THE CONTACT METASOMATIC MAGNETITE DEPOSITS
OF SOUTHWESTERN BRITISH COLUMBIA

by

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M.Sc., McGill University, 1961

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Geology

We accept this thesis as conforming to
the required standard

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FACULTY OF GRADUATE STUDIES

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ABSTRACT

Ore zones, skarn, host rocks, and associated intrusions of 12 magnetite deposits were studied in both field and laboratory to determine their mineralogical and geochemical characteristics, origin of the iron, and factors controlling emplacement of iron-bearing minerals. This study seeks a better understanding of the origin and mode of occurrence of contact metasomatic magnetite deposits which in turn may provide better guides to their exploration and evaluation.

Local folds and faults are important factors in the emplacement of magnetite in volcanic rocks and limestone of the Vancouver group. Adjacent stocks are of intermediate composition. Post-ore leucodiorite dykes are common in many orebodies.

The author proposes that the process by which skarn is formed be called skarnification i.e. the replacement by, conversion into, or introduction of skarn. The term would include all processes by which skarn may be formed such as contact metamorphism, contact metasomatism, or regional metamorphism. Skarn in the coastal British Columbia region is composed mainly of garnet (andradite-grossularite), pyroxene (diopside-hedenbergite), epidote, and magnetite. Conformity to Gibbs Phase Rule and the non-appearance of incompatible phases is strong evidence that equilibrium was attained during skarnification. Magnetite is the major metallic mineral, but chalcopyrite, pyrite, pyrrhotite, and arsenopyrite are locally abundant.

The temperature of intrusion is estimated to be in the range 800-900°C and stability relations of coexisting minerals indicate a temperature of 700-

550°C during skarnification. The pyrite-pyrrhotite geothermometer applied to eight specimens shows that ore deposition took place within the temperature range 400-550°C. The composition of arsenopyrite coexisting with pyrite and pyrrhotite in one orebody indicates a confining pressure of 2600 ± 1000 bars during ore formation.

The immediate source of iron in these deposits is believed to be nearby intrusions. The ultimate source, however, is very probably underlying volcanic rocks which have been assimilated, in part, by an advancing pluton. Iron is considered to have been derived from plutons adjacent to the orebodies and to have been carried to the sites of deposition as aqueous supercritical solutions of iron chloride. Magnetite was precipitated from the ore-forming fluid by an increase in pH brought about by reaction with limestone.

Changes in the chemical and physical nature of the ore-forming fluid during ore deposition are discussed in terms of temperature, density, pH, partial pressures of oxygen and sulphur, and composition. Hydrothermal processes operative in formation of the deposits were solvate opposition, metasomatism, and cavity filling.

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Field of Study:

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R.M. Thompson

Abstract

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CHAPTER I - INTRODUCTION

Many contact metasomatic magnetite deposits occur on islands that fringe the coast of British Columbia (Fig. 1). Although most of these deposits have been known for several decades and some have been sporadically explored, it was not until 1952, with a rising Japanese demand, that iron ore production became an established industry in British Columbia. The first deposit mined was at Iron Hill, near Campbell River on Vancouver Island. Total production from this deposit, from 1952 to 1957, amounted to 1,998,526 short tons of iron ore concentrate. Since 1957 other mines have opened and production has risen steadily (Fig. 2). In 1962 total iron concentrate production in B.C. reached 1,744,431 short tons, accounting for about 8% of the total value of mineral production in the Province.

This thesis is the report of a petrographical, mineralogical, and geochemical study of contact metasomatic magnetite deposits in western British Columbia. Ore zones, skarn rocks, and associated intrusions were studied in both field and laboratory to determine their mineralogical and geochemical characteristics, origin of the iron, and factors controlling emplacement of iron-bearing minerals.

FIGURE 1

LOCATIONS OF CONTACT
METASOMATIC MAGNETITE DEPOSITS,
SOUTHWESTERN BRITISH COLUMBIA

- 1 Argonaut Company Ltd.
- 2 Texada Mines Ltd.
- 3 Empire Development Co. Ltd.
- 4 Brynnor Mines Ltd.
- 5 Nimpkish Iron Mines Ltd.
- 6 Zeballos Iron Mines Ltd.
- 7 Jedway Iron Ore Ltd.
- 8 Coast Copper Co.
- 9 Wesfrob Mines Ltd.

FIGURE 2

Annual shipments of magnetite concentrate, British Columbia. (From BC Min. of Mines Ann. Rep. 1952-62)

It is expected that this study will lead to accurate descriptions and a better understanding of the origin and mode of occurrence of contact metasomatic magnetite deposits and provide better guides to their exploration and evaluation.

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Summary of previous work

The literature of British Columbia magnetite deposits is extensive, dating back to the last quarter of the 19th Century. Most of the early articles are limited to descriptions of natural exposures and lack detailed mineralogical, petrographical, and structural data.

The author will review those works which, in his opinion, have made significant contributions to an under-

standing of the character and origin of British Columbia contact metasomatic magnetite deposits.

The earliest of these works was by Kimball (1897) in which he recognized the deposits as "hydro-chemical" replacements but distinguished two separate types. In the first type, a morphological replacement of limestone by double decomposition between ferrous salts generated from ferrous silicates and calcium carbonate, Kimball considered the ferrous material to have been derived from an extraneous source "through simple permutations and reactions" (Kimball, 1897, p.13). The second type, a partial replacement of ferrous silicates in weathered basic igneous rocks, is brought about through initial decomposition of ferrous and ferric silicates (chiefly the former). These are then considered to form soluble alkaline carbonates along with ferrous salts. This leads, in turn, to double decomposition and finally to isolation of insoluble residues, presumably magnetite. It was suggested that the "union of magnetite molecules is doubtless promoted also by polarity" although Kimball did not elaborate on this intriguing point (Ibid, p.15). Although it suggests a number of interesting ideas and undoubtedly portrays original thoughts on the origin of contact-type magnetite deposits, Kimball's work suffers from an antiquity of language which is difficult to read and understand. One can, however,

recognize in his first type of deposit an early attempt to describe a process of replacement by hydrothermal solutions emanating from an outside source. The iron is not considered to have been carried as magnetite but, instead, as a soluble salt which reacted with limestone to form magnetite. Kimball's second type of deposit, because it replaces only mafic minerals in the country rock, would result in a low grade disseminated magnetite deposit. This process is considered to be of relatively minor importance.

In a discussion of the genesis of several small magnetite bodies found in southern Vancouver Island, Clapp (1913) concluded that they owed their origin to intrusion of adjacent plutons. Apparently, he was impressed by the broken appearance of rock associated with certain deposits and inferred from this that iron was emplaced as magnetite magma, in analogy to an intrusive breccia such as that produced by a silicate magma. This theory may be substantiated in part by the recent discovery of what appears to be a high-grade (65% Fe) magnetite "flow" in northern Chile (Park, 1961; Ruiz and Ericksen, 1962, pp98-99). The porous texture of the "flow" indicates it was at one time probably highly charged with gaseous materials and these volatiles may have lowered the melting temperature of the mixture. Perhaps Clapp's "magma" was similar to this.

Eleven years after Clapp's memoir, Swanson (1924) discussed at some length the origin of contact metasomatic magnetite deposits, particularly those occurring on Texada Island. He distinguished two effects of intrusion of the Gillies stock; namely, contact metamorphism (mainly recrystallization) and metasomatism (resulting in an extensive replacement of country rock). Concerning the magnetite deposits, three features were considered to be distinctive of these bodies: (1) irregular distribution (2) marked chemical changes which were apparently involved in ore formation and (3) replacement of the parent intrusive rock as well as limestone and volcanic rock. An excellent description of the mineralogy of the host rocks, intrusions, and skarn then followed. Swanson decided that "...replacement was quantitatively the dominant process, and that all the minerals of the deposit are at least partly metasomatic" (Swanson, 1924, p. 131A).

The paper concluded with a summary of factors Swanson believed controlled the character and position of the deposits. His summary, brief, but informative is quoted below:

The deposits were formed by magmatic solutions in which the materials were concentrated by the crystallization of the intrusive. The assimilation of limestone by the magma may have been a factor of great significance, as ferrous and ferric oxides were thereby thrown out of the pyroxene molecule and

made available for segregation as magnetite. The release of the solutions from the intrusive was effected by fracturing caused by cooling of the mass and, as a consequence, the distribution of the deposits is mainly controlled by the structural conditions at the time of replacement, which favoured the movement of solutions along the contact zone. From a consideration of the physical permeability and the chemical composition of the rocks into which the solutions were led, it is concluded that the porphyrite was unfavourable both physically and chemically, that the intrusive was favourable physically but not chemically, and that the limestone was favourable chemically but not physically. The chemical factor was probably in control, as the constant removal of material, facilitated by the addition of calcite to the solutions, was as necessary for the formation of extensive replacements as the access of new material. The heterogeneous parts of the intrusive which included blocks of limestone, and the tongues of limestone projecting into the stock, were especially suited to replacement, both physically and chemically, and contain the largest deposits. Finally, the oxidizing effect of the carbon dioxide was probably effective in the formation of the relatively pure deposits of magnetite. (Swanson, 1924, p. 144A).

Swanson's conclusions were deduced entirely from field and petrographic observations. The present manuscript, however, strongly supports many of his deductions concerning nature of the ore fluids, exchange of material, and physical chemistry of ore-forming environments.

Uglov (1926) was one of the first to note and emphasize the significance of the areal association of magnetite deposits with the Vancouver group. He considered

the magnetite occurrences to be of the "contact metamorphic" type and recognized three sub-types: (1) magnetite deposits in limestone (2) magnetite deposits in volcanic rocks and (3) copper-iron deposits in limestone, volcanics, and diorite. The paragenetic sequence—silicates, magnetite, bornite, and chalcopyrite—suggested by earlier writers and supported in part by later ones, was also given in Uglow's paper. Concerning genesis, Uglow rejected Clapp's hypothesis of a magnetite magma in favour of replacement by "tenuous solutions" and gave several brief descriptions supporting this theory.

In a short description of the Iron Hill deposit which lies in the axial zone of a tight syncline composed of both greenstone and limestone, Black (1952) suggested that during deformation the greenstone became closely fractured whereas the limestone merely flowed and recrystallized. Because the fractured greenstone was more permeable to mineralizing fluids than limestone and offered larger surface areas for reaction, it was preferentially replaced by ore minerals. Black observed, however, that such replacement took place near limestone and suggested that the presence of carbonate was necessary.

In 1952 and again in 1956, Bacon briefly reviewed the nature of contact metasomatic iron deposits in British

Columbia. In these reports, he noted the close association of magnetite with limestone and emphasized that "limestone assumes the utmost significance in any search for new magnetite deposits, and belts of this rock.....should be investigated thoroughly" (1952,p.A218). He further warned that surface exposures of this type of magnetite deposit are misleading and that tonnage estimates should be based on closely-spaced diamond drill holes.

To distinguish deposits of contact metasomatic origin from taconite deposits and magnetite-bearing pyroxenites, all of which occur in B.C., Bacon listed the main characteristics of these three types of deposits (1952,p.125). Contact metasomatic type are distinguished by:

1. Magnetite generally associated with skarn in which brown garnet is commonly the dominant mineral.
2. Occurrence along or close to contacts between folded metamorphosed rocks and Coast intrusive rocks.
3. Proximity to limestone, or in limestone.
4. Highly irregular shape and generally abrupt terminations.
5. Relatively high iron content, ranging from 35% to nearly 60%.

Sutherland-Brown (1962) summarized the observations made during the decade 1952-62 by officers of the B.C. Department of Mines and Petroleum Resources. His summary

stressed the constant association of these deposits with Upper Triassic limestone but pointed out that a majority of deposits are actually found in volcanic rocks either above or below a limestone unit (Quatsino formation). It is perhaps significant to note that Sutherland-Brown considered both iron and calcium to have been added to the volcanics during ore formation. Heretofore most writers had tacitly assumed that calcium had moved out of, rather than in to, the area in which ore was being formed.

Sutherland-Brown considered the relatively common occurrence of Quatsino limestone to be the main reason why magnetite appears to be preferentially associated with it. A thinner, less widely distributed Upper Paleozoic limestone unit is not associated with any known economic magnetite deposit in southwestern British Columbia. A map was included to show the distribution of Upper Triassic (Quatsino) limestone in western British Columbia. In his short summary, Sutherland-Brown was the first to note the abundance of post-ore feldspar porphyry dykes in mine areas and their scarcity elsewhere in similar country rocks, but the significance of this association was unstated.

Physiography and climate

The deposits considered in this report are on islands that fringe the coast of British Columbia. The Queen Charlotte and Vancouver Islands are mountainous, with bordering lowlands on their northeast sides. In the Queen Charlottes, the maximum elevation is about 3000 feet but on Vancouver Island several mountains rise above 6000 feet and one, Golden Hinde, exceeds 7200 feet in elevation. Texada Island is low-lying, the highest peak reaching only 2892 feet.

The mean annual range in temperature of coastal stations is 16°F in the south and 23°F in the northern part of the area. Winter temperatures on the coast average 39°F and summers about 57°F. The frost-free period varies from 150 to 250 days (Chapman, 1952, pp. 10-12).

Heavy precipitation is characteristic of the coastal area although only about 1-6% is snow. Locally, unusually high amounts of rain have been recorded; for instance, Henderson Lake, less than 20 miles east of Brynnor Mines Ltd. on Vancouver Island (Fig. 1) has the highest recorded precipitation in one year (251.3 inches) on the continent. Relative humidity in the coastal area is generally high throughout the year, averaging 70-85% (Chapman, 1952, pp. 10-12).

Regional geology

The geology of the coastal region of British Columbia is not yet fully mapped. Mapping in the area is difficult because of heavy forest cover, precipitous slopes, and generally poor weather during the field season. Some areas were mapped in detail 20 years ago whereas others are only now being explored on a reconnaissance scale.

Stratigraphy

The present state of knowledge of the stratigraphy on Vancouver Island, Texada Island, and Moresby Island (southern Queen Charlottes) is shown in Table 1. Unconformities are shown but intrusive contacts or periods of intrusion are not indicated. These are discussed in the section "Intrusive rocks".

Attention is drawn to that group of Lower Mesozoic rocks collectively referred to as the Vancouver group or its correlative. It is with these rocks that the known economic or sub-economic magnetite bodies are associated and for this reason the group is briefly described below. Any one of the three units which comprise the Vancouver group, together with an associated intrusion, may be host to iron ore deposits.

ERA	PERIOD	AGE	VANCOUVER ISLAND	TEXADA ISLAND	MORESBY ISLAND (QUEEN CHARLOTTE ISLANDS)			
CENOZOIC	TERTIARY	Plioc.	Metchosen volcanics		Masset formation			
		Mioc.						
		Olig.						
		Eoc.						
		Paleoc.						
MESOZOIC	CRETACEOUS	Upper	Nanaimo series		Queen Charlotte group			
		Middle	Unnamed sequence of cong., sst., coal		Longarm formation			
		Lower						
	JURASSIC	Upper	} Bonanza group } Quatsino form'n } Vancouver group Karmutsen group }	Sutton formation Marble Bay formation Texada group	} Yakoum formation			
		Middle						
		Lower						
	TRIASSIC	Upper			} Older volcanics			
		Middle						
		Lower						
	PALEOZOIC	PERM.				Sicker series	} Anderson Bay formation	?
		PENN.						

TABLE I
Bedrock stratigraphy of Vancouver, Texada and Moresby Islands, British Columbia

The Karmutsen group of volcanic rocks is the oldest member of the Vancouver group. The exposed thickness exceeds 10,000 feet although its base has not been observed (Mathews and McCammon, 1957, p.35). Volcanic flows, predominantly basalt and andesite, are characteristic of this group. Amygdaloidal textures and pillow lavas are common in some areas but in others the rock is extremely massive. Alteration is widespread; chlorite commonly replaces ferromagnesian minerals and calcite, sericite, or epidote replaces feldspar. Faint pillow forms or fragmental textures are in many places the only remaining evidence of a volcanic rock.

Fragmental volcanic rocks are irregularly distributed throughout the Karmutsen group. Volcanic breccias, both explosive and flow types, are more common than tuff.

Associated with the lavas and breccias of the Karmutsen group are irregular bodies, dykes, and sills of intrusive rock very similar in composition and appearance to the flows. The relation is notable in the vicinity of Benson Lake, Vancouver Island. Here, both the intrusive bodies and the flows are of similar texture, and in at least one locality a dyke was seen to grade upward into a flow or sill. This would indicate that perhaps many dykes served as feeders to overlying flows which were in turn intruded by later bodies.

The upper part of the Karmutsen group consists of black argillite and crystalline limestone interbedded with lavas. Fossils collected from the argillite are of Upper Triassic age (Hoadley, 1953, p.17). Mathews found fossils indicative of the Karnian stage of the Upper Triassic in the Texada porphyrites (Mathews and McCammon, 1957, p.36).

The Quatsino formation that conformably overlies the Karmutsen group consists almost entirely of limestone as much as 4000 feet thick, forming a conspicuous marker in the Vancouver group. The lower part commonly is intercalated with pyroclastic material or flows but these diminish upward into relatively pure white or grey crystalline limestone. Toward the top the limestone becomes darker and is interbedded with argillite of the overlying Bonanza group. Bedding, except where it is represented by colour banding, is generally poor. Cross-bedding, three or four inches thick, in pure Quatsino limestone, was observed by the writer in a road-cut near the Empire Development Company mine.

Fossils are generally poorly preserved in the Quatsino formation presumably because of recrystallization which is common in this unit. Hoadley placed the age of the upper part of the Quatsino limestone as "probably Noric" (1953, p.21), largely because of similar ages for the under-

lying Karmutsen and overlying Bonanza groups. Limestones occupying a similar stratigraphic position on Texada Island (Mathews and McCammon, 1957, p.36) and Moresby Island (Sutherland-Brown and Jeffery, 1960, p.2) have been identified as being of Late Triassic or Early Jurassic age.

The Quatsino formation grades upward into the Bonanza group and the position chosen as the top of the Quatsino is arbitrary. Gunning (1933, p.34) and Hoadley (1935, p.20) chose slightly different horizons to accommodate mapping in different localities.

The Bonanza group may be divided into two parts, a lower sedimentary division and an upper volcanic division. In some areas the distinction is not clear. The top of the Bonanza group is not known but Hoadley (1953, p.22) decided that in the Bonanza Lake area the group is "certainly more than 3200 feet thick".

