UNIVERSITY OF ALBERTA

HYDROTHERMAL ALTERATION AND FLUID SOURCES ASSOCIATED WITH THE BABINE LAKE PORPHYRY COPPER DEPOSITS, WEST-CENTRAL BRITISH COLUMBIA

by

GERARD ZALUSKI

A thesis submitted to the Faculty of Graduate Studies and Research in partial fulfillment of the requirements for the degree of Master of Science

in

GEOLOGY

DEPARTMENT OF GEOLOGY

EDMONTON, ALBERTA

FALL, 1992

ISBN 0-315-77291-3

UNIVERSITY OF ALBERTA

RELEASE FORM

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MASTER OF SCIENCE

FALL, 1992

229 14th Street
WEYBURN, Saskatchewan
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June 9 1992

date

"...what is manifest and certain is known to us at first confusedly; afterwards we know it by distinguishing its principles and elements."

Aristotle

UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled HYDROTHERMAL ALTERATION AND FLUID SOURCES ASSOCIATED WITH THE BABINE LAKE PORPHYRY COPPER DEPOSITS, WEST-CENTRAL BRITISH COLUMBIA submitted by GERARD ZALUSKI in partial fulfillment of the requirements for the degree of MASTER OF SCIENCE in GEOLOGY.

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ABSTRACT

The Eocene Babine Intrusive Suite of west-central British Columbia hosts a number of porphyry copper deposits, most significantly the Bell, Granisle, and Morrison deposits. All deposits feature a central potassic zone containing the ore zone and a peripheral propylitic zone. In addition, Granisle and Bell feature superimposed sericite-carbonate zones between the potassic and propylitic zones and the Bell deposit also features a superimposed phyllic stockwork zone which hosts most of the ore. Weak supergene alteration overprints the phyllic alteration in the upper levels. The Morrison deposit features late superimposed clay±carbonate alteration along the Morrison Fault and subparallel structures.

Calculated fluid compositions from potassic zone biotites yield a range from $\delta^{18}O_{H2O}$ = 6 to 8‰ whereas plagioclase yield $\delta^{18}O_{H2O}$ = 1 to 7‰. Since plagioclase yield more ^{18}O-depleted fluids than coexisting biotites, and since $\Delta_{plag\text{-}bt}$ values are very small, indicative of disequilibrium, it is evident that isotopically lighter fluids were present in later stages of potassic alteration. Oxygen isotope values from the sericite-carbonate and phyllic alteration zones yield fluid compositions more ^{18}O-depleted than magmatic fluids, requiring the incorporation of isotopically light groundwaters.

Calculated hydrogen isotope values of all potassic, propylitic, and sericite-carbonate fluids fall in the range of -40 to -110‰. This, in combination with the $\delta^{18}O$ data suggest that these alteration events were caused by mixtures of magmatic and isotopically depleted meteoric waters. Phyllic and argillic alteration fluids (δD_{H2O} = -100 to -130 and -140‰ respectively) indicate fluids dominantly of meteoric origin.

The stable isotope data in combination with the petrologic and field relations suggest that early alteration involved fluids derived from the magma and those which interacted with the rocks at low water/rock ratios. Later alteration involved the incorporation of greater amounts of external fluids, further from chemical and isotopic equilibrium with the rocks. These were responsible for the more texturally destructive, cation-leached alterations and remobilization of the ores. The incorporation of external fluids into the hydrothermal systems of porphyry copper deposits may either upgrade or downgrade the deposits but invariably changes the character of the alteration zonation.

ACKNOWLEDGEMENTS

The author wished to acknowledge all those who helped throughout the course of this study. I wish to thank Bruce Nesbitt for supervising this project. I also wish to thank the staff of the Bell mine of Noranda Minerals Incorporated for their logistical support during the fieldwork of this study. Especially, I wish to thank Maggie Dittrick for her help and time in showing me the properties, explaining the geology, and helping in many ways with the sample collection. For financial support, I wish to thank the Natural Sciences and Engineering Research Council of Canada for a Post-Graduate Scholarship and the University of Alberta for the Walter H. Johns Graduate Fellowship which supported me during this study.

I also wish to thank my fellow graduate students who expressed interest and instigated helpfull discussions on this subject. Several of the technical staff of the Department of Geology at the University of Alberta were of great help in the completion of this study. These include Don Resultay for making thinsections, Dianne Caird for x-ray diffraction analyses, and Paul Wagner for microprobe analysis.

I also wish to thank Karlis Muehlenbachs for access to his laboratory, without which the stable isotope analyses would have been impossible. Olga Levner, for great assistance with the oxygen isotope analyses and James Steer for much of the hydrogen isotope analyses are also greatly thanked. Also deserving recognition is T.K. Kyser at the University of Saskatchewan for use of his laboratory to analyze necessary samples and for showing interest and providing valuable insights into this study.

Finally, I wish to thank my family for their support during the past two years. Most of all, I wish to thank my fiancée Mayda Petro for her great support and patience throughout the past two years.

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Porphyry Copper Deposits and Models

Defintion and Characteristics

Porphyry copper deposits are defined as large, low grade, epigenitic, intrusion-related copper deposits that can be mined by bulk mining techniques (McMillan and Panteleyev, 1980). These deposits are spatially and genetically related to igneous intrusions, typically epizonal, felsic (usually quartz monzonitic), and porphyrytic and often featuring multiple intrusive pulses, dike swarms, intrusive breccias, and pebble dikes (McMillan and Panteleyev, 1980). Although deposits may be situated in batholithic or volcanic settings, the "classic", most common deposits are found in high level, subvolcanic, post-orogenic stocks intruding unrelated host rocks (McMillan and Panteleyev, 1980). Porphyry copper deposits are associated with orogenic belts in island arcs and continental margins, especially those of Mesozoic and Cenozoic age (McMillan and Panteleyev, 1980). As stated by Henley and McNabb (1978), "Perhaps the most fascinating aspect of these deposits is the very high frequency by which deposits of such similar geologic character have been generated in the crust, and this frequency in turn implies the common reproduction of a special physiochemical environment of formation through geological time.".

Porphyry copper deposits comprise one half of the world's copper reserves and 60% and 90% of the Canadian and British Columbian copper resources respectively (McMillan and Panteleyev, 1980). According to McMillan and Panteleyev (1980), a typical giant porphyry copper deposit (2 billion tons at 1.5% Cu) could supply Canada's copper consumption for 130 years or the world's copper consumption for three years. In addition, some porphyry copper deposits, including the Diorite Model deposits of Hollister (1978) and some calc-alkaline porphyry deposits (particularly in the Circum-Pacific and Caribbean island arc deposits (Kesler, 1973)) have significant gold grades and relatively high Au/Cu ratios. Included in this group is Granisle, having the highest Au/Cu ratio of western Canada's producing porphyry copper deposits (Kesler, 1973). Bell and Morrison plot within the Cu-Mo-Au class of Sutherland Brown (1976b). It is therefore evident that these deposits are of considerable economic significance as well as being of academic interest.

The typical porphyry Cu deposit as described by Lowell and Guilbert (1970) is hosted by a Laramide age, differentiated quartz diorite to quartz monzonite stock (1225 m x 1850 m) emplaced in Late Cretaceous sediments or metasediments and controlled by regional scale strike-slip faulting. The ore body is pipe-like in shape, with rough dimensions of 1075 m by 1850 m and containing 140 million tons of ore, 70% being in the igneous rocks and 30% being in the country rocks, grading 0.45% hypogene Cu, 0.015% Mo, and

0.35% supergene Cu (Lowell and Guilbert, 1970). Copper mineralization is commonly restricted to specific intrusion phases (Nielsen, 1976). Host rocks are altered by hydrothermal solutions into roughly concentric zoning patterns with an outward progression of potassic, phyllic, argillic, to propylitic alteration although all zones are not necessarily present in any particular deposit (Lowell and Guilbert, 1970). Large deposits generally have more regular and well developed alteration zonation patterns (Hollister, 1978). The ore zone is often annular in shape and overlaps the potassic-phyllic boundary and ranges from disseminations in the high grade core through microveinlets and veinlets to high grade peripheral veins, reflecting the increasing effects of structure towards the periphery (Lowell and Guilbert, 1970).

Porphyry copper deposits can be divided into "Wet" types, having high pyrite/chalcopyrite ratios and being surrounded by enormous pyrite-sericite-quartz halos (such as Bingham and Morency) and "Dry" types with low sericite contents (such as Bethlehem) (Lowell and Guilbert, 1970).

Models and Objectives

A number of models have been developed to explain the genesis of porphyry copper deposits. The orthomagmatic models of Burnham (1967) and Nielsen (1968) postulate that a shallow intrusion (900 to 1500 m) becomes water saturated, hydraulically fractures the outer, crystallized shell of the intrusion and releases a volatile phase which causes the alteration and mineralization. Fournier (1968) modified this model by postulating that the volatile phase evolution is due to fault-induced rupture resulting in a sudden loss of H_20, supercooling, and the consequent alteration and mineralization. The hydrothermal models of White (1968) and Norton (1978) postulate an external source for the mineralizing fluids with thermally driven, convected groundwater forming the sulfur-deficient, base metal-rich, Na-Ca-Cl brines by water-rock interaction.

As explained by Beane and Titley (1981), the two-fluid model involves an early, central magmatic stage responsible for potassic alteration being encroached upon by a later, collapsing, meteoric water stage responsible for the phyllic and argillic alteration zones. This model has various forms invoking different degrees of involvement of meteoric fluids in the ore forming process ranging from meteoric fluids being involved only in a post-ore overprint to meteoric fluids being involved in ore formation by fluid mixing.

It has often been observed that mineralization patterns are related to alteration zoning. Lowell and Guilbert (1970) and Henley and McNabb (1978) stated that the ore zone typically straddles the potassic-phyllic boundary. Carson and Jambor (1974) concluded that in the Babine Lake deposits of west-central British Columbia, the size and grade of each copper zone correspond to the areal extent and quality of hydrothermal biotite. Beane and Titley (1981) stated that the most common association is chalcopyrite and potassic alteration, particularly biotite. The hydrothermal alteration associated with porphyry copper deposits is essentially a base-leaching process controlled by the cation/hydrogen ion ratio in the solution (McMillan and Panteleyev, 1980). Since the silicate alteration reactions control those of the opaque phases (Beane and Titley, 1981), an understanding of ore formation requires an understanding of hydrothermal alteration. As a consequence, the source, composition, and evolution of the alteration fluid are important factors relating to ore formation.

Studies of porphyry copper and molybdenum and other types of intrusion-related ore deposits have postulated a number of sources for hydrothermal fluids, including magmatic, meteoric, formation, connate, and seawaters. The purpose of this thesis is to evaluate the roles of magmatic and meteoric fluids (and other external waters) in the formation of porphyry copper deposits.

As discussed earlier, models for porphyry copper deposits can be divided into two main groups, the orthomagmatic models (postulating that fluids and ores originate from the magma) and the convective models (attributing fluids and ores to an origin outside of the intrusive). Essentially, according to the orthomagmatic model and its variations, ore deposit formation depends on the evolution of a magmatic volatile phase and effective precipitation mechanisms and therefore is controlled internally, by the composition, evolution, and crystallization history of the magma. The convective model by contrast requires convective circulation of external fluids and therefore the production of mineralizing solutions depends on the permeability of the host rock as well as effective precipitation mechanisms.

Apart from the theoretical interests of understanding the genesis of porphyry copper deposits, the different models with their associated controls and characteristics have important implications for porphyry ore deposit exploration. Exploration based on the premise of the orthomagmatic model would focus on the location of areas which host magmas of favorable composition and evolutionary history to produce ore deposits. Exploration according to the convective model would focus on areas of sufficient permeability to allow hydrothermal convection systems to develop and hence key targets would be intrusions in brittle rocks in a structural setting favourable for the development of high permeabilities and convective fluid circulation.

Because different fluid types may have distinguishable stable isotope compositions, stable isotopes have been used to determine the sources of fluids responsible for rock alteration. Various stable isotope studies of porphyry copper deposits (e.g., Sheppard *et al.*, 1969, 1971; Sheppard and Taylor, 1974; Sheppard and Gustafson, 1976; Bowman *et* *al.*, 1987; and Norman *et al.*, 1991) have been undertaken, most of these being focussed on deposits from the southwestern United States. However, as pointed out by Sheppard *et al.* (1969), Sheppard *et al.* (1971), and Criss *et al.* (1991), the most favourable locations in which to determine the importance of meteoric waters in alteration and mineralization are continental interiors, high latitudes, and high altitude areas where the isotopic contrast between meteoric and magmatic fluids is the greatest. Unfortunately, stable isotope data from porphyry ore deposits in the Canadian Cordillera with its distinctive, isotopically light meteoric waters has been very limited.

The Babine Lake Intrusive Suite of west-central British Columbia hosts several economic and subeconomic porphyry copper deposits. The Eocene age of the suite indicates that the latitude and climatic conditions at the time of intrusion and mineralization were similar to those of today and therefore meteoric waters would have been isotopically distinct from magmatic fluids at the time of ore formation. This area is therefore an ideal location for the evaluation of relative roles of magmatic versus meteoric fluids in ore formation.

Carson *et al.* (1976) noted the increase in the development of sericitization from Morrison, with potassic and propylitic alteration zones and only localized, minor sericitization, through Granisle with its locally prominent sericite-carbonate alteration, to Bell with its well developed sericite-carbonate zone and superimposed quartz-sericite alteration. They interpreted this alteration at Bell to be due to an influx of meteoric fluids or "meteoric collapse". This area serves as an ideal study area since the ages and all features of the intrusions are comparable, with the only significant differences being the intensity of the sericite alteration and the associated ore character. By comparing the stable isotope data from these deposits the relative roles of meteoric fluids in different alteration events can be evaluated. This information will be combined with petrographic features, mineral stabilities, and an analysis of the fO_2 and fS_2 evolution of the deposits to determine the conditions under which the Babine porphyry copper deposits formed.

<u>Location and Access</u>

The Babine Intrusive Suite is located near Babine Lake and the town of Granisle in west-central British Columbia at 45° 53' N and 126° 12' W (UTM Zone 9, 679 400, 6 085 500). The town of Granisle is located 280 km west of Prince George, approximately midway between Prince George and Prince Rupert. Access to the area is via the Yellowhead Highway from Prince George to Topley and north 65 km by secondary highway to the town of Granisle. Figure 1 shows the location of the Babine Lake area and the town of Granisle. All three of the deposits studied are situated on the east side of Babine Lake. The Bell mine is situated on the Newman Peninsula, Granisle is located on

McDonald Island, and Morrison is located near the southeast end of Morrison Lake, 22 km north of Bell. Access to the properties is by the Noranda Minerals barge to Bell or Granisle, with numerous forestry roads providing access to Morrison and other locations.

Physiography

The Babine Lake porphyry copper deposits are located on the Nechako Plateau, the northernmost part of the Interior Plateau, an area of low relief with flat to gently rolling topography, largely covered by glacial sediments (Carter, 1982). The Nechako Plateau, shown in Figure 2 (after Carter, 1982) is bounded to the north by the Hazelton and Skeena Mountains, to the east by the Omineca Mountains and to the west by the Coast Mountains.

The Skeena Arch, a prominent early Mesozoic transverse structure (Carter, 1982) cuts across the area from Morice Lake eastward to Babine Lake. Carter (1982) states that the Late Triassic and older rocks exposed along the arch and the gneissic rocks east of Babine Lake along a projection from the arch suggest that it is a zone of uplift. This arch is the approximate boundary between the Bowser successor basin to the north and the Nechako Trough to the south. The major tectonic elements of the region are shown in Figure 3.

Previous Research

Porphyry copper deposits have been the subject of numerous studies. Descriptive, geochemical and geodynamic studies include those of Lowell and Guilbert (1970), Titley and Beane (1981), Beane and Titley (1981), Hollister (1978), and McMillan and Panteleyev (1980), as well as two special volumes edited by Titley and Hicks (1966) and Sutherland Brown (1976a). In addition, many studies of stable isotopes from porphyry copper deposits have been published. These include Garlick and Epstein (1966), Sheppard *et al.* (1969), Sheppard *et al.* (1971), Sheppard and Taylor (1974), Sheppard and Gustafson (1976), Bowman *et al.* (1987), Norman *et al.* (1990), and Kusakabe *et al.* (1990).

Regional mapping in this area was undertakin by Carter (1966, 1967a, 1973, 1974, 1982), Richards, (1973, 1974), Tipper (1972), and Richards and Dodds (1973). Reports of the Babine Lake porphyry copper deposits include Carter (1967, 1971), Fahrni (1967), and Bell (1970). Kirkham (1971) described intermineralization intrusions at Granisle. A detailed study of Babine Lake alteration and mineral zonation patterns was undertaken by Carson and Jambor (1974) (with geochemical data published by Jambor, 1974) and clay mineral variations in Bell were studied by Jambor and Delabio (1975). Studies of the Bell, Granisle, and Morrison properties were undertaken by Carson *et al.* (1976), Fahrni *et al.* (1976), and Carson and Jambor (1976) respectively. Wilson *et al.* (1980) and Quan *et al.* (1987) undertook detailed studies of fluid inclusions from Bell and Granisle and Cuddy and Kesler (1982) studied gold distribution at Bell and Granisle.

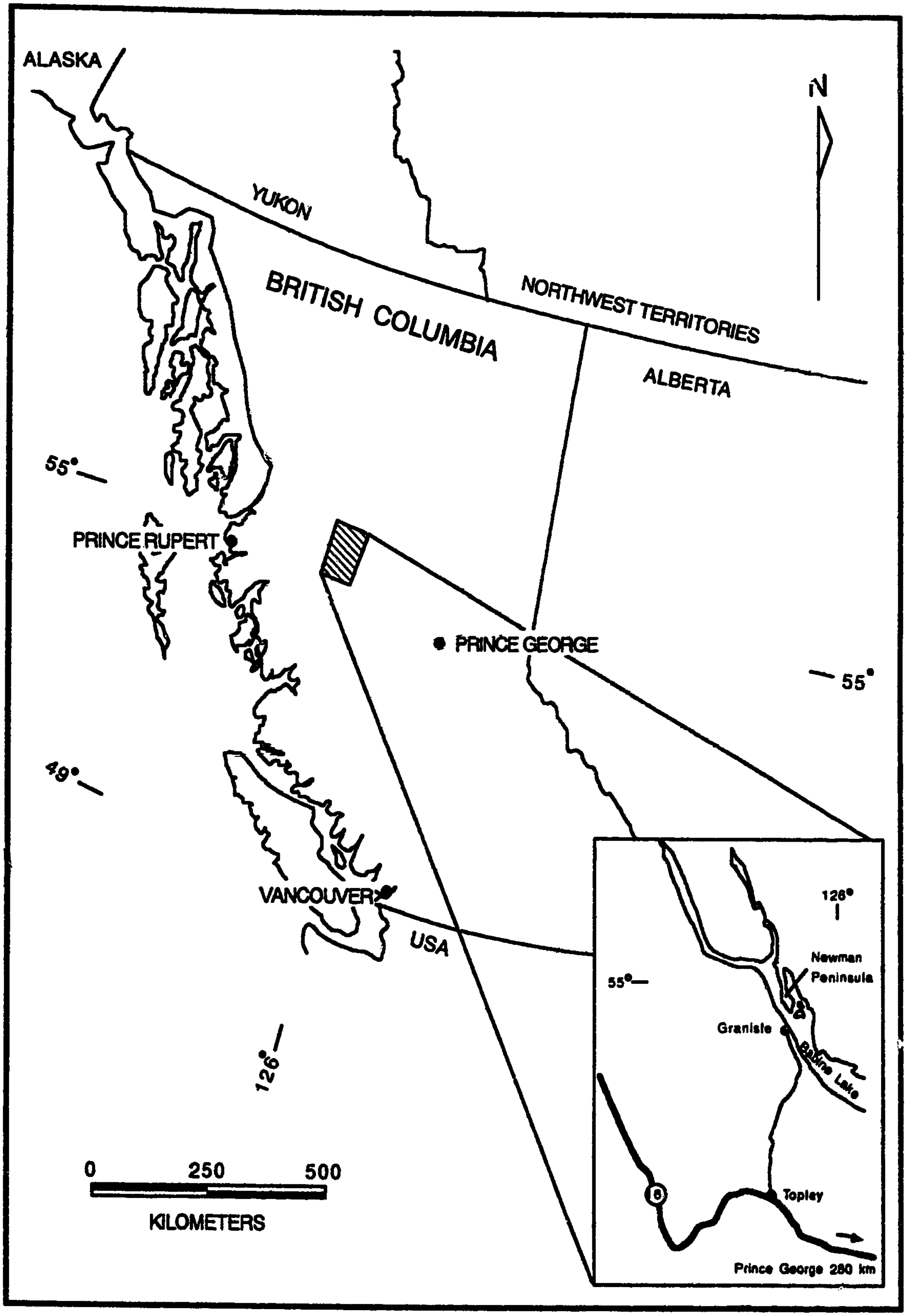

Figure 1. Location and access of Babine Lake and the town of Granisle, British Columbia.

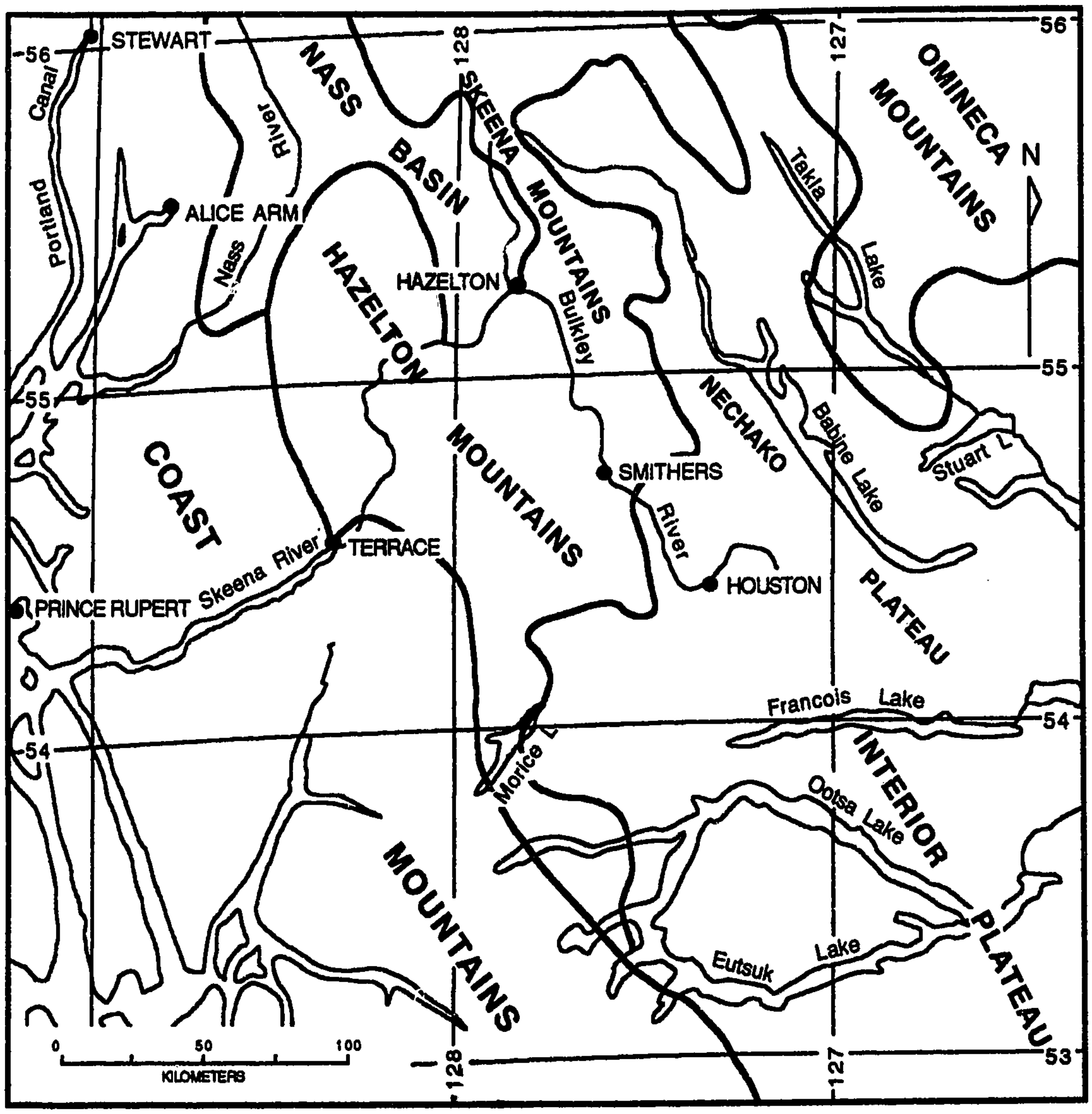

Figure 2. Physiography of the Babine Lake area and west-central British Columbia (after Carter, 1982).

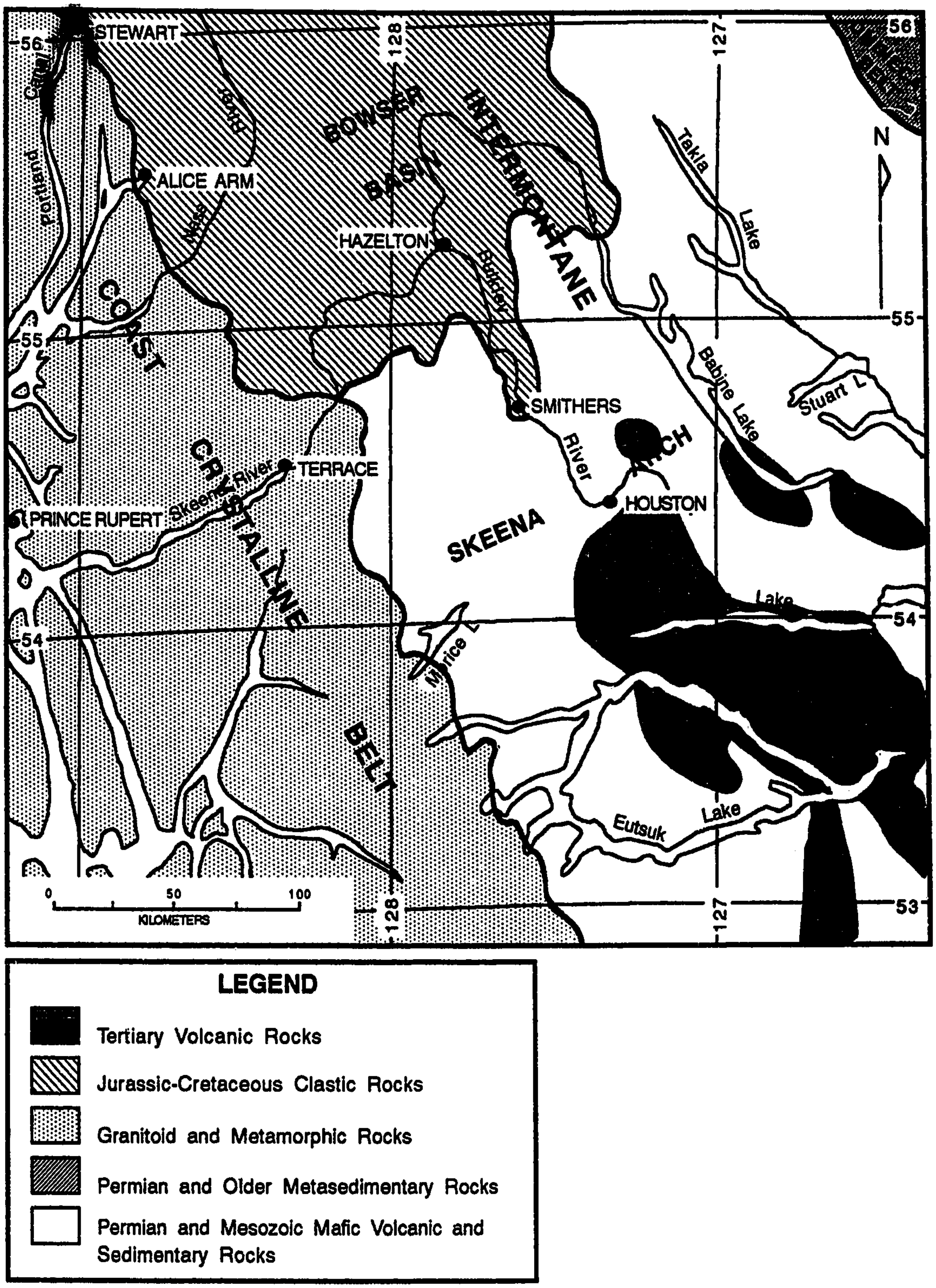

Figure 3. Tectonic elements of west-central British Columbia (after Carter, 1982).

2. REGIONAL GEOLOGY

Introduction

Porphyry copper deposits display a strong association with orogenic belts (e.g., Hollister, 1978; McMillan and Panteleyev, 1980; and Titley and Beane, 1981) and therefore an understanding of the tectonic setting is helpful in understanding porphyry copper deposits. However, most deposits appear to have formed during a lull in batholith emplacement, in a period of strike-slip faulting (Hollister, 1978). Since few deposits actually occur on strike-slip faults, it appears that the strike-slip faulting characterizes the tectonic style in which such deposits develop rather than being an immediate control (Hollister, 1978).

