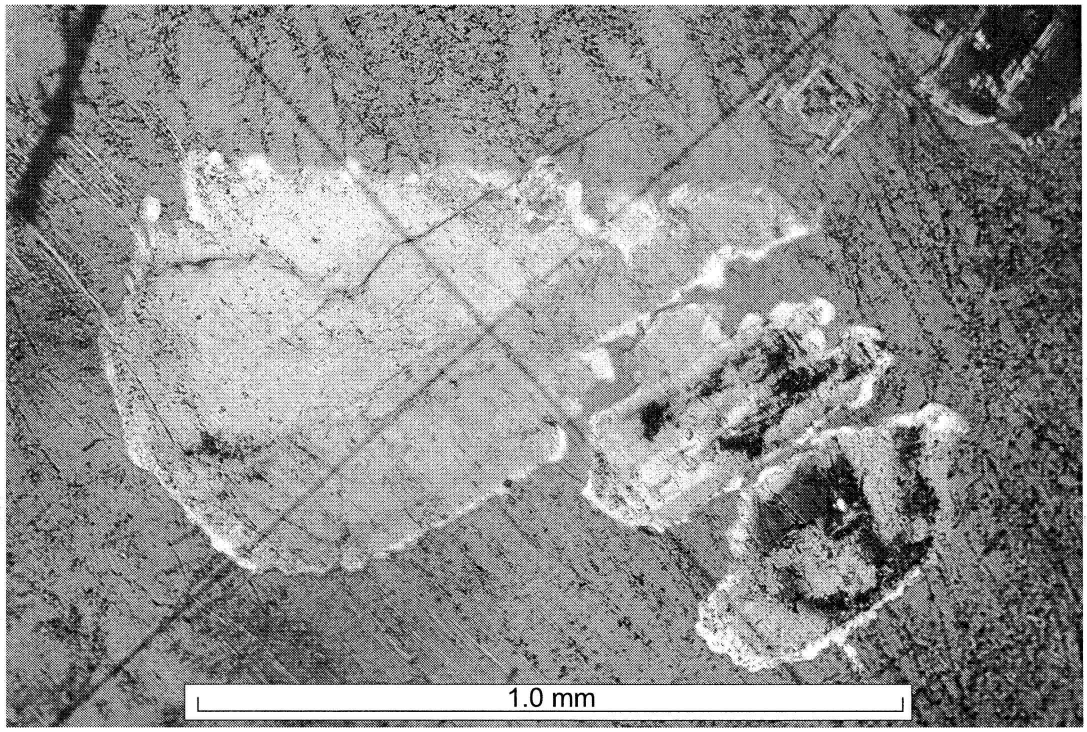


FIGURE 30. Partly replaced plagioclase crystals in a large crystal of K-feldspar. The light-coloured albite rimming the plagioclase is a decalcification effect which commonly precedes the advance of K-feldspar replacement. In places (upper left), trails of albite extend from the plagioclase rims into the K-feldspar. Where the plagioclase is more sodic, the trails become abundant and irregular, resulting in perthitic K-feldspar. From the granodiorite at Caulfeild.

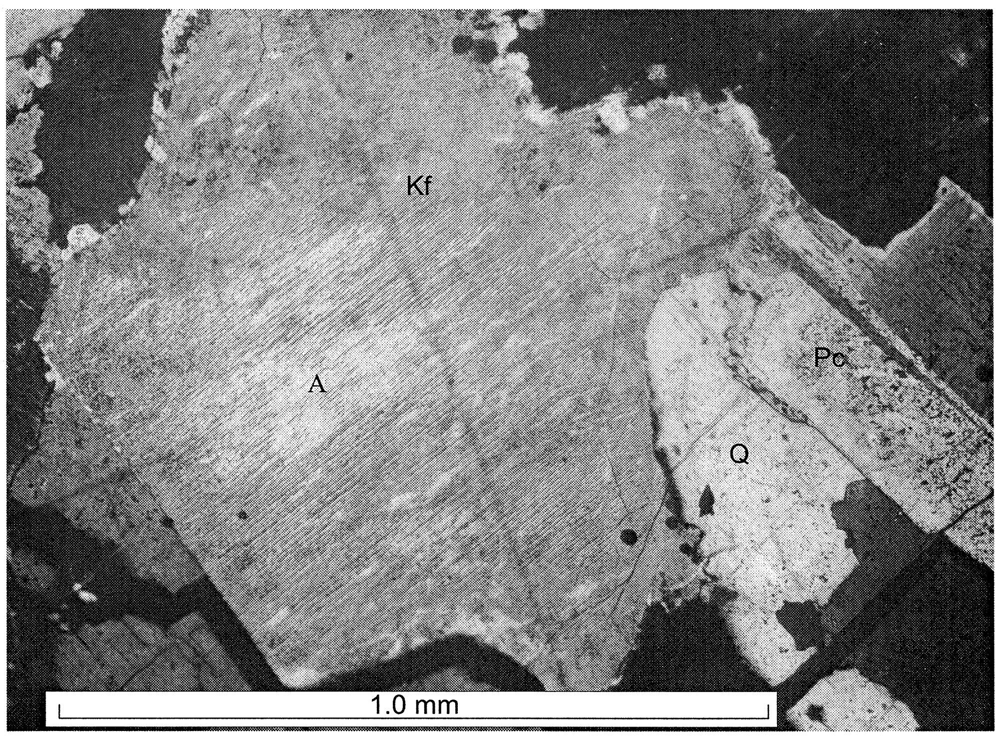


FIGURE 31. Incomplete replacement of plagioclase by K-feldspar commonly results in the preservation in the K-feldspar of lamellar twinning from the plagioclase. The light coloured areas in the K-feldspar (Kf) crystal consist of albite (A). From a granite on the south face of Grouse Mountain.