The sedimentary division consists of thin-bedded sediments and intercalated lava flows. The former consist of carbonaceous limestone, calcareous argillite, argillite, and quartzite.

The volcanic division includes andesitic and basaltic flows, volcanic breccias, and banded tuffs. Intrusive bodies

of irregular size and shape that occur throughout probably served as feeders for overlying flows.

The sedimentary unit of the Bonanza group has been shown to be of Noric age (mid-Upper Triassic) and the volcanic assemblage is thought to be "probably Jurassic" (Hoadley, 1953, p.29).

On Moresby Island in the Queen Charlottes, the Maude formation unconformably overlies the Kunga formation which is predominantly limestone (correlative of the Quatsino formation). The Maude formation probably reaches a maximum thickness of 3000 feet. It is composed largely of shales and siltstones but the basal portion contains greywacke and conglomerate. The age is thought to be Early Jurassic which would correlate with at least part of the Bonanza group.

Structure

The earliest recorded tectonic event to affect the Vancouver-Queen Charlotte-Texada Islands area was the "Cassiar orogeny" which approximately marked the close of the Paleozoic era. This orogeny was characterized by strong uplift, north-west-trending folds, and ultramafic intrusions. No ultramafic bodies, however, have been found in the Insular Region of British Columbia (White, 1959, p.72).

Much has been written about the succeeding tectonic activity, the Coast Range orogeny. White (1959,p.78) described it as "the most complex tectonic event in the history of the Canadian Cordillera, involving regional folding and faulting, strong metamorphism, and a prolonged succession of granitic intrusions". Indications of the impending cataclysm, which probably culminated in the Lower Cretaceous, first appear in the earliest Jurassic and activity persisted, locally at least, into Upper Cretaceous.

Mesozoic folds west of the Coast Range mountains closely follow the direction of the mainland coast. In general, the Coast Range orogeny has produced only mild deformation among the coastal islands. Open folding or broad arches are most common with bedding dipping about 45° or less. Locally, however, it may be overturned.

Major northwest-trending faults are known in southern Vancouver Island and Moresby Island. The latter are described as wrench faults in which the Pacific side moved north (Sutherland-Brown and Jeffery, 1960, p.4).

Coast Range orogeny was closely followed by the Rocky Mountain orogeny of Late Cretaceous-Early Tertiary time. In general, this tectonic activity did not affect the coastal

region although officers of the B.C. Department of Mines have briefly mentioned a Late Cretaceous period of folding on Moresby Island (Sutherland-Brown and Jeffery, 1960, p.4).

One or more epochs of folding and normal faulting which took place during Tertiary and Recent times is collectively termed the "Puget orogeny". Several intrusions on Vancouver Island, in the Garibaldi area of the Vancouver area, and in adjacent Washington state, are known to cut Upper Cretaceous rocks. Folding developed along east or south-east trends and fault movements occurred along old as well as new faults.

Intrusive rocks

Coast Range orogeny was marked by wholesale intrusion of plutonic rocks, chiefly granodiorite and quartz diorite. Once considered to be a single body and therefore termed the "Coast Range batholith", these plutonic rocks now appear to be a product of multiple intrusions and are now known as the "Coast Range complex". Most intrusions are believed to be of Late Jurassic or Cretaceous age. Some, such as the San Cristoval batholith of western Moresby Island, are apparently syntectonic and show well-defined planar structures. Other intrusions are clearly post-tectonic. Most contact metaso-

matic magnetite deposits are associated with the latter type.

Basic sills and dykes are common in the vicinity of iron mines and are generally pre-ore. A later, post-ore, set of leuco-diorite dykes is especially prominent in areas of magnetite mineralization.

CHAPTER II - MINERALOGY AND GEOCHEMISTRY

Intrusions

Intrusions associated with magnetite deposits of Vancouver Island, Texada Island, and Moresby Island have been described by Swanson (1924), Gunning (1930), Stevenson (1950), Hoadley (1953), Sutherland-Brown and Jeffery (1960), Jeffery (1960) and Eastwood (1962) during the course of regional mapping in these areas. Quartz monzonite, granodiorite, and quartz diorite are the most common rock types.

The writer collected and studied specimens of typical granitic rocks, free of inclusions or other contaminants, that are associated with iron ores of the seven mines considered in this report. The specimens were examined petrographically and chemical, normative, and modal analyses made.

The intrusions, in general, are homophanous and have sharp cross-cutting relations with country rocks. According to the classification proposed by Peterson (1961), five intrusions are quartz monzonite, one is granodiorite, and one is gabbro. These modes are shown in Figure 3 plotted

FIGURE 3

Modes of intrusions adjacent to magnetite deposits, British Columbia. Numbers refer to analyses, Table 12. Diagram after Johannsen, 1931.

on a Johannsen diagram. Plagioclase, potash feldspar (some perthitic), quartz, and green hornblende are the major constituents of the intermediate varieties; plagioclase and augite comprise the gabbro. Specimens of granitic rock within a few feet of their external contacts contain augite rather than hornblende. Only one intrusion, the earliest phase of the Gillies stock, contains augite throughout. Later phases of this stock contain hornblende.

Chemical analyses of specimens collected from each of the seven intrusions studied are given in Table 12 and are shown graphically in Figure 4. These rocks are all richer in the group ($\text{Na}_2\text{O} + \text{K}_2\text{O} + \text{Al}_2\text{O}_3 + \text{Fe}_2\text{O}_3$) than world averages of rocks ranging from granite to gabbro. The analyses fall into two divisions: one of intermediate composition (encircled by a dotted line, Fig. 4) and one of more basic composition (No. 3, gabbro from Coast Copper stock). The arithmetical average of the analyses in the first division is represented by point 'X' in Figure 4.

Some plutons are partly rimmed by a more basic phase, possibly as a result of contamination by limestone and/or volcanic country rocks. Contact effects in the country rock are most striking where metasomatic deposits have formed; endomorphism of the pluton is not as noticeable. Because

FIGURE 4

Diagram showing compositions of intrusions adjacent to contact metasomatic magnetite deposits. Numbers refer to analyses, Table 12. Composition 'X' is arithmetic average of analyses encircled by the dotted line. World averages of the granite-gabbro series is also shown for comparison.

contact metasomatic magnetite deposits are spatially, and probably genetically, related to these plutons, it is important to study the chemical changes induced in plutons by the country rocks they invade. The largest contrasts in chemical and physical properties of rock types are found near the margins of plutons and it is in this border zone that chemical activity will be greatest. The endomorphic effects in certain plutons on Vancouver Island are discussed below. Some plutons are directly associated with magnetite deposits whereas others are not, although all are of similar composition and invade similar rocks.

Hoadley (1953) noted that quartz diorite in the Zeballos batholith has a border phase of diorite and gabbro. The Nootka batholith is granite but near its contact with older volcanic rocks changes to syenodiorite or diorite in a border zone 50 to 100 feet wide. The approximate chemical composition of Nootka granite and the border phase diorite was calculated from modes of these rocks (Hoadley, 1953, p. 31) using mineral composition and specific gravities given by Wahlstrom (1955, pp. 88, 89). The composition of Bonanza group andesite is an average analysis of three specimens of green andesite from this group (Hoadley, 1953, p. 25). From a comparison of these compositions (Table 2), a rough estimation of the endomorphic effects in the Nootka granite is possible.

Table 2

Comparison of compositions of Nootka granite, its "contaminated" border phase, and andesite country rock (from Hoadley, 1953).

	Nootka granite	Border phase (diorite)	Bonanza group andesite
SiO ₂	74.3%	49.5%	50.2%
Al ₂ O ₃	14.4	17.4	17.2
Fe ₂ O ₃	0.08	7.22	3.84
FeO	0.09	9.67	8.05
ΣFe	0.13	12.60	8.95
CaO	0.90	9.02	6.59
MgO	0.90	4.98	4.93
K ₂ O	5.18	negligible	0.87
Na ₂ O	3.87	2.25	3.43

The most notable chemical changes accompanying "contamination" of Nootka granite is a decrease in silica and potassium and an increase in total iron, calcium, and magnesium.

The Nimpkish intrusions in northern Vancouver Island all belong to the granite clan (granodiorite, quartz monzonite, granite) except for minor basic phases (Hoadley,

1953, pp. 29-30). The Ehatishat batholith is also of the granite clan and has a more basic border zone (Ibid, p. 32). Eastwood (1962, pp. 118-119) noted that granodiorite in an unnamed batholith in the Kennedy Lake area grades into diorite near its contact with limestone and tuff.

In his study of the geology of the Cowichan Lake area, Vancouver Island, Fyles (1955) found that the Saanich granodiorite pluton changes to a border zone of quartz diorite where it intrudes basaltic country rock. This border phase is usually several tens of feet and locally more than a hundred feet wide. A transitional rock, representing a highly altered phase of the basalt, is found between quartz diorite and basalt. Analyses of the four rocks granodiorite, quartz diorite, transitional rock, and basalt, show the chemical changes across the marginal zone of this pluton where it is in contact with volcanic rock. These analyses are shown in Table 3.

In the marginal zone of Saanich granodiorite there is a relative gain of iron, aluminum, magnesium, and calcium and a relative loss of silica and potassium. Those elements that show a decrease in border granitic rock apparently enrich contiguous altered basalt. This is shown by comparing the composition of the transition rock with that of the

Table 3

Analyses showing changes across the contact of Saanich granodiorite and basalt (from Fyles, 1955).

	Saanich granodt.	Border phase (qtz. diorite)	Transitional rock (altered basalt)	Amygdaloidal basalt
SiO ₂	67.5%	61.4%	50.9%	50.1%
Al ₂ O ₃	14.9	16.7	15.0	12.2
Fe ₂ O ₃	1.59	2.83	3.53	3.39
FeO	2.15	3.21	4.92	5.33
ΣFe	2.78	4.48	6.29	6.43
MgO	1.61	2.43	7.69	9.83
CaO	3.52	4.92	9.18	11.8
Na ₂ O	3.32	3.71	2.35	2.42
K ₂ O	3.28	2.15	1.76	0.88

basalt (Table 3). Both Nootka and Saanich stocks have relatively gained much iron in their border zones adjacent to volcanic rocks. Neither stock is associated with any known magnetite deposits but the Saanich border zone contains considerably more iron than the average Saanich granodiorite.

Acidic intrusions in contact with basic volcanic rock have contaminated border zones as much as 100 feet thick.

These border zones are absent around gabbro intrusions (e.g., some phases of the Zeballos batholith (Stevenson, 1950, p. 27) or parts of the Coast Copper stock). Contacts with limestone are more regular than contacts with volcanic rocks and endomorphism is less intense. Hoadley (1953, p. 33) noted that rock of the Perry Lake stock only ten feet from a contact with Quatsino limestone is almost identical in composition to the average granite of the stock. The intervening contaminated zone is darker than the unaltered granite. From Hoadley's modes of the two rock types the approximate chemical compositions of these rocks have been calculated (Table 4).

In the border phases of both the Saanich and Nootka intrusions silica shows a relative decrease whereas a relative increase of this element is indicated in the Perry Lake stock.

The chemical compositions both of interior and border phases of the Perry Lake stock and the Gillies stock are compared in Tables 4 and 5. Both stocks intrude limestone but the former has no known nearby magnetite deposits whereas the latter is associated with several commercial magnetite deposits. The iron content of the Gillies stock border zone, as is the Perry Lake marginal rock, is practically nil yet skarn only a few feet from the Gillies stock contains approximately 20% total iron and the adjacent ore zone 63%.

Table 4

Composition of Perry Lake granite and associated border phase in contact with limestone (from Hoadley, 1953).

	Perry Lake granite	Border phase
SiO ₂	66.8%	75.7%
Al ₂ O ₃	15.5	11.7
Fe ₂ O ₃	1.24	0.25
FeO	1.63	0.31
ΣFe	2.14	0.41
MgO	3.27	2.82
CaO	4.40	0.76
Na ₂ O	2.93	0.73
K ₂ O	3.22	7.55

Iron, magnesium, and sodium are relatively depleted in the border phases of both the Perry Lake and Gillies stocks. Silica, aluminum, calcium, and potash, on the other hand, do not show consistent trends in the two stocks; aluminum and calcium are relatively enriched and silica and potassium depleted in the Gillies stock border phase. In the Perry Lake border phase, these trends are reversed.

Table 5

Gillies stock quartz monzonite and associated border phase.
In contact with limestone adjacent to the Prescott orebody,
Texada Island. (Analyses by R.M. Staff and A.V. Payne,
 Geological Survey, Canada).

	Quartz monzonite	Border phase
SiO ₂	55.6%	46.3%
Al ₂ O ₃	19.4	26.5
Fe ₂ O ₃	3.6	1.1
FeO	3.3	1.1
ΣFe	5.1	1.6
MgO	3.1	2.2
CaO	7.0	11.5
Na ₂ O	3.6	2.4
K ₂ O	2.2	1.8

A detailed study of the distribution of certain elements was made in the margins of the Coast Copper stock adjacent to the Empire Development Company deposits. This stock was selected because the contact is locally well exposed, outcrops are plentiful, and the rock appears fresh in handspecimen. Five-pound chip samples were collected at

each of 22 stations at approximately 100 foot intervals along two lines. These lines began at or near the contact of Coast Copper gabbro and Bonanza group volcanic rocks and extended roughly perpendicular to this contact for several hundred feet toward the interior of the intrusion. Nine samples were collected along Traverse 'A' adjacent to the Merry Widow deposit; 13 samples were collected along Traverse 'B' which was parallel to 'A' but 1000 feet farther south and remote from any known mineralized zone. The relations are shown in Figure 5. The samples were analyzed spectrographically for the ferride elements (Mn, V, Ti, Ni, Co, Cr, and Fe) and for copper. Weight percent magnetite in each sample was also determined and the results recalculated to weight percent iron. Thus a clear distinction was made between total iron (Fe_t) and iron in magnetite (Fe_m). Results of these analyses are given in Table 6.

Values for Ti, Ni, and Cr are too inconsistent to attempt to correlate with distance from the contact. The method of least squares was used to fit a straight line to values for V, Mn, Co, and Cu. In both traverses, however, the correlation coefficient for these elements was less than 0.5. A "perfect fit" would have a correlation coefficient of 1.0. A coefficient less than 0.5 means that variations in V, Mn,

Table 6

Spectrographic analyses of gabbro in Traverses 'A' and 'B',

Coast Copper stock. (Analyst, W.F. White, Geol. Survey, Canada)

Dist. from con- tact (feet)	Sample number	% Mn	% V	% Ti*	% Ni	% Co	% Cr	% Fe _t	% Fe _m	% Cu
Traverse 'A'										
55	1	0.12	0.042	1.6	NF	0.0043	0.0049	8.4	1.6	0.0064
140	2	0.11	0.037	0.99	"	0.0035	<0.003	6.7	3.2	NF
255	3	0.11	0.086	>2.0	"	0.0058	NF	8.8	3.8	0.022
340	4	0.16	0.092	>2.0	"	0.0072	"	11.1	5.9	0.015
395	5	0.17	0.068	>2.0	"	NF	"	11.2	3.9	0.0069
450	6	0.13	0.067	>2.0	"	0.0037	"	8.3	1.0	<0.002
610	7	0.15	0.044	>2.0	"	0.0051	"	9.2	4.4	0.0050
700	8	0.19	0.030	>2.0	"	0.0046	"	10.4	3.6	0.0059
780	9	0.20	0.016	>2.0	"	NF	"	8.5	1.5	0.0058
Traverse 'B'										
15	1	0.17	0.048	1.8	NF	0.0050	<0.003	9.0	3.7	0.0050
80	2	0.15	0.043	1.7	"	NF	NF	8.8	3.1	0.0039
160	3	0.15	0.074	>2.0	"	0.0068	0.0074	10.8	7.2	0.016
250	4	0.098	0.041	1.0	"	0.0053	<0.003	6.4	3.3	NF
325	5	0.12	0.049	>2.0	"	0.0069	0.0071	9.0	2.0	0.010
390	6	0.15	0.058	1.5	"	0.0049	0.0031	8.7	3.5	0.0029
440	7	0.17	0.072	>2.0	0.013	NF	NF	10.8	5.7	0.0096
540	8	0.15	0.067	1.9	NF	0.0057	"	7.6	4.6	0.0033
600	9	0.20	0.084	>2.0	"	0.0061	"	12.7	8.3	0.013
680	10	0.11	0.050	1.4	"	0.0036	"	7.6	4.0	NF
740	11	0.20	0.063	>2.0	"	0.0039	"	12.2	4.7	0.013
820	12	0.17	0.033	>2.0	"	<0.002	"	10.2	3.2	0.0042
870	13	0.20	0.040	1.6	"	0.0065	"	8.5	1.8	0.0037

*Values for Ti are semi-quantitative only. Upper limit of method approximately 2%.

Co, and Cu cannot be fitted to a straight line relating these values to distance from the contact i.e., there is no statistical change in the concentrations of these elements away from the contact.

In fitting a straight line curve ($y = a + bx$) to values of total iron (Fe_t), however, a correlation coefficient of 0.84 was obtained in Traverse 'A' and 0.67 in Traverse 'B'. These coefficients show that values of Fe_t and distance from the contact can be fitted to a straight line. These values are shown in Figure 6a. The intercept in both traverses is 8.5% Fe_t but the slopes of the lines are small: 0.0017 in 'A' and 0.0021 in 'B'. A statistical test (Student's "t" distribution) showed that such slopes are not significantly different from zero and therefore the average value of Fe_t does not vary significantly from 8.5% at any distance from the intrusive contact in either traverse i.e. although the values can be fitted to a straight line, this line has a zero slope. The zero slope of these lines shows that gabbro of the Coast Copper stock in this area has no significant variation in iron content from the border zone (in contact with volcanic rocks) to the interior of the stock up to 800 feet from the contact. Furthermore, because there is no significant difference in iron content of the stock adjacent

FIGURE 6

Coast Copper stock, Vancouver Island. Graphs showing variations in total iron content (Fe_t)(Fig. 6a) and in the ratio of iron in magnetite (Fe_m) to total iron. For locations of traverses, see Figure 5.

to the Merry Widow deposit (Traverse 'A') as compared to a place remote from the iron deposit (Traverse 'B'), suggests that magnetite of the stock evidently is genetically unrelated to magnetite of adjacent orebodies. A line whose slope is zero is also indicated in the plot of Fe_m/Fe_t versus distance (Figure 6b). This means that the magnetite content of the stock does not change, relative to total iron, whether or not it is in rock adjacent to the Merry Widow deposit.