In order to understand the development of the Babine Lake porphyry copper deposits and the geological history of the area it is helpful first to review briefly the development and setting of this area in the context of the tectonic development of the Canadian Cordillera. This serves to familiarize one with the rock units and their origins and to help one to understand the place of the deposits and porphyry deposits in general with their geodynamic setting.

Tectonic Setting

The Canadian Cordillera is composed of five distinct physiographic or tectonic belts as shown in Figure 4. These include (from east to west) the Foreland Thrust Belt or Mackenzie and Rocky Mountains Belt, the Omineca Crystalline Belt, the Intermontane Belt, the Coast Plutonic Complex, and the Insular Belt (Monger *et al.*, 1982; Monger, 1984; Gabrielse and Yorath, 1989; and Armstrong, 1988).

The Foreland, Intermontane, and Insular Belts consist of unmetamorphosed to low grade metamorphic rocks and make up the suprastructure of the Cordillera (Monger *et al.*, 1982). The Omineca Crystalline Belt and Coast Plutonic Complex are major regional welts consisting of metamorphic and plutonic infrastructure of the Cordillera which was metamorphosed and intruded during collisional and subduction-related orogenies from the Jurassic to the Tertiary (Monger *et al.*, 1982). The Intermontane Belt (in which this study area is located) is situated between the Coast Plutonic Complex to the west and the Omineca Crystalline Belt to the east.

Regional Geology and Tectonics

Tectonic Development

As explained by Monger (1984), the creation of the western Cordillera of North America was a result of three processes: 1) The accretion of terranes to the ancient continental margin, 2) subduction of Pacific Ocean lithosphere, and 3) northward translation of terranes along transcurrent faults. Although these processes overlapped

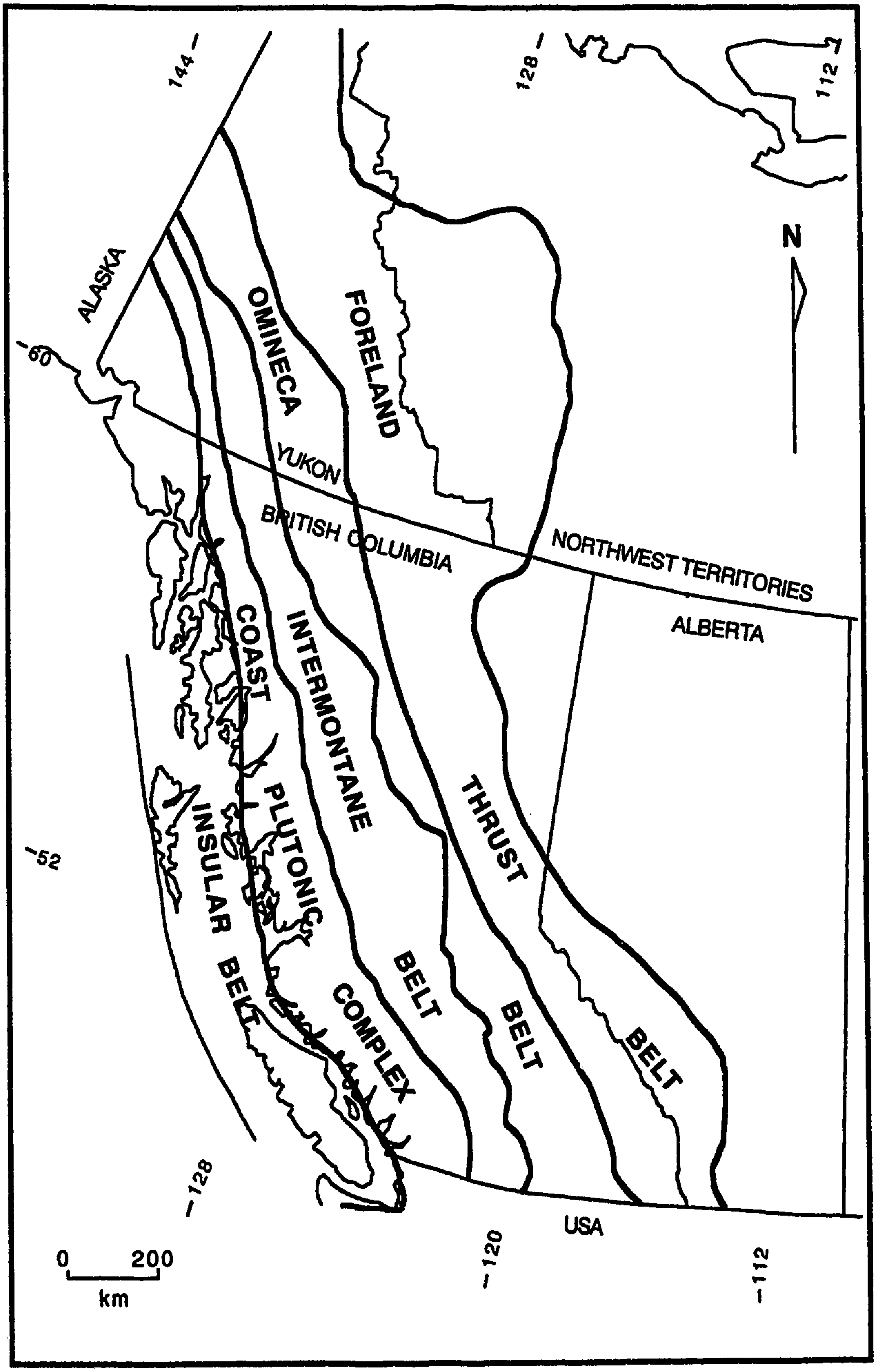

Figure 4. Tectonic belts of the Canadian Cordillera (modified from Armstrong, 1988).

temporally, their dominances in terms of controlling the general tectonic style of the Cordillera were in the order listed. The accretion of two large composite terranes in the Middle Jurassic and Middle Cretaceous created the two structural and metamorphic belts discussed previously during the respective Columbian and Pacific Orogens of Wheeler and Gabrielse (1972). The subduction of thousands of kilometers of Pacific Ocean crust below the western edge of North America from the Late Jurassic to the present resulted in extensive Cretaceous to early Tertiary calc-alkaline magmatism, created the Coast Plutonic Complex, and caused much of the eastern Cordilleran thrusting (Monger, 1984). Post-accretionary northward translation along major transcurrent faults during the Late Cretaceous and early Tertiary resulted in displacements of terranes of over 1000 km (Monger, 1984 and Gabrielse and Yorath, 1989).

These processes outlined by Monger (1984) for the development of the North American Cordillera are responsible for the geologic and tectonic features observed in this study area in west-central British Columbia, either as direct results or indirect consequences of these processes. The tectonic development and the resulting geological features are discussed below.

Island Arc Stage and Composite Terrane I Formation (Late Paleozoic to Late Triassic)

The Babine Lake area is located in the eastern part of the Stikine Terrane near the boundary with the Cache Creek Terrane as shown in Figure 5. The Stikine Terrane is the largest autochthonous terrane in Canada (Coney *et al.*, 1980) and consists of possible upper Precambrian basement, Mississippian and Permian volcaniclastic rocks, basic to acidic volcanic rocks, and carbonates (all locally deformed and intruded in Middle to Late Triassic time), overlain by Upper Triassic basaltic to andesitic volcanic rocks (Monger, 1984). Most of the upper Paleozoic and Triassic rocks of Stikinia are interpreted as products of an island arc setting (e.g., Monger, 1984; Armstrong, 1988; and Mortimer, 1986) but a 400 km long belt of Mississippian to Permian limestone on the northwest side of the terrane indicates that intervals of quiescence and stability also existed (Gabrielse and Yorath, 1989). The Cache Creek Terrane presently located east of the Stikine Terrane is interpreted as Tethyan oceanic crust with overlying seamounts, atolls, and accretionary prism melanges of Mississippian to Triassic age (Monger, 1984).

A Permo-Triassic unconformity is present in all of the island arc terranes in keeping with the global sealevel lowstand at that time (Gabrielse and Yorath, 1989). Extensive mafic volcanism took place in the Late Triassic (230 to 214 Ma) after a distinct hiatus from the older rocks (Muller, 1987, cited by Armstrong, 1988). This volcanism is recorded in the Wrangel, Quesnel, and Stikine Terranes and is represented in this study area by the Takla Group volcanic rocks and related sediments west of Babine Lake. The pre-Jurassic

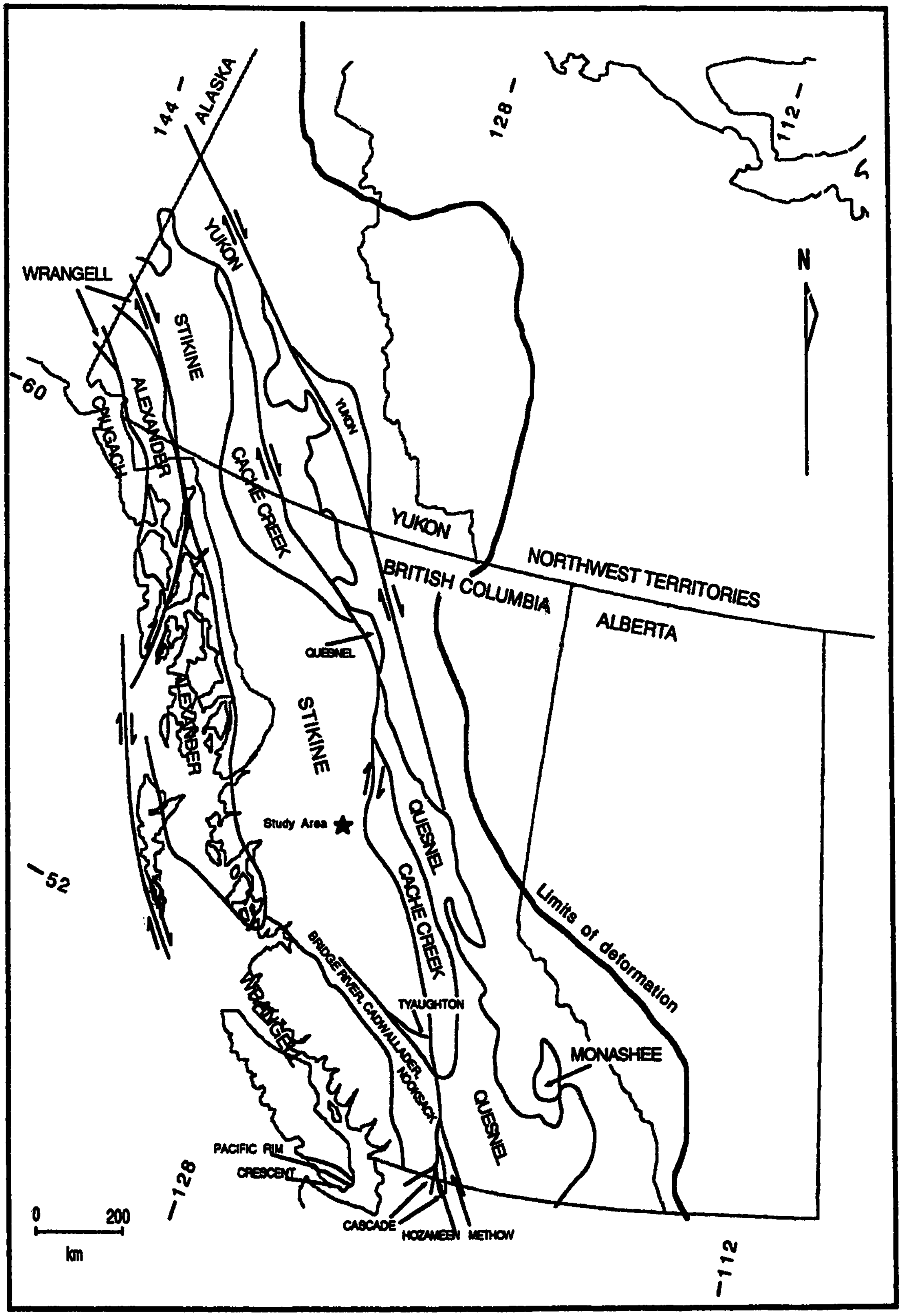

Figure 5. Simplified terrane map of the Canadian Cordillera (modified from Armstrong, 1988 and Gabrielse and Yorath, 1989).

arcs were separated from the North American continental margin possibly by deep back-arc basins similar to the arcs of the southwest Pacific (Monger, 1984), however it is uncertain as to which direction the arcs faced (Coney *et al.*, 1980).

The Quesnellia and Stikinia volcanic arcs and the Cache Creek and Slide Mountain oceanic terranes accreted together to form the composite Terrane I of Monger *et al.* (1982) in the Late Triassic to Early Jurassic before collision with the continental margin (Monger, 1984). Final assembly of the terranes did not take place however until at least the Middle Jurassic (Monger, 1984).

A northeast trending belt of Late Triassic to Early Jurassic (206 to 173 Ma) stocks and small batholiths (the Topley Intrusions) extends from the eastern margin of the Coast Plutonic Complex near Morice Lake to Babine Lake in the core of the Skeena Arch (Carter, 1982) as shown in Figure 3. They range in composition from quartz diorite to quartz monzonite and are thought by Carter (1982) to represent eruption centres of the Lower Jurassic Hazelton Group volcanic rocks.

During and following the formation of Terrane I, the Early Jurassic featured a lull in arc magmatism in Stikinia, indicated by a hiatus between the Upper Triassic Takla Group (mafic) and the Lower to Middle Jurassic Hazelton Group calk-alkaline to silicic volcanic rocks (Armstrong, 1988). The Hazelton Group volcanic and sedimentary rocks which are very widespread in this region form a nearly continuous sequence from the Early to Late Jurassic. They consist of marine and nonmarine volcanic flows and fragmentals and marine sediments derived from older volcanic terranes and reworked contemporary volcanic rocks (Carter, 1976 and Carter, 1982).

Terrane I Accretion (Middle Jurassic)

The Early Jurassic saw a transition from terrane specific volcanism, plutonism, and sedimentation to the development of overlap and post-accretionary assemblages in the Middle Jurassic as Terrane I was accreted to the North American continental margin (Gabrielse and Yorath, 1989). Arcs from the Middle Jurassic onward were probably similar to those of the present eastern Pacific, being built across the terranes accreted to the continental margin (Monger, 1984). Magmatism in the late Early Jurassic was extensive in the accreted terranes (Armstrong, 1988). Hazelton Group marine and nonmarine andesitic volcanism and marine sedimentation continued into the Middle Jurassic when it was replaced by marine and nonmarine sedimentation, especially in the Bowser Basin (Carter, 1976).

As explained by Frazier and Schwimmer (1987), during arc accretion and creation of a new subduction zone with its associated magmatism, the old arc with its volcanic, volcaniclastic, and platform carbonate rocks is deformed by the collision. Cooling following cessation of magmatism results in rapid subsidence and the development of a successor basin on top of the old arc with sediments being derived from the erosion of the sedimentary and volcanic rocks deformed in the orogen and from the new arc. The Bowser successor basin northwest of Babine Lake was the site of shale deposition in the Early Jurassic (Armstrong, 1988) and molassic marine and nonmarine conglomerate sedimentation in the Late Jurassic (Gabrielse and Yorath, 1989).

Continental Arc Plutonic and Subduction Stage (Late Jurassic to Late Cretaceous)

The Cretaceous was characterized by a combination of Andean-type arc and accretionary tectonics (Frazier and Schwimmer, 1987). Little magmatism and regional gaps in the stratigraphic record throughout the Cordillera indicate that the latest Jurassic to Early Cretaceous was a period of emergence, erosion, and external drainage (Armstrong, 1988). Upper Jurassic to Lower Cretaceous sediments are preserved in the Bowser Basin northwest of Babine Lake, consisting of an overlap assemblage of continental (Omineca) and oceanic (Cache Creek) sediments, both derived from the east (Coney *et al.*, 1980). These sediments were deposited as a deltaic wedge prograded over the older marine turbidites of the Bowser Basin (Gabrielse and Yorath, 1989).

Plutonism during the latest Jurassic and Early Cretaceous interval was limited to the Francois Lake Intrusions (Armstrong, 1988) (133 to 155 Ma in the Endako area (White *et al.*, 1970, cited by Carter, 1982)) in a belt extending 160 km southeast from Babine Lake (Carter, 1982). They are dominantly quartz monzonite with lesser amounts of diorite, granite, and adamellite (Carter, 1982). Kitsault intrusions of quartz diorite, augite porphyry, and feldspar porphyry are exposed north of Alice Arm (approximately 100 km northwest of Babine Lake) and are thought to represent volcanic centres of the Middle Jurassic and younger fragmental volcanic rocks of the Hazelton Group (Carter, 1982).

Unconformably overlying the Upper Jurassic to Lower Cretaceous sedimentary rocks is the Skeena Group. It consists of a variety of sedimentary and volcanic rocks, basically black marine shales overlain by and interbedded with volcanic tuffs and breccias (Carter, 1982). In some areas continental sediments are also interbedded with and overlying the volcanic and marine sedimentary rocks (Richards, 1974 and Richards and Dodds, 1973).

Offshore from the continental margin, Wrangellia and the Alexander Terrane amalgamated to form Terrane II of Monger *et al.* (1982) in the Late Jurassic (Monger, 1984). Accretion of Terrane II onto the continental margin took place in the Middle Cretaceous, with linkage being complete by 130 Ma (Armstrong, 1988). This collision resulted in the development of an Andean style magmatic arc as continental arcs were formed across the previosly accreted terranes (Terrane I) on the continental margin (Armstrong, 1988) and caused granulite facies metamorphism in parts of the forming Coast

Plutonic Complex (Monger, 1984). Mid-Cretaceous magmatism was widespread in the Coast Plutonic and Omineca Belts from 110 to 90 Ma (Armstrong, 1988). Some Intermontane Belt intrusions and volcanic rocks, including the Kasalka volcanics and parts of the Skeena Group formed at this time but these were volumetrically small (Armstrong, 1988).

From Barremian to Cenomanian time (Early to Late Cretaceous), clastic marine and continental sediments derived from the Omineca Belt were deposited in the fault bounded troughs of the Sustut and Skeena Basins (north of Babine and Takla Lakes) (Gabrielse and Yorath, 1989). These consist of continental clastic deposits of sandstone, conglomerate, shale, and mudstone (Carter, 1982).

Bulkley Suite plutons intruded the Intermontane Belt across the Skeena Arch in the Late Cretaceous (Armstrong, 1988). These intrusions occur as stocks and small batholiths of granodiorite and quartz monzonite porphyry and have K-Ar ages of 84 to 70 Ma (Carter, 1982). They form an approximate north-south belt with some intrusions being localized by north to northwest striking faults (Carter, 1982). Late Cretaceous to Paleocene Ootsa Lake Group volcanic rocks discordantly overlie older Mesozoic rocks (Carter, 1982). These include rhyolite, dacite, andesite, and basalt flows with some related quartz porphyry intrusions, fragmental rocks, and sedimentary rocks (Carter, 1982). These are best preserved in the Francois Lake - Ootsa Lake areas to the south.

<u>Post-Accretionary Transpression, Uplift, and Magmatism (Late Cretaceous to Recent)</u>

The Late Cretaceous to Paleocene was a time of transpressional tectonics due to the oblique convergence of the oceanic Kula Plate with the continental margin (Monger, 1984). During this time large tracts of the Canadian Cordillera were displaced northward along strike-slip faults, with displacements of up to 450 km on individual faults and 1000 km for the entire system (Monger, 1984). At this time, east-dipping subduction was taking place along the west coast, resulting in extensive calc-alkaline magmatism (Monger, 1984).

By the mid-Tertiary, a tectonic regime of uplift and extension developed due to changes in the relative motions of the Farallon, Kula, Pacific, and North American Plates (Gabrielse and Yorath, 1989). A short-lived, spectacular magmatic event occurred from approximately 64 to 40 Ma (Paleocene to middle Eocene) throughout the Cordillera (Armstrong, 1988). This period of intense magmatism marked the end of a period of rapid uplift and was related to rapid west to southwest relative motion of North America and the resulting dextral shear and eastward subduction (Gabrielse and Yorath, 1989).

Mid-Tertiary volcanic rocks in this area consist of flat to gently dipping andesite and basalt flows and pyroclastics of the Endako Group in the Houston and Francois Lake areas south of Babine Lake (Carter, 1982). The abundant Upper Cretaceous and Tertiary intrusions were divided by Carter (1982) into five suites on the basis of areal distribution, age, whole rock geochemistry, major and trace element contents of biotites, and associated metal content. The Cretaceous Bulkley intrusions were already discussed and the four Eocene age suites (including the Babine Lake Intrusive Suite) are discussed in the next chapter.

Most of these stocks and small batholiths appear to be localized by north-northwest and northeast trending faults and fault intersections (Carson and Jambor, 1974 and Carter, 1982). By the early Cenozoic, the crust was thick enough and radiogenic enough for significant crustal involvement in magmatism to take place and to be indicated by Sr isotopes (Armstrong, 1988). Low initial strontium isotope ratios ($^{87}Sr/^{86}Sr < 0.704$) are restricted with most being in the transitional 0.704 to 0.707 range, indicating an addition of crustal strontium to most of the Cenozoic Coast Plutonic and Intermontane Belt magmas (Armstrong, 1988). This has implications for the metallogeny of the plutonic suites, and will be discussed in the next chapter. Armstrong (1988) also suggests that higher temperatures in the lower crust during the Eocene facilitated crustal incorporation into the magmas.

By the mid-Tertiary, a tectonic regime similar to that of today had developed, with two thirds of the western margin of North America being bounded by transform faults and the remaining third being convergent (Monger, 1984).

Pleistocene and younger plateau basalts, flows, and cinder cones are restricted to north to northeast trending fault zones near Alice Arm (Carter, 1982).

3. CRETACEOUS-EOCENE MAGMATISM AND METALLOGENY

Introduction

The Babine Intrusive Suite is one of five suites of small porphyritic stocks which intruded the Intermontane Belt in west-central British Columbia during the Late Cretaceous and early Tertiary. This plutonism consists of a central north to northwest trending belt of Cretaceous stocks and small batholiths (the Bulkley Intrusions) with small Eocene intrusions situated on the east and west (Carter, 1982).

All suites contain porphyry-type ore deposits and like all porphyry ore deposits of the Canadian Cordillera younger than 85 Ma, they are of the stock or phallic type and were emplaced during a period of strike-slip tectonics (Ney and Hollister, 1976). Although the intrusions and associated deposits overlap extensively in age and were all intruded in the same general tectonic setting, the contained metal compositions vary significantly. This feature is important in the study of these ore deposits because it has implications for metal sources and therefore ore deposit genesis. Moreover, any model for the formation of porphyry ore deposits must account for this spatial variation in ore types.

Intrusive Suites

Numerous small igneous intrusions of Cretaceous to Tertiary age are present in west-central British Columbia. These intrusions have been divided by Carter (1976, 1982) into five suites on the basis of areal distribution, age, whole rock chemistry, major and trace element compositions of biotites, and associated metal compositions. The distribution of the suites is shown in Figure 6. The plutons of these suites are generally small in diameter (<4 km) and are localized by fault intersections (Carter, 1982). The following classification and descriptions are from Carter (1982). Figure 7 shows the distribution of the suites on a quartz-plagioclase-orthoclase ternary diagram.

Bulkley Intrusions

The Cretaceous Bulkley Intrusions consist of stocks and small batholiths (~0.6 to 3 km in diameter) of porphyritic granodiorite and quartz monzonite with ages of 84 to 70 Ma. These intrusions were forcefully emplaced into north and northwest trending faults and are situated in a broad north-south belt passing through Hazelton and Smithers.

Goosly Lake Intrusions

The Goosly Lake Intrusions are situated south of Houston and Francois Lake and north of Houston at Grouse Mountain. They consist of small plugs of porphyritic gabbro and monzodiorite and are interpreted to be Eocene volcanic centres. They have ages of 53 to 49 Ma.

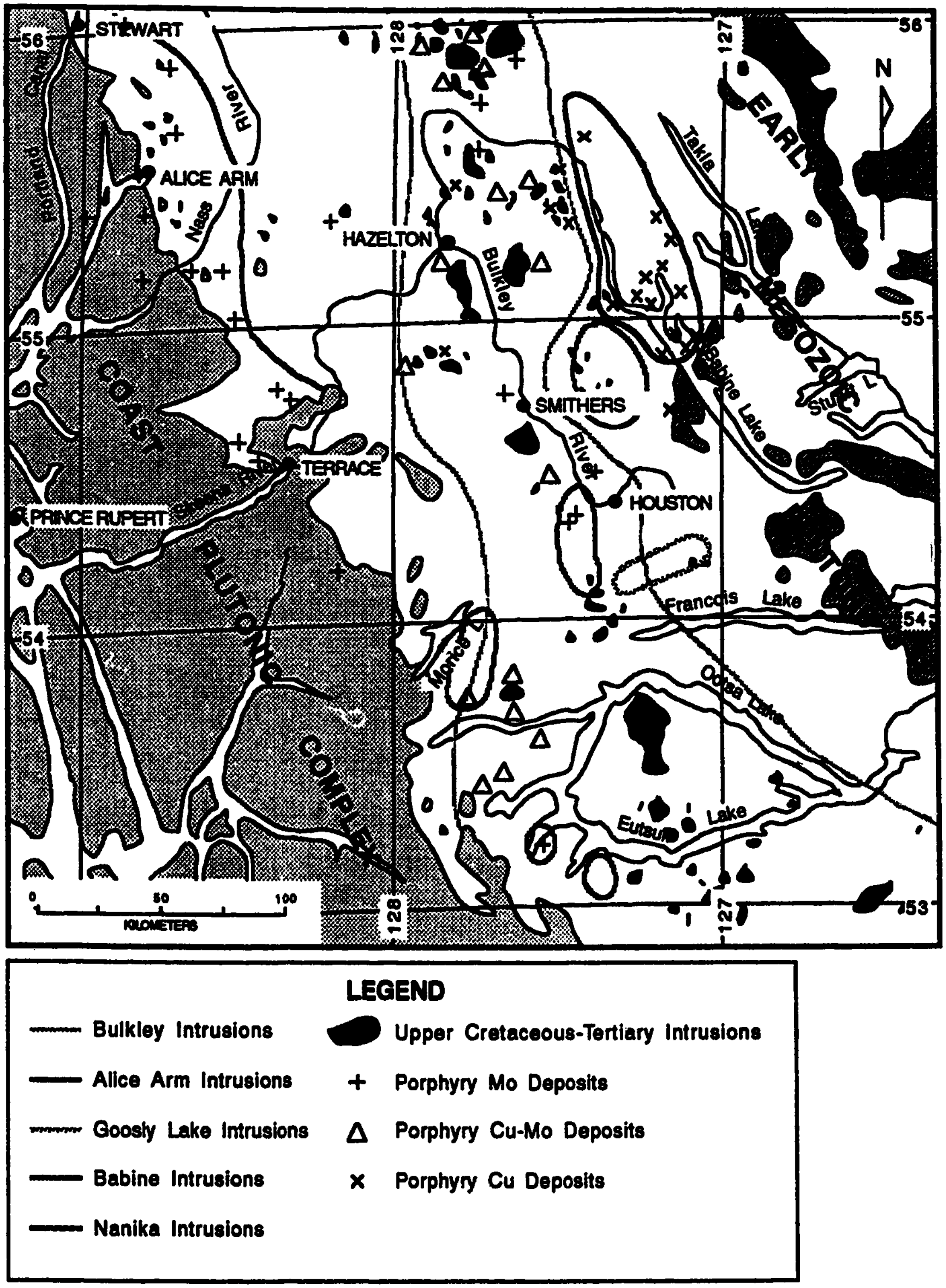

Figure 6. Cretaceous-Tertiary intrusions in west-central British Columbia (after Carter, 1982).

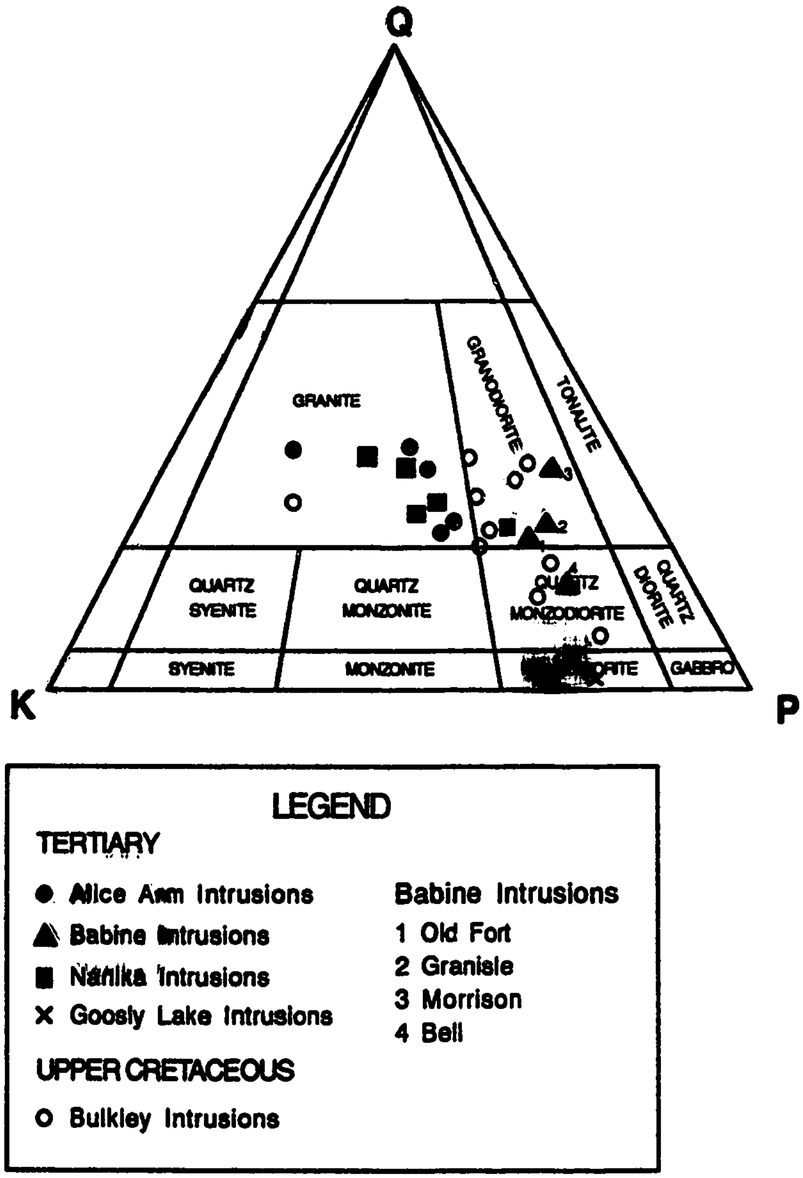

Figure 7. Distribution of Cretaceous-Tertiary intrusions on the basis of normative quartz-plagioclase-orthoclase compositions (after Carter, 1982, according to the classification of the International Union of Geological Sciences, 1973).