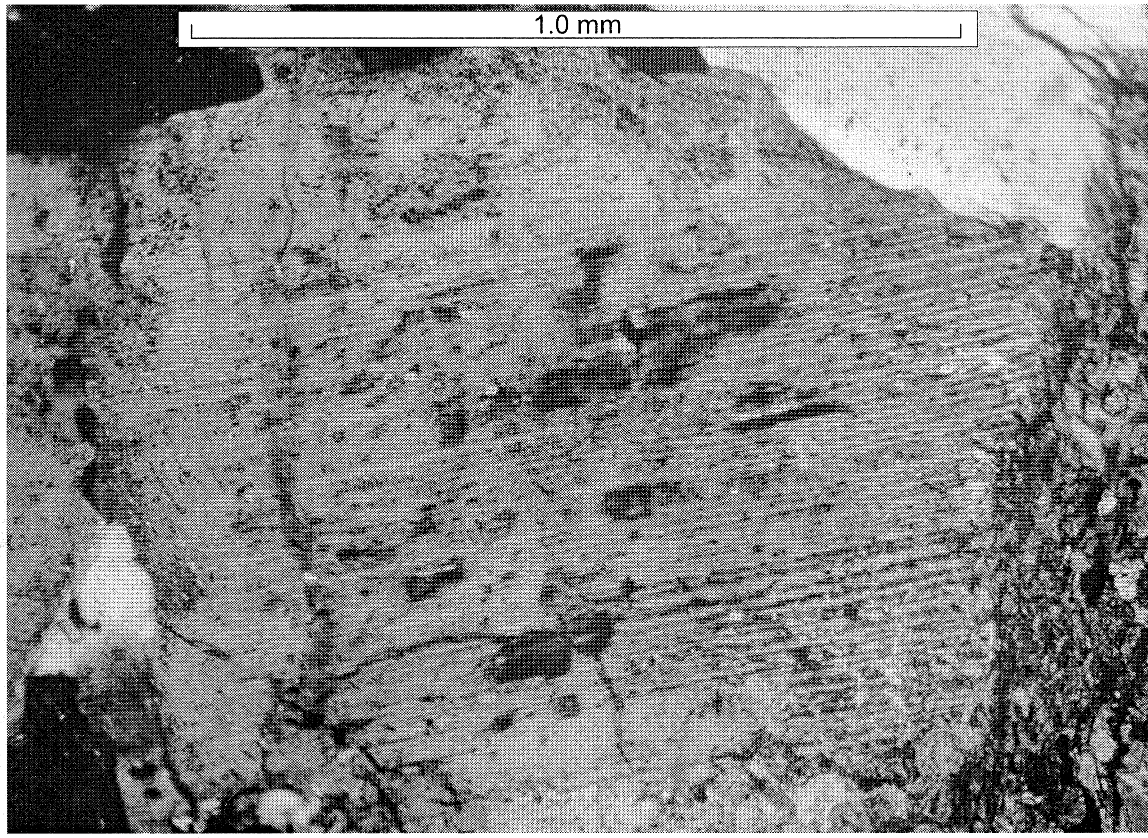


FIGURE 32. As shown here, replacement of plagioclase by K-feldspar does not always advance from the outer plagioclase surfaces inward, but may begin as internal patches which grow and eventually consume the plagioclase. The dark patches in this large crystal of plagioclase are K-feldspar, through which the lamellar twinning of the plagioclase is faintly preserved. From a granite on the south side of Sisters Creek, a tributary of Capilano River.

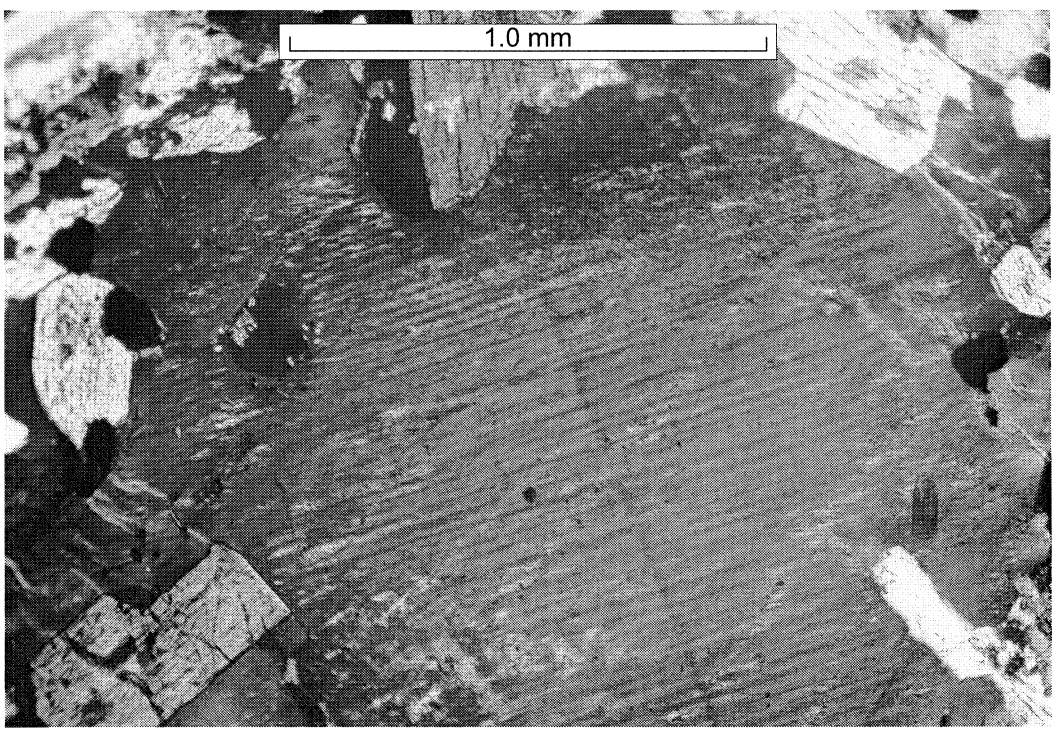


FIGURE 33. This slide shows the faint preservation of plagioclase lamellar twinning in the large K-feldspar, and also how the metasomatic encroachment of K-feldspar may bring out the crystal form of older crystals (plagioclase crystal in the lower left corner). From granodiorite at Caulfeild.