Magnetite constitutes between 1 and 11 weight percent of the Coast Copper gabbro in this area. It replaces augite, the only other major iron-bearing mineral. Without further evidence there is no way of knowing whether such magnetite formed as a late magmatic mineral from the melt which produced the gabbro or whether it has replaced pyroxene as a result of a later period of iron metasomatism, such as that which formed the nearby Merry Widow and Kingfisher magnetite deposits. Magnetite in four samples from the gabbro has a titanium content ranging from 5.6 to 8.4% Ti. In contrast, one sample from the Kingfisher deposit contains no titanium and magnetite from another small deposit in the area analyzed 0.0074% Ti. The contrast in titanium content of accessory and ore magnetite suggest different conditions of formation. Buddington et al. (1955) showed that Ti-rich magnetites are characteristic of rocks formed directly from a melt at high temperature and that,

in these cases, magnetite formed as a primary differentiate. Pegmatites, on the other hand, were shown by the same authors to carry Ti-poor magnetite. It is possible, therefore, that magnetite in the Coast Copper gabbro was produced as a late magmatic mineral and magnetite in the orebodies as a later hydrothermal mineral. The consistent Fe_m/Fe_t ratio shows that as the melt cooled and began to crystallize, a fixed proportion of iron was taken up as augite and the remainder formed magnetite which then, in part, replaced earlier-formed augite.

Volcanic rocks

Of magnetite deposits that are in volcanic rocks, most are in the Karmutsen group or its correlative. Analyses of three volcanic rocks from this group are given in Table 13 and are shown graphically in Figure 7. Average iron content of these Karmutsen rocks is 7.90%. The composition of these rocks is not significantly different from world averages of andesite and basalt (Nockolds, 1954).

Two deposits, Merry Widow and F.L., occur in pyroclastics and flows of the Bonanza group. Two chemical analyses of Bonanza tuffs are given by Stevenson (1950, pp. 20-21); average iron content of these rocks is 8.3%. Red and green siliceous tuffs are intercalated with flows and

sedimentary material on Tahsis Mountain, Zeballos map-area, and an analysis of green tuff is given by Hoadley (1953,p26). He divided the flows into two broad groups, red andesites and green andesites. Two analyses of the red and three of the green andesite show that the red have a higher ferric:ferrous iron ratio, a condition which the red colour would suggest. Average iron content of red andesite is 6.24% and of the green, 8.95% (Hoadley,1953,p.25). A specimen of Bonanza group andesite from the F.L. mine area has an iron content of 7.6% (see analysis 3, Table 13). This analysis is shown graphically in Figure 7 (no.3). An analysis of green andesite collected by the present writer from the Bynnor mine area, tentatively correlated with the Bonanza group (Eastwood,1962, p.113), has a total iron content of 7.29% (No. 2, Fig. 5).

Limestone

Limestone of the Quatsino formation, or its correlative, is associated with all known Coastal British Columbia economic magnetite deposits. This unit is characterized by its (1) massive nature (2) widespread recrystallization (3) high purity and (4) scarcity of fossils. Chemical analyses (Table 14) show that such limestone is more than 85% CaCO_3 . Near some granitic intrusions, dolomitization of limestone

FIGURE 7
Diagram showing compositions of volcanic rocks associated with contact metasomatic magnetite deposits. Numbers refer to analyses, Table 13.

has been recognized (Stevenson, 1950, p.17) and in the Brynnor mine magnesium metasomatism, in the form of serpentine, has increased locally the magnesia content of the associated limestone. Also, the upper part of the Marble Bay formation on Texada Island is magnesium-rich (Mathews and McCammon, 1957, pp.52-55). No known magnetite deposits occur, however, in dolomite, dolomitic limestone, or magnesian limestone. Other impurities, such as alumina and iron, are negligible.

Skarn

The original meaning of the term skarn is lost in antiquity. Lindgren (1933, p.705) mentioned that it is an old Swedish mining term for the garnet-pyroxene-epidote rock accompanying many Scandinavian magnetite deposits. The American Geological Institute Glossary (1957) defines skarn as "..... a rock nearly entirely composed of lime-bearing silicates and derived from nearly pure limestones and dolomites into which large amounts of Si, Al, Fe, and Mg have been introduced". Tactite has been suggested for those skarns of contact metamorphic origin (Hess, 1919; Schmitt, 1948). These definitions, however, are unsatisfactory in that many skarns are derived from rocks other than limestones and dolomites. Shaw (1957) noted that the mineralogy of skarns is complex

and that some are of contact metamorphic origin, some are caused by simple metamorphism, some are metasomatic, and some are veins. He referred to skarn as "a rock consisting mainly of lime silicates with or without carbonates and commonly containing magnetite or other iron-bearing minerals". This definition cannot be applied rigorously because to do so would include rocks such as magnetite-bearing anorthosites, hornblendites, and other rocks not normally considered to be skarns. Minerals such as garnet, diopside, hedenbergite, epidote, ilvaite, vesuvianite, etc. are common constituents of skarns and in this report these minerals constitute 80% or more of the rock referred to as skarn. The remaining 20% may be unreplaced host rock or other non-calcium-silicate minerals. Various adjectives, used in conjunction with the term skarn, identify skarn of a particular mineral composition (e.g. garnet-skarn, epidote-skarn, etc.). The author proposes that the process by which skarn is formed be called skarnification i.e. the replacement by, conversion into, or introduction of skarn. The term is analogous to others such as granitization, chloritization, sericitization, etc. The term would include all processes by which skarn may be formed such as contact metamorphism, contact metasomatism, or regional metamorphism.

Skarn in western British Columbia magnetite deposits

is mineralogically and chemically simple. Four phases, garnet, pyroxene, epidote, and magnetite comprise more than 95% by volume of skarn. Chemical analyses further show that seven components, CaO , FeO , Fe_2O_3 , SiO_2 , Al_2O_3 , MnO , and MgO comprise more than 95% (by weight) of two samples of typical skarn (Table 15). The number of phases, therefore, is not more than the number of components. This conformity to the Mineralogical Phase Rule suggests, although it does not prove, that equilibrium was reached during formation of skarn. Water, carbon dioxide, oxygen, and other volatiles are considered to be perfectly mobile components.

A number of rock types have been replaced by skarn in the area considered by this study. The volume of andesite, basalt, and tuffaceous rocks which have been altered to skarn probably exceeds the volume of all other rock types combined, including limestone. Deposits in volcanic rocks are commonly enclosed in a wide skarn zone extending outward from the central ore zone as irregular tongues, shoots, or patches. Deposits in limestone, in contrast, are surrounded by only a narrow rim of skarn. Where plutonic rocks occur adjacent to magnetite deposits, they too are replaced, to a minor extent, by skarn.

In most cases, skarnification has obliterated

textures and structures originally present in the host rock so that it is difficult to determine with certainty what rock has been replaced. The homogeneous nature of skarn is probably largely caused by the massive texture and structure of the host rocks. Karmutsen volcanics, Quatsino limestone, and granitic plutons are all locally homophanous and of homogeneous composition. Some small-scale inhomogeneities however, do occur in these rocks and even these heterogeneities have been completely transformed into massive red-brown garnet skarn which is macroscopically identical with that formed by replacement of other homogeneous rocks. Examples of such inhomogeneities are aplite dykes in the Gillies stock and bedded tuff in the Merry Widow and Brynnor mines.

In spite of the obliterating effect of skarnification, certain features in some skarns have been recognized as occurring more or less consistently with particular rock types. For example, skarnification of aphanitic rocks, particularly dykes or flows, usually has resulted in a fine-grained compact skarn. There is a general tendency for such skarn to contain a relatively large proportion of pyroxene and/or epidote, rather than garnet. Skarn formed in limestone, in contrast, is medium- or coarse-grained and vuggy. These vugs are filled with calcite, sulphides, quartz or, near the ore zones, magnetite. Garnet is the predominant mineral in skarn of

this type. Where skarnification has altered plutonic rocks, the granitic texture of the host is retained in rocks bordering the massive skarn zone (Plates II and XVII). In such border zones, secondary calc-silicate minerals form only 20-30% of the bulk composition and the rock cannot properly be classified as skarn, although it may grade into compact skarn within a few feet.

Under the microscope, several features of skarn become evident which cannot be seen in handspecimen, even in coarse-grained varieties. Quartz, prehnite, ilvaite, specularite and vesuvianite are frequently found to be relatively abundant in thin-section. Cross-cutting veinlets which establish age relations are also more apparent and many replacement features are visible only under the microscope. Garnet skarn, of the compact variety replacing volcanic rocks, is frequently almost cryptocrystalline.

Replacement by magnetite is the dominant process in the formation of compact, fine-grained ore in which magnetite is disseminated throughout garnet skarn. Skarn formed in limestone, however, is coarse-grained and vuggy. Magnetite fills small vugs and extends outward along crystal boundaries of large garnet crystals (Plate XXII). Replacement textures are subordinate in rock of this type and it appears that

magnetite, which is everywhere later than skarn, first filled these open spaces and then began replacing garnet outward from them (Plate XXIII).

Garnet and diopside are the first skarn minerals to form and appear to be contemporaneous. These minerals are commonly disseminated in each other and cross-cutting veins of garnet and diopside show conflicting age relations. Diopside veins in garnet, however, are more frequent than garnet veins in diopside. From this it can be inferred that in some deposits at least formation of diopside overlapped and outlasted formation of garnet.

Epidote (variety pistacite) is about as abundant as diopside in skarn. Macroscopically epidote is more prominent but much diopside is present as small grains disseminated through otherwise compact garnet. At least two stages of epidote mineralization can be recognized in most deposits. The earliest of these stages is found along joints or faults of small displacement and as disseminations or irregular patches in intrusive granitic rocks or volcanic country rocks. Epidotization of this type has long been recognized in the Coast Range and has generally been interpreted as an alteration product of plagioclase and ferromagnesium minerals or as late-stage vein-fillings (e.g. Bacon, 1957, p.23). These

epidote veins are frequently cut by garnet skarn and do not appear to be genetically related to contact metasomatism (Plate XIX). A later epidote, often found as veins transecting contacts of garnet skarn and country rock, is considered by the writer to be an integral part of skarnification even though such epidote is usually later than the predominant garnet-diopside metasomatism. In those deposits which show zoning, epidote forms the outermost skarn zone and grades outward into relatively unaltered country rock. Epidote of one stage is identical to that of the other.

Magnetite is everywhere later than skarn minerals. It occurs as fracture-fillings, vug fillings, or as a replacement of skarn or country rocks.

Veins and open-space fillings of calcite, quartz, and sulphides are the last of the relatively common skarn minerals to form. Euhedral quartz and calcite crystals are frequently found occurring together in vugs or veins. Their co-existence shows that the rocks in which they were formed had cooled to at least 450°C (Barth, 1951, p. 287). Minor amounts of ilvaite, specularite, chlorite, and actinolite frequently accompany quartz and calcite.

Several other minerals were found in skarn but these

generally occurred too infrequently or in amounts too small to establish their age relative to the main skarn minerals. These include prehnite, wollastonite, vesuvianite, chlorite, tremolite, and actinolite.

A description of skarn-forming minerals encountered in this study follows. They are given in order of decreasing abundance.

Garnet group

Members of the garnet group are by far the most abundant skarn minerals. The compositions of these garnets were closely studied because they provide an excellent guide to the general composition of skarn. Refractive index and cell edge were used to determine composition, using the values published by Skinner (1956). Specific gravity determinations were meaningless because of the many inclusions in these garnets.

Refractive indices were matched in oil in white light and the index of the oil was measured on a Leitz Micro-Refractometer. Determinations on this apparatus can be read directly to 0.01 and estimated to ± 0.002 units. Results were reproducible within the limits of reading error. The

Leitz Refractometer was calibrated against an Abbe Refractometer and results checked within 0.002 to 0.007 units to the upper limit of the Abbe apparatus ($n = 1.70$).

Cell-edge determinations were made by measuring interplanar spacings of X-ray powder patterns and then correcting for film shrinkage. A camera of 5.73 cm. diameter was used and 20 measurements were made with an accuracy of $\pm 0.05^\circ$. Manganese-filtered iron radiation was used throughout.

The limits of error in refractive index and cell edge determinations together result in a garnet composition error of $\pm 5\%$. The compositions of 60 garnets determined in this manner are shown in Figure 8 on a diagram modified from Winchell (1958). The directions of hydrogrossularite (Flint, McMurdie, and Wells, 1941) and the fictive "hydroandradite" are also shown in relation to the grossularite and andradite parameters. Ford (1915) found that two-thirds of all analyzed garnets could be represented by three formulae; two formulae are always predominant and the third averages about five percent and is not known to exceed 20%. Figure 8 shows that garnets in skarn deposits considered in this study fall close to the grossularite-andradite join with small amounts of almandite and/or spessartite appearing in most specimens. The preponderance of grandite garnet (a simple designation for

the grossularite-andradite series) is in accord with the findings of Wright (1938) who showed that garnet of this series constitutes more than 90% of the garnet in calcareous contact rocks.

Several determinations plot beyond the point representing pure andradite (see Fig. 8) and for these garnets, cell-edge measurements are all less than or equal to that of andradite. The refractive indices, however, of these garnets are too high for the corresponding cell edges. Garnets of anomalously high refractive index show no correlation with the type of rock replaced, colour of garnet, or birefringence. A high TiO_2 content (5% TiO_2 or more) is known to occur in garnets of high refringence (Winchell, 1959, p. 493). Thirteen specimens, many obtained from single garnet crystals, were analyzed spectrographically for titanium but averaged only 0.02% Ti. The reason for the high refringence remains unknown.

Garnets studied in this project are various shades of reddish-brown; yellow-green, green, dark brown and black varieties have also been found. Garnets of two or even three distinct colours may occur in the same hand specimen. Where coexisting garnets of two colours, such as red and green, were analyzed in the manner described above, the difference in composition was less than the analytical error of 5%.

FIGURE 8

Composition of garnet in skarn as determined by cell edge and refractive index. Values for end members taken from Skinner (1956).
Diagnostic components shown beside each sub-species.

Complete crystals are seldom found. Several crystal faces frequently are found projecting into vugs or other cavities but the remainder of such grains is generally massive. Common crystal forms are the dodecahedron and trapezohedron or a combination of these. Faces are commonly striated.

Many garnet crystals show concentric colour zoning accentuated by striations parallel to the zones. Not all crystals in the same hand specimen show zoning nor do all zoned crystals show the same kind or degree of zoning.

Garnet has a variety of textures that depends to some extent on the rock replaced. Compact, fine-grained garnet has very commonly replaced volcanic rocks, dykes, or plutonic rocks. Round or sub-round clusters of garnet grains are frequently found enclosed in massive magnetite in an aggregate resembling glomeroporphyritic texture in igneous rocks.

Many garnet grains are markedly birefringent and frequently such anisotropism occurs in concentric straight-sided zones, parallel to the crystal faces, and alternating with otherwise similar isotropic zones (Plate XXI). In some specimens the anisotropic variety forms irregular patches within massive, isotropic, almost cryptocrystalline garnet; in other specimens, birefringent garnet transects the boundary between two crystals. In a few specimens, the anisotropic

type is lighter in colour than the isotropic variety but otherwise no correlation with colour was noted; neither is there a tendency for birefringent garnets to be found as a replacement of a particular rock type or to be a particular composition. Inclusions of pyroxene, magnetite, calcite, epidote, and other minerals are seen in most isotropic garnets under the microscope. Birefringent garnets, however, are commonly more coarse-grained and free of inclusions than isotropic garnets.

Many birefringent zones give off-centre uniaxial interference figures; some are optically positive, others are negative. A few grains show what appear to be off-centre biaxial interference figures but attempts to confirm this with the universal stage were unsatisfactory. Biaxial andradite has been reported by Stose and Glass (1938) and biaxial grossularite by Taylor (1935).

Many investigators have noted the anomalous birefringence displayed by garnets in contact deposits but few writers have offered to explain this phenomenon. Yoder (1950) suggested the anisotropism is caused by a change in the crystal structure brought about by substitution of ferric iron for trivalent aluminum. Yoder considered this to result in a sub-group of symmetry lower than that of the basic grossularite

structure. It is difficult to understand how such substitution would cause birefringence, as ferric iron and aluminium are the main metallic elements in andradite and grossularite, respectively. As a continuous solid solution series exists between these two end members anomalous properties such as birefringence would not be expected.

Birefringence in a normally isotropic material is evidence of a lowering of its optical symmetry. To determine whether there was a corresponding decrease in the crystal symmetry of birefringent garnet, X-ray powder diffraction photographs of several doubly refracting garnets were taken and compared with isotropic garnets of comparable composition. In every instance, the patterns were identical within the limits of reading error. An example of such a pair is shown in Table 7.

Other minerals which show anomalously low optic symmetry are biaxial quartz and calcite. Neither of these minerals have been shown to be crystallographically different from the normal uniaxial variety. Frondel (1962,p.133) found that in quartz "a biaxial character can be produced artificially by compression in directions inclined to the optic axis, as well as by a thermal gradient, or by application of an electric field". Biaxial calcite, frequently showing bent twinning

lamellae, has been noted by many investigators. Paulitsch (1951) reviewed data on the biaxial nature of calcite and concluded that the phenomenon had not been satisfactorily explained. If the reason for the subnormal optical symmetry displayed by these minerals could be explained it might explain also birefringent grandite garnets.

Table 7

Comparison of interplanar spacings for birefringent and isotropic garnets of composition $An_{55}Gr_{37}Al_{08}$

U.B.C. 3761 - Isotropic		U.B.C. 3671 - Anisotropic	
	<u>d (meas.)</u>		<u>d (meas.)</u>
1.	2.98		2.98
2.	2.66		2.67
3.	2.53		2.55
4.	2.43		2.44
5.	2.34		2.34
6.	2.17		2.18
7.	1.934		1.937
8.	1.753		1.757
9.	1.719		1.723
10.	1.652		1.656
11.	1.593		1.597
12.	1.488		1.492
13.	1.299		1.299
14.	1.270		1.270
15.	1.109		1.109
16.	1.090		1.090
17.	1.056		1.056

Fronde1 (1962) noted that a thermal gradient produced biaxial quartz. Wright (1915) discovered that garnet lost its birefringence after being heated for a few hours at 800°C and did not regain it after several hours at 600°C. A similar experiment was performed on andradite by Stose and Glass (1938) and they found it became isotropic at 860°C. and did not invert to its original anisotropism on cooling. Specimens of andradite from the Pewabic mine, N.M. lost their birefringence at 860°C. (Allen and Fahey, 1957, p. 894). Grossularite has been shown to lose its birefringence at temperatures between 700°-800°C. (Stone, 1959). These workers proposed that birefringent garnet could be used as a rough geologic thermometer, based on the experimental evidence cited above. They concluded that birefringent garnet has formed below, and isotropic garnet above, a temperature of about 750°-850°C.