Alice Arm Intrusions

This suite is located along the eastern margin of the Coast Plutonic Complex north and south of Alice Arm and consists of small stocks (<1 km in diameter) of quartz monzonite porphyry. They have ages of 54 to 48 Ma and intrude sediments along major fault zones (trending east-northeast and north-northwest).

Nanika Intrusions

The Nanika Intrusions are small (<1 km in diameter) porphyritic granodiorite to granite intrusions with ages of 56 to 47 Ma. They are situated within and along the margins of the belt of Bulkley Intrusions and are localized at intersections of northeast and northwest trending faults and cores of anticlines. Many show evidence of forceful emplacement.

Babine Intrusions

The Babine Intrusive Suite located near Babine Lake consists of small plugs, dikes, and dike swarms of fine-grained biotite-feldspar porphyry (BFP) of quartz diorite to granodiorite composition. They have K-Ar ages of 55 to 49 Ma. The intrusions are thought to be volcanic centres and volcanic equivalents are locally preserved. The magmatism was localized at the intersections of northeast and northwest trending faults, with most of the suite being preserved in three northwest trending grabens; 1) Granisle-Bell-Old Fort, 2) Morrison-Hearne Hill, 3) Dorothy-Nak Lake-Trail Peak.

Mineralization

Tectonic Associations and Metal Distributions

The Coast Plutonic Complex (CPC) is composed of three zones, the western zone of quartz diorite (140 to 84 Ma in the west and 79 to 64 Ma in the east), a central zone (45 Ma) of migmatite gneiss, quartz diorite, and granodiorite, and an eastern zone of granodiorite and quartz diorite (50 to 40 Ma) intrusive into the Mesozoic volcanic and sedimentary rocks (Carter, 1982). Carter (1982) theorized that the Cretaceous-Eocene magmatism of west central British Columbia was related to the CPC and therefore to the eastward migration of the subduction zone beneath the western edge of North America from the late Jurassic to early Tertiary. Carter (1982) classified the porphyry mineralization in west-central British Columbia into four roughly north-south trending belts. From west to east these are: 1) Alice Arm molybdenum deposits (50 Ma), 2) Bulkley copper-molybdenum deposits (84 to 70 Ma), 3) Nanika copper-molybdenum deposits (50 Ma), and 4) Babine copper deposits (50 Ma). These porphyry deposit suites are characterized as follows.

Bulkley Cu-Mo

According to Carter (1982), the Bulkley Intrusions correspond in age (84 to 70 Ma) to the quartz diorite and granodiorite of the eastern part of the western zone of the CPC and may have been generated by the same underthrusting event, although being emplaced at a shallower level and a greater distance above the subduction zone. The Bulkley Intrusions host copper-molybdenum and molybdenum-tungsten mineralization, with best mineralization near the stock contacts and better Mo grades within the stock. Silver-lead-zinc veins are found peripheral to the stocks in many cases.

Alice Arm Mo

Renewed underthrusting in the Eocene (50 to 40 Ma) resulted in the emplacement of the central and eastern zones of the CPC and the related Alice Arm Intrusions on the eastern flank (Carter, 1982). These intrusions host molybdenum mineralization, likely because of their felsic nature, their sedimentary country rocks (Carter, 1982), and the greater crustal thickness near the CPC.

Nanika Cu-Mo

The Nanika copper-molybdenum deposits (~50 Ma) are also thought by Carter (1982) to be related to the subduction zone. As explained by Carter (1982), they are similar in composition to the Alice Arm Intrusions, and are located along the eastern flank of the CPC in the west and are localized farther east by the same fault systems as are the Bulkley Intrusions.

Babine Cu

As explained by Carter (1982), the Babine Intrusions (55 to 49 Ma) are very different from the Alice Arm and Nanika Intrusions. They are typical subvolcanic intrusions, emplaced as dikes, dike swarms, necks, and plugs. They are found farther east than the other porphyry suites. According to Carter (1982), their volcanic nature indicates emplacement a greater distance above the subduction zone. Their copper mineralization may reflect a dominance of mafic volcanic host rocks (Carter, 1982) or their typical calc-alkaline granodiorite to quartz diorite composition and a smaller crustal source component.

More than a dozen porphyry copper deposits are located in the northern Babine Lake area (Carson and Jambor, 1974). Mineralization is dominantly chalcopyrite with bornite being locally important (e.g., Granisle), and only minor molybdenum. Quartz-carbonate veins with pyrite, galena, sphalerite, and chalcopyrite are present peripheral to Granisle and Bell (as are commonly found around porphyry deposits). These were explored 40 years before the porphyries themselves were prospected (Carter, 1982).

Goosly Lake Ag-Cu

The small plugs of the Goosly Lake (53 to 49 Ma) porphyritic gabbro and monzodiorite are hosts of Cu-Ag mineralization of massive sulfide character, such as the Sam Goosly deposit and the Equity Silver Mine south of Houston (Carter, 1982). They are thought to be centres of Eocene volcanism (Carter, 1982).

Summary

The Late Cretaceous to Tertiary intrusions of west-central British Columbia are related to subduction beneath the western edge of North America (Carter, 1982). This magmatism is likely related to the same general processes responsible for the major magmatism of the CPC discussed by Armstrong and Ward (1991). The greater distance from the centre of the magmatic belt likely resulted in a stronger structural control on the magmatism, with intrusions and volcanism being focused locally by strike-slip fault systems as discussed by Carter (1982). The type of porphyry mineralization reflects the magma composition and crustal conditions, with molybdenum mineralization reflecting dominantly sedimentary source rocks and acidic intrusions. Copper mineralization reflects more intermediate magmas relative to the molybdenum deposits and a dominance of volcanic country rocks. Copper-molybdenum deposits reflect a mixture of the crustal types. This results in a general trend in metal zonation from west to east of Alice Arm Mo deposits, Bulkley and Nanika Cu-Mo deposits, and Babine Cu deposits (Carter, 1982) suggesting a greater component of crustal source material farther west. This is in accordance with the proximity to the tectonically thickened crust of the CPC.

4. LOCAL GEOLOGY AND ORE DEPOSITS

Introduction

The statigraphy of the Babine Lake area consists of Mesozoic to early Cenozoic volcanic and sedimentary rocks which are essentially unmetamorphosed. These are intruded by minor Mesozoic intrusions and numerous Eocene volcanic plugs and dikes of the Babine Intrusive Suite. This latter intrusive suite hosts several economic and subeconomic porphyry copper deposits including the mines of Bell and Granisle and the subeconomic Morrison property. It is these deposits which are the subject of this study.

This chapter will introduce the local geology and the general features of deposits which will be discussed in greater detail in subsequent chapters.

Local Geology of the Babine Lake Area

The local geology of the northern Babine Lake area is shown in Figure 8. Units of Late Triassic to Eocene age are exposed and are represented as follows.

Takla Group (Upper Triassic)

The oldest rocks exposed in the area, Late Triassic Takla Group rocks, are exposed in a block along the west shore of Babine Lake (Carson *et al.*, 1976). The Takla Group in this area includes green, submarine tuffs, breccias, and flows, with lesser amounts of limestone, argillite, greywacke, and augite porphyry (Richards, 1974). Outcrop exposures of Takla Group mudstone and greywacke are located near the Bell mine barge landing on the west side of Babine Lake. One sample of soapy, black mudstone contains thin calcite stringers and grains, zeolites, and fine-grained disseminated pyrite, pyrrhotite, and marcasite with traces of chalcopyrite and sphalerite.

Hazelton Group (Lower and Middle Jurassic)

The Lower Jurassic Hazelton Group rocks are the most widespread units in the area. The Hazelton Group consists chiefly of continental and marine calc-alkaline volcanic rocks (mainly pyroclastics) (Richards, 1974), with lesser amounts of clastic sediments and limestones (Richards and Dodds, 1973).

The oldest unit, the Telkwa Formation consists of light green flows, aquagene tuff, lapilli tuff, and breccia overlain by mid-Lower Jurassic, green, tuffaceous argillite and siltstone (Carson *et al.*, 1976). Agglomerates, abundant on Newman Peninsula, contain 50 to 80 percent breccia fragments 2 mm to 2.5 cm in diameter. These rocks weather greenish-brown to reddish-brown, often with a mottled appearance because of differential weathering of the various clast types and matrix. Fine-grained hematite disseminations are abundant and minor pyrite, chalcopyrite, sphalerite, and covellite are also present. All samples are cut by thin calcite or calcite + quartz ± chlorite veinlets and minor epidote veins. Telkwa Formation rocks have undergone subgreenschist grade metamorphism, with

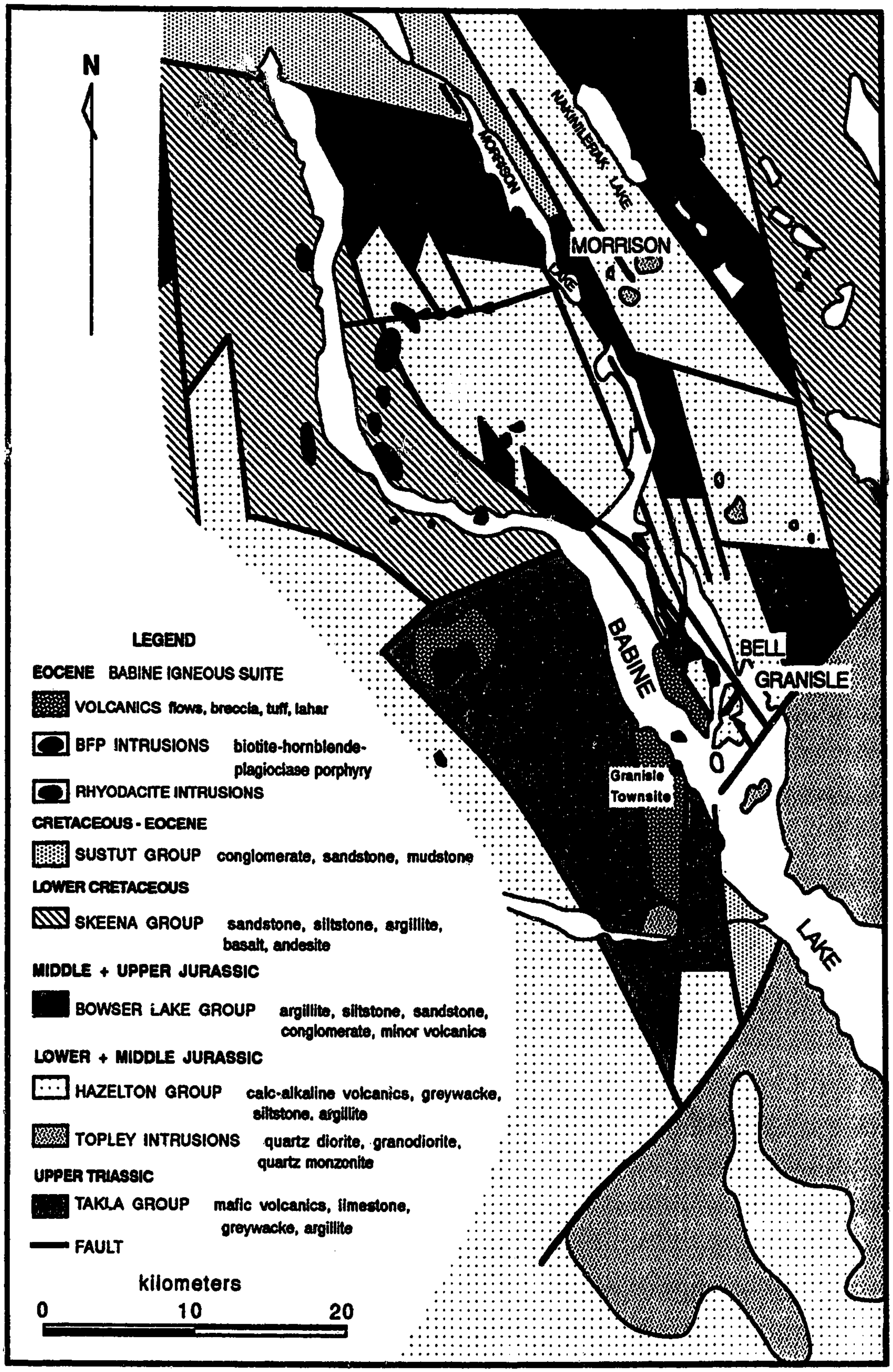

Figure 8. Geology of the northern Babine Lake area (after Carson *et al.*, 1976)

a typical assemblage of epidote, chlorite, prehnite, albite, calcite, and actinolite (Carson *et al.*, 1976). The Telkwa Formation is exposed on the eastern side of the Newman Peninsula, north of the Newman Peninsula, and on the eastern side of Hagan Arm.

The Telkwa Formation is overlain in fault contact by the Red Tuff Member of the Nilkitkwa Formation, which consists of subaerial red, maroon, and purple andesitic to dacitic lapilli tuffs (Carson *et al.*, 1976). This unit is exposed at the north end of the Newman Peninsula (Carson *et al.*, 1976) and in a small outcrop approximately 5.5 km northwest of Granisle. The Nilkitkwa Formation is overlain by a fossiliferous marine greywacke, the Bajocian Smithers Formation (Carson *et al.*, 1976).

Topley Intrusions (Upper Triassic to Lower Jurassic)

Coeval with the Hazelton Group volcanic rocks are the quartz diorite to quartz monzonite Topley Intrusions (Carson *et al.*, 1976) which were discussed previously. Only one Topley intrusion was observed in this area, a quartz monzonite plug 5 km south of Granisle along the lakeshore at Red Bluff Provincial Park.

Bowser Lake Group (Upper Jurassic to Lower Cretaceous)

The next oldest unit is the Middle to Upper Jurassic Bowser Lake Group, exposed between Babine and Morrison Lakes (Carson *et al.*, 1976) and hosting the Morrison deposit. The Bowser Lake Group, a successor basin sequence, consists of a lower unit (the Ashman Formation) of fine-grained clastic sediments of a distal siltstone and argillite facies and an upper unit of coarse-grained clastic marine and nonmarine deltaic sediments (Carson *et al.*, 1976).

Skeena Group (Mid-Cretaceous)

Mid-Cretaceous strata in the Babine Lake area consist of marine and nonmarine sediments and volcanic rocks of the Skeena Group (Carson *et al.*, 1976). Biostratigraphic dating of shale units indicates an Albian age (Carson *et al.*, 1976) and Armstrong (1988) reports a mid-Cretaceous (130 to 84 Ma) age for volcanic units. Volcanic rocks are found as a number of disconnected volcanic piles of basic augite-plagioclase porphyry and rhyolite in the Babine and Takla valleys (Richards, 1974). These include flows, breccias, and agglomerates and are interbedded with sedimentary units (Richards, 1974). The sedimentary units are generally fine-grained, gritty shales with sandy lenses and laminations and contain scour marks, bioturbation features, and crossbedding typical of a tidal or prodeltaic environment (Carson *et al.*, 1976). One unaltered Skeena Group sample near the west shore of the Newman Peninsula is a brownish grey, medium-grained, clast supported lithic arenite. Monomineralic grains include quartz (10%; some strained grains and aggregates), plagioclase (10%), and minor microcline and muscovite. Volcanic fragments (50%) include partially sericitized plagioclase microporphyries, volcanic glass fragments and tuff, and fine-grained sericitized volcanic fragments. Other grain types include granophyric textured quartz-feldspar clasts (5%), metaquartzite (some cut by quartz veins), shale, phyllite (5%), and siltstone (10%) clasts. This assemblage indicates a wide range of source materials and the presence of metamorphic and felsic igneous rock fragments indicates that the source of at least part of the grains was to the east. The well winnowed nature, with a very minor matrix component is consistent with a tidal or shoreline origin. Minor opaque phases (5%) include interstitial hematite and hematite coatings on some grains and rare magnetite grains.

Most Skeena Group samples in this area consist of very fine-grained gritty siltstones. Highly contorted greywackes and fine tuff are exposed on the western shore of the Newman Peninsula (Carson *et al.*, 1976). The Skeena Group also contains green andesitic tuffs and breccias exposed south of the Bell Mine along the road to the barge landing which are very similar to the Telkwa Formation.

Sustut Group (Upper Cretaceous to Paleocene)

The Upper Cretaceous to lower Tertiary Sustut Group underlies part of the Babine Valley and most of the Takla Valley and consists of continental alluvial sediments (Richards, 1974). It is comprised of quartz-feldspar-chert sandstone with interbedded mudstone and conglomerate (Richards, 1974). The Sustut Group along with the Bowser Lake Group is preserved in downfaulted blocks in linear grabens related to the northwest trending block faults (Carson *et al.*, 1976).

Babine Intusive Suite (Eocene)

The compressional event of the mid-Cretaceous was followed by extensive block faulting in the early Tertiary (Richards, 1974). This episode was responsible for the north-northwest structural grain of the area and the faults and fault intersections served as loci for the Babine Lake porphyry intrusions (Richards, 1974). The Babine Intrusions are located in a belt extending from south of Fulton Lake to the northwest arm of Takla Lake (Ogryzlo, 1975). They include the more abundant rhyodacite porphyries as well as the porphyry copper-related hornblende-biotite-plagioclase porphyries (locally known as BFP) (Carson *et al.*, 1976). The rhyodacite and BFP intrusions are both approximately 51 Ma (Ogryzlo, 1975) but locally the order of emplacement may vary. For example, the rhyodacite intrusion is older than the BFP at the Bell mine (Carson *et al.*, 1976) but is coeval with the BFP at Morrison (Carson and Jambor, 1974).

i) Rhyodacite Porphyry

The rhyodacite porphyry intrusions and flows of the Babine Igneous Suite are widespread on the Newman Peninsula and are also found in the Bell pit. They consist of white, light brown, pale greyish-green, and buff coloured coarsely to finely porphyrytic flows and shallow intrusions with quartz and albite phenocrysts (Carson *et al.*, 1976). Flows may be dense, massive, or flow banded and breccia bodies also occur (Carson *et al.*, 1976). Plate 1 shows flow banded rhyodacite in the Bell mine.

Most rhyodacites are massive grey flows with sparse (<10%), fine-grained plagioclase phenocrysts in a light brown aphanitic groundmass. Some samples are extensively stained orange by the presence of limonite and siderite along microfractures. The groundmass consists of fine-grained plagioclase (70%), quartz (10%), and calcite (<10%) with accessory zircon and sphene (some of which is altered to calcite + rutile).

Some coarser-grained varieties are present, consisting mainly of 1 to 3 mm clouded plagioclase phenocrysts and minor potassium feldspar. The minor groundmass is medium-grained, consisting mainly of plagioclase (40%), quartz (10%), minor K-feldspar, and minor alteration chlorite, calcite, siderite, and limonite.

ii) Biotite-Feldspar Porphyry

The Babine Intrusions range in composition from quartz diorite to granodiorite to quartz monzonite and include small stocks, dikes, dike swarms, and plugs (Carter, 1982). Unaltered BFP is a medium grey crowded porphyry with abundant 0.25 to 5 mm phenocrysts of zoned oligoclase-andesine, biotite, and hornblende in an aphanitic groundmass of the above minerals plus quartz and K-feldspar (Carson *et al.*, 1976). It weathers greyish-brown. The volcanic equivalents appear very similar except that hornblende is markedly more abundant than biotite. Few virtually unaltered samples are present, with most showing weak propylitic alteration.

A virtually unaltered sample of BFP (B84) is shown in Plate 2. Plagioclase phenocrysts average 30% of the rock, consisting of 0.1 to 3 mm euhedral to subhedral zoned crystals of An_{35-40}. Some show minor traces of sericite. Biotite phenocrysts (~5%) are 0.1 to 1 mm books and plates. Pleochroism is from golden yellow to dark brown or black. Hornblende (~10% in volcanic BFP samples) is present as 0.1 to 1 mm needles with yellowish brown or yellowish green to reddish brown pleochroism. Some extrusives contain Ti-rich amphiboles with deep orange to greenish orange pleochroism and very high birefringence. Amphiboles are commonly zoned and in many cases have black reaction rims. Some show minor carbonate or epidote alteration.

The groundmass which usually accounts for 50 to 60% of the rock is microcrystalline to cryptocrystalline and consists of plagioclase (some samples show trachytic textures), fine-grained biotite, hornblende, quartz, K-feldspar, and devitrified glass. Accessory phases include zircon, apatite, and sphene. All samples show minor alteration in the form of zeolites in the groundmass and some show limonite and siderite staining along fractures,

Plate 1. Flow banded rhyodacite exposed in the Bell mine.

Plate 2. Photomicrograph of unaltered biotite-feldspar porphyry (BFP) (sample B84). The euhedral hornblende phenocryst in the top centre is 400 μm across. Plane polarized light.

carbonate alteration (up to 10%), and minor epidote, indicating a gradation into propylitic alteration.

Opaque phases comprise only a few percent of the rock and consist mainly of hypidiomorphic to xenomorphic magnetite and in some cases hypidiomorphic ilmenite (often with prominent lamellar twinning and as small grains in amphiboles and amphibole reaction rims). Some samples show alteration of magnetite to hematite, sometimes to complete replacement. Hematite is also present as minor blades and as fine grains in amphibole reaction rims. Other minor phases include disseminations of pyrite, chalcopyrite, and galena.

Extrusive equivalents include columnar jointed flows (at the south end of the Newman Peninsula), flow breccias, tuffs, agglomerates, and lahars. These extrusives are fine- to medium-grained with a granodioritic to andesitic composition and amphibole as the major mafic mineral (Carter, 1982). Two types of lahars are present on the peninsula. One type has angular, light-coloured, flow banded quartz-albite porphyry clasts (probably equivalent to the rhyodacite) while the other contains mauve, purple, and light green fine grained biotite-hornblende-plagioclase porphyry clasts up to 0.6 m in diameter and is probably the extrusive equivalent of the BFP (Carson *et al.*, 1976).

Post-Eocene faulting has preserved the extrusive equivalents in downfaulted blocks which are exposed on the southern part of the Newman Peninsula and north of the town of Granisle (Carter, 1982). Post-Eocene weathering and Pleistocene glaciation have eroded and scoured the area, resulting in the final form and the extensive burial by glacial till (Carson *et al.*, 1976).

Geology of the Ore Deposits

Bell

The Bell mine is located on the Newman Peninsula at the northeast end of Babine Lake. The generalized geology of the central part of the peninsula is shown in Figure 9.

Eocene block faulting juxtaposed Cretaceous Skeena Group shales, siltstones, and volcanic rocks against Lower Jurassic Hazelton Group marine tuffs and siltstones (Ogryzlo, 1975) with a stratigraphic separation of 700 to 1300 m across the Newman Fault (Carson *et al.*, 1976). During the later stages of this block faulting episode (from 55 to 49 Ma), Eocene volcanic rocks were extruded along the fault zones (Carter, 1982). The first volcanism is represented by the flow banded and massive white, light brown, pale greyish-green, and buff coloured rhyodacite (quartz-albite) porphyries (Carson *et al.*, 1976). Closely following the rhyodacite volcanism was the BFP volcanism, represented by breccias, lahars, and columnar jointed flows preserved on the west side and south end of the Newman Peninsula (Carson *et al.*, 1976). Intrusive equivalents of the BFP phase are

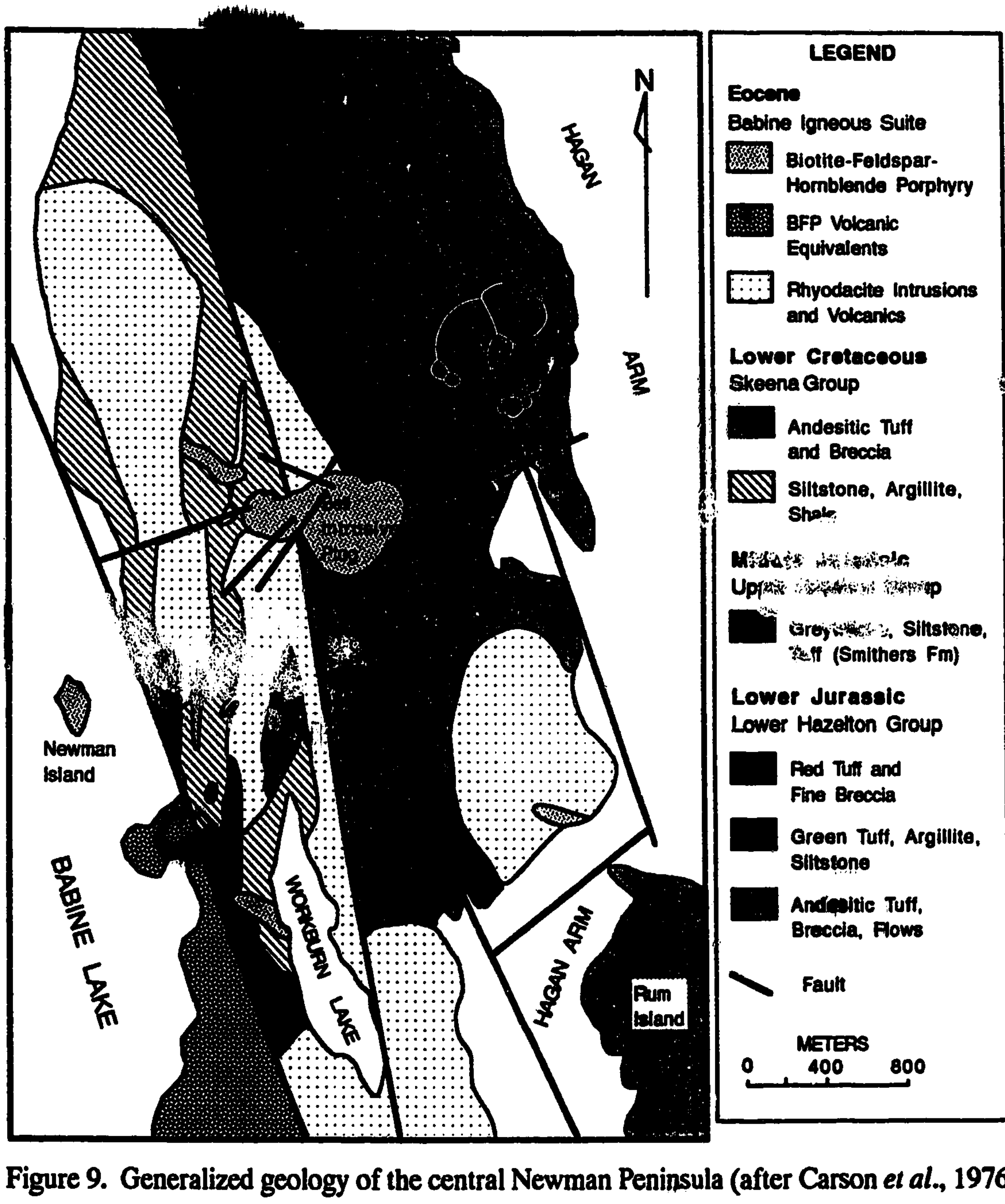

Figure 9. Generalized geology of the central Newman Peninsula (after Carson *et al.*, 1976).

preserved as plugs, dikes, and volcanic necks, including the intrusive plug at the Bell mine. As shown in Figure 10, this volcanic neck is localized along the fault contact between the Skeena Group siltstones to the west and Hazelton Group volcanic rocks to the east (Carter, 1982). Associated radial dikes are situated on the west side of the intrusion (Carter, 1982).

The pear-shaped BFP plug, 200 m wide in the west and 600 m wide in the east (Carson *et al.*, 1976) intrudes the volcanic and sedimentary country rocks and also the slightly earlier rhyodacite porphyry. A small, slightly younger phase intrudes the southeast corner of the pit (Dittrick, pers. com. 1991 and personal observation). Several high angle faults crosscut the BFP plug and country rocks in the Bell pit. Ogryzlo (1987) suggested that the structures at Bell could be related to cauldron subsidence of the Bell volcano.