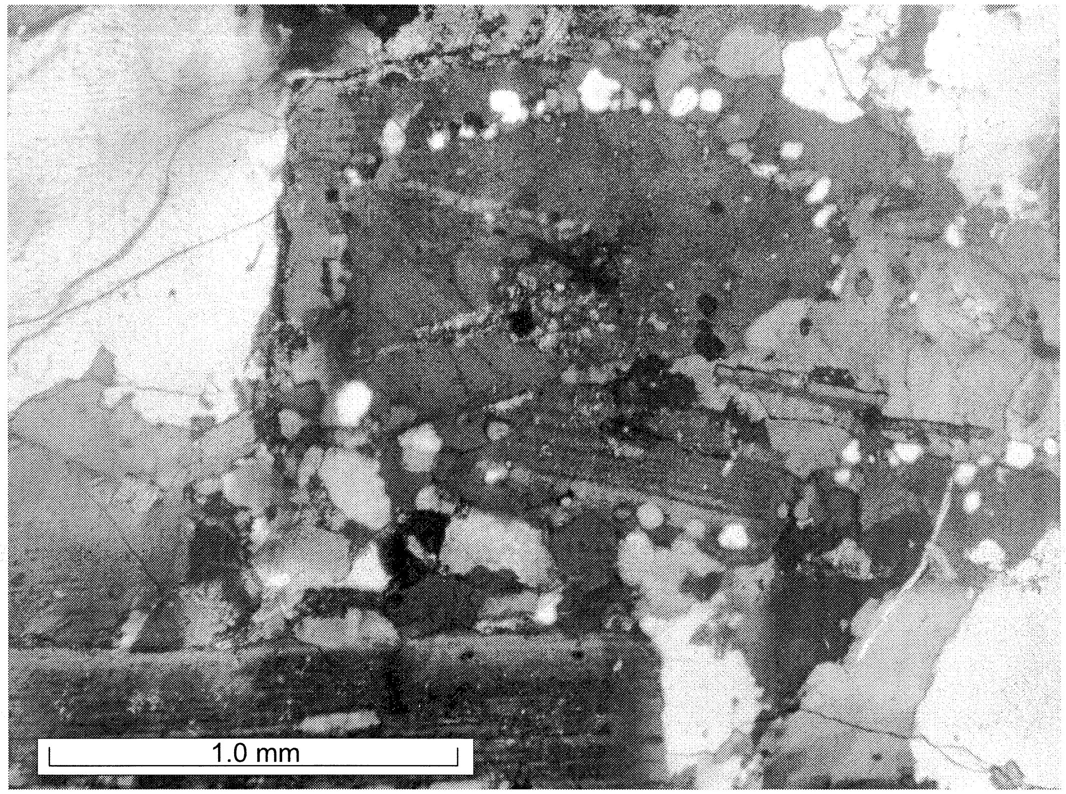


FIGURE 34. The former shape of the large plagioclase porphyroblast is outlined by a string of quartz granules, survivors probably from the original protolith. Similar relic shapes are commonly preserved in garnet porphyroblasts in some metamorphic rocks. From a granodiorite on the ridge between Indian Arm and Coquitlam Lake.

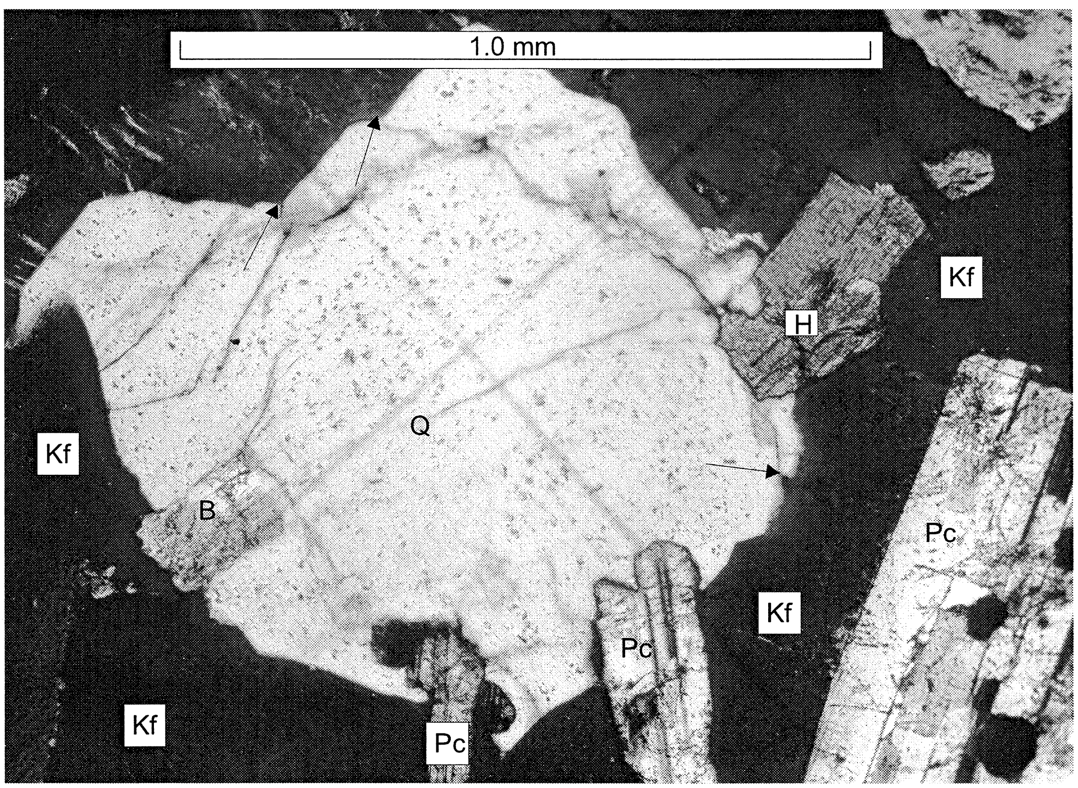


FIGURE 35. The white quartz crystal (Q) with protruding crystals of plagioclase (Pc), hornblende (H) and biotite (B) lie in a dark matrix of K-feldspar (Kf), which has replaced the quartz in preference to the calcic plagioclase, hornblende, and biotite. Note reentrants of Kf into quartz where fractured prior to the metasomatism (arrows). From a granodiorite on the south slope of Hollyburn Ridge.