Slight variations in composition of garnet could possibly produce anomalous optical properties. In a study of mixed crystals of simple salts, Bunn (1933, p. 569) found that cubic crystals of lead nitrate and barium nitrate became birefringent and pleochroic when coloured with methylene blue. The contaminated nitrate crystals, when first formed, were a single solid phase. Optically, each cube was divided diagonally into six pyramids with their apices at the centre of the

crystals. The pyramids opened out to six cube faces, each pyramid having the same optical orientation with respect to its own cube face. These pyramidal forms are similar to those found in birefringent garnets showing biaxial interference figures with optic angles of $50-90^{\circ}$ (Winchell, 1959, p.490).

In another set of experiments, Bunn (p.571) found that solutions of ammonium chloride containing the divalent chlorides of Mn, Fe, Co, Ni, or Cu crystallized into birefringent cubic crystals. These cubes were divided diagonally into six sectors, each of which was uniaxial with its optic axis normal to the cube face.

When allowed to rest for three years in the dry state, the contaminated ammonium chloride crystals, which were originally physically homogeneous, separated into two solid phases.

Small birefringent crystals appeared on the surface and the birefringence of the main ammonium chloride crystals had decreased practically to zero. The rate of this "exsolution" is apparently not the same for all metallic chlorides; tetragonal crystals of ammonium cupric chloride "exsolved" within three months, the other chlorides took longer to separate (Bunn, 1933, p.574). Bunn concluded that, in the examples studied, the optical effects showing the presence of impurities are stronger when the crystals have grown rapidly than when they grew slowly. Slow-growing crystals are apparently able to reject impurities arriving at their faces whereas, when

growing rapidly, they are unable to do this completely. In time, however, even the rapidly-crystallized crystals were able to reject their impurities as a second solid phase (Bunn, 1933, p. 572).

The effects produced by impurities in these chemically and crystallographically simple salts are strikingly similar to those observed in birefringent garnet. Mention has been made elsewhere that birefringent garnets are more common as replacements of limestone. Such garnets could conceivably have formed by rapid precipitation from solutions brought into contact with the highly reactive carbonate. Rapid growth of crystals would also be enhanced by high temperatures. These rapidly-grown crystals would not be able to reject impurities incorporated into their structure and would therefore be birefringent. These impurities would be retained in the crystal and would not separate into a second phase. It has been previously noted that birefringent garnets are generally free of solid inclusions. Isotropic garnet, on the other hand, is more common in volcanic rocks and is generally crowded with minute inclusions of a variety of minerals. Such garnet could have formed slowly as a result of solutions brought into contact with these chemically less-reactive volcanic rocks. Lower temperature could also slow the growth

of garnet of this type. These more slowly-growing grains would reject impurities now seen as solid inclusions in isotropic garnet. This proposal that birefringent garnet formed at a higher temperature and more quickly than isotropic garnet contrasts with the conclusions of Wright, Stose and Glass, Stone, and Taylor reached from their experiments on heating birefringent garnet. The difference can perhaps be explained in this way: when birefringent garnet is subjected to prolonged heating, the impurities present in the crystal have time to arrange themselves into discrete grains. These are rejected by the garnet which then becomes isotropic. Because of the more complex chemistry and crystallography of garnet, prolonged heating is required to produce in this mineral what time alone produced in the simple salts studied by Bunn. The inclusions, once they have been rejected by the garnet structure, would not be re-incorporated at lower temperatures and therefore garnet would not regain its birefringence upon cooling. Alternate isotropic and anisotropic zones, or irregular birefringent "patches" in grandite, if the process outlined above even approximates what actually takes place in nature, could therefore be caused by different rates of crystal growth in garnet. Depositional rates can change with time, temperature, composition of the depositing solutions, composition of the host rocks, or other factors.

It is concluded that because temperature is only one factor which can cause birefringence in isotropic compounds, anisotropism in garnets cannot be correlated with temperature of formation.

Pyroxene

Green, anhedral pyroxene is next in abundance to garnet. Pyroxene occurs as veins, irregular patches, or fine-grained microscopic disseminations in garnet skarn. Locally, it can form a major proportion of skarn but generally it is subordinate to garnet.

The most common form is medium-grained, anhedral, granular aggregates or veins of green crystals in garnet or altered volcanic rocks. In limestone, however, euhedral crystals up to five inches long and one inch thick have been found. Crystals from the Prescott mine, Texada Island, are "perched" on a stem of pyroxene attached to a layer of magnetite (Plate X).

The intermediate refractive index (n_y) of five specimens ranged from 1.689 to 1.695 ± 0.002 . Optic angle measurements showed $2V = 55 \pm 2^\circ$. These properties classify the mineral as salite, an intermediate member of the diopside-

hedenbergite series (Hess, 1941; Winchell, 1959, Fig. 289, p. 410). According to correlation charts published by Kennedy (1947, p. 565) these refractive indices correspond to compositions ranging from $\text{Di}_{74}\text{He}_{26}$ to $\text{Di}_{65}\text{He}_{35}$.

Long, brown pyroxene crystals with a dull bronze-like lustre coexist with garnet and ilvaite in skarn at Iron Hill. Refractive index measurements show: $n_x = 1.721$, $n_y = 1.729$, $n_z = 1.747$. These correspond to pyroxene of composition $\text{He}_{84}\text{Di}_{16}$ (Kennedy, 1947, p. 565) and classify the mineral as hedenbergite (Hess, 1941).

Epidote group

Epidote is ubiquitous in skarns of southwestern British Columbia. It is the last of the major calc-silicate minerals to form and, in zoned deposits, predominates in the outermost zone.

The most common variety is green pistacite occurring in veins, patches, disseminations, or aggregates of euhedral to subhedral crystals. In thin section, the mineral is pleochroic from colourless to straw yellow and has inclined extinction. The intermediate refractive index (n_y) of three specimens was measured and gave the following: $n_y = 1.750$,

1.758, and 1.775. These correspond to compositions of 27, 33, and 47 molecular percent of $\text{Ca}_2\text{Fe}_3(\text{OH})\text{Si}_3\text{O}_{12}$, respectively (Winchell, 1959, Fig. 343, p. 449).

A small amount of clinozoisite ($\text{Ca}_2\text{Al}_3(\text{OH})\text{Si}_3\text{O}_{12}$) was identified in a few samples and is found on the outside border of the epidote zone (Zone 3; sub-zone 1(b); Table 11).

Calcite

Calcite is most common in skarn formed as a replacement of limestone. In this type of skarn it fills vugs or small openings between euhedral garnet crystals (Plate XX) and cleavage rhombs in these openings are frequently three to four inches wide. A large vug in skarn at the Iron Hill mine contained inward-projecting crystals of translucent Iceland spar, some of which showed concentric zones of a red dust-like material believed to be hematite. Rhombs of transparent calcite have been reported also from the Merry Widow mine.

Calcite, accompanied by sulphides, quartz, or other minerals, cuts garnet, diopside, and epidote and is one of the first of the non-calcium-silicate minerals to form in skarn. Disseminated calcite inclusions common in garnet and diopside probably were incorporated by these crystals during growth.

Quartz

Quartz is found in small quantities in skarn and is everywhere a late-forming mineral. Prismatic crystals with pyramidal terminations occur with calcite in veins or vugs. Many examples of such coexisting calcite and quartz were examined in this section and nowhere was there evidence of reaction. In several specimens, quartz crystals which project through scalenohedrons of calcite into open vugs have sharp, uncorroded contacts with calcite.

Amphibole group

Clusters of acicular green actinolite crystals are frequently found with epidote on the margins of skarn zones. Bodies of garnet skarn are, in places, rimmed by a thin discontinuous zone of actinolite. Much interstitial actinolite disseminated with chlorite in diopside may be an alteration product. Optical properties of one specimen of actinolite are: $N_x = 1.655$, $N_y = 1.662$, $N_z = 1.680$; $(-)2V = 83 \pm 2^\circ$. These parameters are also similar to those for intermediate members of the cummingtonite series but X-ray powder diffraction patterns showed the mineral to be of the tremolite-ferrotremolite series. The optical properties described above place this mineral almost exactly midway in composition

between the two end members; that is, the composition of this actinolite is $\text{Ca}_2\text{Mg}_{2.5}\text{Fe}_{2.5}\text{Si}_8\text{O}_{22}(\text{OH})_2$ (Winchell, 1959, Fig. 323, p. 433). An amphibole approaching the composition of grunerite was identified from Texada Island by Swanson (1924, p. 120A) on the basis of refractive index determination. However, no members of the cummingtonite (kupfferite-grunerite) series were found by the present writer.

Prehnite

Narrow veins of prehnite, some showing poorly-developed "bow-tie" structure, transect most primary skarn minerals. It is not known for certain whether such prehnite is a late-forming primary skarn mineral or whether it is an alteration of pre-existing minerals.

Vesuvianite

Anhedral to subhedral vesuvianite was identified in specimens from the Iron Hill mine. Most grains occur as disseminations in garnet-pyroxene skarn but several interstices between euhedral garnet crystals are completely filled with vesuvianite.

Wollastonite

This is one of the less common minerals found in skarns considered in this study. It is found with garnet and calcite in the Prescott mine, Texada Island.

Ilvaite

This mineral was found only at the Iron Hill mine. At this locality, long black ilvaite crystals occur with garnet and hedenbergite in skarn. Quartz crystals, projecting through calcite into a hollow vug, are covered with a drusy coating of ilvaite and magnetite. Quartz-calcite veins in this mine also carry minor amounts of ilvaite.

Axinite

Light purple crystals of axinite are found in a quartz-calcite vein in the Brynnor Mines Ltd. open pit. The vein appears to follow a contact between andesite and skarn.

Metallic minerals

Magnetite, pyrite, and chalcopyrite are the main metallic minerals in the magnetite deposits considered in

this study. In most mines, the proportion of sulphides to oxides is very low. Magnetite is, of course, the main oxide mineral and sulphides occur as disseminations in magnetite or as irregular masses. The age relation between magnetite and sulphides is not consistent even within a single deposit; there is evidence of sulphides forming contemporaneously with or later than magnetite but none which would indicate that they formed before magnetite. Pyrite and chalcopyrite are the most abundant sulphides; pyrrhotite, sphalerite, arsenopyrite, and cobaltite occur in minor amounts.

Magnetite

Granular, subhedral, medium-to coarse-grained, blue-black magnetite is the most common habit of this mineral. Coarse-grained vuggy magnetite occurs in limestone and a fine- or medium-grained variety replaces volcanic rocks. Magnetite is skarn, particularly in garnet skarn, appears to have first filled cavities and then proceeded along garnet crystal boundaries and then into the crystals themselves. Crystal boundary replacement has resulted in rounded garnet crystals in a matrix of magnetite (Plate XXIII). Complete replacement of garnet then followed and several examples of magnetite apparently pseudomorphous after garnet were seen.

Fracture-filling followed by replacement is the dominant process of deposition in volcanic rocks (Plate XII); complete replacement along an advancing chemical "front" resulting in bulbous forms and sharp contacts of magnetite deposits predominates in limestone (Plate XI).

Under the microscope the magnetite usually forms a granular aggregate of interlocking, isotropic anhedral grains. Magnetite pseudomorphous after garnet and specularite was noted although it is not common. Some samples show an interlocking aggregate of magnetite grains which are distinctly anisotropic but non-pleochroic. Anisotropism varies in intensity from area to area in a single section but the polarization colours are the same: blue-gray to reddish-brown. A texture resembling polysynthetic twinning is seen in some magnetite of this type. Anisotropic magnetite has been reported from the contact metasomatic deposit at Cornwall, Pa. (Hickock, 1933) but the reason for the anomalous optical property is unstated. An X-ray powder diffraction pattern of anisotropic magnetite from the Iron Hill mine, Vancouver Island, showed that such magnetite had normal cubic symmetry. Magnetite with rhombohedral symmetry at -158°C . has been reported by Rooksby and Willis (1953). However, Abrahams and Calhoun (1953) believe the transition is to orthorhombic symmetry. To determine whether or not anisotropic magnetite

contains excessive titanium a specimen from the Yellow Kid deposit was analyzed. It contained only 0.027% Ti, about average for isotropic magnetite from the same deposit.

Contact metasomatic magnetite is characteristically low in common trace element impurities found in the magnetite of other types of deposits. Elements such as titanium and phosphorous, relatively abundant in many magmatic magnetite deposits, are found in negligible quantities in contact deposit. Manganese is common in many sedimentary iron ores but is a very minor constituent of contact magnetite deposits. Analyses for P, Ti, Mn, and other elements in concentrates from several British Columbia deposits are given in Table 8, using data from company files and from Elver (1963):

Table 8

Partial analyses of magnetite concentrates, British Columbia
(Analyst, W.F. White, Geological Survey, Canada)

Mine	Elements (%)				
	P	Mn	Ti	Cr	As
Brynnor	0.05	--	0.03		
Jedway	0.02	0.17	0.04		
Nimpkish	0.013	0.16	--		
Texada	0.011	0.13	0.09	0.0014	0.007
Zeballos	0.02	0.1	0.08	0.0008	not detec'd

(other elements for Zeballos - Zr 0.04, Sn 0.0005,
Ni 0.002, Co 0.001, V 0.003)

Masses of magnetite with colloform texture have been found in the Kingfisher, Merry Widow, Shamrock-Blackjack, Yellow Kid, and Bacon Lake deposits. These masses range from a fraction of an inch to one foot in diameter and are roughly hemispheroidal in shape (Plate XVI). Sections through larger hemispheroids or "nodules" show a radiating structure and concentric banding. In smaller nodules the radiating structure is lacking. The banding is caused by layers of magnetite of differing grain size. Some layers are composed of acicular magnetite crystals up to $\frac{1}{2}$ inch long; in others, the magnetite is extremely fine-grained and exhibits an irregular fracture pattern or "crackled" texture. The different layers frequently spall off when struck and some are separated by thin parallel layers of calcite. The radiating structure is produced in two ways: (1) parallel elongate magnetite crystals arranged perpendicular to the banding (2) thin, radial, bifurcating cracks extending inward from the outer surface of each hemispheroid. Some cracks only penetrate about one quarter of the way into the mass, others extend to the centre. Some cracks are filled by calcite. The outer surfaces of the hemispheroids are parallel to the curved inner bands.

Textures such as banding, bifurcating radial cracks, and "crackled" appearance, together with a spheroidal shape, have been attributed by Bastin (1950) to deposition from a

colloidal dispersion. A colloidal dispersion is a system in which extremely fine-grained particles are suspended, more or less indefinitely, in a fluid medium. Particle size in the colloidal range is about 5 μ to 200 μ (Hartman, 1947). The following points are characteristic of colloidal dispersions: (1) extreme buoyancy (2) particles are affected by Brownian movement (3) the particles are easily absorbed (4) some substances, when in the colloidal state, carry an electrical charge which repels the particles and stabilizes the dispersion. Flocculation is brought about by neutralizing this charge (5) the large surface area exposed is favourable to chemical interaction between the solid and the enclosing fluids and (6) the dispersion has little effect on the physical properties of the solution (this is particularly true of lyophobic colloids such as a suspension of a non-hydrous metallic substance in a hydrous medium).

Colloform texture is common in many minerals and these are considered to have been deposited from colloidal dispersion (Bastin, 1950; Boydell, 1924-25; Lindgren, 1924). Evidence of colloidal deposition, however, merely indicates that the mineral was once in the colloidal state and need not have been carried to its present position as a colloid. Also, all substances precipitating from electrolytic solutions must

pass through the colloidal size range even though the final particles may be larger than those of a colloid. Rapid precipitation is known to produce stable colloids by crystallization about many centres. A mineral having colloidal texture could have been brought to its site of deposition in electrolytic solution and then deposited as a colloid by rapid precipitation. It is significant to note that magnetite showing such texture in the Merry Widow, Kingfisher, Yellow Kid, and Klaanch deposits is found only in limestone. The close association with this "highly reactive" rock may have been instrumental in the formation of colloidal magnetite by promoting rapid precipitation.

Colloform magnetite from the Sudbury district, Ontario, was described by Moore (1932) although he did not suggest its origin. Jeffery (1960b) and Jeffery and Stevenson (1961) described "nodules" of colloform magnetite found in talus in the Merry Widow and Kingfisher deposits and suggested a colloidal origin for such magnetite although they did not elaborate. The present author observed botryoidal magnetite in place not only in the Kingfisher but also in the Klaanch deposit. In both deposits it occurs in limestone in or near a post-ore fault in which magnetite fragments have been cemented by calcite. In most cases the fragments are

surrounded by a rim of coarsely crystalline magnetite. It is suggested that solutions which deposited the calcite cement also dissolved and reprecipitated magnetite around the fragments. Colloidal magnetite could have formed at this time because there is apparently no reason why colloidal aggregation and electrolytic precipitation cannot take place at the same time (Bastin, 1950, p.22). The external surfaces of colloform magnetite are always convex toward surrounding calcite. Colloidal particles of magnetite may have coagulated in open spaces between earlier brecciated magnetite and then the remaining space was filled with calcite. Deposition of calcite sometimes overlapped that of magnetite as shown by "conformable" layers of calcite in colloform magnetite.

Coagulation of magnetite particles was probably brought about by neutralizing an electric charge on the particles. This could have been effected by electrolytes in a hydrothermal solvent. The magnetic properties of magnetite particles in some way may have also aided coagulation. The effect of temperature on colloidal dispersions is not fully understood but Hartman (1947, p.283) noted that, in general, the stability of a solution increases with rising temperature i.e. the tendency to coagulate is greater at lower temperatures.

The properties of magnetite sol have not been fully investigated. Several texts on colloid chemistry do not mention magnetite nor is the mineral discussed in many papers by Boydell and by Lindgren on colloids as applied to ore deposits. Colloidal ferric hydroxide is well known but in the samples of colloform magnetite studied there is no indication that such magnetite has replaced an earlier hydroxide; no unreplaced remnants of goethite, hematite, or other mineral were found in more than two dozen specimens examined. Lindley and Brown (1958) discovered colloidal-sized magnetite (50 mu) in the oil system of a steam turbine. In a mixture of colloidal magnetite and oil, no deposition took place. However, when only 1% brine was added as an emulsion, the magnetite particles formed aggregates around the fine droplets of water in the medium. The NaCl electrolytic solution probably acted as a pectizing agent which flocculated magnetite by neutralizing the charge on colloidal particles. In some respects, then, magnetite sol behaves as an ordinary lyophobic sol.