Post-ore movement along the faults was minimal since no offset of the BFP plug or alteration zones is present (Carson *et al.*, 1976). Post-Eocene faulting did occur in other locations however, preserving the extrusive equivalents on the southern end of the peninsula and on the west side of Babine Lake (Ogryzlo, 1975).

Prior to glaciation, extensive weathering of the ore zone resulted in supergene alteration and oxidation to depths of 50 to 70 m (Carson *et al.*, 1976). This was followed by glaciation and burial by 5 to 30 m of glacial till (Carson *et al.*, 1976).

Granisle

The Granisle deposit is located on McDonald Island just south of the Newman Peninsula. Host rocks for the Eocene intrusions are Hazelton Group volcanic and sedimentary rocks consisting of two members: 1) green to purple waterlain andesitic tuffs and breccias with intercalated chert pebble conglomerates in the central and eastern parts of the island; and 2) massive and amygdaloidal andesitic flows and thinly bedded shales overlying the former unit on the western side of the island (Fahrni *et al.*, 1976). The host rocks and Eocene intrusive phases are shown in Figure 11. The intrusives were localized by the intersections of the northwest and northeast trending faults, within the same graben as the Bell and Old Fort deposits (Carter, 1982).

The earliest intrusive phase at Granisle is a dark grey, fine-grained quartz diorite microporphyry with 1mm plagioclase phenocrysts in a biotite-quartz-plagioclase matrix (Carter, 1982). It contains xenocrystic inclusions of metavolcanic and metasedimentary rocks of three types: 1) a breccia with 1 to 3 cm rounded chert and volcanic rock fragments in a fine-grained dioritic matrix; 2) a breccia with chert fragments in a white felsic matrix with 4 mm clasts of quartz and chloritized biotite (representing recrystallized and metasomatized country rock); and 3) fine-grained light to dark grey hornfelsed volcanic and sedimentary country rocks (Carter, 1982).

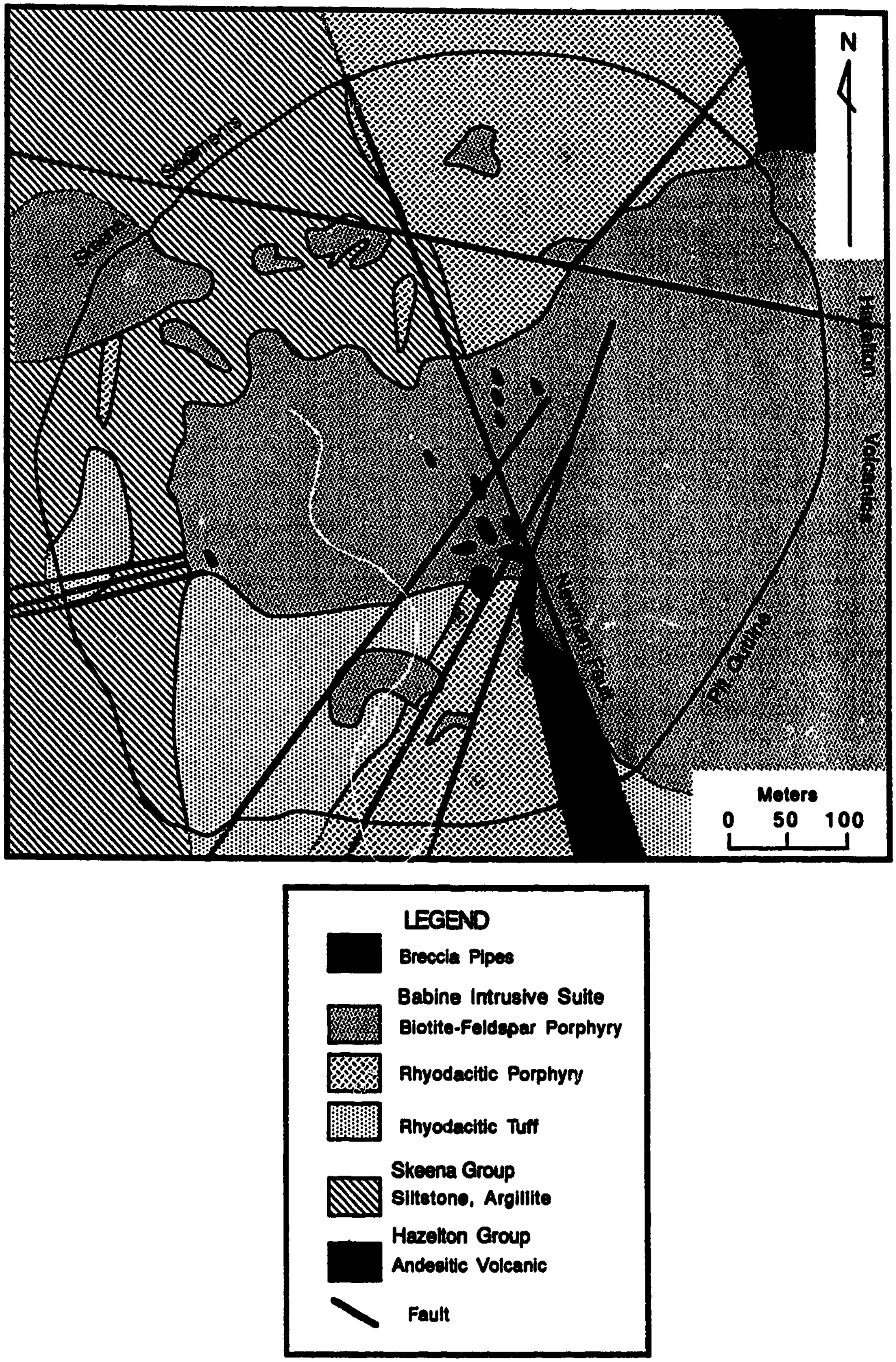

Figure 10. Generalized geology of the Bell Mine (modified from Carson *et al.*, 1976).

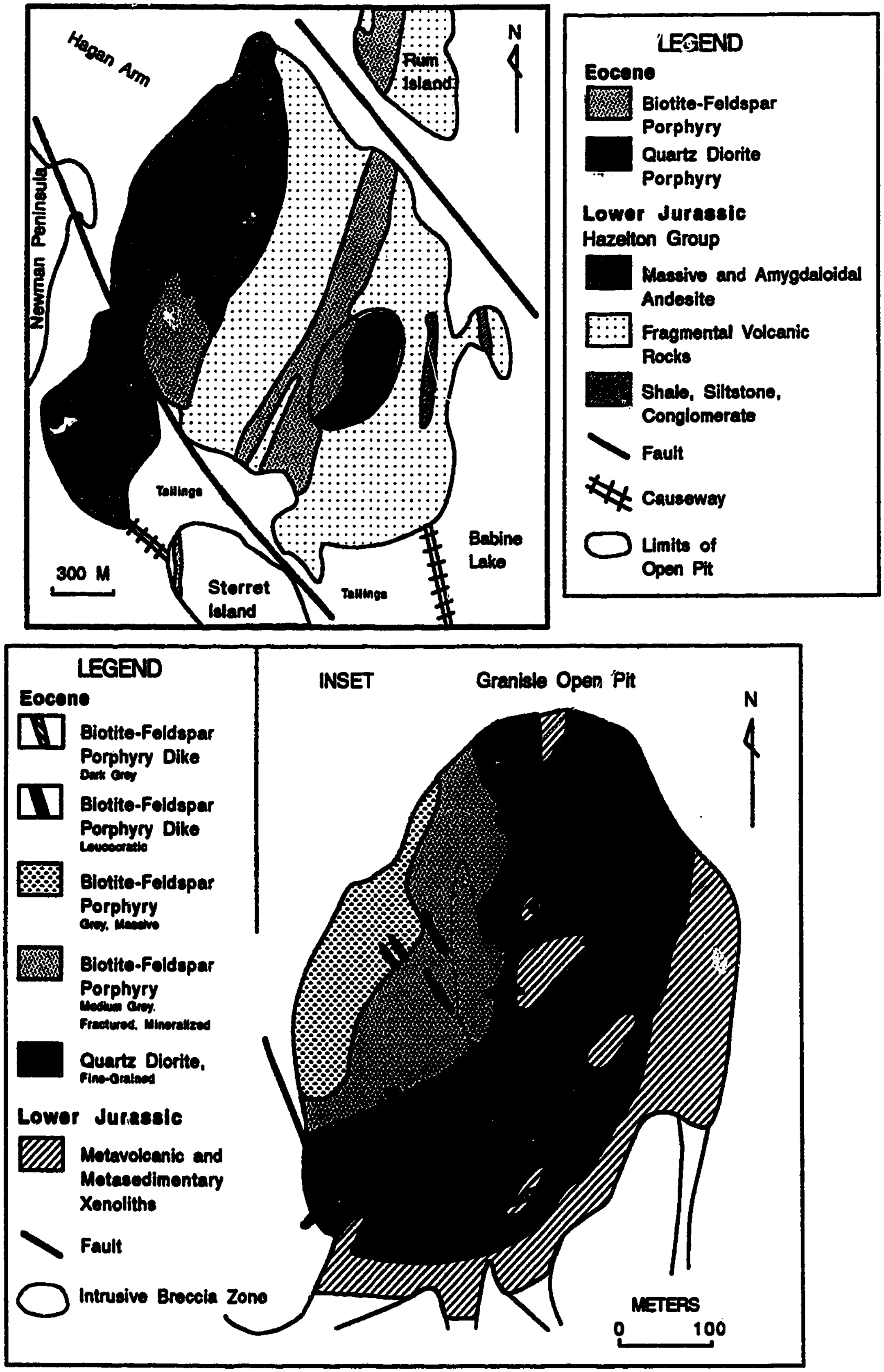

Figure 11. Geology of McDonald Island and the Granisle deposit (modified from Fahrni *et al.*, 1976).

This early quartz diorite intrusive phase was in turn intruded by a 120 to 200 m wide BFP dike striking northeast across the island (Fahrni *et al.*, 1976). This dike intruded along the western edge of the quartz diorite with smaller dikes radiating outward from it (Carter, 1982). This BFP phase is a light to dark grey quartz diorite to granodiorite porphyry (Carter, 1982) with 35 to 50% plagioclase (oligoclase-andesine) and biotite phenocrysts in a fine-grained matrix of quartz, plagioclase, biotite, and amphibole with lesser amounts of potassium feldspar and apatite (Fahrni *et al.*, 1976). Several different BFP phases are present within the intrusive body, with the latest phase being post-mineralization, consisting of dikes of dark grey BFP with sparse plagioclase phenocrysts and very fine-grained biotite and magnetite and is located along the east side of the pit (Fahrni *et al.*, 1976). A dacite porphyry phase, slightly older than the BFP is exposed on the north side of McDonald Island and is characterized by abundant pyrite and the absence of mafic phenocrysts (Fahrni *et al.*, 1976).

The BFP dike (which hosts most of the porphyry mineralization) is bounded by two northwest trending parallel faults (as shown in Figure 12); on the west, a topographic lineament crossing the island south of the mine and on the east by a fault within the channel between the island and the east shore of Babine Lake (Fahrni *et al.*, 1976).

<u>Morrison</u>

The Morrison deposit is a subeconomic deposit located 22 km north of the Bell mine near the south end of Morrison Lake. This deposit was first described by Woolverton (1964, unpublished) and Carter (1967, published). The deposit is hosted by the Middle to Upper Jurassic Bowser Lake Group siltstones, silty argillites, and minor conglomerates (Carson and Jambor, 1976). As described by Carson and Jambor (1976), these rocks are dominantly composed of very fine- to medium-grained quartz, feldspar, and volcanic and lithic fragments. The intrusions hosting the deposit were localized in a northwest trending graben bounded on the west by a fault along the eastern edge of Morrison Lake and on the east by a fault 800 m east of the property (Carson and Jambor, 1976). It is within this graben that the host rock Bowser Lake Group rocks are preserved. Carson and Jambor (1976) also suggested that the BFP plug could have been localized by a north-northwest trending isoclinal fold.

In Eocene time, the sediments were intruded by a small plug and peripheral biotite-feldspar-hornblende porphyry (BFP) and feldspar porphyry [rhyodacite porphyry] dikes (Carter, 1982). These intrusions metamorphosed the country rock sediments to biotite hornfels facies (Carter, 1982). The BFP intrusive plug and dikes are dark grey quartz diorite containing 25 to 33% of 2 to 3 mm plagioclase (oligocene-andesine) phenocrysts, plates and books of biotite, and hornblende which is partially altered to biotite (Carter, 1967). The main plug is approximately 500 m in diameter, was circular before faulting, and contains many offshoots of 1 to 500 m in length (Carson and Jambor, 1976). As in the other deposits, many different phases of the BFP are present (Carson and Jambor, 1976). Comagmatic with the BFP are tan coloured rhyodacite dikes consisting of aplitic textured quartz, albite, and potassium feldspar (Carson and Jambor, 1976). The geology of the Morrison property is shown in Figure 12.

After emplacement of the plug and development of the alteration zones discussed in a later section, the plug was bisected and offset by the north-northwest trending Morrison Fault, a linear zone of parallel shears and fractures 25 m wide (Carson and Jambor, 1976). Movement along this vertical fault is judged to be approximately 300 m of dextral slip with considerable vertical displacement; the west side being uplifted relative to the east (Carson and Jambor, 1976).

Erosion and weathering continued after the emplacement of the intrusions and faulting but post-glacial weathering was minimal (Carson and Jambor, 1976), as shown by the presence of a thick layer of glacial cover over most of the property.

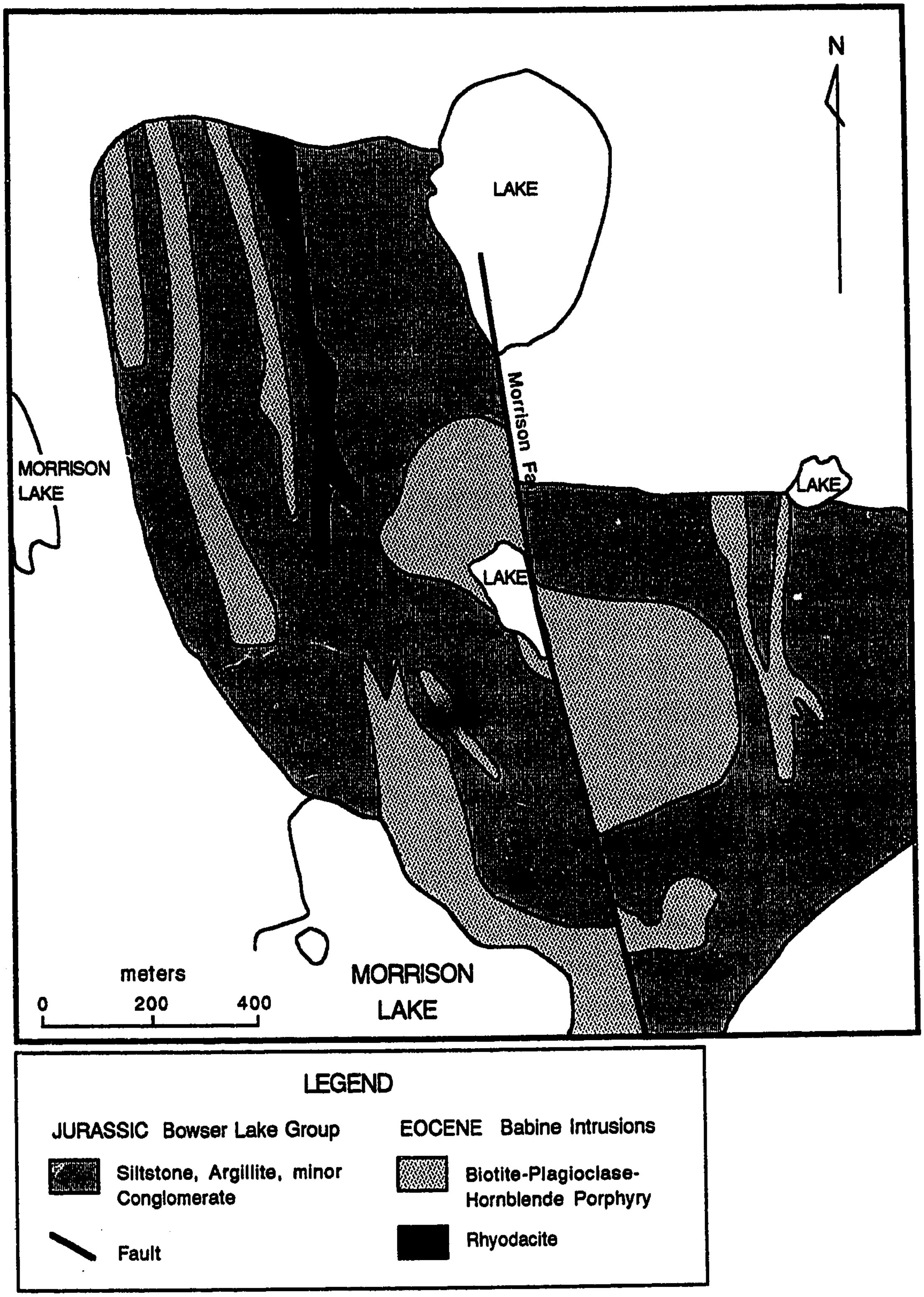

Figure 12. Geology of the Morrison property (after Carson and Jambor, 1976)

5. HYDROTHERMAL ALTERATION AND ORE PETROLOGY

Introduction

The relationship between hydrothermal alteration and ore deposition in porphyry copper deposits has long been recognized (e.g., Lowell and Guilbert, 1970; Beane and Titley, 1981; Rose, 1970; Beane, 1982; Soregaroli, 1975; and Carson and Jambor, 1974). As stated by Beane (1982), "The fact that certain phases persist in composition, association, and paragenesis across the porphyry copper deposits as a group suggests that some combination of environmental conditions is common to the deposits and is optimal for development of economic sulfide mineralization.".

Therefore, studies of porphyry copper deposits invariably involve analysis of hydrothermal alteration patterns. As with other porphyry copper deposits, the Babine Lake deposits are closely associated with hydrothermal alteration patterns. In order to understand the origin of these deposits and to interpret the stable isotope data in a geologically meaningful and realistic manner, it is imperative to first understand the features of the hydrothermal alteration patterns and their significances.

This section will first review the general features of hydrothermal alteration associated with porphyry copper deposits and then examine the Morrison, Granisle, and Bell deposits. A paragenetic sequence will be constructed for each deposit in order to constrain the physical and chemical conditions of hydrothermal alteration and ore deposition and to identify any differences in their development.

General Character of Porphyry Copper Hydrothermal Alteration

The classic hydrothermal alteration zonation pattern of Lowell and Guilbert (1970) involves the outward progression of potassic, phyllic, argillic, and propylitic alteration zones although all zones are not necessarily present in any deposit. This general zonation pattern along with the sulfide zonation is shown in Figure 13. As indicated by Beane (1982), although porphyry copper deposits exhibit strong similarities and the model of Lowell and Guilbert (1970) is based on the characteristics of 27 deposits, few of even these deposits are "average" in every respect.

Despite the individual characters of deposits, certain general, important similarities exist. Most porphyry copper deposits consist of coaxially cylindrical alteration zones (Lowell and Guilbert, 1970). Potassic and propylitic alterations are generally considered to be early stages, with phyllic and argillic alterations being later overprints (Carson *et al.*, 1976; McMillan and Panteleyev, 1980; Beane and Titley, 1981; and Beane, 1982). Deep level alteration zoning consists of an outer zone of chlorite + sericite + epidote + magnetite grading to an inner zone of quartz + K-feldspar + sericite + chlorite (Lowell and Guilbert, 1970). The alteration types are described as follows.

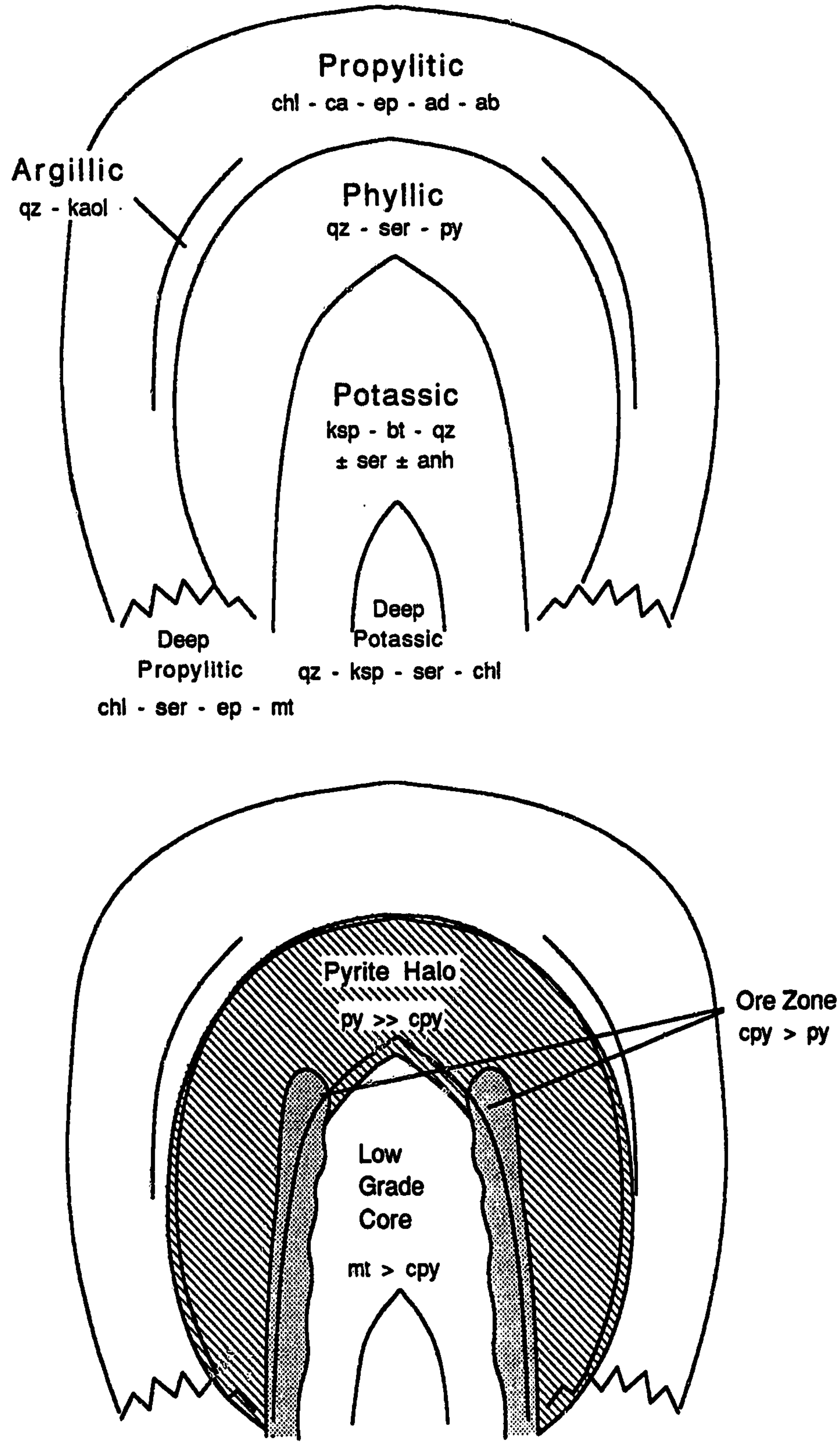

Figure 13. Hydrothermal alteration and sulfide zonation patterns of typical porphyry copper deposits (modified from Lowell and Guilbert, 1970).

Potassic Alteration

Potassic alteration is defined as the addition or replacement of minerals by potassium-bearing phases and is characterized by the assemblage orthoclase + biotite + quartz ± sericite ± anhydrite ± albite ± apatite with opaque phases including some of chalcopyrite, pyrite, magnetite, molybdenite, and bornite (Lowell and Guilbert, 1970 and Beane and Titley, 1981). This is the innermost alteration zone and often contains a barren core and part of the ore shell (Lowell and Guilbert, 1970). Its high temperature of formation and high salinity hydrothermal fluid have been interpreted as indications of a late magmatic origin (e.g., McMillan and Panteleyev, 1980 and Wilson *et al.*, 1980).

Potassic alteration involves the exchange of K for Ca and Na (Titley, 1982). It is most commonly identified by conversion of igneous hornblende to biotite but also involves biotization of the groundmass, K-feldspar and/or biotite in or enveloping quartz veinlets, and potassium feldspar replacement of plagioclase phenocrysts and /or the groundmass (Titley, 1982). In the strictest sense, igneous biotite and K-feldspar are not stable but undergo compositional readjustment, with hydrothermal K-feldspar containing less Na and biotite less Ti and Fe than equivalent minerals in unaltered igneous rocks (Beane, 1982).

Potassic alteration is a selectively pervasive alteration type, affecting specific minerals and therefore enhancing rock texture (Titley, 1982). It is likely that the minerals of the host rock rather than the solution composition, control the alteration (Titley, 1982). Figure 14 (from Beane, 1982) shows AKF and ACF diagrams for the various alteration types in comparison to unaltered igneous rocks. The AKF ternary system includes $Al_2O_3 + Fe_2O_3$ (A), $K_2O + Na_2O$ (K), and $MgO + FeO$ (F) whereas the ACF ternary consists of CaO (C) instead of $K_2O + Na_2O$ (K) (Beane, 1982). Both have SiO_2 and H_2O in excess (Beane, 1982). Shown in Figure 14a are the fields of typical intermediate to felsic igneous rocks. Figure 14b shows the field of potassically altered rocks in relation to igneous rocks. It can be seen that the potassic alteration field includes the igneous field and extends to include muscovite (sericite). The stable existence of anhydrite is shown in the ACF diagram. The opaque mineralogy sequence progresses from early magnetite to chalcopyrite to pyrite (Helgeson *et al.*, 1969).

Propylitic Alteration

Propylitic alteration involves the creation of new Ca- and Mg-bearing minerals and is considered to be the equivalent of greenschist facies metamorphism (Beane and Titley, 1981). It is produced from the original rock forming minerals with minor additions of H_2O and CO_2 and is therefore nearly isochemical (Beane, 1982). Indicative minerals include chlorite, epidote, and calcite with minor apatite, anhydrite, ankerite, and hematite (Beane and Titley, 1981). Sulfide minerals include pyrite and minor chalcopyrite (Lowell and

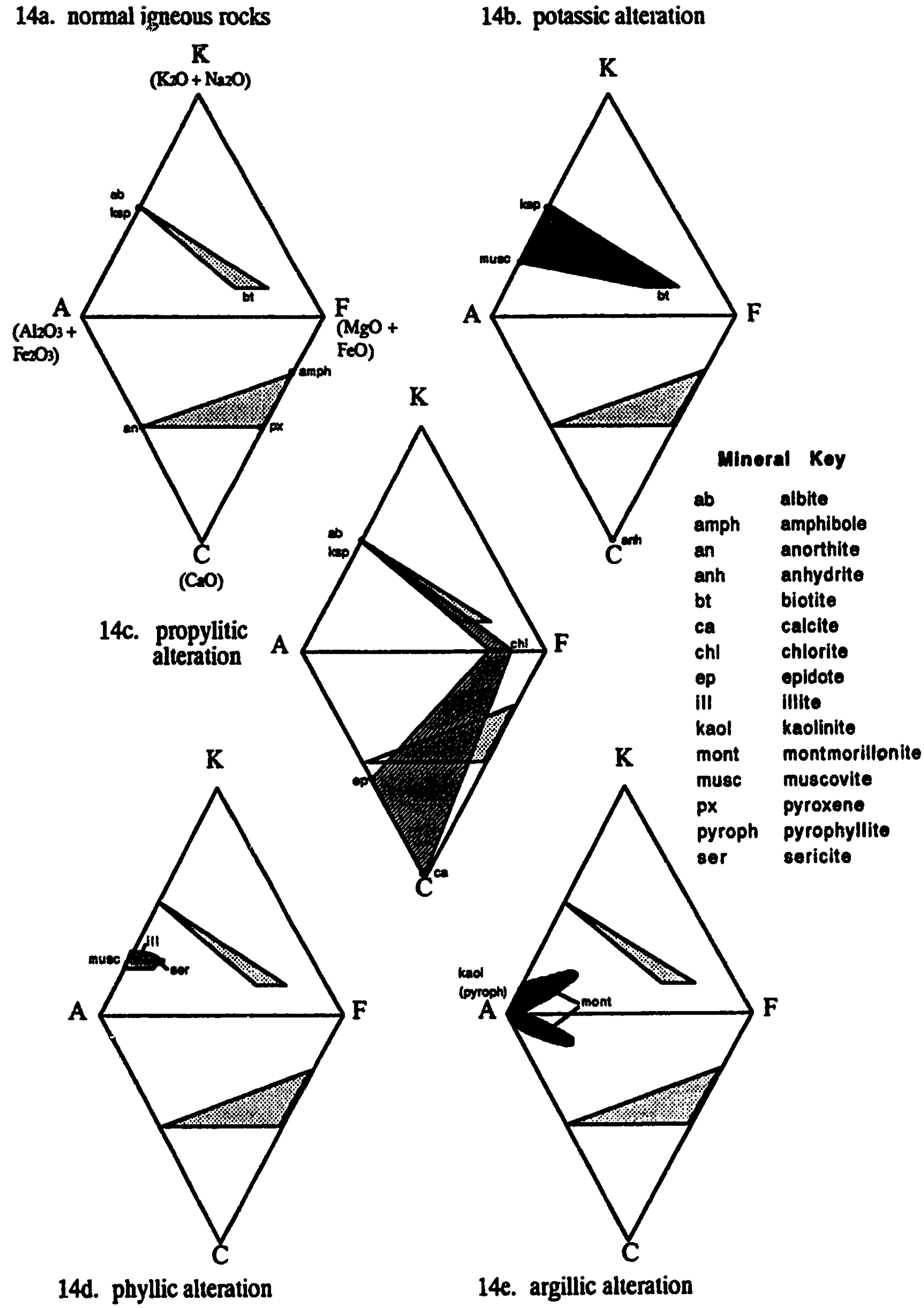

Figure 14. AKF and ACF diagrams of igneous rocks and alteration assemblages associated with porphyry copper deposits (after Beane, 1982).