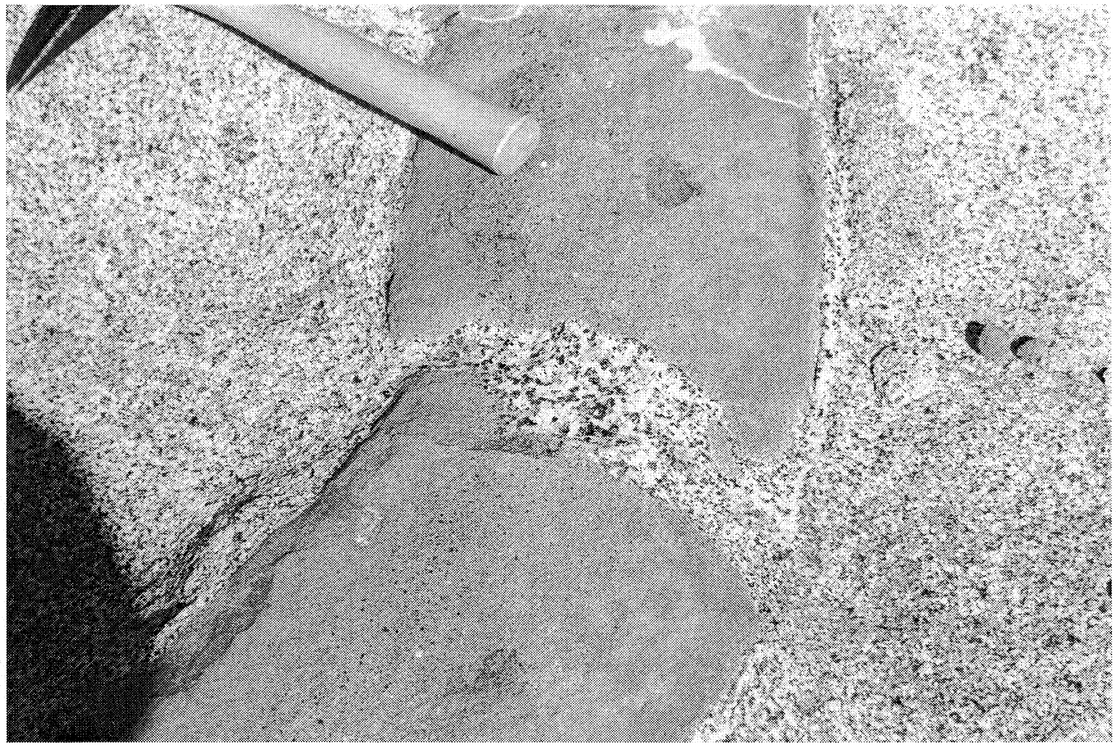


FIGURE 36. This synplutonic dyke has been torn apart by plastic movement within the pluton. The material filling the gap may have got there by plastic infill, but it is distinctly coarser suggesting accelerated mineral growth spurred by concentration of hydrous fluid there. The host rock is a quartz diorite, from the ridge between Toba River and Montrose Creek (south of Montrose Peak), Bute Inlet map-area.

FIGURE 37. A more extreme case of pulling apart of a synplutonic dyke by plastic movement within the pluton, leaving a trail of angular inclusions. Its total isolation in the pluton indicates that the fragment train is not a layer of older amphibolite. From a quartz diorite in northern Bute Inlet map area, south of the head of Stafford River. Photo by G.J. Woodsworth, 1970.

FIGURE 38. This dyke (synplutonic) has been drag-folded by plastic deformation (arrows) within the pluton. Concentration of hydrous fluid in the resulting zone of tension led to the development of coarse-grained (pegmatitic) material (mainly quartz, K-feldspar and biotite), and obliteration of that part of the dyke. From a quartz diorite outcrop in Grand Creek, which drains into the northern part of Indian Arm from the northeast.