Colloform aggregates of other "high temperature" minerals such as garnet and molybdenite have been described by Vakhrushev (1961) in ferruginous skarns of western Siberia. Colloform garnet of the andradite-grossularite type is the main constituent of this skarn. Vakhrushev concluded

that colloform garnet formed as the result of activity of the residual hydrothermal solutions after the formation of skarn. Crystalline garnet was dissolved and redeposited in the form of fine-grained and cryptocrystalline aggregates. The process of formation of colloform garnet described by Vakhrushev is almost identical to that proposed by the present author for colloform magnetite.

The writer concludes: (1) colloform magnetite in the Kingfisher and Kilauea deposits was deposited from a colloidal dispersion (2) colloform magnetite is a secondary feature brought about by redeposition of earlier magnetite in a post-ore fault now cemented by calcite (3) the solutions which carried this calcite dissolved some of the crystalline magnetite of the brecciated ore in the fault and (4) this dissolved magnetite was precipitated both as an electrolytic crystalline aggregate and also as a colloid.

Pyrite

Pyrite occurs in a variety of forms and is relatively abundant in most rock associated with contact metasomatic magnetite deposits. Anhedral pyrite is scattered throughout massive magnetite and skarn; euhedral forms are more abundant in limestone. Striated cubes, octahedrons, pyritohedrons, or

combinations of these have been found. Several deposits contain minor amounts of pyrite as roughly circular plates about 1/16 inch in thickness and one inch in diameter. These plates have both a random and an oriented habit. In polished section, these plates are seen to be composed of several large pyrite grains. Many are anisotropic but X-ray diffraction patterns showed the grains to be pyrite and not marcasite.

Chalcopyrite

Anhedral masses of chalcopyrite are found in almost every deposit. In some deposits disseminated blebs and patches in magnetite appear to have been deposited essentially contemporaneously with the iron oxide; in other deposits, or in different parts of the same deposit, chalcopyrite can be seen veining magnetite and skarn or filling vugs in skarn. Very minor amounts are exsolved in sphalerite.

Chalcopyrite in the Texada Island, Merry Widow, and Tassoo deposits carries considerable amounts of silver and minor amounts of gold. Eastwood and Merrett (1961, pp.96-97) suggested that, in the Merry Widow deposit, silver is preferentially concentrated in chalcopyrite but gold is related to the total sulphide content of the ore. It probably also occurs in other sulphide minerals.

Pyrrhotite

This mineral occurs in minor amounts in much the same manner as chalcopyrite. It is found with chalcopyrite, pyrite, and arsenopyrite.

Sphalerite

Small amounts of sphalerite are found with other sulphide minerals in several deposits. It also occurs as exsolution bodies in chalcopyrite.

Arsenopyrite

Arsenopyrite is found only in the Merry Widow deposit and is associated with other sulphide minerals in that deposit.

Cobaltite

Cobaltite occurs in very small amounts in magnetite and skarn in the Merry Widow deposit.

CHAPTER III - ENVIRONMENT OF ORE DEPOSITION

Confining pressure

An estimate of the confining pressure during skarnification and ore deposition can be deduced from the depth of intrusion of the adjacent pluton.

From a study of granitic plutons, particularly of their internal and external structures, Buddington (1959) defined three depth zones of granite emplacement. These zones and their approximate depth and temperature limits are:

(1) epizone - extends from the surface to six miles, commonly four miles. The maximum temperature of country rocks before intrusion is about 300°C . (2) mesozone - about four to ten miles in depth; a temperature range before intrusion of $250\text{-}500^{\circ}\text{C}$. (3) catazone - seven to twelve miles (or more); temperature range $500\text{-}700^{\circ}\text{C}$. Buddington placed the entire "Coast Range batholith" in the mesozone. However, this "batholith" is now considered an intrusive complex composed of widely different compositions and ages. The isolated granitic stocks and small batholiths on Vancouver, Texada, and Moresby Islands must each be assigned to a depth zone on its own merits. In

consequence, the writer will attempt to classify individual intrusions adjacent to magnetite deposits according to Buddington's criteria. Data have been used from the author's observations in mine areas and from information gathered by officers of the B.C. Department of Mines during regional mapping. Buddington discussed several criteria for each depth zone, but not all of these criteria carry equal weight. Those considered by the present writer to be fairly diagnostic and to be applicable in the present study are given below (paraphrased from Buddington, 1959):

Epizone

1. The plutons are largely or wholly discordant.
2. The plutons are closely connected with volcanic rock of the same age.
3. Country rock, outside of the contact metamorphic zone, is relatively unmetamorphosed.
4. Earlier members of the intrusions show chilled borders against country rocks.
5. Miagrolitic texture is common.
6. Granophyre is present.
7. The pluton is accompanied by explosion or intrusion breccia.
8. The intrusions are accompanied by contact metamorphism or metasomatism.
9. Deformation of country rock is in part caused by forceful intrusion of the magma.
10. The intrusive rocks are homophanous.

Mesozone

1. Metamorphism of the country rock has not exceeded the green schist or epidote-amphibolite facies.
2. There is a complex emplacement relationship between the intrusion and country rock. Contacts are in part concordant, in part discordant.

3. Foliation is common in the outer zone of the pluton.
4. Schistose structure is common in bordering country rock.

Catazone

1. Country rock is metamorphosed to granulite or orthogneiss.
2. There is an extensive migmatite zone bordering the pluton.
3. Gneissic structure is widespread within the intrusion.

In Table 9 these criteria are applied to plutons adjacent to magnetite deposits. The Coast Copper and Jedway stocks, the smallest, appear to be of the epizone type. Evidence is inconclusive for the Zeballos and Kennedy Lake intrusions but the Quinsam, Gillies, and Nimpkish intrusions have characteristics transitional between those of the epizone and mesozone. Consequently, a reasonable estimate of depth of intrusion of these seven plutons and their magnetite deposits is 4.5 miles. The lithostatic pressure at this depth is near 2000 bars.

An independent estimate of confining pressure for the Coast Copper stock during ore deposition is made possible by studying phase relations of coexisting minerals in the Merry Widow deposit. Pyrite, pyrrhotite, and arsenopyrite coexist with magnetite and other sulphides. Clark (1959/60) has shown that the composition of arsenopyrite is a function of temperature and confining pressure. If the temperature of

Table 9

Classification of seven intrusions into the three depth zones
of Buddington (1959)

		Quin- sam	Gillies	Coast Copper	Kennedy Lake	Zeb- allos	Nimp- kish	Jedway
Epizone character- istics	1.		x	x				x
	2.							
	3.	x	x	x				x
	4.							
	5.							x
	6.							
	7.			x?				
	8.			x				x
	9.			x			x	
	10.	x	x	x	x	x	x	x
Mesozone character- istics	1.			x	x	x	x	
	2.	x			x	x	x	
	3.							
	4.							
Catazone character- istics	1.							
	2.							
	3.							

formation of arsenopyrite can be estimated by independent means, then the arsenopyrite composition (as deduced by the (131) interplanar spacings) yields an estimate of confining pressure. An average temperature of 515°C was obtained from two samples of pyrite and pyrrhotite coexisting with arsenopyrite in this orebody (Table 10, nos. 6 and 7). The arsenopyrite has a (131) interplanar spacing of 1.6310 $\pm$

0.0005⁸Å. If the temperature derived from the pyrite-pyrrhotite relation is the temperature of formation of coexisting arsenopyrite, the confining pressure by Clark's method would be 2600 bars. Clark's curves for low (131) interplanar spacings in arsenopyrite are tentative so that limits of error may be ± 1000 bars (Clark, written communication, 1964).

In the discussions to follow, a confining pressure of 2000 bars will be used for the purpose of calculating limits of other variables.

Temperature of the pluton

The melting temperature of an igneous rock is a function of many variables. Two of the more important are rock composition and water vapour pressure. Smith (1963,p.201) calculated that, at constant confining pressure, the temperatures of complete melting of dry igneous rocks of the granite-gabbro series vary in a linear manner with the percentage of silica. The effect of pressure on dry igneous rocks is to raise slightly their melting temperatures (Smith,1963,p.202). When heated under water vapour pressure, however, rocks melt at considerably lower temperatures but it is reasonable to assume that minimum melting temperatures of "wet" rocks, at constant water

pressure, also will vary lineally with silica content.

Beginning-of-melting curves of granite (containing 72% SiO_2) and olivine tholeiite (containing 49% SiO_2) in the presence of water vapour are shown in Figure 9 (Yoder and Tilley, 1962).

These curves are very nearly parallel and the author has drawn an intermediate melting curve for a rock of intermediate composition (62% SiO_2). Six plutons adjacent to magnetite deposits (point 'X', Fig. 4) have an average SiO_2 content of 62.5%.

Their average beginning-of-melting curve would be expected to fall close to the intermediate melting curve shown in Figure 9.

Assuming this "average" magma to be saturated with water at the time of intrusion and that water was the major volatile constituent, then water pressure in the magma would equal confining pressure. This pressure has been estimated as about 2000 bars and, using the intermediate curve of Figure 9, rock of composition 'X' would have a minimum melting temperature of 780°C . Large variations in water vapour pressure do not

appreciably alter melting temperature. For example, if the pressure given above varies as much as ± 1000 bars, this will change the minimum melting temperature by only $\pm 40^\circ\text{C}$. The gabbro phase of the Coast Copper stock has a silica content of about 47% and was not included in calculating average 'X'.

Using the basalt curve in Figure 9, a minimum melting temperature of 880°C and 2000 bars water pressure is indicated

FIGURE 9

Beginning-of-melting curves of granite and olivine tholeiite (from Yoder and Tilley, 1962). Intermediate curve (dotted line) obtained by interpolation.

for such gabbro. Data from Yoder and Tilley (1962, pp. 449 et seq.) indicate that pyroxene in mafic rocks is stable only at water pressures lower than 1400 bars. At higher water pressures, amphibole is formed. Consequently, pyroxene gabbro of the Coast Copper stock may have formed at water pressures slightly less than 2000 bars.

Temperature of skarnification

It seems reasonable to conclude that the maximum temperature of skarnification is the temperature of the intruding magma. Other evidence, to be described below, is required to establish a minimum temperature.

Intrusions adjacent to magnetite deposits in coastal British Columbia are not in contact with pelitic "indicator" rocks which would provide a guide to the degree of metamorphism reached in the skarn contact aureoles. Silica-free limestone and andesitic volcanic rocks are the country rocks and, other than being recrystallized, have been only slightly affected by thermal metamorphism.

Considering their proximity to intrusions, however, contact skarns probably formed in the innermost zone of contact metamorphism, the temperature range of which

is largely determined by the melting temperature range of adjacent intrusions. The innermost zone of a granite contact aureole, the hornblende hornfels facies (Fyfe, Turner, and Verhoogen, 1958, pp. 205-211), approaches the lower limit of magmatic temperatures. The lower and upper temperature limits of the hornblende hornfels facies are about 300°C. and 500°C., respectively, under "dry" conditions. With six kilobars water pressure, these limits increase to about 500°C. and 700°C. (Ibid, p. 237). Contact aureoles surrounding gabbro stocks, on the other hand, are commonly of the pyroxene hornfels facies with temperature limits of about 500-1000°C. ("dry") and 700-1100°C. ($p_{H_2O} = 6$ kb.) (Ibid, pp. 211-213; 237). The "average" intrusion considered in this study is of quartz diorite composition (point 'X', Fig. 4). A contact aureole adjacent to such an intrusion would have a temperature range intermediate between that of hornblende hornfels and pyroxene hornfels facies. If an average confining pressure of 2000 bars is accepted (see preceding section), then skarn formed in this contact aureole at about 700°C. (Fyfe, Turner, and Verhoogen, 1958, Fig. 107, p. 237).

Experimentally-determined stability relations of the individual minerals comprising skarn will also aid in establishing temperature limits of skarnification. Garnet

of the andradite-grossularite series is the most abundant skarn mineral. Andradite and grossularite have been formed hydrothermally from their constituent oxides at 500°C and 400 atmospheres (Flint,McMurdie, and Wells, 1941). Attempts to synthesize these minerals at lower temperatures failed and experiments were apparently not made at temperatures higher than 500°C. Further work showed that a complete solid solution series exists between grossularite, hydrogrossularite, "hydroandradite", and andradite.

Pistorius and Kennedy (1960) showed that, in the presence of excess silica, grossularite is stable above 780°C and above 15 kilobars water pressure (Fig. 10a). In a silica-poor environment, however, the pressure of formation of all anhydrous garnets is lower (Fyfe,Turner,and Verhoogen, 1958,p.156). For example, Yoder (1950,p.245;1954,p.120) showed that grossularite without quartz is stable at 800°C under only 2000 bars water pressure (Fig. 10b).

A small mol percent of almandite is present in skarn garnets (Fig. 8) and at 2000 bars water pressure, the thermal stability limit of this mineral is 830°C; at atmospheric pressure almandite is not stable above 785°C. Above its stability range, this mineral breaks down to iron-cordierite, fayalite, and hercynite. The lower stability range of

FIGURE 100

Grossularite-hydrogrossularite stability relations in the presence of excess silica (from Pistorius and Kennedy, 1962)

FIGURE 10b

Stability relations in the system grossularite - H_2O
(from Yoder, 1950; 1955)

almandite is not known (Yoder, 1955). As neither hydrogarnets nor decomposition products of almandite and grossularite have been found in the skarns studied, the temperature of formation of pure grossularite at 2000 bars water pressure would be about 800°C. High P-T data are not available for andradite and these figures are tentatively extrapolated to include all members of the grandite series.

Grandite garnet in contact zones is generally birefringent. Several investigators (Wright, 1915; Stose and Glass, 1938; Allen and Fahey, 1957; Stone, 1959), finding that such garnet became isotropic upon heating to about 800°C., have suggested that birefringent garnet could be used as a crude geologic thermometer. They concluded that birefringent garnet had formed below, and isotropic garnet above, 800°C. The origin of birefringent grandite garnets has been discussed elsewhere in this paper and their use as geologic thermometers has been questioned.

Diopside is another common constituent of skarn and its upper limit of thermal stability is about 600°C. at atmospheric pressure. Above this temperature diopside reacts with calcite to form akermanite (Bowen, 1940). With increasing pressure, the equilibrium temperature of this reaction increases steadily to 980°C. at 1000 atmospheres (Weeks, 1956, p. 259, Fig. 4),

the highest P,T-conditions for which thermodynamic data are available. The absence of akermanite in the skarns studied places an extreme upper limit of 980°C. to the temperature of formation of these skarns.

Epidote is a widespread skarn mineral and is generally the last calcium-silicate to form. It predominates in the outer (and presumably cooler) fringes of zoned deposits. The approximate temperature of formation of epidote was estimated to be 400°C. (1 atm.) by Stringham (1952) based on published information available to 1951. Recently Merrin (1960) has extended the "apparent stability curve" of epidote to 700°C. at 5000 bars pressure. At 2000 bars, the upper stability limit of this mineral, as deduced from his data, is 550°C. This temperature, lower than that estimated for garnet and diopside at the same pressure, is in accord with both the position of epidote in skarn paragenesis and its peripheral abundance.

Wollastonite was found only in the Prescott deposit, Texada Island. The mineral occurred in a block of limestone surrounded by garnet skarn about 70 feet from quartz monzonite. The presence of wollastonite indicates the temperature at this place reached at least 480°C. This is the minimum temperature of formation of wollastonite from quartz and

calcite (Barth, 1952, pp. 286-287). With increasing carbon-dioxide pressure, the temperature of formation rises so that at 5000 atmospheres the equilibrium temperature is about 900°C. In contact aureoles, reactions which produce metasomatic rocks in limestone probably take place with rapid production of CO₂. A relatively high CO₂ pressure would raise the equilibrium temperature of the quartz-calcite reaction. Barth (1950, p. 287) has suggested, however, that, because CO₂ might be expected to percolate rapidly away through semi-permeable rock cover, pressure would have little effect on the stability temperature of wollastonite. Because of the uncertainty in estimating CO₂ pressure in metasomatic rocks, the presence of wollastonite can only indicate a minimum temperature of about 500°C. In the foregoing discussion the assumption is made that carbon dioxide is the only volatile present during metasomatism. The presence of other fluids, particularly water, and their effect on equilibrium temperatures of minerals will be discussed below.

Lack of evidence of thermal dissociation of calcite adjacent to intrusions is of little use in estimating the temperature of the contact aureoles in which skarn has formed. Less than 30 atmospheres CO₂ pressure prevents dissociation at 1100°C. (Harker and Tuttle, 1955). The equilibrium

temperature at one atmosphere is 885°C . (Birch et al., 1942; Weeks, 1956). In view of the rapid increase of dissociation temperature with small increase in CO_2 pressure, it is unlikely that calcite would dissociate in an aureole adjacent to a granitic pluton.

Temperatures of formation of minerals (and of mineral assemblages) from calculated and experimental equilibrium curves must be considered maximum values. Weeks (1956) pointed out that the vapour phase present during metamorphism and metasomatism of limestone is probably a complex mixture of water, carbon dioxide, and other constituents. The partial pressure of any component in solution cannot normally be greater than the pressure of the pure component under the same P,T-conditions. A mixture of two or more components in the vapour phase depresses the reaction temperature at any given pressure. The effect of binary gas mixtures on crystal-vapour equilibrium temperature has been discussed further by Greenwood (1961/62). He has shown that, in the presence of a vapour phase consisting of carbon dioxide and water at constant vapour pressure, the equilibrium temperature of reactions which release these fluids as products will vary with the composition of the vapour phase. The equilibrium temperature of reactions in which CO_2 is released will rise as the mol

fraction of CO_2 in the vapour phase increases; whereas in those reactions in which H_2O is released, this equilibrium temperature will decrease as the CO_2 content of the volatile phase increases. For example, with a vapour phase consisting of 0.25 mol fraction CO_2 and 0.75 mol fraction H_2O at 1000 bars, calcite and quartz are in equilibrium with wollastonite and the vapour phase at 580°C . If the composition of this vapour phase is changed to 0.75 mol fraction CO_2 and 0.25 mol fraction H_2O , the equilibrium temperature of the reaction rises to 650°C . Consideration of the stability of grandite garnet suggests a temperature of 800°C . for skarnification but in view of Weeks' and Greenwood's arguments, the temperature of 700°C , deduced from Fyfe, Turner and Verhoogen, seems more reasonable. In any case, skarn appears to have formed over a temperature range from about 700°C . (for garnet-diopside) to 550°C . (for epidote).