Guilbert, 1970). Propylitic alteration involves alteration of mafic minerals to chlorite and conversion of the anorthite component of plagioclase to epidote and calcite with preservation of albite (Beane, 1982). Like the potassic assemblage, this alteration assemblage is selectively pervasive although less so than the former and encroaches inward with time (Titley, 1982).

As shown in Figure 14c, propylitic alteration overlaps the unaltered field extensively although slightly to the Al_2O_3 side of the igneous field so as to avoid the biotite stability field, indicating slight K^+ and Mg^{2+} leaching (Beane, 1982).

Phyllic Alteration

Phyllic or quartz-sericite-pyrite alteration is a pervasive, texturally destructive alteration which results in a complete change of rock composition, mineralogy, and texture (Titley, 1982). There is no strong hostrock control on the effects of the solutions, indicating that the rocks have been overwhelmed by the effects of the fluid composition (Titley, 1982). Phyllic alteration is a superimposed alteration and is centred on the intrusive-country rock contact where fracturing is most intense (Beane, 1982).

Phyllic alteration involves a leaching of Na, Ca, and Mg and replacement of original silicates by sericite and quartz (Beane and Titley, 1981). The abundant pyrite involves Fe in excess of original mafic minerals, indicating the net addition of both sulfur and iron (Beane, 1982).

In addition to the characteristic quartz-sericite-pyrite assemblage, illite, chlorite, apatite, and rutile may be present but carbonate and anhydrite may not (Lowell and Guilbert, 1970). Chalcopyrite is variable in abundance, usually present as disseminations and often in the sites of sericite altered pseudomorphs (Lowell and Guilbert, 1970). Pyrite is present as veinlets and disseminations, ranging in abundance from 2 to 30% but averaging 5 to 10% (Lowell and Guilbert, 1970).

As shown in Figure 14d, phyllic alteration plots near the K_2O-Al_2O_3 edge of the AKF diagram and involves removal of Na, Ca, and Mg during intense hydrogen and weaker potassium metasomatism (Beane, 1982) The separation of the igneous and phyllic fields indicates the dominant control of the hydrothermal fluid over the host rock.

Argillic Alteration

Argillic alteration involves extensive base leaching under acidic conditions. Intermediate argillic alteration results from incomplete cation leaching, with the remaining cations forming montmorillonite, hydromica, and chlorite, whereas advanced argillic alteration involves complete leaching, with a resulting assemblage of diaspore, quartz, andalusite, corundum, and alunite (Beane and Titley, 1981). The most important minerals are kaolinite (as well as halloysite and dickite), pyrophyllite, montmorillonite, and quartz (Beane, 1982). Opaque minerals are dominated by pyrite although minor chalcopyrite, bornite, enargite, and tennantite may also be present (Beane and Titley, 1981).

Figure 14e shows the argillic alteration field on AKF and ACF diagrams located near the Al_2O_3 apex. Since Na^+ and K^+ are leached more extensively than Ca^{2+}, this assemblage usually plots only on the ACF diagram (Beane, 1982). Like phyllic alteration but to a greater extent, argillic alteration represents extreme dominance of the rocks by the hydrothermal solution.

Babine Lake Porphyry Deposits

General Characteristics

Carson and Jambor (1974) undertook an extensive study of mineralization and alteration of economic and subeconomic Babine porphyry deposits in order to determine the relationships between ore and alteration zoning and to outline features of value to exploration geologists. The economic and subeconomic deposits, including those studied by Carson and Jambor (1974) are shown in Figure 15 and their findings are summarized as follows.

Characteristic features of ore grade Babine porphyry copper deposits are: 1) an altered BFP intrusion; 2) a circular or elliptical zone 2,150 to 2,450 m in diameter in which all rocks are visually altered and most contain anomalous amounts of pyrite; 3) a zone of hydrothermal biotite alteration several hundred meters in diameter containing the ore zone and surrounded by a chlorite alteration zone; and 4) a pyrite halo greater than 300 m wide overlapping the outer edge of the biotite zone containing from 5 to 10% pyrite. In contrast to the Lowell and Guilbert (1970) model in which best ore grades are located in the phyllic zone (especially where it overlaps the potassic zone), ore in Babine Lake porphyry deposits is mostly associated with the biotite (potassic) zone.

The alteration zonation pattern identified by Carson and Jambor (1974) includes an outer chlorite-carbonate zone and and inner biotite alteration zone with an intervening sericite alteration zone being present in some deposits. The chlorite-carbonate and biotite alteration zones are equivalent respectively to the propylitic and potassic alteration zones of Lowell and Guilbert (1970). The sericite zone may be divided into quartz-sericite alteration as observed at Bell, which is equivalent to the phyllic alteration of Lowell and Guilbert (1970) and sericite-carbonate alteration, which is not exactly equivalent to any zone of the Lowell and Guilbert (1970) model, but appears in some respects to include characteristics of the propylitic, phyllic, and argillic zones.

i) Potassic (biotite) alteration

Potassic alteration in the Babine deposits is characterized by the occurrence of hydrothermal biotite, with only minor amounts of hydrothermal K-feldspar. With the

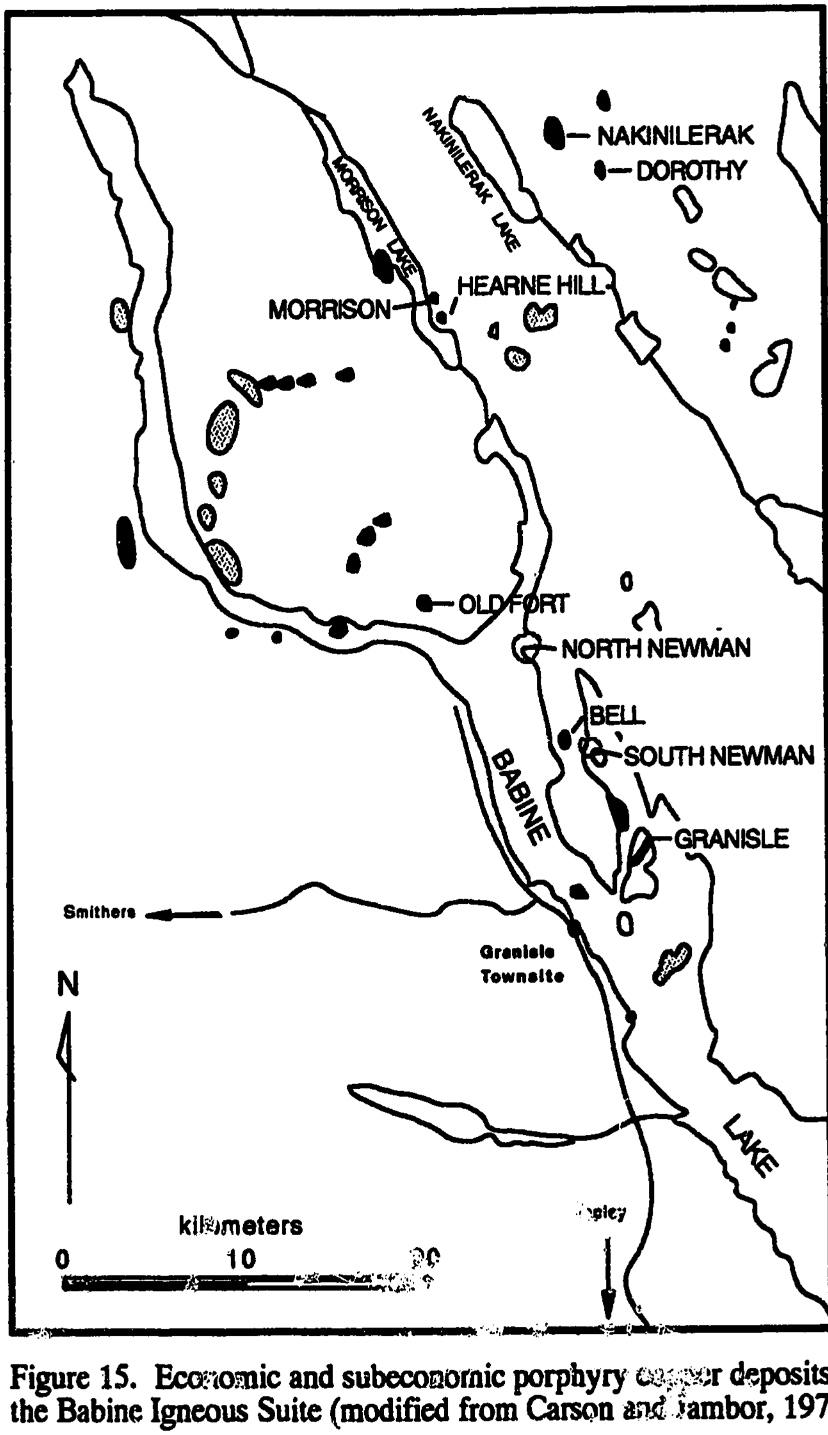

Figure 15. Economic and subeconomic porphyry copper deposits of the Babine Igneous Suite (modified from Carson and Jambor, 1974).

exception of Bell, the ore bodies are contained entirely within the potassic zones and the quality of biotite ranges from sugary, coarse-grained secondary biotite in the ore zone to fine-grained shreddy biotite in the outer part of the biotite zone (Carson and Jambor, 1974 and personal observation).

Biotite alteration is most prominently manifested as the replacement of igneous hornblende by hydrothermal biotite. This has also been observed in many other studies (e.g., Beane, 1982 and Brimhall *et al.*, 1985). Plate 3 shows the replacement of hornblende phenocrysts by well crystallized hydrothermal biotite in the potassic zone at Bell. Biotite alteration of hornblende takes place by reactions such as:

$$NaCa_2Mg_5(AlSi_7O_{22})(OH)_{2(edenite)} + K^+_{(aq)} + 8H^+_{(aq)} + 2SO_4^{2-}{}_{(aq)} = KMg_3AlSi_3O_{10}(OH)_{2(phlogopite)} + 4SiO_{2(quartz)} + 2CaSO_{4(anhydrite)} + Na^+_{(aq)} + 2Mg^{2+}{}_{(aq)} + 4H_2O_{(l)} \quad (1)$$

Secondary biotite is generally more Mg-rich than igneous biotite (Mason, 1978 and Hendry *et al.*, 1985) and contains less Ti (Carson and Jambor, 1974). The compositional adjustment of biotite phenocrysts towards more Mg-rich compositions during potassic alteration has been explained by the reactions (from Beane, 1982):

$$KFe_3AlSi_3O_{10}(OH)_{2(annite)} + 1/2O_{2(g)} = KAlSi_3O_{8(orthoclase)} + Fe_3O_{4(magnetite)} + H_2O_{(l)} \quad (2)$$

and

$$3KFe_3AlSi_3O_{10}(OH)_{2(annite)} + 2H^+_{(aq)} + 3/2O_{2(g)} = KAl_2(AlSi_3O_{10})(OH)_{2(muscovite)} + 3Fe_3O_{4(magnetite)} + 6SiO_{2(quartz)} + 2K^+_{(aq)}\ 3H_2O_{(l)} \quad (3)$$

or the sulfidation reaction:

$$KFe_3AlSi_3O_{10}(OH)_{2(annite)} + 3S_{2(g)} = KAlSi_3O_{8(orthoclase)} + 3FeS_{2(pyrite)} + H_2O_{(l)} + 3/2O_{2(g)} \quad (4)$$

Whereas biotite phenocrysts contain an average of 4.3 wt% TiO_2, hydrothermal biotite contains 2.8 wt % TiO_2.(Carson and Jambor, 1974). The more Mg-, Ti-, and Fe^{3+}-rich hydrothermal biotites in the centre of the potassic zones are attributed to higher temperatures and a higher fO_2 at the time of alteration (Carson and Jambor, 1974).

In addition to biotite alteration of hornblende phenocrysts, secondary biotite is also present as groundmass disseminations, especially in the central part of the potassic zone. Rarely biotite phenocrysts are replaced by hydrothermal biotite along the edges. Minor biotite is also present in quartz veinlets in the potassic zone.

Hydrothermal potassium feldspar is rare in the Babine deposits although it is present (detectable by staining) in the +0.3% Cu zone (Carson and Jambor, 1974), mainly in quartz veins or in alteration envelopes of quartz veins. Plagioclase phenocrysts are

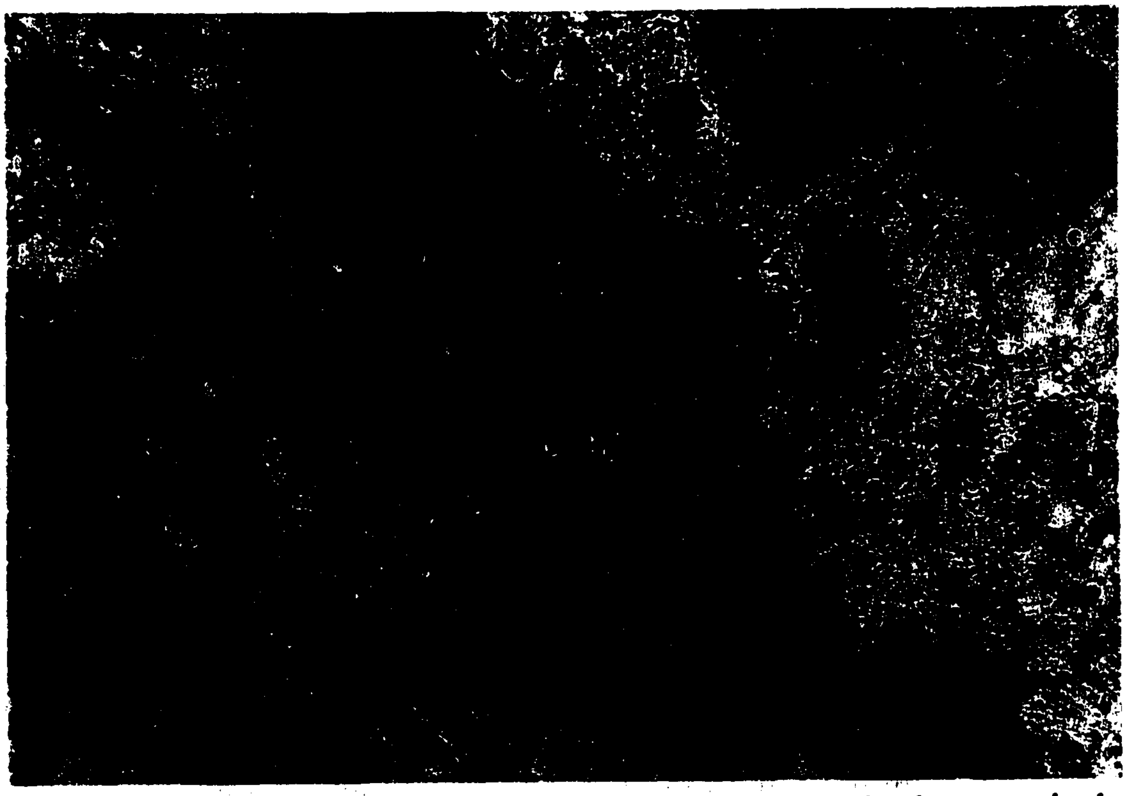

Plate 3. Well crystallized secondary biotite replacing a hornblende phenocryst in the potassic alteration zone of Bell. The small opaque grain in the centre is 375 μm long. Plane polarized light

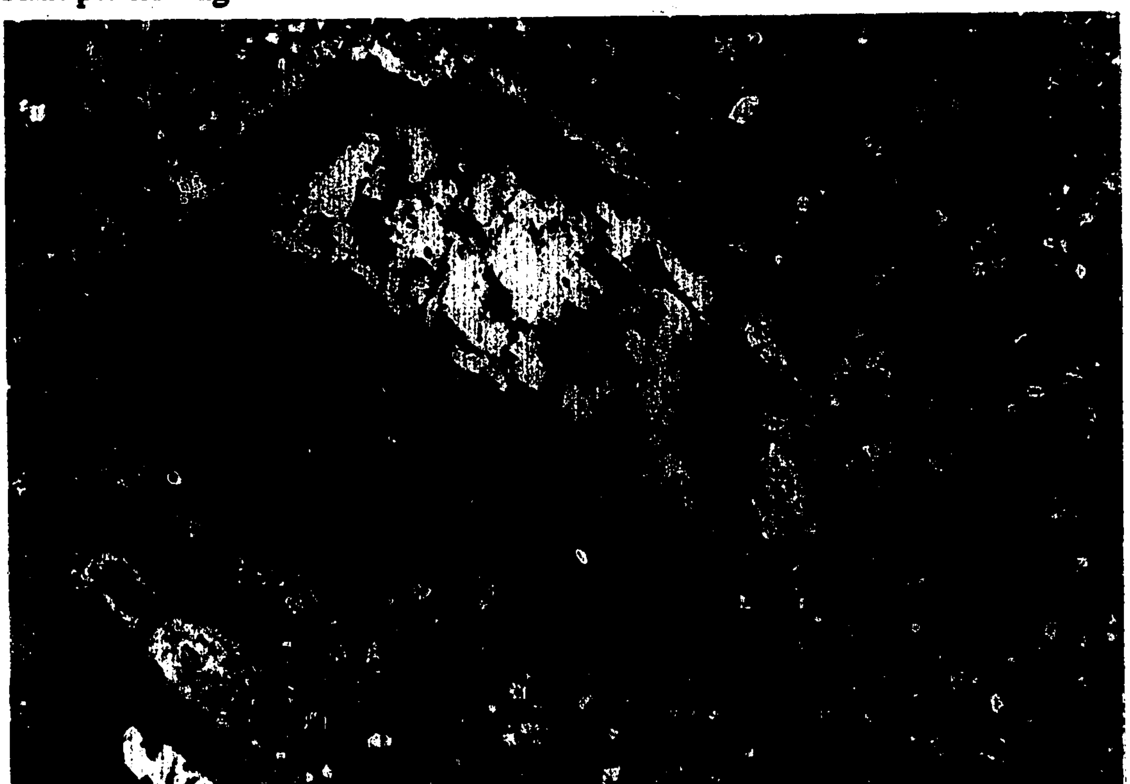

Plate 4. Alteration of anorthite-rich zones of a plagioclase phenocryst to sericite + carbonate from the potassic zone of Bell. The plagioclase phenocryst is 1250 μm long. Plane polarized light.

generally stable in the potassic alteration environment of the Babine porphyry copper deposits although minor sericite or carbonate alteration may take place along fractures or anorthite-rich zones as shown in Plate 4.

As shown by reaction (1), potassic alteration of Ca-bearing phases such as hornblende along with SO_4^{2-} in solution produces anhydrite but this may later be remobilized (Brimhall *et al.*, 1985) or converted to gypsum.

Biotite altered samples may contain abundant magnetite disseminations with or without chalcopyrite and bornite. Chalcopyrite and bornite disseminations and fracture coatings are also important since the potassic zone usually contains the ore zone.(Carson and Jambor, 1974). Pyrite is generally low in abundance in the ore zone. At Granisle, Carson and Jambor (1974) reported that nearly all ore samples studied lacked disseminated pyrite.

ii) Propylitic alteration

The other early hydrothermal alteration assemblage in the Babine deposits is chlorite-carbonate or propylitic alteration. The propylitic alteration zone has a gradational contact with unaltered rock, with the first evidence of alteration being the presence of carbonate in the groundmass and chlorite + carbonate mixtures replacing hornblende phenocrysts (Carson and Jambor, 1974). Carbonate is dominantly calcite but also includes dolomite and siderite (Carson and Jambor, 1974 and personal observation). Epidote may be present as small grains in the groundmass or in hornblende pseudomorphs. Farther inward from the periphery, carbonate and pyrite become more abundant and chlorite becomes coarser grained (Carson and Jambor, 1974 and personal observation). Although epidote is present in the outer propylitic zone, it is lacking in the inner parts of the zone. Carson and Jambor (1974) interpreted this to be due to a relatively low $f\,CO_2$ in the outer parts of the zone and higher $f\,CO_2$ values inward. This destabilizes the epidote according to the reaction:

$$Ca_2Al_2FeSi_3O_{12}(OH)_{(epidote)} + CO_{2(g)} = CaAl_2Si_2O_{8(anorthite)} + 1/2Fe_2O_{3(hematite)} + CaCO_{3(calcite)} + SiO_{2(quartz)} + 1/2H_2O_{(l)} \qquad (5)$$

Whereas chlorite in the outer, weakly altered zones has blue-grey interference colours, chlorites from the inner parts of the zone have green to brown interference colours (Carson and Jambor, 1974). Since propylitic alteration is essentially isochemical, the host rock likely exerts a stronger control over the chlorite compositions than it does for the biotite (Carson and Jambor, 1974).

Alteration of hornblende to chlorite + carbonate probably takes place by reactions such as:

$$2NaCa_2Mg_5(AlSi_7O_{22})(OH)_{2(edenite)} + 12H^+_{(aq)} + 4CO_{2(g)} = Mg_5Al(AlSi_3O_{10})(OH)_{8(clinochlore)} + 4CaCO_{3(calcite)} + 11SiO_{2(quartz)} + 2Na^+_{(aq)} + 5Mg^+_{(aq)} + 4H_2O_{(l)} \qquad (6)$$

Titanium from the original hornblende forms small needles or aggregates of rutile in the chloritized pseudomorphs or more rarely, ilmenite. Hematite and magnetite are also alteration products of the hornblende, present as disseminations or reaction rims. Plate 5 shows the relative persistence of biotite in the propylitic zone compared to hornblende, which is completely altered to chlorite with disseminations and fine needles of hematite and rutile.

Although biotite is generally persistent in the propylitic zone, it is locally altered (especially in the inner part of the propylitic zone) to chlorite, sericite, or carbonate along cleavage planes and may even be completely replaced. Chloritization of biotite takes place by reactions such as:

$$2KMg_3AlSi_3O_{10}(OH)_{2(phlogopite)} + 4H^+_{(aq)} = Mg_5Al(AlSi_3O_{10})(OH)_{8(clinochlore)} + 3SiO_{2(quartz)} + 2K^+_{(aq)} + Mg^{2+}_{(aq)} \quad (7)$$

or the iron analogue:

$$2KFe^{2+}{}_3AlSi_3O_{10}(OH)_{2(annite)} + 2H^+_{(aq)} + H_2O_{(l)} + 1/2O_{2(g)} = Fe_5Al(AlSi_3O_{10})(OH)_{8(chamosite)} + 1/2Fe_2O_{3(hematite)} + 3SiO_{2(quartz)} + 2K^+_{(aq)} \quad (8)$$

Reactions such as (8) produce the Fe-oxides often found in chloritized mafic pseudomorphs in the propylitic zone.

Plagioclase phenocrysts in the outer propylitic zone show minor carbonate alteration along fractures as well as flecking by sericite + carbonate mixtures (Carson and Jambor, 1974 and personal observation). Alteration is essentially of the anorthite component (Beane, 1982) according to the reaction:

$$3CaAl_2Si_2O_{8(anorthite)} + 2K^+_{(aq)} + 3H_2O_{(l)} + 3CO_{2(g)} = 2KAl_2(AlSi_3O_{10})(OH)_{2(muscovite)} + 3CaCO_{3(calcite)} + 2H^+_{(aq)} \quad (9)$$

Alteration increases inward to a maximum near the inner edge of the propylitic zone and sericite-carbonate zone where present, coinciding with the most intense part of the pyrite halo (Carson *et al.*, 1976). Selectively pervasive alteration of plagioclase phenocrysts such as this is most commonly observed in rocks with aphanitic groundmasses although the reasons for this are not clear (Titley, 1982).

Groundmass alteration in the outer parts of the propylitic zone consists of epidote grains and patches and veinlets of carbonate. In the inner part of the zone, the groundmass is more intensely altered to carbonate and chlorite with variable amounts of clays. Clays are more abundant in the inner part of the propylitic zone, where locally plagioclase phenocrysts are completely altered to kaolinite by reactions such as:

$$CaAl_2Si_2O_{8(anorthite)} + 2H^+_{(aq)} + H_2O_{(l)} = Al_2Si_2O_5(OH)_{4(kaolinite)} + Ca^{2+}_{(aq)} \quad (10)$$

$$NaAlSi_3O_{8(albite)} + 2H^+_{(aq)} + H_2O_{(l)} = Al_2Si_2O_5(OH)_{4(kaolinite)} + 4SiO_{2(quartz)} + 2Na^+(aq) \quad (11)$$

Plate 5. Propylitically altered BFP from Morrison showing a relatively stable biotite phenocryst adjacent to two hornblende pseudomorphs completely altered to chlorite with minor rutile needles and pyrrhotite blebs. Plane polarized light. The biotite phenocryst is 400 μm on a side.

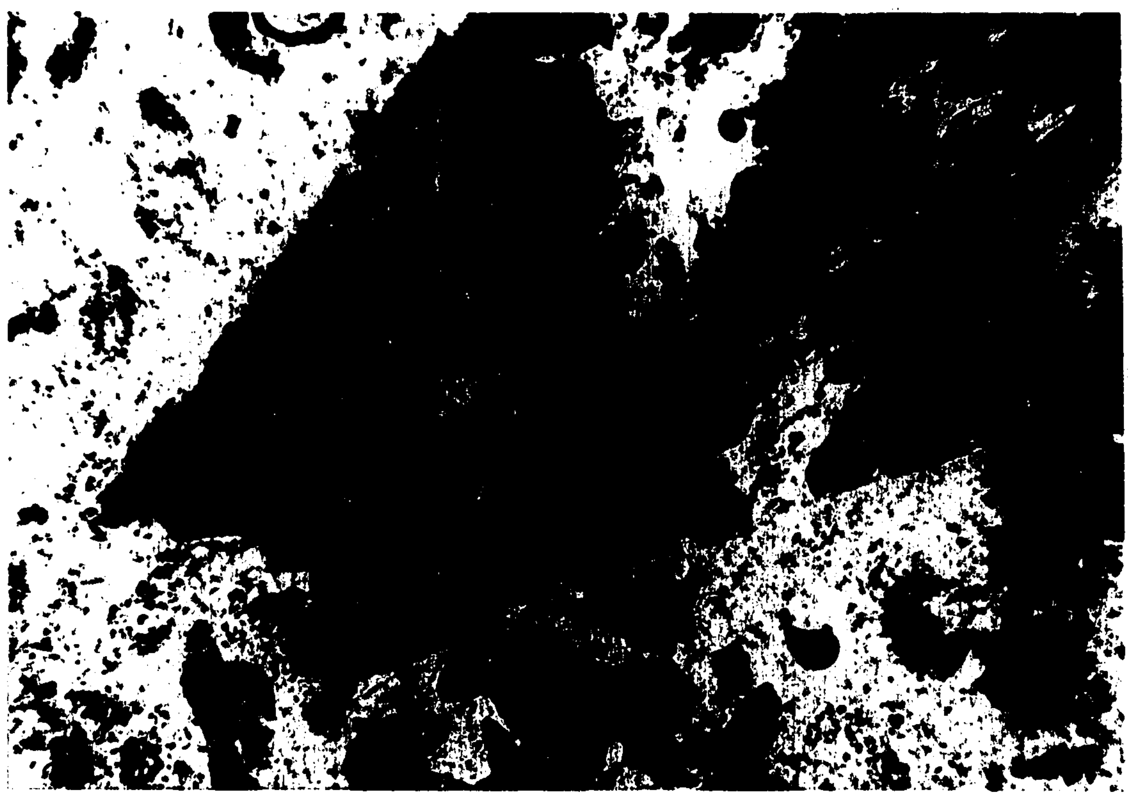

Plate 6. Pyrite replacing hornblende phenocryst, showing relict cleavage planes, from the sericite-carbonate zone of Bell. The field of view is 800 μm across. Plane polarized light.