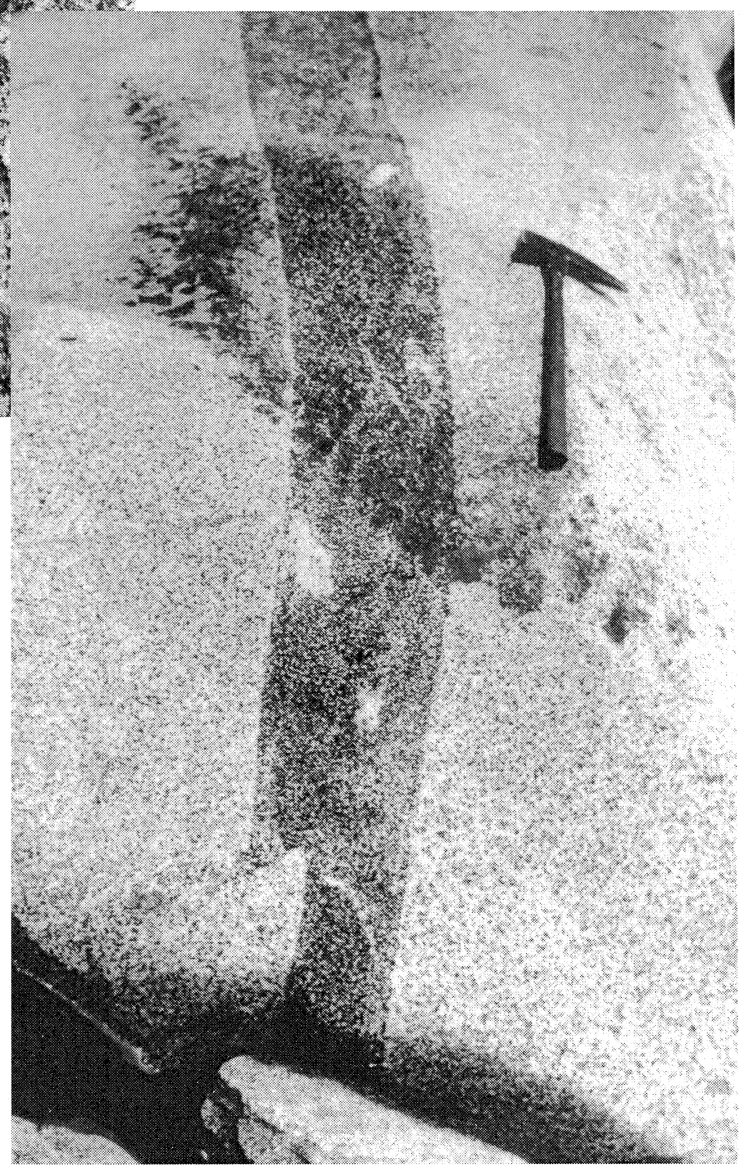


FIGURE 39. This synplutonic dyke has been frozen in the process of being obliterated by the growth of plagioclase porphyroblasts (similar in composition, size, and shape to those in the host pluton. The crystal growth is irregularly distributed, concentrated in places (white patches), sparse in others, reflecting the irregular penetration of hydrous fluid. A thin layer of hydrous fluid that gathered at the dyke surface led to the destruction of mafic minerals there by feldspathic metasomatism, giving the dyke a whitish border. Vancouver North map area.

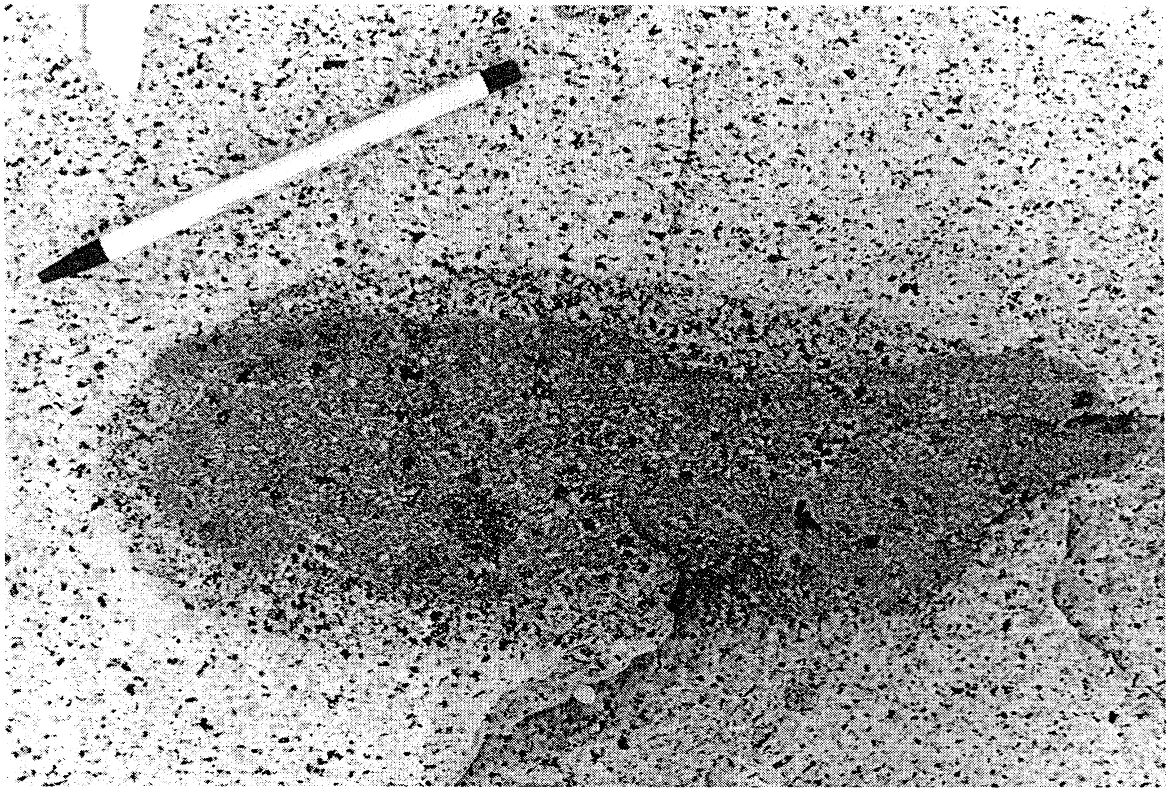


FIGURE 40. Partly replaced amphibolite inclusion in quartz diorite. Metasomatic encroachment has been frozen here at a stage where a former outline of the inclusion is preserved. Growth of plagioclase and hornblende porphyroblasts within the inclusion is also evident. South of Penrose Island, Rivers Inlet map area.