Temperature of formation of metallic minerals

Sulphide minerals, in general, were deposited contemporaneously with, or later than, magnetite. Temperature determinations based on coexisting sulphides, then, will give an estimate of the minimum temperature of magnetite deposition.

Pyrite and pyrrhotite coexist in several deposits and offer an opportunity to apply the pyrite-pyrrhotite geothermometer as described by Arnold (1962). By this method, the metal content of pyrrhotite coexisting in equilibrium with pyrite is related to the temperature and pressure of crystallization. No correction of crystallization temperature is necessary for pressures up to 2000 bars at temperatures below 670°C. This is the pressure estimated for deposition of the magnetite deposits considered so that corrections for pressure were not necessary. In applying this geothermometer, certain assumptions are made: (a) that pyrite and pyrrhotite coexisted in equilibrium (b) that the pyrrhotite composition has not changed significantly during subsequent geological events and (c) that small concentrations of elements in solid solution do not significantly affect the equilibrium relations. It is not essential that the two minerals be deposited simultaneously. If they were deposited at different times and were in a position to react with each other, it is assumed that equilibrium was attained at the conditions of deposition of the later phase and the temperature measured will be that at which the later phase crystallized (Arnold, 1962). In the specimens studied, pyrite and pyrrhotite are in contact with one another and it is assumed therefore that equilibrium was attained during deposition.

The composition of pyrrhotite from eight specimens was determined by an X-ray spacing method which relates $d(102)$ to composition (Arnold, 1962). The $d(102)$ values were determined using a North American Philips diffractometer with Fe-radiation. The (220) reflection of fluorite ($d = 1.931$; Swanson and Tatge, 1953) was used as standard. The diffractometer was set to oscillate three to five times over the region to be measured with a goniometer rotation of $\frac{1}{2}^\circ$ /minute and chart speed of one inch/degree. The centres of gravity of the standard and unknown reflections were measured to the nearest $1/30$ inch ($\pm 0.05^\circ 2\theta$) and the average of three to five measurements was used in calculating temperature. The uncertainty of the method is $\pm 50^\circ\text{C}$. Metal contents and estimated temperatures of crystallization of eight pyrrhotite specimens are given in Table 10.

These temperatures, derived from co-existing pyrite and pyrrhotite in four different contact metasomatic deposits and from different mineralogical zones, have a total difference of only 160°C . All of these temperatures are lower than the most probable temperature range of skarnification. Furthermore, they are in accord with field observations that magnetite and sulphides were formed later than skarn.

Table 10

Estimated crystallization temperatures of pyrrhotite coexisting with pyrite

<u>No.</u>	<u>Sample</u>	<u>Location</u>	<u>% metal</u>	<u>Temp. (°C.)</u>
1.	NP-5*	Paxton mine	47.1	380
2.	NP-1*	" "	47.1	380
3.	SP-351B	Brynnor mine	47.1	380
4.	L-71*	Lake mine	46.6	490
5.	1-103*	" "	46.5	510
6.	SP-282B	Merry Widow mine	46.5	510
7.	SP-282D	" " "	46.4	520
8.	SP-141A	Paxton mine	46.3	540

*Specimens kindly supplied by R. Patterson,
Geologist, Texada Mines Ltd.

Note:

1-3 - Specimens collected from Zone 3

4-5 - Locality unknown

6-7 - Zone 1, Merry Widow mine

8 - Thin sulphide layer between skarn (Zone 2) and
limestone (Zone 4).

(Zones correspond to those given in Table II)

In summary, the temperature of intrusion seems to have been 900-800°C., of skarnification 700-550°C., and deposition of metallic minerals 550-400°C.

Partial pressure of oxygen

O₂ pressure in the pluton

The partial pressure of oxygen in a rock melt is a function of total vapour pressure, temperature, and gas composition. If the silicate melt was in equilibrium with its volatile constituents at the time it crystallized, the pressure of O₂ must have been equal to the partial pressure of O₂ of the total volatiles. The partial pressure of oxygen in the melt determines the ratio of ferrous to ferric oxide at a given temperature. This oxygen pressure can be computed if three variables are known: (a) the temperature of the silicate melt at the time of crystallization (b) the relative amounts of Fe₂O₃ and FeO in the system and (c) the equilibrium constant (K) for the reaction

Relative amounts of ferric and ferrous oxide are determined by normal analytical methods and Kennedy (1948) has measured values of K at various temperatures for a melt of basaltic composition. The average of six specimens from plutons adjacent to contact magnetite deposits is shown by point 'X' in Figure 4. Rock of this composition at 2000 bars water pressure

begins to melt at $780 \pm 80^{\circ}\text{C.}$ and contains 2.9 mol percent FeO and 0.73 mol percent Fe_2O_3 . Assuming a melting temperature of 800°C. for purposes of calculation the dissociation constant (k) is 302×10^{-9} (Kennedy, 1948, p. 540). Under these conditions, the partial pressure of oxygen in melt of composition 'X' is 21.4×10^{-12} atmosphere. Using values of 10.5 mol percent FeO, 1.92 mol percent Fe_2O_3 , $K = 138 \times 10^{-7}$, melting temperature = 900°C. , the Coast Copper gabbro cooled in equilibrium with a partial oxygen pressure of 41.9×10^{-12} atmosphere. Oxygen pressures such as those given above are many orders of magnitude less than the partial pressure of oxygen produced by the dissociation of water. For example, at 900°C. , the partial oxygen pressure is 3.62×10^{-4} atmosphere (Kennedy, 1948, p. 545).

O₂ pressure in the ore-forming fluids

The calculation of partial pressures of gases in equilibrium with ore fluids has been discussed by several authors including Holland (1959), Krauskopf (1959), and Barnes and Kullerud (1957a). Such calculations generally involve determination of equilibrium pressures for various limiting reactions at various temperatures.

During the formation of magnetite deposits, the upper limit of oxygen pressure is defined by the reaction:

An increase in O_2 pressure above the equilibrium pressure will result in formation of hematite. The extreme lower limit of oxygen pressure in magnetite deposits is governed by the reaction:

At constant O_2 pressure, increasing temperature results in reduction of iron oxides. Consequently, if the temperature of magnetite deposition can be estimated by independent means, the absence of native iron or hematite in a deposit will define the maximum possible range of O_2 pressure at that temperature. The fields of stability of iron oxides have been investigated by Holland (1959) and from his graphs the oxygen pressure during deposition of contact magnetite deposits can be estimated. The temperature of deposition of ore in these deposits has been estimated to fall in the range 400-550°C. (Table 10). This temperature range, together with the absence of hematite or native iron in magnetite deposits, permits a possible oxygen pressure range of 10^{-17} to 10^{-34} atmosphere in equilibrium with magnetite. This

wide range in oxygen pressure may explain the widespread occurrence of magnetite as the main iron-bearing mineral in contact deposits. Partial pressures of oxygen in the ore fluids are considerably lower than those in magmatic fluids in equilibrium with an adjacent pluton.

In those deposits containing coexisting pyrite, pyrrhotite, and magnetite, the oxygen pressure is fixed, at a given temperature, by equilibrium in the reactions

and

At 400°C., the equilibrium oxygen pressure is 10^{-27} atmosphere; at 550°C., it is 10^{-18} atmosphere. These pressures fall within the maximum possible range determined for magnetite in the absence of native iron or hematite. Coexistence of the three phases magnetite, pyrite, pyrrhotite, narrows the range of possible oxygen pressure in equilibrium with these minerals during ore formation.

Partial pressure of sulphur

As most contact magnetite deposits commonly contain pyrite, the minimum sulphur pressure in such deposits, in the temperature range 400-550°C., must have been 10^{-8} to 10^{-3} atmosphere (Holland, 1948, p.189). Lower sulphur pressures would result in formation of pyrrhotite according to the reaction

For those deposits that do contain coexisting pyrite and pyrrhotite, the range of sulphur pressures is limited to the equilibrium pressures of the pyrite-pyrrhotite reaction. At 400°C., this equilibrium pressure is 10^{-8} atmosphere; at 550°C. it is 10^{-3} atmosphere.

CHAPTER IV - GENESIS

Origin of iron compounds

The iron of contact metasomatic deposits might have any one or a combination of the following sources:

- 1) Ferriferous meteoric waters heated by an invading magma.
- 2) Bonanza limestone.
- 3) Vancouver group volcanic rocks.
- 4) Nearby granitic plutons.

Meteoric waters Under certain conditions groundwaters are capable of transporting and depositing iron compounds and silica. Should such waters be heated by an invading magma, the contained iron and silica conceivably could be fixed as magnetite and skarn. However, as contact metasomatic deposits are associated only with intrusions of intermediate composition, and not highly acidic or ultrabasic ones, such a mechanism does not seem valid.

Quatsino limestone Iron might have been leached in some unknown way from the Quatsino limestone so prevalent with contact metasomatic deposits of the Pacific

coast. Inasmuch as such limestone has very little iron (0.06%; Table 14), prohibitive volumes of rock would be involved.

Vancouver group volcanic rocks

Most of the deposits of the Pacific coast replace either Karmutsen or Bonanza volcanic rock which may contain up to 6-8% Fe as an original constituent. Such iron would be a source of iron for contact metasomatic deposits if some mechanism of "lateral secretion" could be invoked. Boyle (1959) suggested lateral secretion of iron and silica from surrounding metasediments took place during metamorphism in the Yellowknife area, N.W.T. These elements, among others, were then deposited as veins in dilatant zones. Mineralization was shown to be accompanied by widespread alteration extending outward from the mineralized areas for many hundreds of feet. Hagner et al. (1963) concluded that magnetite of the Scott mine, N.Y. could have been derived from amphibolite host rock by metasomatism, regional metamorphism, and granitization. Highly dispersed ferriferous material was considered by the authors to have been released from amphibolite and later concentrated into orebodies in low pressure zones.

The writer believes that the Karmutsen and/or Bonanza volcanic rocks are not the immediate source of iron in contact metasomatic deposits for the following reasons:

1) These rocks have not been extensively metamorphosed, altered, or granitized.

2) There is no evidence of low pressure zones such as described by Boyle and Hagner as favourable sites of ore deposition.

3) Material moving from the volcanic rocks to form an orebody adjacent to an intrusion would be migrating against a rising temperature gradient. Furthermore, if the intrusion were discharging water, carbon dioxide, and other volatiles from its margins, material migrating toward the pluton would be moving also against a rising pressure gradient.

4) Barnes (1959) showed qualitatively that it is impossible for zinc of the Hanover, N.M. contact metasomatic base metal deposits to have been derived from surrounding sediments by lateral secretion. Nearby unaltered Devonian shale, containing sufficient zinc to contribute significantly to the metal content of these deposits, has the same metal content as its recrystallized and silicified equivalent in the contact metamorphic aureole (excluding skarn). Barnes' evidence indicates that metals do not migrate toward a contact aureole.

Nearby granitic plutons

Clapp (1913) suggested that several small contact magnetite bodies in

southern Vancouver Island originated in adjacent plutons as a magnetite "magma". Magmatic magnetite deposits are known in several localities. The iron ore of Kiruna, Sweden, has a high phosphorous content and a distinctive texture suggesting a eutectic intergrowth of apatite and magnetite (Lindgren, 1933,p.792). Titanium-rich magnetite bodies are common in basic and ultrabasic rocks of the Grenville sub-province of the Canadian Shield. Pure magnetite melts at 1600°C although some reduction of melting temperature could be caused by a small percentage of volatiles. Even at slightly reduced temperatures, however, molten magnetite would crystallize at much higher temperatures than those calculated from coexisting pyrite and pyrrhotite (Table 10). Consequently, magnetite of contact metasomatic deposits does not appear to have been deposited as a "magma".

On the other hand, selective replacement along favourable horizons, pseudomorphs of magnetite after garnet, gradational contacts between skarn and unaltered host rocks, and the abundance of hydroxyl-bearing minerals such as epidote, sericite, and chlorite suggest emplacement of magnetite and skarn by hydrothermal fluids emanating from a nearby granitic pluton.

Margins of several plutons in contact with Karmutsen and Bonanza rocks are relatively enriched in iron which could be carried by outward-migrating solutions and redeposited as an orebody. Hamilton et al. (1964) have demonstrated experimentally that in a basalt melt under a partial pressure of oxygen equal to that calculated for intrusions in the present study, iron oxide crystallizes at a lower temperature than coexisting plagioclase and pyroxene. Thus, iron would be concentrated residually for later transport in a hydrothermal fluid.

In a series of experiments, Ovchinnikov (1959) proved that iron can be separated from a silicate melt which is assimilating limestone. It was found that, by fusing granite or basalt with limestone and cooling the melt gradually, iron separated from the silicate melt and was transported upwards. Depending on the partial oxygen pressure, metallic iron, magnetite, or hematite was formed. Ovchinnikov's experiments illustrate a possible mechanism whereby an iron-rich segregation might be concentrated in the marginal zone of a pluton.

A simple radioactivity survey was made by the author at the Iron Hill mine. A Precision Radiation Instrument, Inc. Scintillometer, Model 111B was used. Readings were taken

every 20-25 feet on all levels of the open pit. At each station the instrument was placed on the bench face and the radioactivity read in milliroentgens per hour. Eleven readings were taken on fresh granodiorite, 19 on fresh andesite, 21 on limestone, and 47 on ore. It was found that granodiorite gave the highest readings, ore gave intermediate, and limestone and andesite gave the lowest readings. The range, in mr/hr for each rock type, and the average (in brackets) are as follows: granodiorite 2.5-6.5 (4.2), ore 1.0-6.0 (3.0), andesite 1.5-3.0 (2.1), and limestone 1.0-3.0 (1.9). As no consistent difference could be detected in readings taken on massive magnetite from those on massive garnet skarn, the two sets were combined. Neither was there a detectable difference in ore formed by replacement of andesite from that replacing limestone. The intermediate readings found in ore might be interpreted as evidence that radioactive tracer elements accompanied iron and silica in hydrothermal emanations from the Quinsam intrusion.

The Prescott and Paxton mines, in particular, have a crude concentric zoning around a central ore zone (Maps 2 and 4). These centres are on the country rock side of the intrusive contact and the pluton itself actually lies within peripheral zones. Such zoning is expressed by a series of concentric circles about a centre slightly removed from the

granitic contact. This arrangement suggests that the mineralizing solutions did not migrate "horizontally" to the ore zone from the adjacent pluton but rather along the intrusive contact. When a favourable ore-forming environment was reached, deposition took place in the Vancouver group host rocks and zoning developed around the site of initial deposition. Analyzed specimens of the Coast Copper stock adjacent to the Merry Widow deposit show no statistical change in iron content from the interior of the intrusion toward its margin (Fig. 5 and 6). This evidence also supports the theory that iron did not migrate horizontally from the intrusion.

The writer concludes that the immediate source of iron in contact metasomatic magnetite deposits is a nearby intrusion. However, the ultimate source is very probably the Karmutsen volcanic rocks. Granitic plutons, intruding and in part assimilating Karmutsen rocks, thereby acquired additional iron in their margins. This iron subsequently was fixed as magnetite deposits in and near Quatsino limestone.

Composition of the ore-forming fluids

Statements regarding composition of ore fluids must be of a qualitative, or at best semi-quantitative, nature---

a limitation imposed by our lack of knowledge of important variables which determine composition of such fluids. Mineral assemblages in contact metasomatic deposits are a result of interaction of several variables in a medium which is very likely above its critical point. The relation between composition of ore minerals and ore solutions is complex. At least four independent variables (pressure, temperature, ionic species, and activity coefficients) must be qualitatively evaluated in order to deduce the composition of an ore solution from the composition of precipitated minerals. Even if these variables were known, however, application of empirical functions is limited to the cases in which equilibrium during precipitation and absence of post-depositional changes can be demonstrated (Barnes and Kullerud, 1957a). Under certain conditions the distribution of minor elements in vein minerals could be used to define the temperature, pressure, and chemical composition of the fluid from which precipitation took place (Holland, 1956).

Because many of the variables mentioned above are unknown in the deposits studied, the following discussion will be limited to an outline of the most probable composition and physical nature of the fluid which transported iron and deposited magnetite in contact metasomatic deposits.

If a magma has been in place long enough to reach diffusion equilibrium, the partial pressure of water will be the same throughout the melt. Kennedy (1955) concluded that this could be true only if the actual quantity of water is greater in cooler parts of the melt and in those parts having the lowest confining pressure. These conditions would exist in margins or upper parts of the magma chamber i.e. the cupolas and apophyses of a pluton. As water migrates toward the margins of a magma chamber, it will be accompanied by other volatile material. Osborne (1959) has shown that rocks of the calc-alkali trend, such as those of plutons adjacent to the deposits studied, are produced by fractional crystallization at constant partial oxygen pressure. Osborne states that as water dissociates in the melt, the partial oxygen pressure will either remain constant or increase slightly. The partial O_2 pressure for melts of two compositions has been calculated as 21.4×10^{-12} atmosphere and 41.9×10^{-12} atmosphere (p.99); hence, plutons adjacent to magnetite deposits of this study crystallized under almost identical oxygen pressures and, according to Osborne, this is possible in the presence of abundant water. At $1100^{\circ}C$ the solubility of water in basalt at 2000 bars is 4.6 weight percent; in andesite at the same P,T-conditions the solubility is 6.0 weight percent (Hamilton et al., 1964). There is a small but important variation in solubility with

temperature — the increase in solubility at constant pressure amounts to roughly 0.5 weight percent for each 100°C decline in temperature (Kennedy, 1955, p. 491). Andesite at 800°C therefore would contain 7.5 weight percent water. It is apparent from these figures that abundant water is available in the plutons associated with magnetite deposits and that such water, migrating toward the margins, could transport and later deposit metal compounds as peripheral orebodies.