Opaque phases in the propylitic zone include abundant pyrite disseminations and veinlets in the inner part of the zone (the pyrite halo) with lesser pyrrhotite and minor chalcopyrite (often in mafic pseudomorphs). Also found in mafic pseudomorphs are abundant magnetite, hematite, and rarely ilmenite as well as fine-grained rutile. The outer propylitic zone contains fine-grained disseminations and veinlets of sphalerite and lesser galena (Carson and Jambor, 1974 and personal observation).

iii) Sericite-carbonate alteration

A zone of sericite-carbonate alteration is present at the Bell and Granisle deposits, superimposed between the potassic and propyiitic zones and intrusive-wall rock contact and coinciding with the most intense part of the pyrite halo (Carson and Jambor, 1974; Carson *et al.*, 1976; and personal observation). The sericite-carbonate altered BFP is a massive, bleached, chalky rock and is cut by abundant pyrite-coated fractures, some of which are silicified along the margins. Although the sericite-carbonate zone contains abundant sericite and minor quartz, it is not equivalent to the phyllic zone of Lowell and Guilbert (1970), which contains abundant quartz and sericite with only very minor carbonate. It is a pervasive alteration, with the entire rock being altered to sericite + carbonate + pyrite ± rutile ± quartz ± clays. Minor chalcopyrite disseminations arre present; however, these appear to be largely remnant from earlier potassic and propylitic alterations. Within this zone, plagioclase phenocrysts are entirely altered to sericite + carbonate + clays ± albite and hornblende is generally altered to medium-grained sericite with relict cleavage planes being preserved. Biotite phenocrysts are somewhat more resistant to alteration, with first stages of alteration being carbonate or sericite alteration along cleavage planes. More advanced alteration involves complete replacement by medium-grained sericite ± carbonate with relict cleavage planes. This conversion results from reactions such as:

$$3KMg_3(AlSi_3O_{10})(OH)_{2(phlogopite)} + 20H^+_{(aq)} = KAl_2(AlSi_3O_{10})(OH)_{2(muscovite)} + 6SiO_{2(quartz)} + 2K^+_{(aq)} + 9Mg^{2+}_{(aq)} + 12H_2O_{(l)} \quad (12)$$

and:

$$3KFe^{2+}_3(AlSi_3O_{10})(OH)_{2(annite)} + 38H^+_{(aq)} + 18SO_4^{2-}{}_{(aq)} = KAl_2(AlSi_3O_{10})(OH)_{2(muscovite)} + 9FeS_{2(pyrite)} + 6SiO_{2(quartz)} + 2K^+_{(aq)} + 21H_2O_{(l)} + 63/2O_{2(g)} \quad (13)$$

The groundmass is altered to fine-grained sericite + carbonate + quartz + clays + albite.

Rutile is present as fine aggregates formed from Ti of mafic phenocrysts or ilmenite. Pyrite is present as disseminations and often replaces mafic phenocrysts as shown in Plate 6. Minor chalcopyrite is usually present as fine-grained disseminations after mafic phenocrysts.

This alteration zone has features similar to those of the propylitic, phyllic, and argillic zones of Lowell and Guilbert (1970) and on the basis of field observations and thin section analyses, appears to be a later, overprinting alteration. Although it contains abundant sericite and clays, it differs from the argillic and phyllic zones respectively by the abundance of carbonate and sparsity of quartz.

iv) Phyllic (quartz - sericite - pyrite) alteration

Only the Bell deposit features a strong phyllic or quartz + sericite + pyrite ± chalcopyrite alteration zone which includes much of the ore zone. This zone features an intense quartz stockwork and has a strong structural control (Carson *et al.*, 1976) The phyllic alteration is a pervasive, texturally destructive alteration. Rocks of this zone are more glassy and compact than are the chalky, massive rocks of the sericite-carbonate zone (Carson *et al.*, 1976). The phyllic alteration overprints the earlier potassic, propylitic, and sericite-carbonate alterations (Carson *et al.*, 1976). It overlaps the contact of the intrusive plug and country rocks, altering both the BFP and rhyodacite.

Rocks of this alteration zone are altered almost completely to quartz + sericite with minor carbonate and clays. Pyrite and chalcopyrite are present as disseminations, as fracture coatings, and in the quartz stockwork veins. Bornite, although minor is widespread and is intergrown with chalcopyrite. Some samples also contain abundant specular hematite, especially along the margins of the quartz stockwork veins. Plate 7 shows sericite altered hornblende and plagioclase phenocrysts in a groundmass of quartz and sericite with pyrite disseminations.

Alteration reactions responsible for the formation of this zone include the sericitization reactions discussed previously (e.g., equations (12) and (13)). However, because of the domination of the system by the hydrothermal fluid, extreme silica and minor potassium metasomatism have taken place and most other cations (e.g., Na, Ca, and Mg) have been leached.

v) Summary

As first recognized by Carson and Jambor (1974), the Babine porphyry copper deposits typically feature a central potassic zone containing the ore zone and an outer propylitic zone. In some cases, including the two economic deposits of Bell and Granisle, an intervening sericite-carbonate zone is superimposed. At Bell, the largest deposit, a phyllic alteration zone exhibiting strong structural control is superimposed on both the early potassic-propylitic and the later sericite-carbonate alterations.

Carson and Jambor (1974) also concluded that the quality of alteration and in a sense, the degree to which a porphyry Cu or Mo deposit resembles the hypothetical model is directly correlated with the size and grade of the deposit. They explained that most higher

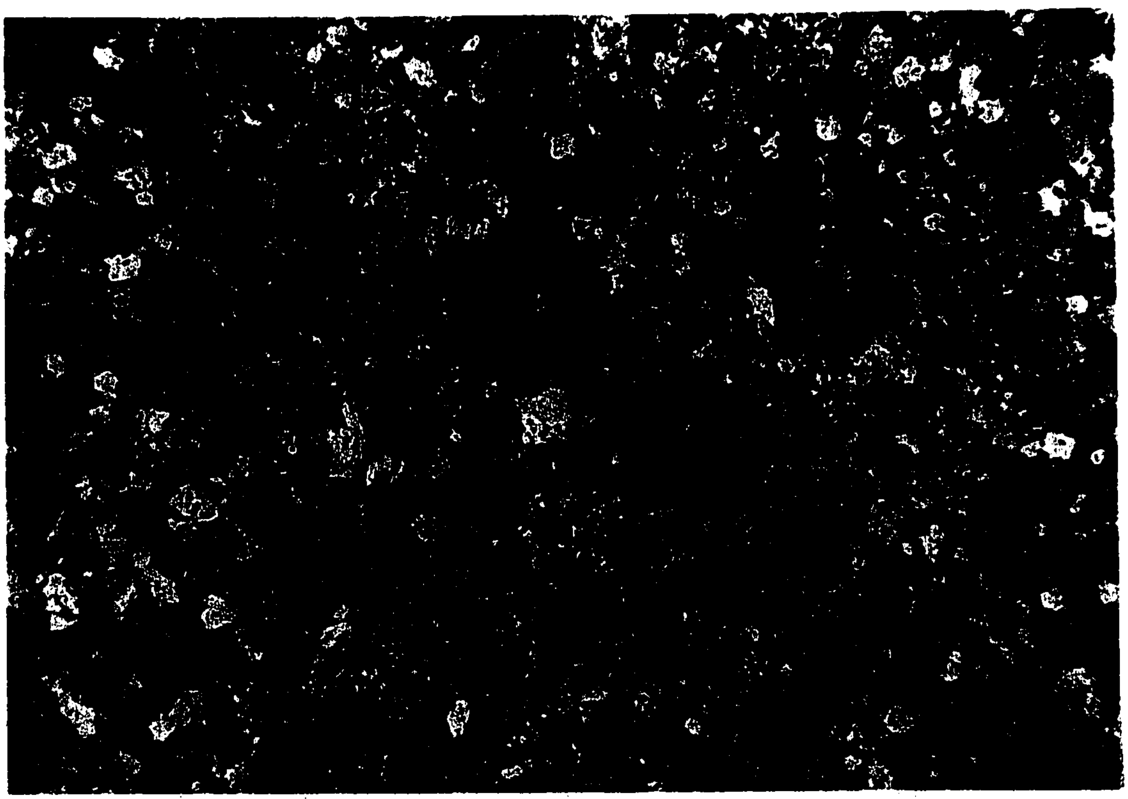

Plate 7. Quartz-sericite alteration of BFP at Bell. The rock is almost completely altered to quartz + sericite + pyrite. Plagioclase phenocrysts are usually replaced by fine-grained sericite whereas biotite and hornblende phenocrysts are replaced by medium-grained sericite. Large hornblende pseudomorph is 800 µm long. Crossed polars.

grade (>0.6 wt % Cu) primary (hypogene) porphyry deposits have strong potassic alteration zones (e.g., San Manuel and Bingham) but in some cases, an intense quartz-sericite alteration may overprint and obliterate this potassic zone (such as at Bell). Slightly lower grade deposits (such as Granisle or Bougainville) have a widespread, moderate potassic alteration zone but this is not as pervasive as in the larger deposits. Lower grade deposits (such as Morrison and Brenda) have propylitic alteration zones closely associated with the higher grade potassic alteration. Marginal deposits have poor quality, patchy potassic alteration and therefore hydrothermal systems without potassic minerals likely do not contain significant Cu or Mo mineralization.

Sulfide zoning at the Babine porphyry deposits features chalcopyrite ± pyrite [± bornite] in the biotite altered core grading outward to pyrite ± chalcopyrite and to pyrite ± pyrrhotite in the pyrite halo (Carson and Jambor, 1974). In addition, the most distinct and annular shaped pyrite halos are associated with the higher grade deposits, and the quantity of pyrite in the pyrite halo is directly related to the quantity of copper in the copper zone (Carson and Jambor, 1974). Bell and Granisle have pyrite halos containing 10% pyrite, Morrison has zones of 10% pyrite, and Dorothy, North Newman, and Nakinilerak have lower amounts of pyrite. These pyrite halos are strongest at the potassic-propylitic transition or where present, in the sericitic alteration zones.

Morrison

i) Exploration history

The exploration history of Morrison is summarized by Carson and Jambor (1976) as follows. Anomalous geochemical stream samples collected in 1962 were followed in 1963 by the discovery of Cu-bearing float and outcrop in the stream flowing over the ore zone. This in turn was followed by trenching. From 1963 to 1973, 95 diamond drill holes were bored and induced polarization and magnetic surveys were undertaken. Geological mapping was undertaken in 1963 and 1967 and by 1967 a well developed biotite-chlorite zonation was defined, with biotization being shown to be related to copper mineralization. Mineable reserves at Morrison are estimated at 86 million tons grading 0.42% Cu.

ii) Hydrothermal alteration and sulfide zonation

a) Potassic alteration

Potassic (biotite) alteration at Morrison extends beyond the intrusive rocks into the Bowser Lake Group sedimentary rocks. The hydrothermal zonation patterns are shown in Figure 16. As with the other Babine Lake deposits, the quality of the hydrothermal biotite is directly related to copper grades and to the proximity to the ore zone (Carson and Jambor, 1974 and personal observation). However, the biotite alteration is not as intense as at Bell or Granisle, in keeping with the observation of Carson and Jambor (1974) that

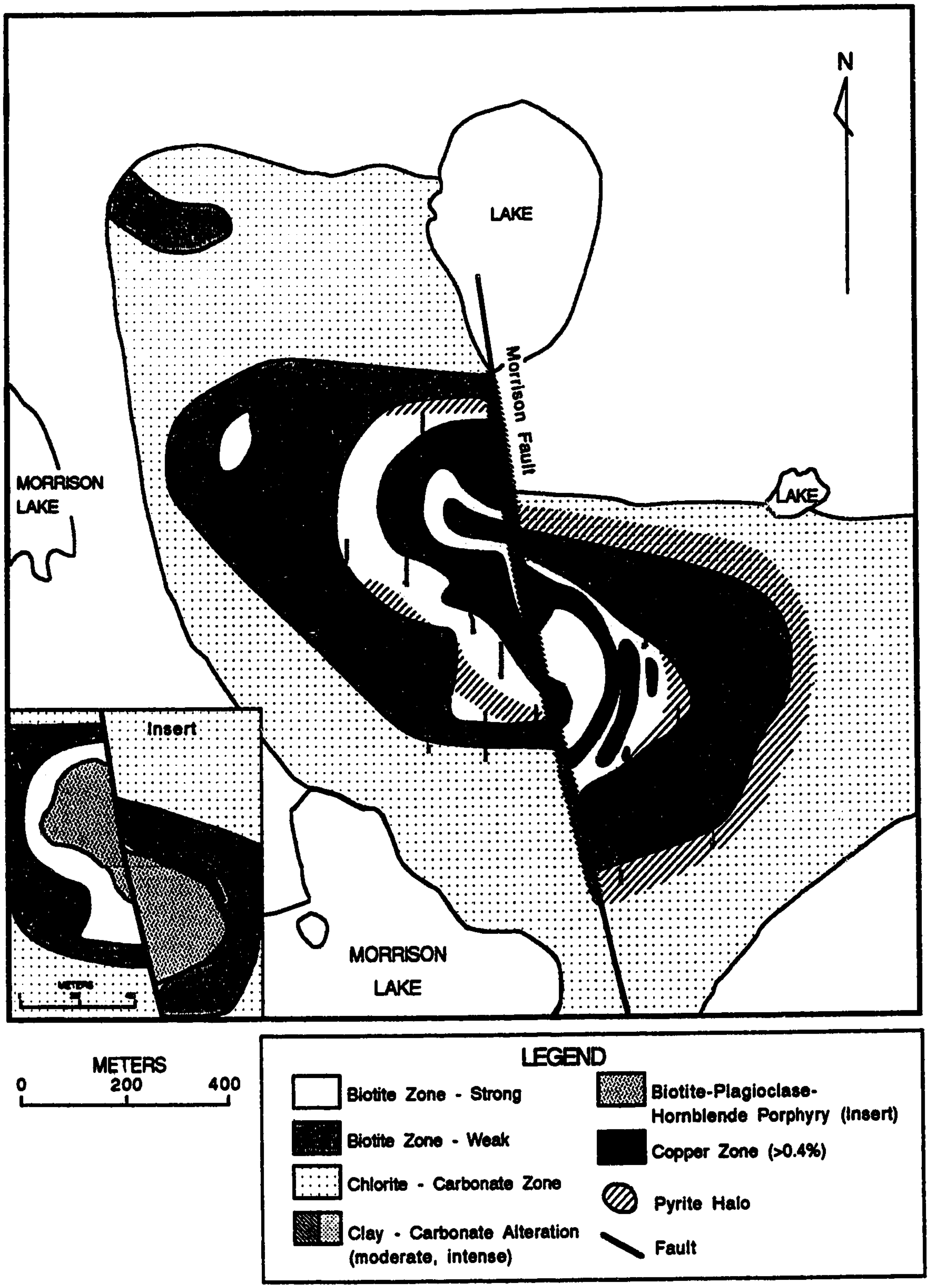

Figure 16. Hydrothermal alteration and sulfide zonation at Morrison (after Carson and Jambor, 1976).

potassic minerals in marginal deposits are of inferior quality. In addition, unaltered cores are present in some amphibole phenocrysts.

The most intensely altered samples feature abundant disseminated and veinlet secondary biotite in the groundmass and well crystallized biotite replacing hornblende phenocrysts. In some cases, biotite phenocrysts are replaced along the edges by hydrothermal biotite. Plagioclase phenocrysts are generally only weakly altered to sericite ± carbonate along fractures, edges, and Ca-rich zones. Minor hydrothermal amphibole of the tremolite-actinolite series is locally present and is thought to be a result of the low $f\,CO_2$ of the hydrothermal fluids (Carson and Jambor, 1974). This may be explained by reactions such as:

$$2KMg_3AlSi_3O_{10}(OH)_{2(annite)} + 4SiO_{2(quartz)} + 3CaCO_{3(calcite)} + 4H^+_{(aq)} = Ca_2Mg_5Si_8O_{22}(OH)_{2(tremolite)} + CaAl_2Si_2O_{8(anorthite)} + 3H_2O_{(l)} + 3CO_{2(g)} + 2K^+_{(aq)} + Mg^{2+}_{(aq)} \quad (14)$$

Minor sericitization of the groundmass is present as is local chloritization of the groundmass and mafic phenocrysts. Medium-grained quartz is present as an alteration product of the groundmass in some sections.

Magnetite disseminations are abundant in some samples and are commonly accompanied by fine-grained chalcopyrite disseminations. Locally magnetite is altered to hematite and in some cases hematite may be a primary alteration mineral.

The dominant ore mineral at Morrison is chalcopyrite, found as disseminations, fracture coatings, and thin narrow veinlets with or without quartz or biotite. As described by Carson and Jambor (1976), Morrison features a central low grade core (0.15 to 0.20% Cu) of chalcopyrite, pyrite, magnetite, and minor bornite surrounded by an annulus 15 to 150 m wide grading 0.5% Cu. The deposit also averages 0.01% Mo, 0.3 ppm Au, and 3 ppm Ag. The northwest side of the ore zone is larger than the southeast side, with a larger 0.5% Cu zone and a 0.3% Cu zone extending into the sedimentary country rocks.

Pyrite is present in some thin sections as very fine-grained disseminations, larger disseminations overprinting chalcopyrite alteration, or as thin veinlets with chalcopyrite, calcite, quartz, and clays. The sulfide zoning at Morrison is not as distinct as at Granisle or Bell and the >0.3% Cu zone contains moderate amounts of pyrite (Carson and Jambor, 1974). Some chalcopyrite is younger than the pyrite since it is often present as overgrowths. In some thinsections sphalerite veinlets and overgrowths on pyrite and chalcopyrite are present. As noted by Carson and Jambor (1976), the Cu zone contains significant but uneconomic levels of Pb and Zn. Locally, thin sphalerite veinlets and gypsum veinlets crosscut quartz veinlets.

Thin calcite veining and later alteration of plagioclase cores to kaolinite are locally present. This is confined to the bleached BFP in the Morrison Fault zone and in dike contacts, bedding planes, and fractures subparallel to the Morrison Fault (Carson and Jambor, 1976). Some samples feature complete alteration of plagioclase phenocrysts to kaolinite and calcite and kaolinite alteration envelopes are present around some quartz-calcite-pyrite veinlets.

The pyrite halo overlaps the outer, weaker part of the potassic zone, the potassic-propylitic transition, and the inner part of the propylitic zone and is stronger, wider, and more continuous in the southeastern half of the deposit (Carson and Jambor, 1976). These samples contain abundant disseminations of pyrite and variable but minor chalcopyrite. Pyrrhotite is more abundant than at Bell and Granisle (Carson and Jambor, 1976 and personal observation) and is present as disseminations with pyrite, chalcopyrite, and often marcasite. The marcasite is often present as fine-grained masses rimming or locally replacing pyrrhotite and pyrite and appears to be a late stage alteration product.

b) Propylitic alteration

Propylitic alteration at Morrison is characterized mainly by chloritization of mafic phenocrysts. Hornblende is extensively altered to chlorite, carbonate, very fine-grained rutile, and magnetite or pyrite. Most of the chlorite is pale yellow to green, with anomalous blue, brown, or greenish brown interference colours but chlorite from samples containing sericite is colourless. One sample contains radiating masses of brownish to greenish-blue tourmaline. Biotite phenocrysts are generally unaltered but may show chloritization along zones or cleavage planes and may be altered to colourless chlorite where associated with sericite. Some samples from the peripheral portion of the propylitic zone contain epidote alteration. They also contain disseminated magnetite which is locally altered to hematite.

Plagioclase phenocrysts are weakly to strongly altered to sericite and carbonate along fractures, with alteration increasing inward and reaching a maximum in the pyrite halo (Carson and Jambor, 1976 and personal observation). Plagioclase locally has undergone complete alteration to kaolinite and carbonate.

c) Sericite-carbonate alteration

Only minor sericite alteration is present at Morrison. One sample of sericite-carbonate altered Bowser Lake Group siltstone from Morrison was collected. This sample was altered by sericite + feldspar + carbonate ± quartz ± pyrite ± pyrrhotite veins with distinct, bleached sericite alteration envelopes. Pyrrhotite and pyrite show local alteration to marcasite and minor Fe-oxides after pyrrhotite were observed.

d) Clay-carbonate alteration

Intense argillic or more properly, clay-carbonate alteration is confined to the Morrison Fault zone, a linear zone of parallel shears and fractures averaging 75 m wide and bisecting the deposit (Carson and Jambor, 1976). Weaker clay-carbonate alteration is locally found in minor shears and structures parallel to the Morrison Fault (Carson and Jambor, 1976).

One BFP sample from the Morrison Fault zone features intense kaolinization of plagioclase phenocrysts and alteration of mafic phenocrysts to chlorite and/or montmorillonite. Groundmass feldspar (medium-grained) is locally clouded and pitted. Fine-grained disseminated rutile is present in mafic pseudomorphs from liberation of Ti from the igneous biotite and hornblende. Minor disseminated pyrite and marcasite are also present.

This clay-carbonate alteration appears to be a late stage phenomenon (Carson and Jambor, 1976 and personal observation) and locally Pb and Zn sulfides and carbonate-cemented gouge and breccia are superimposed (Carson and Jambor, 1976). The Morrison Fault has grades of 0.20 to 0.25% Cu from dilution and leaching of copper during the faulting, hydrothermal activity, and groundwater leaching (Carson and Jambor, 1976). It is marked by a distinct magnetic low due to the leaching of oxide, sulfide, and mafic minerals (Carson and Jambor, 1976).

iii) Paragenetic sequence

The paragenetic sequence of the Morrison deposit is shown in Figures 17, 18, and 19. Figure 17 shows the alteration in the core of the system, Figure 18 represents the alteration in the periphery and Figure 19 shows the overall sequential development of the system. In all cases, the relative amounts of the introduced minerals are shown by the respective line widths. The relative stabilities of the minerals are also shown. It should be noted that the extent of alteration varies with location in the deposit and therefore, earlier phases may be locally preserved depending on the intensity of the alteration although not indicated as such in the paragenetic diagrams. For example, although hornblende is unstable in all alteration zones, cores may by preserved locally where alteration was weak. Also, the depiction of "stages" of alteraton is merely for illustrative purposes and in reality, alteration and zonation are more of a continuum. As cautioned by Hemley and Hunt (1992), paragenetic diagrams should be used conservatively because they cannot depict spatial variations and therefore certain phases may be unstable in one part a zone at a given time and stable in another location. In this study, paragenetic diagrams are used to show the relative stabilities of minerals in the core and peripheral regions and the generalizations made therein should not be used without the discussions of the zoning and the figures showing both the lateral and temporal zonation patterns.

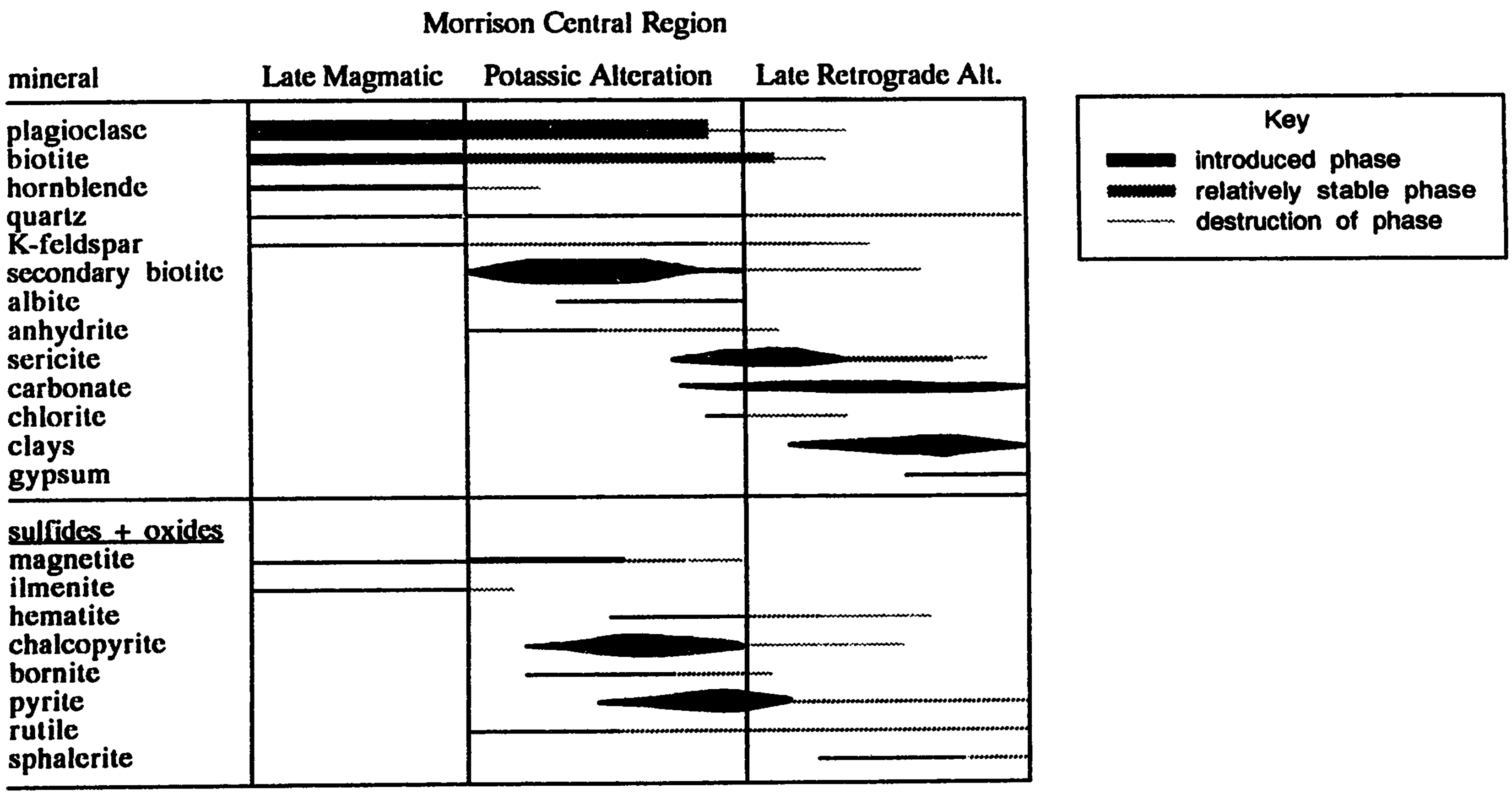

Figure 17. Paragenetic sequence of the central region and ore zone of the Morrison deposit.

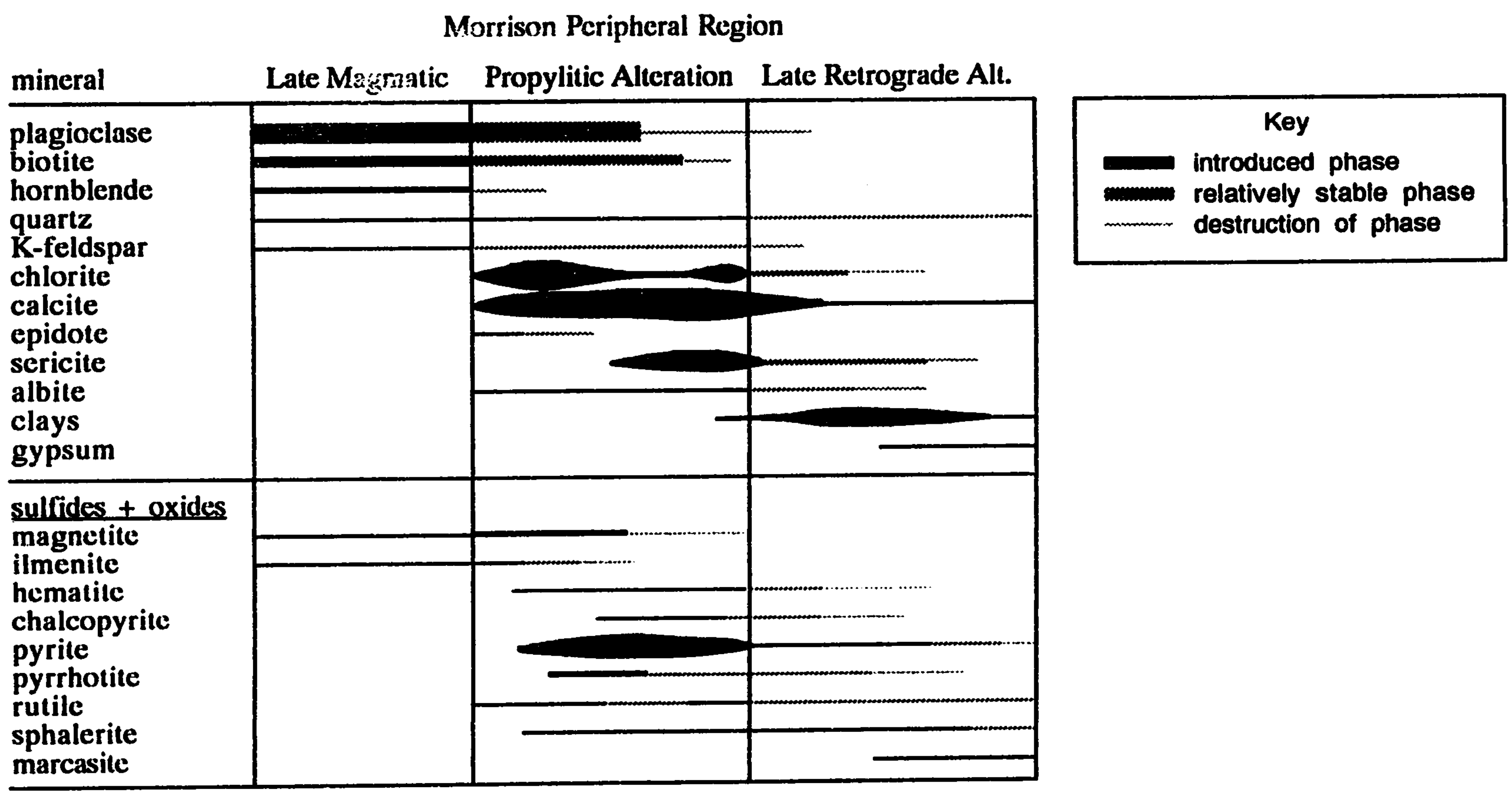

Figure 18. Paragenetic sequence of the peripheral region and pyrite halo of the Morrison deposit.