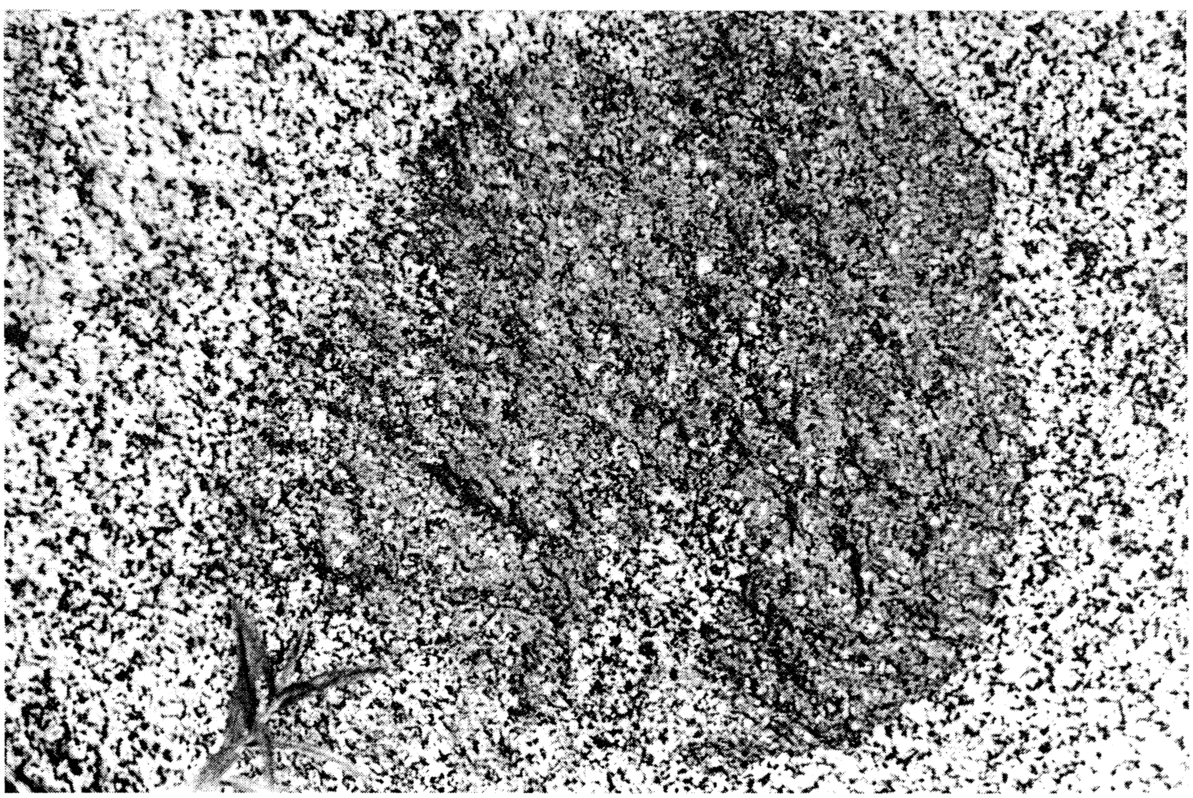


FIGURE 41. This amphibolitic inclusion is well on its way to becoming quartz diorite. From a quartz diorite pluton on the west side of Capilano River valley, north of Sisters Creek.

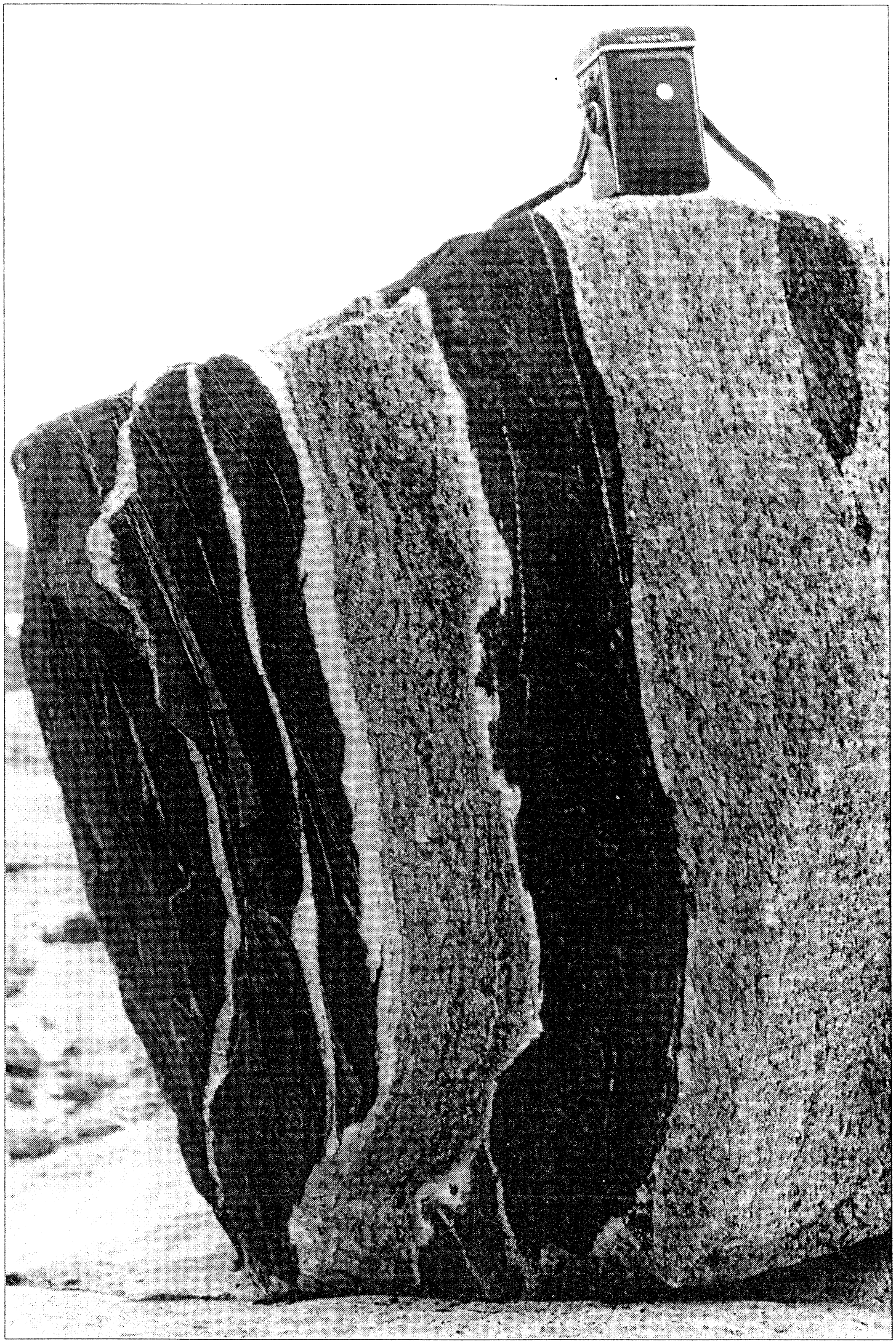


FIGURE 42. This boulder of Kemano gneiss illustrates quartzo-feldspathic growth that results from the concentration of hydrous fluid along anisotropic surfaces, in this case, layers of amphibolite (and also, of course, in fractures). Douglas Channel map area.