As the critical temperature of water is 374°C at a pressure of 220 atmospheres, pure water would be in the supercritical state at the P,T-conditions estimated for ore formation. Ore-forming fluids, however, would be a complex solution containing many solutes and solvents, each of which would affect the P-V-T relations of the system. During contact metasomatism near limestone, for example, CO₂ would be a dominant constituent and its principal effect would be to lower the critical temperature of water. On the other hand, solutions of metal chlorides (the principal compounds found as fluid inclusions in vein minerals) would raise the critical point of such solutions above that of pure water. Because salinity and CO₂/H₂O ratios vary in an unknown manner, it can only be assumed that the P-V-T relations of aqueous ore-forming fluids are similar to those of pure water. At 850°C and

2000 bars pressure (the P,T-conditions during intrusion) the density of an aqueous ore fluid would be less than half that of water under standard conditions (Smith, 1963, p. 80).

Field observations of contact deposits in limestone show that large volumes of CaCO_3 have been replaced by magnetite and skarn. Because thermal dissociation of calcite is unlikely under the conditions of intrusion, removal of carbonate must have been effected by solution. This would be possible only if the pH of the ore-bearing solutions was less than 7.8 (the minimum stability pH of calcite). It may possibly have been below 7 as the solubility of calcite in pure water at 500°C and 1000 bars is only 120 ppm (Morey, 1957). Evidence supporting acidic ore fluids is the presence of sericitic alteration of plutons adjacent to deposits. At temperatures above 350°C , sericite is stable only in an acid environment (Stringham, 1952). A small sericite stability field at lower temperature and slightly alkaline conditions might indicate that sericite in the plutons did not form at the time of ore deposition but rather during a later period at lower temperature and higher pH. Such conditions are unlikely, however, because zeolite, which occurs as post-sericite veins in several plutons (such as near the Klaanch, Iron Hill (Plate I), and F.L. deposits), is stable only below $350\text{--}400^\circ\text{C}$.

(Stringham, 1952). Assuming temperature declined as mineral deposition progressed, sericite must have been deposited above 400°C from acidic solutions.

At 25°C the pH of pure water is 7.0. Under pressure, however, the pH of a neutral solution is slightly less than 7 and if it contains a dissolved salt its neutral point is slightly greater than 7 (Barton, 1959). The effects on pH of pressure and salt are small, however, when compared to that of temperature. With increasing temperature, the pH of neutrality decreases steadily to a minimum of 5.6 at 250°C . At the critical temperature of water, the pH of neutrality has risen to about 7.0. In the supercritical state the effect of temperature on pH is much greater and, at 500°C and 2000 bars, the pH of neutrality is close to 5.0 (Barnes and Kullerud, 1961). It is difficult to predict the effect of temperature on the pH of complex solutions such as ore fluids but it is assumed their pH approaches that of water under similar P,T-conditions.

Several lines of evidence indicate that iron was carried as chloride. Surface incrustations at most high-temperature vents in the Katmai area contain abundant magnetite as well as chlorides and fluorides. Although Zeis (1929) concluded that fumarolic magnetite and hematite formed from iron halide vapour, simple volatility of iron chlorides

cannot contribute substantial amounts of iron to dilute ore solutions (Krauskopf, 1964). At 500°C ferric chloride is a liquid with a vapour pressure of only 9.7 bars. The vapour is composed of 98% Fe_2Cl_6 gas and 2% FeCl_3 gas (Holser and Schneer, 1961, p. 378). At this temperature a small amount of ferric chloride dissociates to crystalline FeCl_2 and gaseous Cl_2 . The vapour pressure of FeCl_2 at 500°C is less than 10^{-5} bar (Holser and Schneer, 1961, p. 378). Although Fe_2Cl_6 is the most volatile of the iron chlorides, its concentration in the vapour phase is restricted by the presence of water with which it reacts to form ferric oxide and hydrogen chloride. Holser and Schneer calculated that, at 500°C and 500 bars, the concentration of HCl necessary to mobilize 1 ppm Fe as Fe_2Cl_6 is 2.7N. Concentrations of HCl in volcanic emanations are commonly of the order 0.01-0.1N and Krauskopf (1957, p. 793) estimated 0.06N HCl for magmatic vapour at 600°C. Consequently, the high concentrations of HCl necessary to mobilize Fe_2Cl_6 gas are unlikely in natural fluids. Probably solubility of iron chlorides in an ore-forming fluid is more important than volatility because the density of such fluids, under the P,T-conditions of ore formation, would be about 0.8 and consequently are more liquid-like than gas-like. At 100°C the solubility of Fe_2Cl_6 is 5350 g/litre of water and of FeCl_2 is 1050 g/litre of water. In comparison, NaCl at the

same conditions has a solubility of 390 g/l (Hodgman, 1958). In a theoretical and experimental study of the solubility of magnetite in dilute aqueous solutions of HCl at high temperature and pressure, Holser and Schneer (1961) concluded that iron mobilized in their experiments was in true solution and that volatility of iron halides did not account for any significant mobilization of iron. In addition to simple chlorides, iron can be carried as complex ions. For example, sodium, a common constituent of fluid inclusions in ore minerals (Smith, 1954; Roedder, Ingram, and Hall, 1963) forms complexes such as $\text{Na}(\text{FeCl}_4)$ or $\text{Na}_2(\text{FeCl}_4)$ (Kalnin, 1962).

When an ore fluid reaches a site of deposition its chemical and physical nature changes. The temperature of this supercritical fluid has decreased from 800-900°C to about 400-550°C (Table 10) and its density has increased to about 0.8. In this temperature and density range, most of the iron chloride is carried in true solution and some ferric chloride dissociates to ferrous chloride and chlorine. Consequently, solutions reaching a site of ore formation contain both ferrous and ferric chloride.

Zeis (1929) suggested the formation of magnetite in the Katmai area was brought about by the reaction:

For the complex ions, the following reaction has been suggested (Kalnin, 1962):

In the Katmai fumaroles, HCl would be swept away as a gas, thereby permitting reaction (1) to proceed to the right with deposition of magnetite. In contact metasomatic deposits near limestone the HCl would be neutralized by calcite, permitting precipitation of magnetite.

Magnetite has been synthesized from ferrous chloride and calcite at 100°C and atmospheric pressure (Kuhara, 1925) and at high temperatures and low pressures (Vinogradov and Dontsova, 1952). Only recently have qualitative experiments confirmed that magnetite can be deposited by neutralization of iron chloride solutions at geologically significant temperatures and pressures. Holser and Schner (1961) prepared iron chloride solutions of 1000 ppm ferric iron and 500 ppm ferrous iron with a pH of about 2. This solution was allowed to react both with crushed limestone and a calcite cleavage rhomb at 320°C and 1100 bars. In both cases the product was crystalline magnetite. After reactions were

completed the solutions contained less than 20 ppm Fe^{+3} and less than 200 ppm Fe^{+2} and had a pH slightly greater than 5 (Holser and Schneer, 1961, p. 376).

In the absence of a strong acid, it is unlikely that much solution of calcite would take place. In carbon dioxide solutions, the solubility of calcite decreases steadily with increasing temperature. At about 374°C the solubility is less than 10^{-5} m for partial pressures of CO_2 of 1-100 atmospheres. The solubility might increase slightly above 400°C but experimental evidence is inconclusive (Ellis, 1959). The solubility of CO_2 in water is considerable — about 20 weight percent at 400°C and 2300 bars (Smith, 1963, p. 330). Carbon dioxide in excess of that dissolved in an ore-forming fluid would exist as gaseous CO_2 . Because of the lack of CO_2 -bearing minerals in the deposits, the partial pressure of CO_2 can only be estimated within wide limits. The equilibrium pressure at 885°C (the approximate temperature of intrusion) for the reaction $\text{CaCO}_3 \rightleftharpoons \text{CaO} + \text{CO}_2$ is one atmosphere (Birch, 1942; Weeks, 1956). At 500°C , calcite is stable to at least a partial CO_2 pressure of 10^{-3} atmosphere (Holland, 1959, p. 218). During ore formation, however, solution of limestone by ore-forming fluids would release CO_2 and increase the partial pressure of carbon dioxide. An upper limit to the possible CO_2 pressure during ore deposition can be estimated by the

absence of siderite in metasomatic magnetite deposits. This mineral is relatively unstable at elevated temperatures and at 327°C, the highest temperature for which thermodynamic data are available, slightly more than 100 atmospheres CO₂ pressure are necessary to enter the siderite stability field (Holland, 1959,p.199). At 500°C the necessary CO₂ pressure could well be over 1000 atmospheres. Consequently, the absence of siderite in these relatively high temperature deposits may be attributed to a partial pressure of CO₂ less than about 1000 atmospheres.

The supposed oxidation of magnetite by CO₂ (Butler, 1923) has been shown to be extremely unlikely between 200-500°C (Hawley and Robinson, 1948). Between these temperatures the ratio CO:CO₂ at equilibrium in the reaction $2\text{Fe}_3\text{O}_4 + \text{CO}_2 \rightleftharpoons 3\text{Fe}_2\text{O}_3 + \text{CO}$ ranges from about 1:10¹² to 1:10⁷. Hawley and Robinson consider these ratios very unlikely in contact deposits. The absence of native iron in magnetite deposits on the other hand, places the upper limit of the CO:CO₂ ratio at about 2:3 (Lasky, 1931).

Fluids in equilibrium with crystalline magnetite would be probably nearly neutral or slightly alkaline supercritical aqueous solutions. Supercritical carbon dioxide would be present but mainly as a mixture rather than in true

solution (Smith, 1963, p. 81). For the assemblage magnetite, pyrite, pyrrhotite, plus vapour between 500-550°C, the equilibrium partial pressure of oxygen is 10^{-18} to 10^{-27} atmosphere and of sulphur is 10^{-8} to 10^{-3} atmosphere (Chap. III). The partial pressure of other volatiles in the vapour phase is unknown but Barnes and Kullerud (1961) have calculated that, for the conditions stated above, the vapour pressure of a mixture of S_2 , S_4 , S_6 , S_8 , SO_3 , and O_2 is between 10^{-5} and 10^{-1} atmosphere. As the partial pressure of S_2 alone is 10^{-8} to 10^{-3} atmosphere, divalent sulphur is far more abundant than other species of sulphur.

The foregoing discussion indicates that great changes in the physical and chemical nature of an ore-forming fluid must take place during the process of ore deposition. While still in the magma chamber, the fluid was at a temperature about 800-900°C with a partial oxygen pressure of about 10^{-12} atmosphere. Total pressure on the system was about 2000 bars and, because water is assumed to be the major volatile at this stage, this would approximate the partial pressure of H_2O . As it left the magma the fluid had a density of about 0.4, a pH less than 7.8, and carried substantial amounts of iron and lesser amounts of silica, magnesium, and aluminum necessary to form skarn in limestone. Where skarnification took place

in volcanic rocks, these host rocks no doubt contributed such elements to the fluid system.

At a later stage, the temperature of the ore fluids declined to about 400-550°C. The silica content decreased relative to iron; the partial oxygen pressure decreased to within limits of 10^{-17} to 10^{-34} atmosphere; and magnetite was precipitated as a stable phase replacing limestone and volcanic rocks as well as earlier skarn.

In the final stage of sulphide deposition, the partial pressure of sulphur increased to 10^{-8} to 10^{-3} atmosphere, allowing iron and copper-iron sulphides to form. The partial pressure of oxygen had dropped to a minimum range of 10^{-18} to 10^{-27} atmosphere. The pH of the fluid had now become nearly neutral mainly by reaction with limestone. Removal of Si, Al, K, Na, Ca, CO_2 , Cl_2 , etc. was facilitated by an increase of fluid density to about 0.8.

Ore Controls

Recognized ore controls of contact metasomatic magnetite deposits can be divided into two main types: 1) physical ore controls and 2) chemical ore controls. Some have both physical and chemical characteristics and may be

regarded as physico-chemical controls.

Physical ore controls

No contact metasomatic iron ore deposits occur in the Upper Paleozoic limestone and volcanic rocks of Vancouver and Texada Islands. Perhaps this is merely because the exposed area of such rocks is much less than the exposed area of Upper Triassic rocks (50 sq. mi. of U. Paleozoic vs. 600 sq. mi. of Vancouver group; Canada, Geol. Survey Map 932A, 1962; Mathews and McCammon, 1957, Fig. 3). On the other hand, some chemical factor may favour deposition of contact metasomatic iron ore deposits in Upper Triassic rocks. One possible chemical factor is discussed in the section "Chemical ore controls" below.

All magnetite orebodies studied are found in folded rocks, either on a regional or local scale. Broad, open folds are characteristic of Triassic rocks on Vancouver Island. Superimposed on these are small, generally closed, local folds which increase in intensity toward plutons and are probably the result of forceful intrusion. Of the 12 deposits studied, 10 are in synclines and the other two may be in anticlines. During folding of the Vancouver group, volcanic rocks which were locally brecciated provided more open space for ingress

of ore-forming fluids. Shattering of these volcanic rocks is also a physico-chemical control whereby replacement of such rocks was enhanced by increased surface area. Limestone, on the other hand, flowed during folding and remained impermeable to ore-forming fluids.

Both stratigraphic and igneous contacts are other important ore controls. In the Paxton deposit, for example, the best ore is along the contact of Gillies intrusion and country rocks. From this contact skarn and magnetite extend along the contact of Marble Bay limestone and Texada group andesite. The Jessie and Merry Widow deposits are tabular and parallel to an intrusive contact.

For those deposits which are not directly in contact with an intrusion, pre-ore faults were probably channelways along which mineralizing fluids travelled to sites of deposition. The Kingfisher orebodies, for example, are at the intersection of a steeply-dipping normal fault and a gently-dipping thrust fault. There is some evidence that the Yellow Kid and Tassoo orebodies also were controlled, at least in part, by pre-ore faults.

The local configuration of a contact between country rock and pluton is an important control; irregular tongues or cupolas of intrusion projecting into country rocks are favour-

able sites of ore deposition. At the Klaanch deposit, a cupola of quartz monzonite projects westward into limestone and andesite and magnetite has replaced these rocks near the nose of the projection. Vancouver group rocks partly surrounded by granitic intrusions are favourable sites of ore deposition at the F.L., Iron Hill, and Brynnor mines.

Although all contact metasomatic deposits are related to igneous contacts, they may range up to several hundred feet distant from the actual exposed contact (i.e. Tassoo, Coast Copper, Kingfisher, Lake, and Brynnor deposits). Other controls are required to localize ore deposits.

Selective replacement of favourable rock units is a relatively minor ore control in the majority of the deposits studied. All three units of the Vancouver group are themselves relatively homogeneous and lack individual beds favourable to selective replacement. A large part of the Coast Copper orebody has replaced what is thought to be a narrow limestone bed in otherwise massive volcanic rock (Eastwood and Merrett, 1961). As there is no conclusive evidence of limestone in the mine, ore conceivably could have replaced a volcanic bed which possessed favourable physical or chemical properties. The Jessie and Tassoo deposits which are roughly conformable tabular bodies in volcanic rocks may

be selective replacements of former units of unknown composition.

Chemical ore controls

The presence of limestone is the most important chemical factor of deposition of magnetite in contact metasomatic deposits. It is a necessary constituent for two reasons: 1) the separation and concentration of iron in an adjacent silicate melt and 2) the neutralization of iron-bearing solutions.

Experiments by Ovchinnikov (1959) have demonstrated that iron, among other metals, separates from a basaltic or granitic melt fused with limestone. From a study of silicate melts in general, Ovchinnikov concluded that assimilation of limestone results in coalescence of iron atoms dissolved in the melt. Existence of atomic, as well as ionic, particles in a silicate melt has been demonstrated by several investigators (see discussion in Ovchinnikov, 1960). The particles of iron coalesce by assimilation of limestone, are carried to the upper parts of the melt and, depending on the partial oxygen pressure, form crystals of either metallic iron, magnetite, or hematite (Ovchinnikov, 1959). Dissociation of limestone liberates gas bubbles which carry iron particles

upward in a process analogous to froth flotation.

Ovchinnikov (1960) found that when basalt and limestone were fused, the basalt became pale green at its contact with a piece of limestone. Analysis of this green zone showed that it was impoverished in silica, alumina, iron, magnesium, sodium, and potassium. Calcium, on the other hand, increased more than two-fold in the green zone. These chemical changes are almost identical to those found across the contact between Gillies quartz monzonite and limestone (Table 5). Moreover, magnetite formed in the fusion experiments contained no titanium, but copper increased twenty-fold. The experimentally-produced relative concentrations of these elements are very similar to their relative concentrations in contact metasomatic magnetite orebodies.

Iron which has been separated from a silicate melt is carried toward the margins by gas bubbles where it comes in contact with aqueous solutions. These can then dissolve the iron and carry it to the country rock as chlorides. As previously discussed, these solutions have a pH less than 7.8 and are probably acidic. By neutralizing these solutions, limestone promotes deposition of magnetite. As more carbonate is dissolved, the pH rises and the stability of magnetite increases. The reactions which involve complex ions

(equation (2)), magnetite is similarly stabilized by the presence of excess carbonate. For those deposits in volcanic rocks, neutralization is brought about in a similar manner (e.g. orebodies such as Iron Hill, F.L., and Klaanch, although in contact with limestone, have largely replaced brecciated volcanic rock). If kept in contact with limestone through fissures or intergranular boundaries, ore fluids which fill fractures in, or replace, andesite will be neutralized by diffusion. An analogy is a long acid-filled tube in contact with an unlimited supply of carbonate at one end. The solution is neutralized first at the end nearest the carbonate but in time will become neutral throughout. In nature, the process of neutralization of fluids remote from a limestone would be aided by the "stirring" action of CO_2 migrating through the rock. Precipitation would be most rapid in limestone and slowest in volcanic rocks. Rapid precipitation has been suggested previously as a contributing cause of birefringence in grandite garnets, the majority of which are found in limestone (pp.60-62).

In the absence of limestone, iron chloride-bearing ore fluids may be partly neutralized by alumino- or magnesio-silicates or by carbonate-bearing volcanic rocks. Small deposits of disseminated magnetite, such as are found scattered throughout Karmutsen group volcanic rocks on Vancouver Island, may have formed in this manner.

Mention has been made of the occurrence of magnetite orebodies exclusively in the Vancouver group (p.122). This may have, however, a chemical explanation. In order for a pluton to be in contact with Quatsino limestone and to form magnetite deposits, it first must pass through the Karmutsen group. This unit is at least 10,000 feet thick (Mathews and McCammon, 1957, p.35) and contains about 8% iron (Table 13). Consequently, an intrusion forcing its way through such rock would have ample opportunity to become at least locally enriched in iron, particularly in its margins. Such thick sequences of iron-rich volcanic rocks are lacking among Paleozoic rocks and, in consequence, an intrusion in Permian limestone might not have assimilated sufficient excess iron to produce a magnetite deposit.