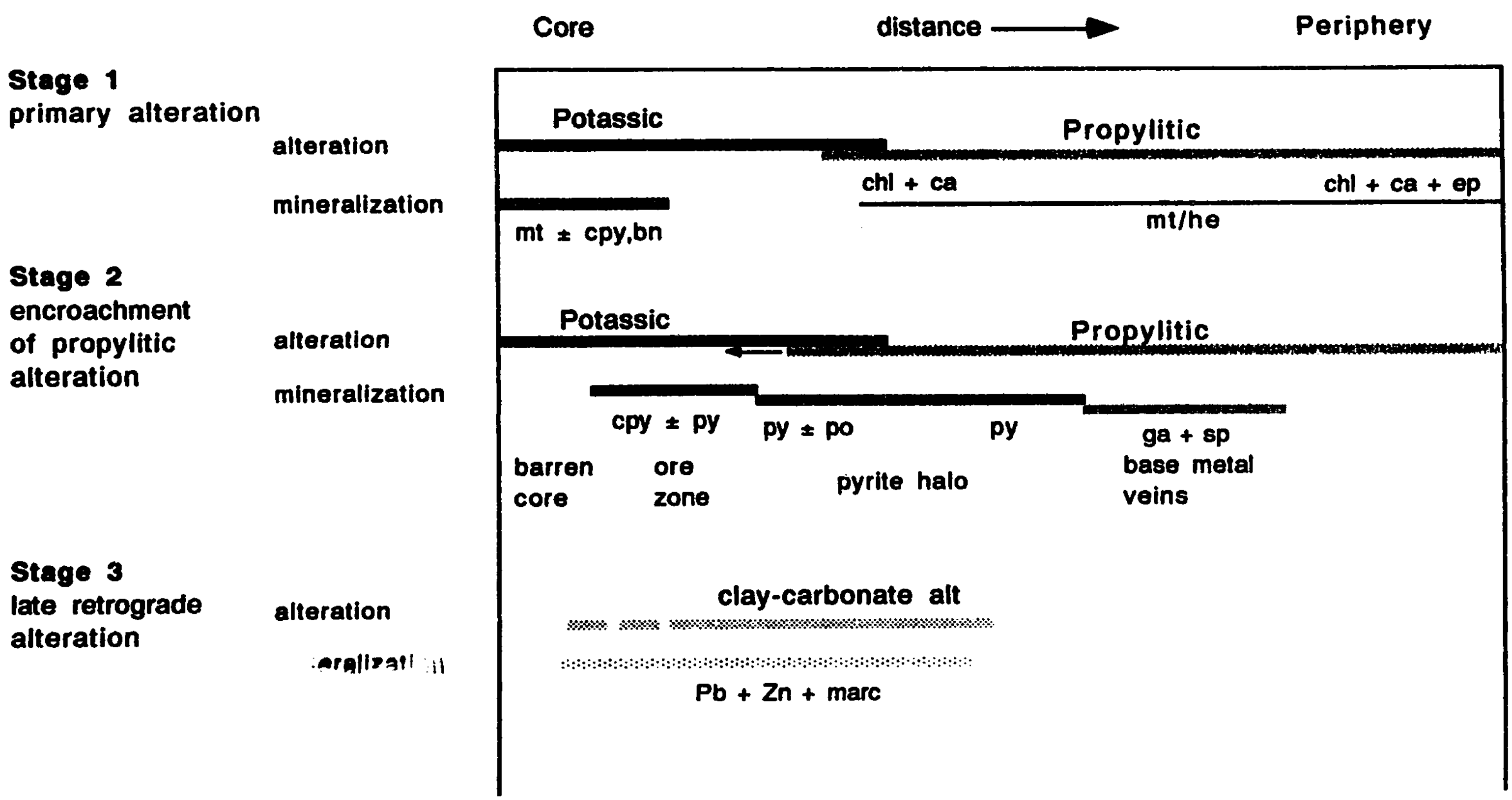

Figure 19. Lateral and temporal alteration and mineralization zonation development at Morrison.

a) Early alteration

Early alteration in the core region consisted of potassic alteration, with secondary biotite replacing hornblende phenocrysts, biotite alteration of the groundmass, and minor potassium feldspar alteration associated with thin quartz veins. Accompanying this silicate alteration was magnetite (dominantly in the barren core), chalcopyrite, and minor bornite and pyrite. This constitutes the mineralized zone and is interpreted by Carson and Jambor (1976) to be coeval with the potassium silicate alteration.

However, this author believes that although ore deposition was associated spatially with biotite alteration, the bulk of the ore deposition likely took place after the most intense potassic alteration. Evidence for this is the presence of chlorite in chalcopyrite-bearing veinlets, indicating that during deposition of at least part of the ores, alteration conditions were changing from those accompanying the intense biotite alteration. Other lines of evidence include the observations from Bell and Granisle discussed below and the observation of ore deposition late in the potassic alteration phase in other studies (e.g., Titley, 1982 and Beane, 1982).

Minor sericitization of plagioclase phenocrysts took place during potassic alteration, most likely during later stages. Anhydrite deposited during the potassic alteration event was largely redissolved or hydrated to form gypsum at a later time.

Development of the pyrite halo took place peripherally to the ore zone in the outer potassic and inner propylitic zones. This consisted of pyrite and pyrrhotite (the latter more abundant than in either Bell or Granisle). Magnetite deposited during the early stages of potassic alteration was partially altered to hematite. The relative lack of pyrrhotite is likely due to the relatively high fO_2 of porphyry copper deposits, stabilizing Fe-oxides in early stages and pyrite later (Hemley and Hunt, 1992).

Peripheral to the potassic alteration was the development of propylitic alteration, which slowly encroached inward with time. Sericite and carbonate alteration accompanied this alteration zone, being most prominent in the inner part of the zone, corresponding with the pyrite halo. Peripheral to the pyrite halo are disseminations and veinlets of sphalerite and galena (Carson and Jambor, 1976 and personal observation). As the system cooled, pyrite, galena, and sphalerite were deposited in minor amounts in the central part of the system.

As explained by Hemley and Hunt (1992), metal zoning is a function of both relative metal solubilities and also the manner of intersection of transport paths with the individual metal saturation surfaces. As a result, early magnetite forms at the highest temperatures followed by chalcopyrite which is in turn followed by sphalerite and galena. As the system cools, the saturation surfaces are intersected closer to the core of the system and therefore lower temperature assemblages overprint earlier assemblages (Hemley and Hunt, 1992). The relative positions of the solubility curves change with temperature and since iron remains in solution longer at low temperatures relative to other metals, the development of late pyrite is a result of cooling of the system and results in the pyrite (along with sericite) overprint of many porphyry copper deposits (Hemley and Hunt, 1992).

b) Late overprinting

Overprinting the primary hypogene zonation pattern is the localized clay-carbonate alteration near the Morrison Fault and parallel fractures as discussed previously. This alteration locally overprints the potassic and propylitic zones and is accompanied by marcasite, pyrite, arsenopyrite, galena, sphalerite, geochronite, and boulangerite in vuggy quartz-carbonate veinlets (Carson and Jambor, 1976). During late stage alteration accompanying cooling of the hydrothermal system, minor gypsum veining occurred and gypsum replaced earlier formed anhydrite.

Granisle

i) Exploration history

The exploration history of Granisle is summarized by Fahrni *et al.* (1976). McDonald Island was first prospected in the early 1900's, with the discovery of several veins in what later became the Granisle mine. In 1929, Cominco Ltd. drilled five holes. The property stood idle until 1955 when Granby acquired it and drilled holes on a 60 m spacing. Drilling and a feasibility study were undertaken in 1964, the minesite was constructed in 1965, and production began in 1966. Reserves at Granisle totalled 85 million tons grading 0.43% Cu with negligible molybdenum (Drummond and Godwin, 1976) and 0.2 ppm Au (0.0048oz/ton) (Cuddy and Kesler, 1982). The mine operated until closing in the early 1980's.

ii) Hydrothermal alteration and sulfide zonation

a) Potassic alteration

Alteration at Granisle consists of a central potassic or biotite alteration zone surrounded by a sericite-carbonate zone and a peripheral propylitic zone. The alteration patterns are shown in Figure 20. Samples from the potassic zone appear essentially unaltered in hand specimen, the colour typically being medium grey to black and weathering orange or grey.

The ore body is entirely contained within the biotite zone with sugary, coarse-grained biotite replacing hornblende and abundant fine-grained, shreddy biotite in the groundmass (Carson and Jambor, 1974). Outside the ore zone secondary biotite is finer grained, less abundant, and may be intermixed with chlorite (Carson and Jambor, 1974). Biotite phenocrysts are generally persistent throughout the biotite alteration zone, with replacement by hydrothermal biotite taking place in only the most intensely altered samples and even

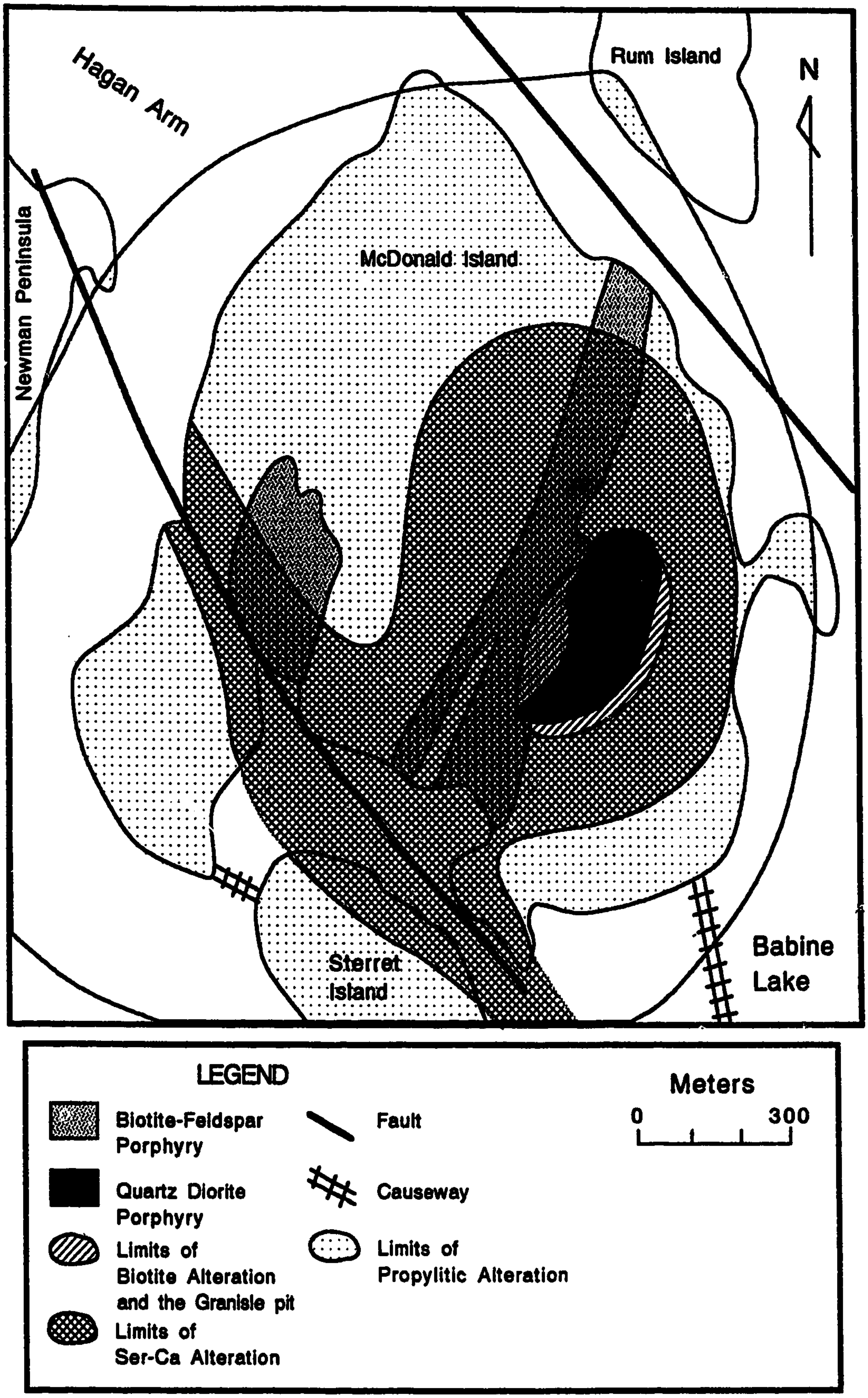

Figure 20. Hydrothermal alteration zonation pattern at Granisle and McDonald Island (after Fahrni *et al.*, 1976).

then only along the edges (Carson and Jambor, 1974). Hydrothermal amphibole is rare but was observed in one sample, with brown hornblende being altered to actinolite and biotite.

Plagioclase phenocrysts remain relatively unaltered in the biotite zone but alteration ranges from minor sericitization and carbonatization along fractures and Ca-rich growth zones to complete replacement. Minor clay alteration of plagioclase phenocrysts is present, especially in the outer parts of the biotite zone (Carson and Jambor, 1974 and personal observation). Compared to Bell, carbonate alteration and veining is minimal. Minor amounts of potassium feldspar are present at Granisle, usually associated with quartz in thin veinlets and restricted to the +0.3% Cu zone (Carson and Jambor, 1974 and personal observation). Other alteration phases are anhydrite and gypsum (Carson and Jambor, 1974 and personal observation). Gypsum occurs widespread in the ore zone and pyrite halo, entirely within the biotite zone (Carson and Jambor, 1974).

The ore zone is located near the contact between the BFP and the earlier quartz diorite porphyry phase and consists of chalcopyrite and bornite with lesser amounts of pyrite in the central potassic alteration zone (Fahrni *et al.*, 1976). Nearly all ore samples lack disseminated pyrite and significant bornite is present only in the higher grade portions of the deposit (Carson and Jambor, 1974). Carson and Jambor (1974) also noted that the inward sulfide zonation pattern of pyrite - chalcopyrite - chalcopyrite±bornite is classical in porphyry copper deposits.

The chief ore mineral is chalcopyrite, mostly found in quartz filled fractures, especially those striking at 035° to 060° and 300° to 330° and dipping steeply (Fahrni *et al.*, 1976). In contrast, horizontal fractures are only weakly mineralized (Fahrni *et al.*, 1976). Sulfide fracture coatings without quartz are also important and include chalcopyrite, chalcopyrite + bornite, and chalcopyrite + bornite + pyrite. Bornite is found in the southern half of the ore zone in quartz-filled fractures especially in the upper 75 m of the deposit (only in higher grade portions of the deposit (Carson and Jambor, 1974) and in veins and breccia zones (<1 m wide) of coarse-grained bornite, chalcopyrite, quartz, biotite, and apatite (Fahrni *et al.*, 1976). Disseminated opaques in the most intensely biotized rocks consist of chalcopyrite, hematite, magnetite (often altering to hematite) with minor bornite (intergrown with chalcopyrite) and molybdenite. Samples with weaker biotite alteration have disseminations of chalcopyrite (± bornite), pyrite, rutile, hematite, and magnetite.

Molybdenum is locally present in drusy quartz veinlets (Fahrni *et al.* ,1976), as disseminations, and in thin fractures and appears to be later than the main mineralization stage. Magnetite and specular hematite are present in the north half of the ore zone in chalcopyrite- and pyrite-bearing fractures (Fahrni *et al.*, 1976). Covellite is locally found as an alteration product of chalcopyrite.

b) Propylitic alteration

Propylitic altered samples at Granisle are dark grey to greenish grey and weather orange-brown. Plagioclase phenocrysts often weather out, enhancing the texture. The propylitic zone is gradational outward to unaltered BFP (Carson and Jambor, 1974). It features initial replacement of hornblende phenocrysts by chlorite, carbonate, and ilmenite, magnetite or rutile and partial replacement of biotite phenocrysts along cleavage planes by carbonate (Carson and Jambor, 1974 and personal observation). Minor epidote is also present (Carson and Jambor, 1974). Chlorite in the inner part of the zone is well crystallized and has green to brown interference colours whereas in the outer zone, it is finer grained with blue to grey interference colours (Carson and Jambor, 1974 and personal observation). Plagioclase is partially replaced by sericite, carbonate, and chlorite, reaching a maximum in the pyrite halo and decreasing outward (Carson and Jambor, 1974). The BFP matrix is altered with sparse carbonate patches in the outer part of the zone and with chlorite, carbonate, sericite, pyrite, kaolinite, and quartz in the inner part of the zone (Carson and Jambor, 1974 and personal observation). Carbonate in the propylitic and sericite-carbonate zones is largely ankerite and ferroan dolomite (Carson and Jambor, 1974). Chlorite and carbonate veins are common in the outer part of the propylitic zone.

Ilmenite and magnetite exist as xenomorphic to idiomorphic grains of probable igneous origin and as minor alteration grains in hornblende pseudomorphs. Minor chalcopyrite is present associated with magnetite and occasionally with ilmenite in pseudomorphs. Minor hematite is also present.

A quartz + carbonate + pyrite + galena + sphalerite + chalcopyrite vein with silver has been found at the south end of the island, following a northeast striking fault (Carson and Jambor, 1974). This is consistent with the frequent occurrence of zinc-rich veins peripheral to porphyry deposits (Lowell and Guilbert, 1970 and Carson and Jambor, 1976).

The pyrite halo is located mostly within the sericite-carbonate zone and is described in detail in the next section.

c) Sericite-carbonate alteration

The sericite-carbonate zone forms a partial ring around the deposit and merges with alteration along a regional fault to the southwest (Carson and Jambor, 1974). This zone overprints the earlier potassic and propylitic zones near their contact and is best developed on the east side of the Granisle pit (Carson and Jambor, 1974).

Rocks from this zone are light grey to buff and plagioclase, hornblende, and biotite phenocrysts display varying degrees of alteration to sericite, calcite, sulfides, and clays depending on the intensity of the alteration. Although minor quartz is introduced during this alteration phase, the zone has significant differences from the classic phyllic alteration assemblage. Plagioclase phenocrysts are first replaced by fine-grained sericite, carbonate, and clays along fractures and Ca-rich zones with alteration continuing to complete replacement where alteration is intense. Hornblende phenocrysts are generally completely replaced by fine- to medium-grained sericite (muscovite) + carbonate + pyrite + rutile ± quartz but locally hydrothermal biotite or chlorite from previous alteration assemblages remains. In areas of weak alteration, biotite phenocrysts show carbonate alteration along cleavage planes but with increasing alteration, the biotite phenocrysts are replaced by fine- to medium-grained oriented sericite, often preserving the biotite cleavage traces. Lesser amounts of carbonate are present and pyrite pseudomorphs of mafic phenocrysts are common. The groundmass is largely altered to sericite + quartz with lesser amounts of carbonate but in less intensely altered samples, plagioclase and chlorite may remain.

Opaque minerals are dominated by pyrite. The pyrite halo is mostly contained within the sericite-carbonate zone. It is 150 to 250 m wide with pyrite disseminations, blebs, fracture coatings, and stringers (Fahrni *et al.*, 1976), averaging 10% pyrite by volume (Carson and Jambor, 1974). Fine-grained rutile is ubiquitous in mafic and oxide pseudomorphs and as groundmass disseminations. It is likely an alteration product formed from Ti remaining after alteration of igneous biotite and hornblende and to a lesser extent, ilmenite and magnetite.

Minor, small chalcopyrite disseminations are locally present, probably as a result of incomplete dissolution by the sericite-carbonate-pyrite alteration fluids. Some pyrite blebs also contain small inclusions of pyrrhotite and chalcopyrite, some exhibiting myrmekitic intergrowth textures. These pyrrhotite inclusions are probably remnant from an earlier alteration halo similar to that of Morrison. Trace sphalerite and covellite are also present associated with chalcopyrite.

This alteration zone is a pervasive, texture- and ore-destructive alteration type in which chalcopyrite is not only not precipitated but likely is dissolved. Although minor chalcopyrite may be present with pyrite disseminations or in veinlets, there is a net reduction of chalcopyrite and bornite within the sericite-carbonate alteration zone. This is observed on the thinsection scale as a scarcity of chalcopyrite disseminations in proximity to overprinting pyrite veinlets and intense sericite-carbonate-pyrite alteration.

Sericite-carbonate alteration begins firstly as alteration along fractures and progresses outward to meet other alteration envelopes until the entire rock mass is altered. Three samples of this type were observed at the northwest end of the Granisle pit and one is shown in Plate 8. Here thin sericite-carbonate alteration envelopes with central pyrite ± quartz ± carbonate veinlets were located in biotite altered BFP. Within the alteration

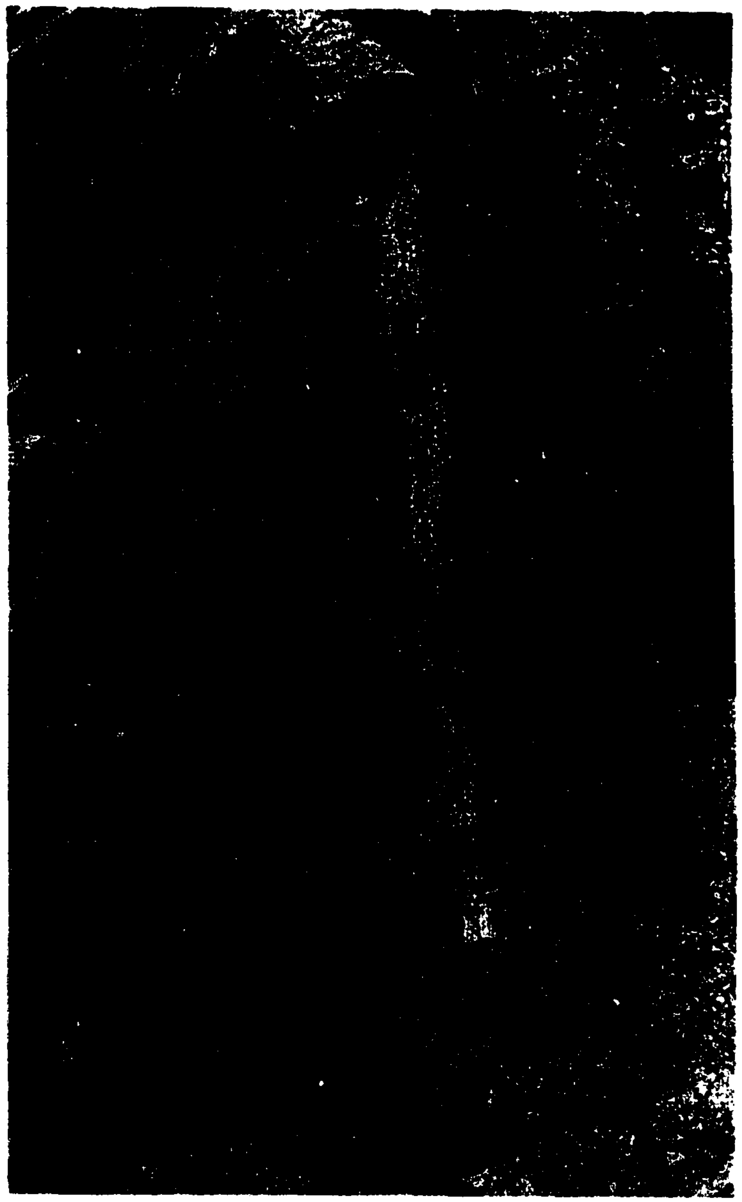

Plate 8. Sericite-carbonate alteration, often with pyrite veins overprinting potassic alteration in BFP from the Granisle mine.

envelope plagioclase phenocrysts are altered to fine-grained mixtures of carbonate + sericite ± clays and mafic phenocrysts are altered to sericite + chlorite, with alteration increasing toward the vein. The quartz within some of the central veinlets was found to contain high salinity fluid inclusions with multiple daughter salts (personal observation) and therefore at least part of the sericite-carbonate alteration took place under the influence of high salinity fluids. The presence of kaolinite with quartz sets an upper temperature limit of 265°C for at least part of the alteration (Beane, 1982). Outward, chlorite abundance increases and chlorite is intergrown with significant orange-brown, fine-grained stilpnomelane. The alteration envelopes grade into biotite altered BFP within less than 3 cm.

Minor coarse-grained carbonate + sulfide + quartz veins occur in the sericite-carbonate zone. One sample features coarse-grained chalcopyrite and pyrite in a carbonate (ankerite) ± quartz vein and features thin veinlets of molybdenite crosscutting the chalcopyrite and pyrite, indicating the relatively late introduction of the molybdenite mineralization. Chalcopyrite is locally altered to covellite, bornite, and chalcocite.

iii) Paragenetic sequence

The paragenetic sequence of the Granisle deposit is shown in Figures 21, 22, and 23. Figures 21 and 22 show the development of alteration in the central and peripheral portions of the system respectively. As in the Morrison study, the paragenesis diagrams give only the sequential development of alteration assemblages and do not indicate the intensity of the alteration nor the spatial variations. The lateral extents and relative intensities of the alteration types are shown in Figure 23. As shown, the sericite-carbonate alteration overprints the potassic and propylitic zones mostly at their mutual boundary.

Alteration zonation is believed to have been the same as Morrison in early stages, in keeping with the interpretation of Carson *et al.* (1976). Subsequent to the development of a central potassic alteration zone and peripheral propylitic zone, a sericite + carbonate + pyrite ± quartz alteration was superimposed near the potassic-propylitic transition. This coincides with the most intense part of the pyrite halo and likely contains pyrite from two sources; the pyrite associated with the primary alteration zonation and pyrite introduced during the sericite-carbonate-pyrite alteration. This phase was ore-destructive, indicated by a reduction in the copper content of these rocks on both macroscopic and microscopic scales.

As discussed previously, the sericite-carbonate alteration began along fractures and was accompanied at least in part by high salinity fluids. The frequent presence of kaolinite and other clays in this zone suggests that temperatures were below 265°C for at least part of the alteration. The sericite-carbonate alteration is interpreted by Carson *et al.* (1976) to have been caused by residual magmatic fluids, cooled since the potassic alteration but as discussed later, is interpreted by this author to have involved surficial fluids.

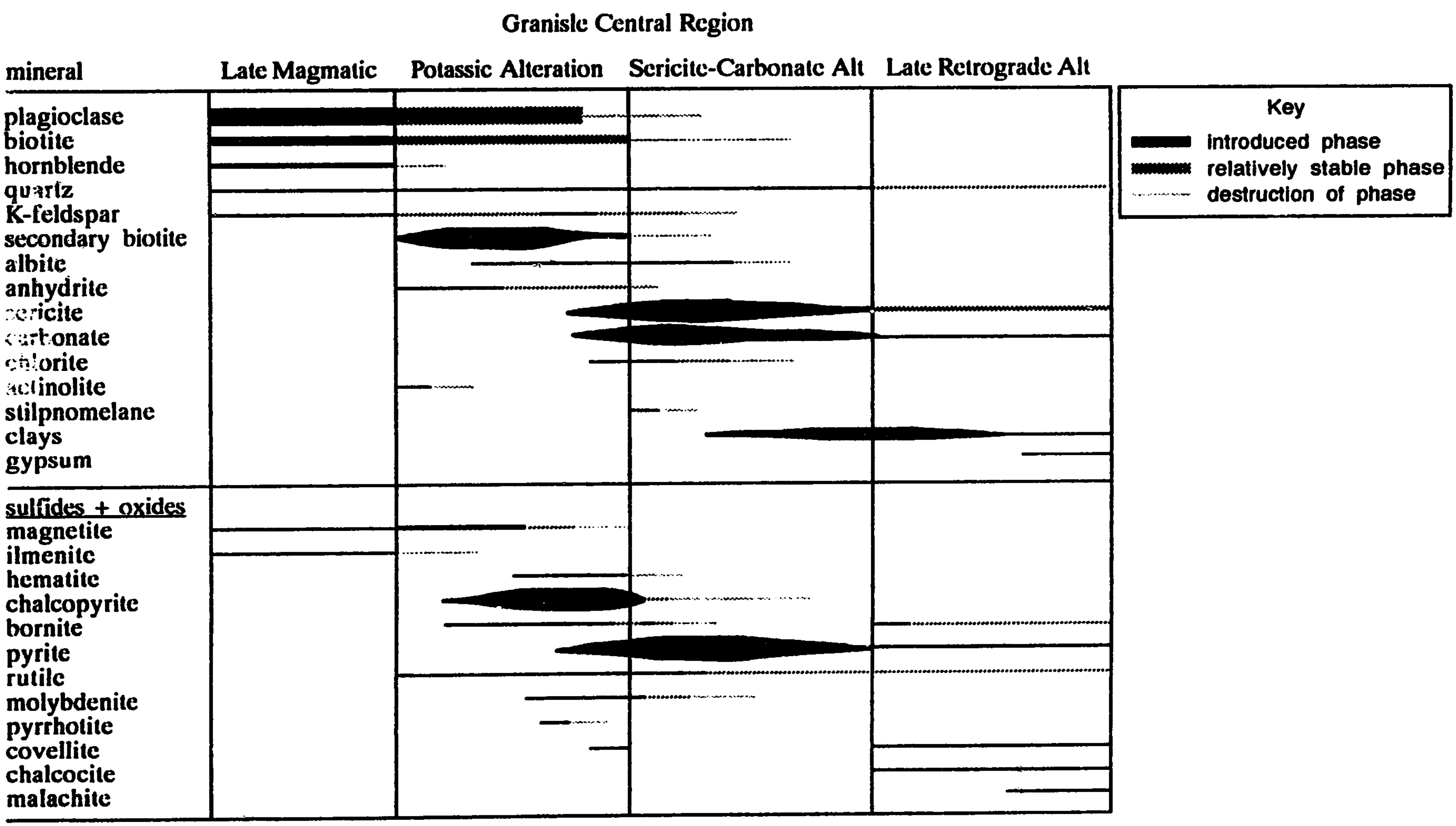

Figure 21. Paragenetic sequence of the central region and ore zone of the Granisle deposit.