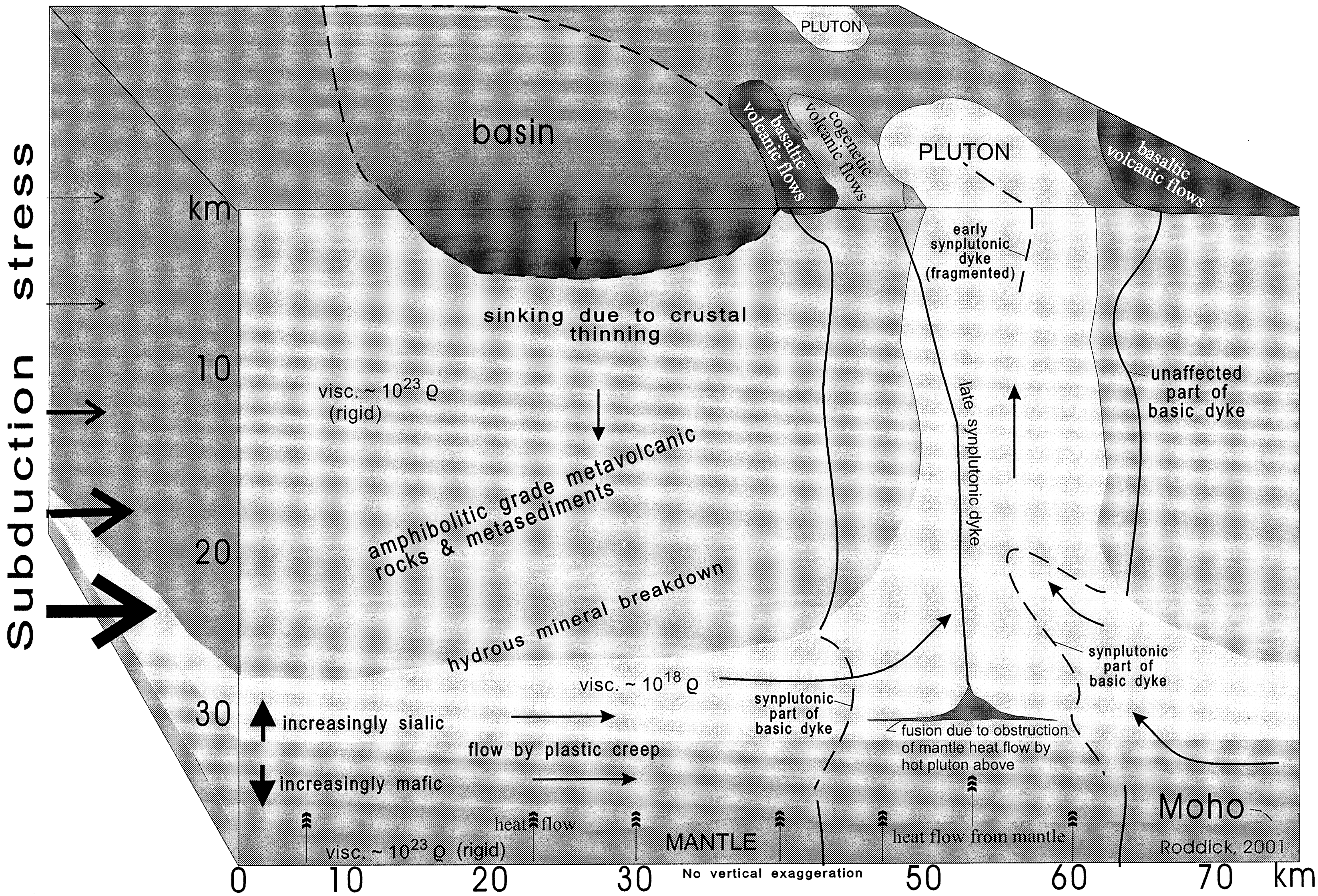


FIGURE 43. Pluton emplacement: When external stress is great enough, part of the material in the lower crust which has a lower (but still high) viscosity than the confining upper crust and upper mantle is forced upwards as plutons. The main body of a pluton must at all times maintain a high viscosity to prevent it from dyking like volcanic magmas (which does happen occasionally when water from the country rock penetrates the outer margin of a pluton).

REFERENCES CITED

Adhémar, J. A., 1842, Révolutions de la mer: privately published, Paris.

Agassiz, L., 1840, Etudes sur le glaciers: privately published, Neuchâtel.

Armstrong, J.E., 1990, Vancouver Geology: Roots, Charlie and Staargaard, Chris, eds.: Geological Association of Canada, 128 p.

Blunden, R.H., 1971, Vancouver's downtown peninsula urban geology: M.Sc. thesis, unpublished, 48 p., University of British Columbia, 1971.

Bailey, R.C., 1998, Tectonic Implications of very low mid-crustal viscosities: Eos ,Transactions, American Geophysical Union, v. 79, n. 17, Spring Meeting Supplement, T51A.

Balk, R., 1949, Structure of the Grand Saline salt dome, Van Zandt County, Texas: American Association of Petroleum Geologists Bulletin, v. 33, p. 1791-1829.

Brunhes, B., 1906, Recherches sur la direction d'aimantation des roches volcaniques: Journal de Physique Théorique et Appliquée, Series 4,5, p. 705-724.

Clague, J.J., 1977, Quadra Sand, Geological Survey of Canada, Paper 77-17, 24 p.

Croll, J., 1875, Climate and time: Appleton & Co., New York.

Daly, R.A., 1914, Igneous rocks and their origin: McGraw-Hill Book Co., New York

Dott, Robert H., and Batten, Roger L., 1971, Evolution of the earth: McGraw-Hill Book Company, 649 p.

Edwards, J.W.F., 1984, Interpretations of seismic and gravity surveys on the eastern part of the Cornubian platform, *in* D.H.W. Hutton and D.J. Sanderson, eds., Variscan Tectonics of the North Atlantic Region: Geological Society, London, Special Publications, v. XX, p. 119-124.

Ewing, M. and Donn, W.L., 1956, A theory of ice ages: Science, v. 123, p. 1061-1066.

Heywood,, W.W., 1957, Isachsen area, Ellef Ringnes Island, District of Franklin, Northwest Territories: Geological Survey of Canada Paper 56-8, 36 p.

Hoffman, Paul F., Halverson, Galen P., Schrag, Daniel P., Kaufman, Alan J., 1998, Neoproterozoic snowball Earth and early animal evolution: Geological Society of America Annual Meeting, Abstracts with Programs, p. 232.

Imbrie, John and Imbrie, K. P., 1979, Ice Ages, solving the mystery: Enslow Publishers, Hillside, New Jersey, 213 p.

Jackson, L.E., and Clague, J.J., 1991, The Cordilleran Ice Sheet: one hundred and fifty years of exploration and discovery: Géographie Physique et Quaternaire, v. 45, no. 3, p. 269-280.

Johnston, W. A., 1923, Sedimentation of the Fraser River Delta map-area: Geological Survey of Canada, Memoir 135, 87 p.

Köppen, W., and Wegener, 1924, Die Kimate der Geologischen Vorzeit: Gebrüder Borntraeger, Berlin.

Magaritz, Mordeckai, and Taylor, Hugh P. Jr, Oxygen 18/oxygen 16 and D/H studies of plutonic granitic and metamorphic rocks across the Cordilleran batholiths of southern British Columbia: Journal of Geophysical Research, v. 91, p. 2193-2217.

Mathewes, R.W., 1979, A paleoecological analysis of Quadra Sand at Point Grey, British Columbia, based on indicator pollen: Canadian Journal of Earth Sciences, v. 16, p. 847-858.

Matthews, D. H., 1987, Can we see granites on seismic reflection profiles?: Annales Geophysicae, v. 5B, no. 4, p. 353-356.

Matuyama, M., 1929, On the direction of magnetisation of basalt in Japan: Imperial Academy of Japan Proceedings, v. 5, p. 203-205.

Mooney, W.D. and Meissner, R., 1992, Multi-genetic origin of crustal reflectivity, *in* D.M. Fountain, R. Arculus and R.W. Kay, eds., Continental Lower Crust: Elsevier, Developments in Geotectonics 23, p. 45-79.

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

Paterson, S.R., Miller, R.B., Anderson, J.L., Lund, Steve, Bendixon, John, Taylor, Nick, and Fink, Tom, 1994, Emplacement and evolution of the Mt. Stuart batholith: Geological Society of America Annual Meeting, p. 2F1-2F47.

Phemister, T.C., 1945, The Coast Range Batholith near Vancouver, B.C: Quarterly Journal, Geological Society of London, v. 101, p. 37-89.

Roddick, J.A., 1965, Vancouver North, Coquitlam and Pitt Lake map areas: Geological Survey of Canada, Memoir 335, 276 p.

Roddick, J.A., 1983, Geophysical review and composition of the Coast Plutonic Complex, south of latitude 55° N: *in* J.A. Roddick, ed., Circum-Pacific Plutonic Terranes: Geological Society of America Memoir 159, p. 195-212.

Roddick, J.A., and Hutchison, W.W., 1972, Plutonic and associated rocks of the Coast Mountains of British Columbia: Guidebook for field excursion AC04, 24th International Geological Congress, Montreal, Quebec, 71 p.

Ronov, A.B., and Yaroshevsky, A.A., 1969, Chemical composition of the earth's crust: *in* P.J. Hart, ed., The earth's crust and upper mantle: American Geophysical Union, Geophysical Monograph 13, p. 37-57.

Rouse, G.E., 1962, Plant microfossils from the Burrard Formation of western British Columbia: Micropaleontology, v. 8, p.187-218.

Rouse, G.E., Mathews, W.H., and Blunden, R.H., 1975, The Lions Gate Member; a new Late Cretaceous sedimentary subdivision in the Vancouver area of British Columbia: Canadian Journal of Earth Sciences, v. 12, p. 464-471.

Rubin, Allan M., 1995, Getting granite dikes out of the source region: Journal of Geophysical Research, v. 100, no. B4, p. 5911-5929.

Stewart, D.B., Unger, J.D., Phillips, J.D., Goldsmith, R., Poole, W.H., Spencer, C.P., Green, A.G., Loiselle, M.C., and St. Julien, P., 1986, The Quebec-Western Maine seismic reflection profile, *in* Barazangi, M., and Brown, L.D., eds., Reflection Seismology: the Continental Crust: American Geophysical Union, Geodynamics Series 14, p.189-199.

Streckeisen, A.L., 1976, To each plutonic rock its proper name: Earth Science Reviews, v. 12, p. 1-33.

Turner, R.J.W., Clague, J.J., Groulx, B.J., 1996, Geoscape Vancouver: Geological Survey of Canada, Open File 3309.

Varsek, J.L., Cook, F.A., Clowes, R.M., Journeay, J.M., Monger, J.W.H., Parrish, R.R., Kanasewich, E.R., and Spencer, C.S., 1993, Lithoprobe crustal reflection structure of the southern Canadian Cordillera: Tectonics, v. 12, p. 326-360.

Warner, M., 1990, Basalts, water, or shear zones in the lower crust?, *in* J.H. Leven et al., eds., Seismic Probings of Continents and their Margins: Tectonophysics, v. 173, p. 163-174

Wooton, A.E., 1959, The Tertiary eruptives at Vancouver, British Columbia: B.Sc. thesis, unpublished, University of British Columbia, 36 p

Zielinski, Gregory A., 1999, Use of paleo-records in determining variability within the volcanism-climate system: Quaternary Science Reviews, v. 19, p. 417-438.