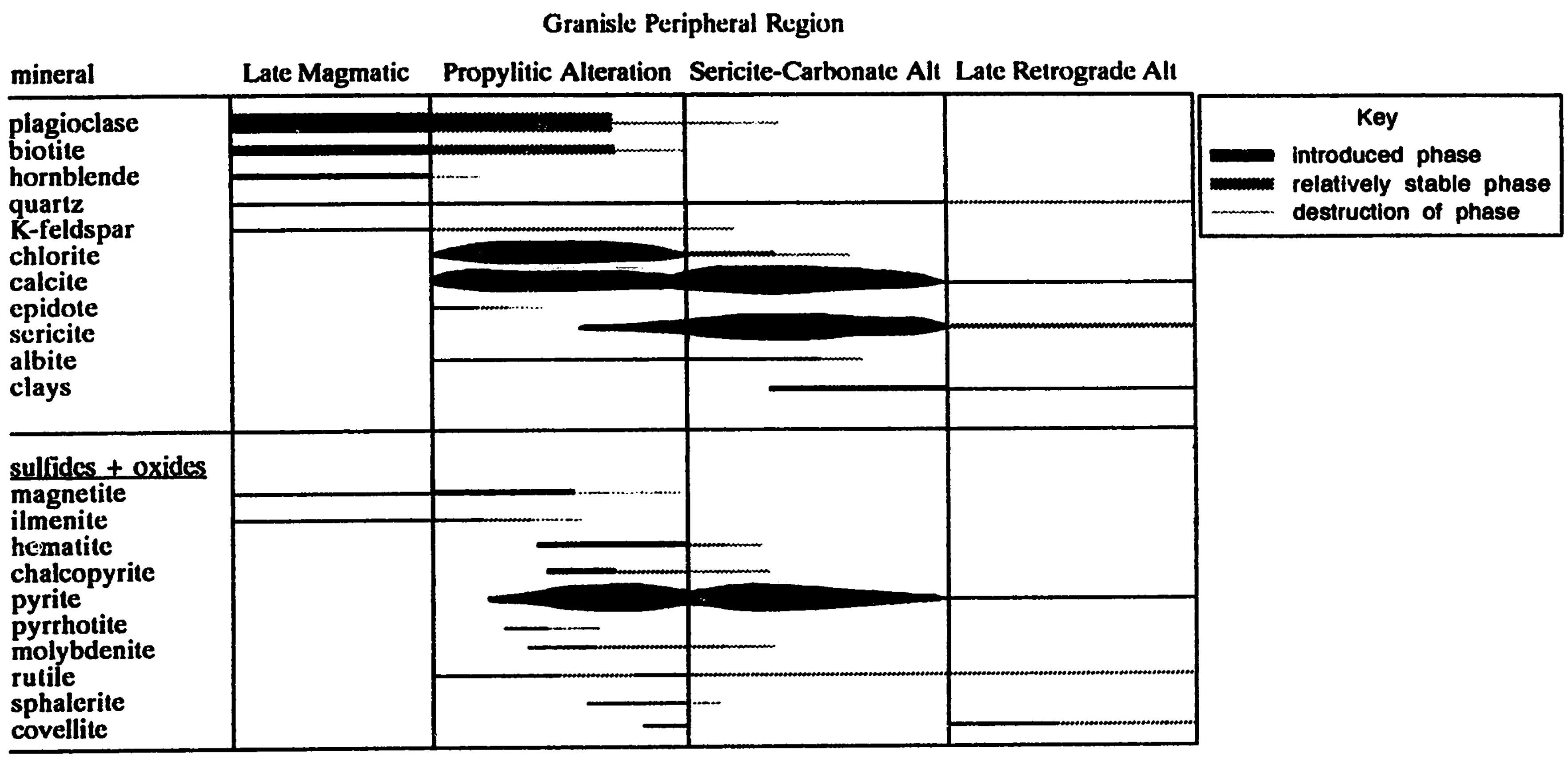

Figure 22. Paragenetic sequence of the peripheral region and pyrite halo of the Granisle deposit.

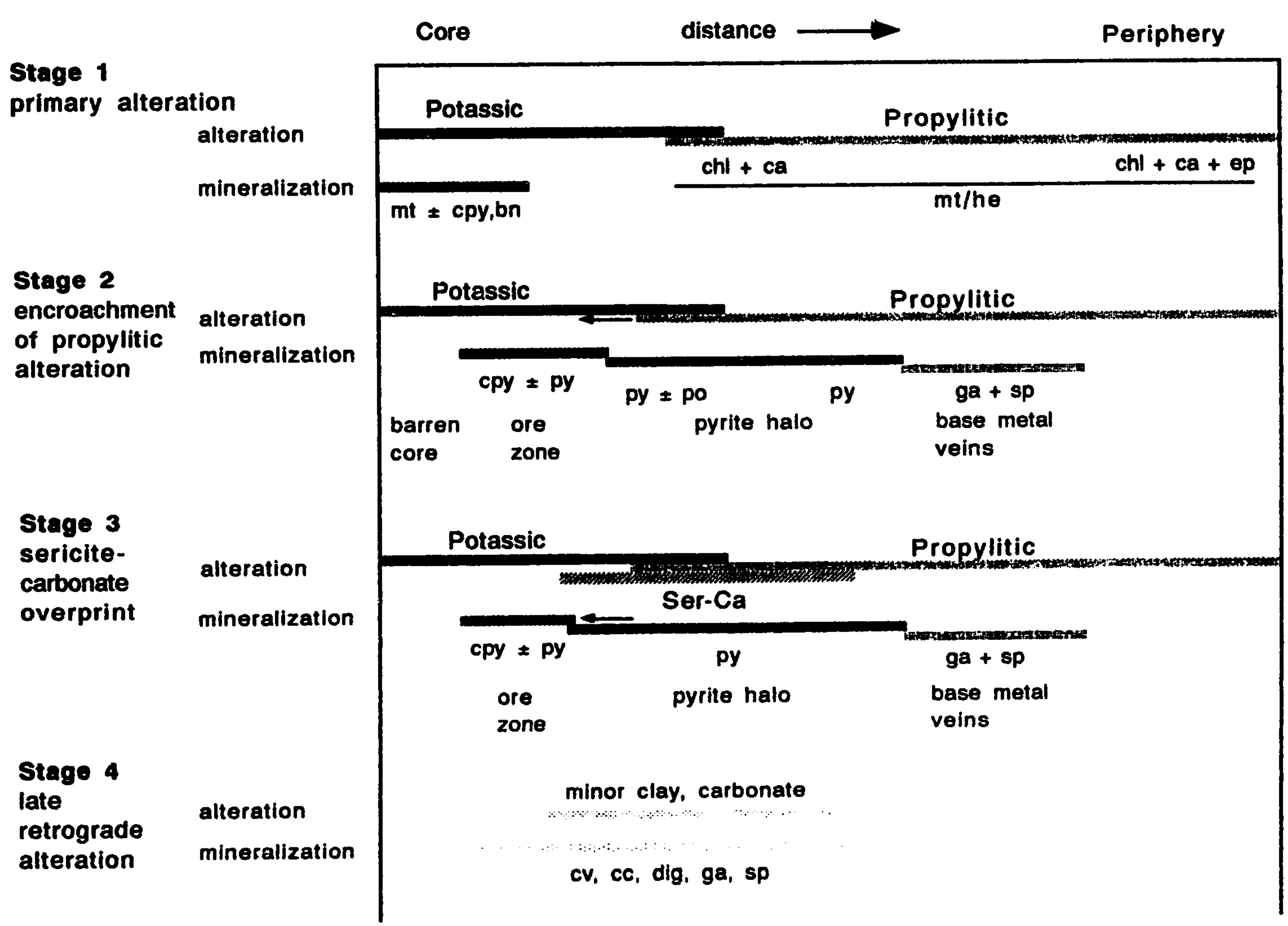

Figure 23. Lateral and temporal alteration and mineralization zonation development at Granisle.

Later retrograde alteration involving carbonate and clay was minor but Carson and Jambor (1974) reported minor galena and sphalerite in veins cutting the ore zone and minor marcasite. Covellite and bornite alteration of chalcopyrite are also present. Late gypsum fracture fillings are common in the potassic zone, especially on the northeast side of the deposit.

Bell

i) Exploration history

The first exploration work on the Bell property was the 1913 discovery of sulfide veins on the west shore of the Newman Peninsula (Carter, 1982). In 1927, Charles Newman drove three adits into the shore at the site of the Pb-Zn veins (Ogryzlo, 1975). Apart from renewed exploration activity during World War II, the next work on the property was in 1962 when Noranda Minerals Inc. undertook soil sampling, geophysical surveys, and diamond drilling, defining ore reserves of 46 million tons grading >0.30% Cu (Ogryzlo, 1975). The Bell mine went into production in 1972 with reserves of 116 million tons grading 0.48% Cu and 0.35 ppm Au(Carson *et al.*, 1976) and despite a closure in the early 1980's, has been in production ever since. Although slated for closure in the summer of 1992, remaining geological reserves are 120 million tons grading 0.43% Cu (Dittrick, 1991, pers com). Although the cylindrical mineralized zone continues beyond the depth of drilling (>615 m) (Dittrick, 1991, pers com), total proven and probable reserves to 154 m elevation (approximately 615 m depth) with a 0.20% Cu cutoff grade are 288 million tons grading 0.42% Cu (Ogryzlo, 1987). Contained metals are 1.21 million tons of Cu and 2.9 million ounces Au, making Bell as large as a large volcanogenic massive sulfide deposit of the Canadian Shield (Ogryzlo, 1987).

ii) Hydrothermal alteration and sulfide zonation

Hydrothermal alteration at Bell is more complex than at the other Babine Lake deposits. The alteration patterns are shown in Figures 24 and 25.

a) Potassic alteration

Potassically altered BFP is typically dark grey to medium grey. Although virtually all of the potassic zone in the Bell mine is affected to some degree by a sericitic overprint, some samples particularly from the dark grey core zone near the pit floor in the southeast corner of the pit feature the least overprinting. Samples from this zone are typically intensely shattered by several sets of high angle fractures and closely spaced (3 to 10 cm) subhorizontal fractures, as shown in Plate 9.

Near the centre of the alteration zone, well crystallized hydrothermal biotite replaces hornblende phenocrysts and alters the groundmass. In some cases minor replacement of biotite phenocrysts along the edges by secondary biotite is also present. Minor chlorite is

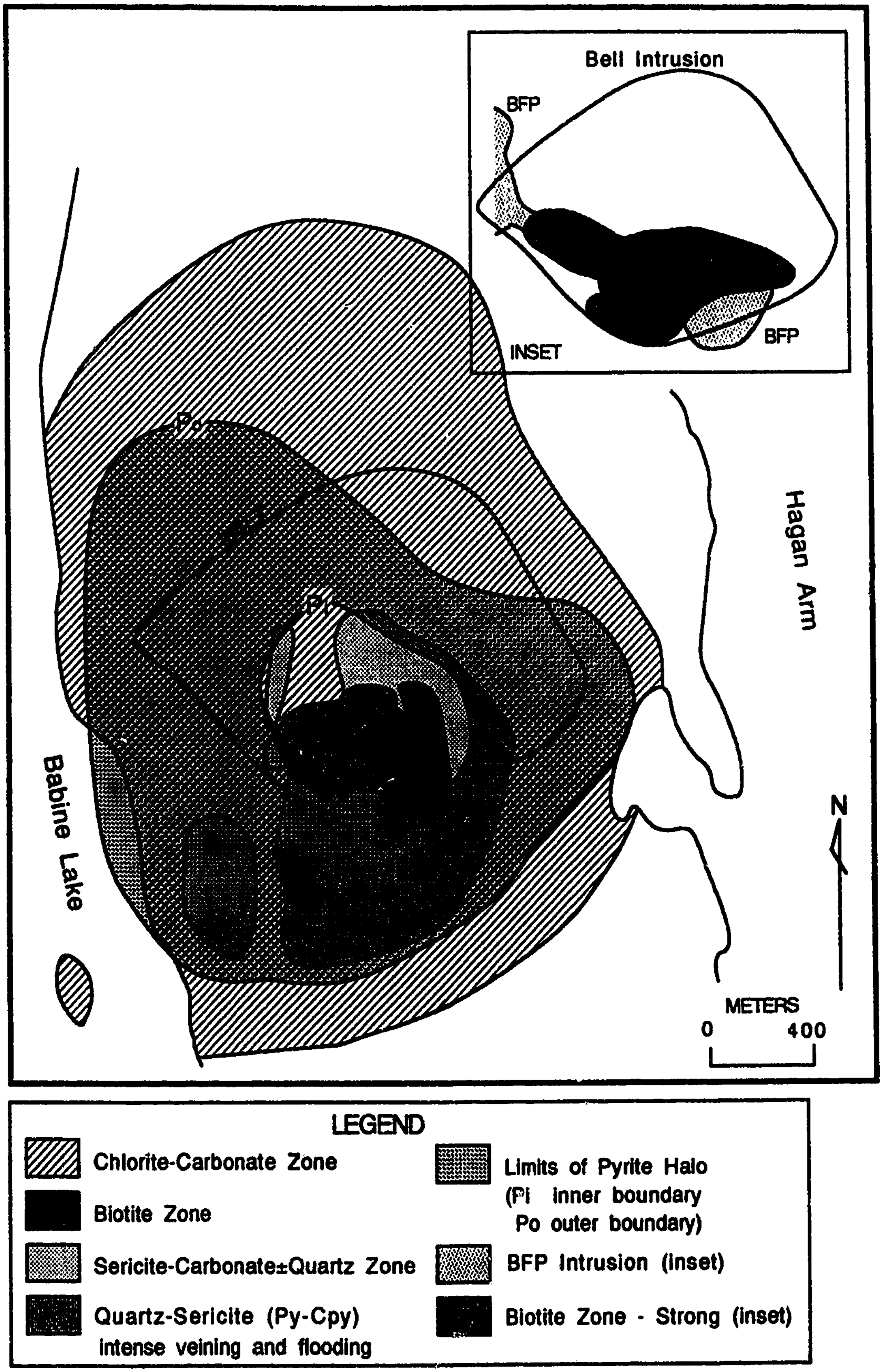

Figure 24. Alteration zonation of the Bell deposit and the Newman Peninsula (modified from Carson *et al.*, 1976).

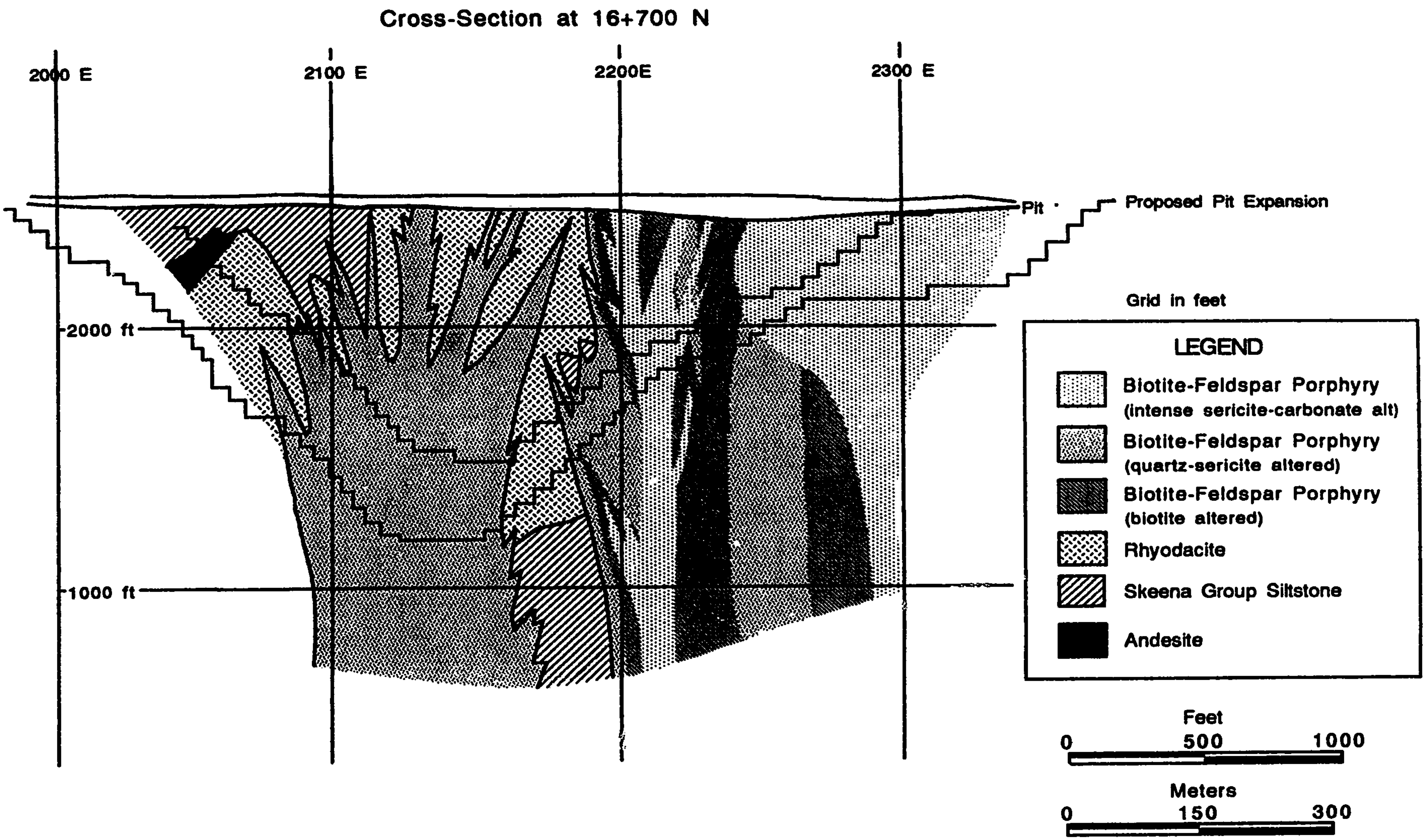

Figure 25. Cross-section of the Bell pit showing rock types and hydrothermal alteration (modified from Noranda Minerals Inc. unpublished map).

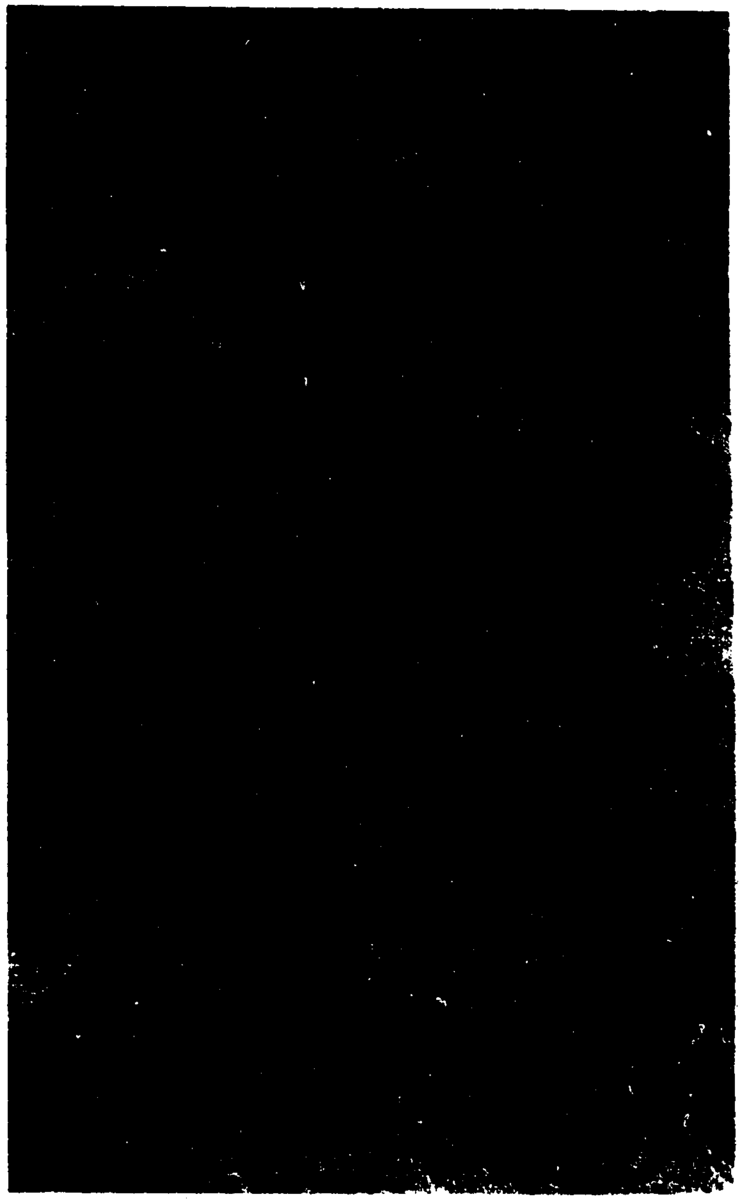

Plate 9. The intensely shattered "barren core" zone of the Bell deposit. Several sets of high angle fractures and closely spaced subhorizontal fractures result in few rocks greater than 10 cm.

locally present, likely of retrograde origin. Potassically altered rocks also feature silicification or recrystallization of the groundmass, with groundmasses typically consisting of medium-grained quartz and plagioclase and frequent but often discontinuous quartz microveinlets. Potassium feldspar is uncommon but is sometimes present in quartz veinlets or their margins. Anhydrite has been reported to be present as disseminations and along grain boundaries in quartz veinlets (Carson *et al.*, 1976) but was only rarely observed.

Magnetite introduced with the biotite and chalcopyrite along with the igneous magnetite is stable in the early stages of potassic alteration. However, during later stages of potassic alteration magnetite is altered to hematite. In addition, some hematite is precipitated as hypidiomorphic blades in mafic pseudomorphs and in the groundmass.

Sulfides occur both as disseminations and fracture coatings. Chalcopyrite typically is deposited early with magnetite and/or hematite, often in mafic pseudomorphs as shown in Plate 10. Minor bornite is frequently intergrown with the chalcopyrite.

In later stages of potassic alteration, thin quartz + sericite + chalcopyrite + hematite ± pyrite ± bornite veinlets are present. In addition, pyrite disseminations become much more abundant. This alteration accompanies increasing sericitization of the groundmass and feldspars in a gradation to sericite-carbonate alteration. Later stage chalcopyrite tends to be associated with pyrite and often minor bornite as disseminations but more importantly, in fracture coatings. Bornite invariably is associated with the chalcopyrite and both are at least in part paragenetically later than some of the pyrite since they often overgrow pyrite or cut across it in fractures. Chalcopyrite associated with the quartz veins also appears paragenetically late, cutting across the veins and optically continuous quartz grains and filling fractures. Other sulfide phases include molybdenite, as small disseminations and covellite, as a minor alteration product or a coprecipitate with chalcopyrite.

Sericitization of plagioclase phenocrysts commonly occurs along fractures; however, more intense sericitization and complete kaolinization of feldspars often accompanies thin, late stage carbonate veins. This is obviously a late stage alteration since kaolinite is present and is likely due to an influx of dilute fluids. The rock is usually intensely sericitized along late stage calcite - pyrite ± quartz veinlets. Thin gypsum veinlets and blebs are often present, replacing original anhydrite or filling late fractures and cavities.

Potassically altered Skeena siltstone is typically a greenish-black colour and weathers a tan-green colour and often has a soapy feel. It is moderately fractured and contains pyrite and chalcopyrite fracture coatings. Locally gypsum fracture coatings are also present. Hydrothermal biotite ranges from 15 to 30% and consists of fine- to medium-grained disseminations, masses of fine grains, and veinlets. Other minor alteration phases are carbonate and chlorite.

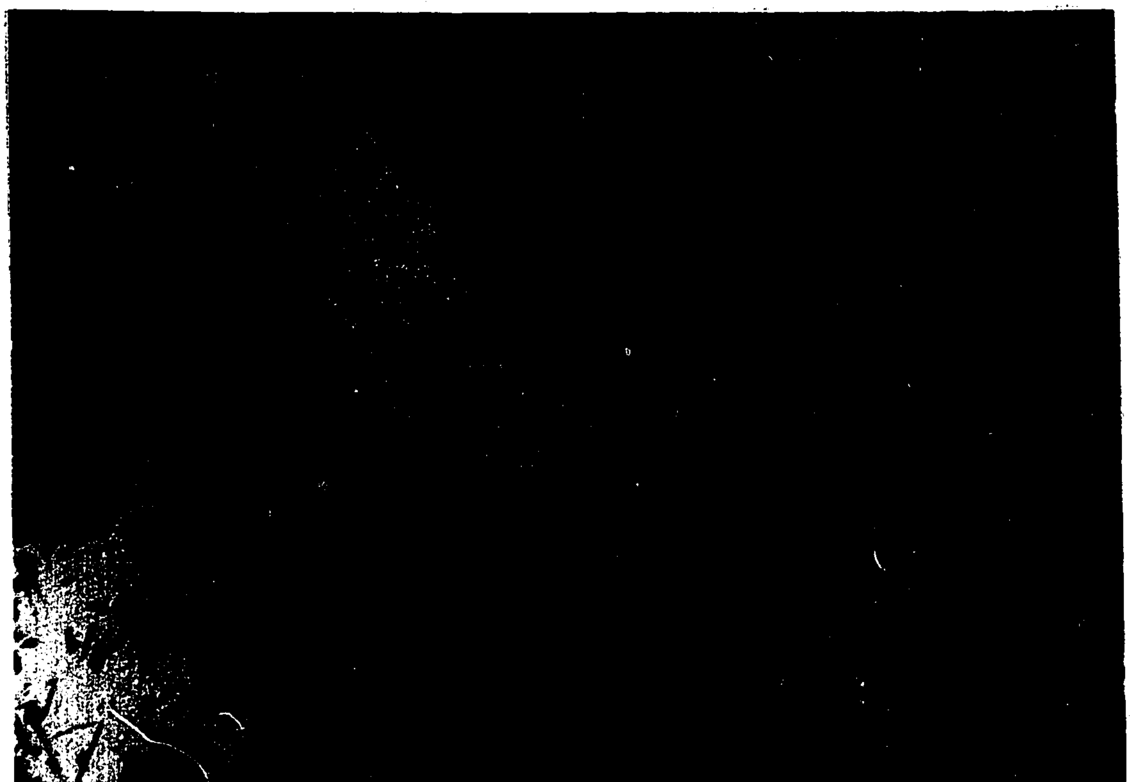

Plate 10. Chalcopyrite replacement of hornblende phenocryst (400 μm) in potassic altered BFP from Granisle. The minor grey mineral is rutile. Plane polarized light.

Plate 11. High salinity fluid inclusions in a quartz stockwork vein from the phyllic zone of Bell. Multiple daughter salts are common, including the reddish brown hematite which gives the quartz veins their pinkish colour. Plane polarized light. The large inclusion with multiple daughter salts in the right centre is 20 μm long.

APPENDIX II. CLOSURE TEMPERATURE CALCULATIONS

Closure temperature, the temperature at which significant diffusional isotopic exchange ceases is determined by the following relation (from Dodson, 1973; cited by Giletti, 1986):

$$Tc = \frac{E/R}{\ln[(-ARTc^2(Do/a^2))/(E(dT/dt))]}$$

where:

Tc = closure temperature (K).

E = activation energy of oxygen diffusion in that mineral (cal $g^{-1}atom^{-1}$).

A = diffusional anisotropy parameter (= 8.7 for a slab, 27 for a cylinder (Graham *et al.*, 1987), and 55 for a sphere).

R = gas constant = 1.98 cal mol^{-1} deg^{-1}.

Do = pre-exponential factor in oxygen diffusion Arrhenius relation.

a = size of mineral grain in cm (radius of sphere or cylinder; half-thickness of slab).

dT/dt = cooling rate (°C s^{-1}).

The equation must be solved iteratively.

Example:

Biotite

Parameter	Oxygen (1)	Hydrogen (2)
E	36 000 cal g^{-1} $atom^{-1}$	27 800 cal g^{-1} $atom^{-1}$
A	8.7	27
Do	1.2×10^{-5}	3.4×10^{-7}
a	0.05 cm	0.05 cm
dT/dt (3)	3.08×10^{-10}°C s^{-1}	3.08×10^{-10}°C s^{-1}
T(K)	814	706
T(°C)	541°C	433°C

(1) from phlogopite data from Giletti and Anderson (1975).
(2) from Suzuoki and Epstein (1976).
(3) based on a life of the geothermal system of 50 000 years during cooling from 800°C to 200°C.

These closure temperatures are based on diffusional exchange. Surface reaction-controlled systems may exchange to lower temperatures. The accuracy of these determinations depends on the accuracy of the diffusional parameters and in some cases, on the validity of extrapolation of the data to lower temperatures (for minerals with rapid diffusion rates and small grainsizes).