Hydrothermal processes

Metasomatism is debated at length in several excellent papers by Lindgren (1925), Duffell (1937), Holser (1947), Garrels and Dreyer (1952), and Hanus (1960) among others. The present author will limit his discussion to the relative importance of the various hydrothermal processes leading to formation of magnetite deposits.

Metasomatism is essentially a process of practically simultaneous capillary solution of a pre-existing mineral or rock and the deposition of new material in its place (Lindgren, 1933,p.91). This implies that metasomatism proceeds in a solid with no supercapillary openings. The volume formerly occupied by the replaced mineral is exactly filled by the metasome. As pointed out by Garrels and Dreyer (1952) and Hanus (1960), however, the existence of large cavities in metasomatic ore-bodies has been noted by several authors. In the present study, large partly-filled vugs were found in skarn at Iron Hill and in magnetite of the Yellow Kid and Shamrock-Blackjack deposits (Plate VI). In coarse-grained skarn, magnetite or calcite-quartz commonly fills former open spaces between euhedral garnet crystals (Plates XX, XXI, and XXII). Open-space filling is not included in the traditional definition of metasomatism (Lindgren,1933); even the existence of open spaces is not accepted. If open spaces do occur and are filled by later minerals, then such minerals cannot be considered metasomatic.

Large cavities can form during hydrothermal deposition when

$$\frac{\text{volume of metasome precipitated}}{\text{volume of host dissolved}}$$

$$< 1$$

Replacement will stop when this ratio is greater than one (Garrels and Dreyer, 1952). At unity, volume of metasome exactly equals volume of host dissolved and metasomatism, in its traditional sense, has occurred. Recognizing that some so-called metasomatic rocks only partly fill the volume formerly occupied by their host, Hanus (1960, p. 73) proposed the term solvate apposition - "the process of formation of solution cavities of supercapillary size and their simultaneous but only partial filling with new minerals. It is a process capable of forming more free space than it can simultaneously fill with newly-formed minerals". Solvate apposition, metasomatism, and cavity filling are considered to be, in order of decreasing importance, the main hydrothermal processes of formation of the contact metasomatic deposits studied.

In brecciated volcanic rocks, narrow veinlets of magnetite with roughly matching walls and sharp contacts (Plate XII) indicate that cavity filling was probably the first process of emplacement of skarn and magnetite in such rocks. As these cavities were filled, replacement became the dominant process in both volcanic rocks and limestone. Active flow of an ore-forming fluid through rock takes place only in openings of supercapillary size. In smaller openings

occupied by such fluid, movement is mainly by diffusion of ions. Experimental evidence of Garrels and Dreyer (1952) indicates that maximum amounts of material can diffuse through rocks only under conditions approximating steady-state diffusion, such as would be expected in nature.

Under the physical and chemical conditions of metasomatism, mineral reactions are assumed to result in equilibrium among the phases produced. Existing criteria of chemical equilibrium in rocks are based on negative evidence: lack of incompatible phases and conformity to Gibbs Phase Rule. As pointed out by Zen (1960) conformity to this Rule is a necessary, though not sufficient, condition for equilibrium. For arbitrary values of temperature and pressure, a condition applicable to a randomly chosen specimen, a restricted version of the Phase Rule is

$$p \leq c' - c''$$

where p is the number of phases, c' is the number of independent components, and c'' is the number of mobile components (Zen, 1960, p. 167). Four phases — garnet, pyroxene, epidote, and magnetite — comprise more than 98 weight percent of skarn. Furthermore, five independent components — SiO_2 , CaO , $(\text{Mg,Fe,Mn})\text{O}$, $(\text{Al,Fe})_2\text{O}_3$, and H_2O — comprise more than

98 weight percent of skarn (Table 15). Assuming CO_2 to be the major mobile component, even the restricted version of the Phase Rule (above) applies. This is strong evidence that chemical equilibrium was attained during skarnification.

Even a rock which conforms to the Phase Rule may be out of chemical equilibrium if phases are found together which have been experimentally or empirically determined to be mutually incompatible (Zen, 1963). It has been shown previously (Chap. III) that skarn minerals, evidently formed contemporaneously, are indeed compatible under the same P,T-conditions.

Fulfillment of the above criteria indicates only that equilibrium could have been obtained during metasomatism but does not prove that equilibrium did, in fact, prevail. Further tests, such as the distribution of trace elements among coexisting skarn minerals (e.g. Kretz, 1960), are required to show that equilibrium was attained during skarnification.

CHAPTER V - CONCLUSIONS

Origin and classification of the deposits

The magnetite orebodies were formed adjacent to plutons of intermediate composition in response to chemical and physical changes in the ore-forming fluid caused in part by a decrease in temperature. The immediate source of the iron in contact metasomatic deposits is a nearby intrusion. The ultimate source, however, is very probably the Karmutsen volcanic rocks which, in part, have been assimilated by an advancing pluton. The ore fluid that originated in a nearby intrusion as a dilute aqueous supercritical solution consisted mainly of iron chlorides with lesser amounts of silica, magnesium, and aluminum. The initial temperature of the solutions approximated that of the intrusion (800-900°C). Movement to the sites of deposition was either along the intrusion-country rock contact, resulting in deposits immediately adjacent to the stock, or along faults, resulting in deposits several hundred feet from the intrusive contact. Some orebodies may also have been formed by solutions migrating outward from the intrusion along a contact between limestone and volcanic rocks.

Skarnification, which took place in the temperature range 700-550°C, generally preceded the main stage of magnetite deposition. Conformity to Gibbs Phase Rule and non-appearance of incompatible phases is strong evidence that equilibrium was attained during skarnification. Neutralization of iron chloride solutions by calcite resulted in precipitation of magnetite in the temperature range 400-550°C. Ore fluids, originally one phase, probably developed into a two-phase system at lower temperatures. These fluids increased in pH by reaction with calcite until they reached at least 7.8, the minimum stability pH of calcite. Magnetite first filled cavities in skarn and brecciated volcanic rocks; then diffusion into, and replacement of, volcanic rocks took place. Where the volume of host rock dissolved exceeded the volume of metasome deposited, cavities were formed, some of which were later filled by magnetite or by post-ore calcite and/or quartz.

The most favourable sites of deposition were along contacts between limestone and volcanic rock, particularly where local folding had brecciated the more brittle igneous rocks. Brecciated volcanic rock, relatively more permeable to ore-forming fluids and having a relatively large surface area, was a particularly favourable site of ore deposition.

Lindgren (1933) classified contact metasomatic deposits as those which were deposited by direct igneous emanations, in contrast to those formed by hot ascending waters of uncertain origin but charged with igneous emanations. The temperatures and pressures which Lindgren considered to be characteristic of hypothermal deposits are very similar to those under which contact deposits were formed. Hypothermal deposits, however, were considered by Lindgren to have formed at great depth whereas contact metasomatism was considered to be a relatively near-surface phenomenon. Recently, however, Ridge (1958) has suggested a modification of Lindgren's classification in which contact metasomatic deposits are regarded as hypothermal deposits chemically concentrated in calcareous rocks by hydrothermal solutions with slow decrease of temperature and pressure. Suggested conditions of formation of this class are: temperature - 300-600°C, pressure - 700-3500 atmospheres, depth - 2-10 miles (Ridge, 1958, pp. 192-193). Inasmuch as the deposits of the present study are thought to have been deposited from aqueous solutions under conditions of temperature, pressure, and depth within the limits suggested by Ridge, they conform to his modification of the Lindgren Classification of Ore Deposits.

Suggestions for further work

1. Studies of the trace element distribution among coexisting minerals would aid in establishing whether or not equilibrium was indeed attained during formation of these deposits.
2. The pressure and temperature limits of ore deposition suggested in the present study could be checked by further use of the pyrite-pyrrhotite-arsenopyrite "geothermobarometer".
3. The use of birefringent grandite garnet as a geologic thermometer has been questioned because included impurities can cause birefringence in normally isotropic simple salts. Detailed chemical and crystal studies are required to determine whether such impurities may contribute also to birefringence of grandite garnet.
4. The author has suggested that physical features such as local faults, folds, and various types of contacts are the main factors required for local ore deposition. These are relatively small-scale features which are recognized only by detailed mapping of host rocks adjacent to igneous contacts. Exploration for contact metasomatic magnetite deposits should consist of detailed geological and geophysical surveys along contacts of plutons and favourable country rocks.

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APPENDIX I

Descriptions of properties

1. The Argonaut Company Limited (Iron Hill Mine)

The Iron Hill property lies southwest of the community of Campbell River on the east coast of Vancouver Island (Fig.1). The mine is on a small ridge rising about 750 feet above the southwest shore of Upper Quinsam Lake. Magnetite was discovered on this ridge early in the century but little was done until 1948 when the Coast Iron Company acquired the property, carried out a limited drill program, and shipped a small quantity of ore. The mine was leased to the Argonaut Company early in 1951; production began in August of that year and continued until December, 1956. During this period, 1,998,526 short tons of iron ore concentrate were shipped.

Local geology

In the Iron Hill area an elongate body of granitic rock intrudes sedimentary and volcanic rocks considered by Gunning (1930) to be part of the Vancouver group. The volcanics, relatively younger than the sediments, are correlative with the Karmutsen group; the sediments, mainly limestone, are correlative with the Quatsino formation. The intrusive rock, named the Quinsam diorite by Gunning, is overlain unconformably elsewhere by Upper Cretaceous sandstone and was considered by

him to be Late Jurassic or Early Cretaceous.

Fine-grained, dark green andesite outcrops on top of Iron Hill and along the ridge to the southwest. In places, the rock is distinctly porphyritic; star-shaped clusters of feldspar crystals are most common but some phenocrysts are dark green hornblende. Pillow-like structures were noted in these rocks along with what appeared to be flow breccia. Faint bedding in tuffaceous layers showed that these rocks dip northwards at a moderate angle.

The rock is fresh; minor alteration is shown by a slight "clouding" of plagioclase (seen only in thin section) and growth of actinolite in hornblende. The origin of "clouded" feldspars has been discussed by several authors (see discussion, Appendix II). A point count gave the following mode: plagioclase (An_{64} to An_{78}) 52.0%, hornblende 37.5%, others (opacite, sphene) 10.0%. A chemical analysis is given in Table 13, no.1.

Alternating grey and white beds characterize limestone overlying the volcanic rocks. Recrystallization has altered the limestone to medium-grained marble. Three chemical analyses show that such marble is more than 97% calcium carbonate (Table 14).

The Quinsam intrusion is seven miles long in a northwesterly direction and three miles wide. Gunning (1930) recognized several phases of the intrusion but at Iron Hill it is a medium-grained, massive granodiorite containing euhedral hornblende crystals. In thin section, the hornblende is seen to be largely poikilolitic and encloses several plagioclase crystals. Plagioclase (An_{57} to An_{67}) and hornblende are generally euhedral. Quartz and untwinned potassium feldspar are interstitial. Accessory minerals are sphene, magnetite (and/or ilmenite), apatite, sericite, chlorite, epidote, and zircon. A modal analysis of Iron Hill granodiorite gave: plagioclase 47%, quartz 22%, amphibole 19%, potassium feldspar 9%, accessories 2% and opacite 2%. A chemical analysis is given in Table 12.

Near its contact with volcanic rocks, the granodiorite contains many dark-coloured inclusions ranging in size from $\frac{1}{2}$ to 10 inches. Most are angular, some are round, but all have sharp contacts with the granitic matrix. Several of the fragments contain the star-shaped clusters of feldspar characteristic of the overlying andesite.

Highly altered green dykes, generally only a foot or two wide, cut through all rock types. The dyke-rock is now almost entirely composed of laumontite and calcite plus

scattered remnants of diopsidic pyroxene. It is difficult to estimate the original rock composition but it may have been not unlike diabase. A fresh diabase dyke outcrops in the bed of Granite Creek a few hundred feet north of the open pit.

Geology of the Iron Hill mine

A detailed description of the Iron Hill mine was given by Black (1952) in which he suggested the major structure is an overturned syncline. The axial plane of this fold strikes northwesterly and dips to the northeast between 45 and 80 degrees. The beds at both ends of the syncline were considered by Black to dip inwards, thereby forming a structural "low" in which the deposit was formed. The original outcrops formed a rough ellipse, the centre of which was composed of limestone. Volcanic rocks surrounded limestone on three sides and between the two an irregular zone of magnetite and skarn was developed. The limestone was thought by Black to mark the approximate location of the axial trace of the syncline. Skarn obscures most of the limestone-volcanic contact but company records show that most of the ore occurred as a replacement of volcanics rather than of limestone.

Map 1 (in pocket) shows exposures of the main rock types at the time of the writer's visit in June, 1961.

The northeasterly dip of the south limb of the fold is shown by bedding in the limestone but attitudes on the north limb are obscured by intense alteration, migmatization, and skarnification. Cross-sections of the deposit, based on company data obtained before and after Black's visit, were constructed by the writer. Figure 11 is a representative cross-section. It shows a large, central mass of limestone surrounded on three sides by volcanics which have been more-or-less replaced by iron ore. Most of this ore has developed along the contact between limestone and volcanic rocks and in the trough formed by the fold. A plan of the deposit (Map 1) shows the control exerted by this contact on the emplacement of both skarn and ore. Figure 11 illustrates how the intrusion of granodiorite has conformed to the pre-existing synclinal structure so that the older rocks seem to form an embayment in the underlying granitic rock. This figure also emphasizes the thickening of the orebody as the zone of maximum flexure in the fold is approached.

A contour map of the top of the granodiorite body (Fig.12) graphically illustrates the saucer-like shape of the contact between sediments and intrusion. The form of this depression is very similar to the trough-shaped structure of the sedimentary rocks above the intrusion as shown by Black, (1952, Fig.17). When Figure 12 is compared with a contour map

FIGURE 12
Contour of top of Quinsam granodiorite stock sub-outcrop, Iron Hill mine.

FIGURE 13
Contour of first appearance of ore zone above Quinsam granodiorite stock, Iron Hill mine.

of the first appearance of skarn and ore above granodiorite (Figure 13), the similarity between the two maps illustrates the commanding influence played by the sedimentary pendant on the emplacement of ore.

Ore deposition at Iron Hill was evidently controlled by the following features:

- (a) The contact between limestone and volcanic rocks.
- (b) The "trough" formed by the folded sediments.
- (c) Emplacement of granodiorite adjacent to the limestone-volcanic contact.
- (d) Preservation of the syncline enclosed by the stock.
- (e) Selective replacement of volcanics adjacent to limestone and along a synclinal axis.

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2. Texada Mines Limited

Texada Island is in the Strait of Georgia about 50 miles northwest of Vancouver. Magnetite deposits occur on the west side of the island (Fig.1) and can be reached by road from Vananda or Blubber Bay. These deposits were among the first to produce iron ore in British Columbia. Magnetite was discovered in the north part of the island in 1873 and was acquired by the Puget Sound Iron Company in 1875. Ore shipments began in 1885 and continued until 1890. The Prescott mine was leased to the Pacific Steel Company in 1901 and shipments to Irondale, Wash., were resumed for about three years. The property was again leased in 1907 and in the following year 1000 tons of ore were shipped from the Lake orebody. Spasmodic activity, mainly exploration, continued until 1916. In 1922 virtually all the buildings were destroyed by fire.

In 1951 Texada Mines Limited purchased the entire holdings of the Puget Sound Iron Company. Production began in 1952 and a high grade concentrate has been shipped since then. In 1956, the milling procedure was altered to recover chalcopyrite as a by-product. Up to 1962, Texada Mines has shipped 3,896,347 short tons of iron ore concentrate.

Local geology

The holdings of Texada Mines Limited include four main orebodies occurring at or near the contact of a granitic intrusion and volcanic and sedimentary rocks. Three main rock types are common to all deposits: Texada formation, Marble Bay formation, and the Gillies stock. The Marble Bay formation lies within the central part of a large northerly to north-westerly plunging syncline and the Gillies stock was emplaced at or near the nose of this structure. (Mathews and McCammon, 1957, p.55).

The Texada formation consists of dark green to black, massive, fine-grained feldspar porphyry (andesite). Phenocrysts of feldspar are generally about one-eighth inch long but occasionally reach one-half inch. Veins and patches of epidote occur almost everywhere but are more common near magnetite deposits. Plagioclase and green hornblende form the groundmass. A modal analysis gave the following: hornblende 62%, plagioclase (An_{46}) 29%, opacite 2%, others (mainly sphene) 7%.

The Marble Bay formation, up to 2500 feet thick, is composed almost exclusively of calcareous rock. On the basis of chemical composition, the formation has been subdivided into three members, each several hundred feet thick.

The lower member is high-calcium limestone (89% CaCO_3), and succeeding members become progressively richer in magnesium (Mathews and McCammon, 1957, p.53). Magnetite deposits are developed in the lower member which is completely recrystallized with bedding appearing only as discontinuous grey or black coloured bands.

The Gillies intrusion is exposed in an area of about 13 square miles between Davis Bay and Gillies Bay on the north-west coast of Texada Island. The intrusion cuts both Texada and Marble Bay formations. Three main rock types have been recognized in the stock, distinguished by differences in colour, composition, and relative age. The oldest and predominant type is medium-grained, grey quartz monzonite. It generally has an equigranular texture but phenocrysts of pyroxene are occasionally seen. Sub-angular inclusions of a darker, finer grained rock are common, probably recrystallized xenoliths of Texada formation. Contacts with older rocks are sharp except where obscured by skarn. A representative specimen of the quartz monzonite was chosen for analysis (Table 12, No.2). A modal analysis gave: plagioclase (An_{47}) 29%, potassium feldspar 24%, quartz 16%, augite 8%, biotite 29%, opacite 5%, accessories 10%.

This quartz monzonite is cut by irregular masses of a lighter-coloured rock, also locally porphyritic, which does not contain angular inclusions. This rock has a bleached appearance in hand specimen and thin section examination shows that both feldspar and pyroxene are much more altered than these minerals in the quartz monzonite. A modal analysis gave: potassium feldspar 43%, plagioclase (An_{40}) 30%, quartz 11%, mafics 16%. Potash feldspar is generally altered to kaolinite, plagioclase is only slightly replaced by sericite, and mafic minerals (augite, hornblende, biotite) are unaltered. The rock is classified as monzonite although the quartz content (11%) places the rock near the border between monzonite and quartz monzonite.

The youngest of the three phases forming the Gillies complex is aplite, readily recognized by its pink colour and sharp contacts with other phases. It appears to have filled fractures in older rocks. These fractures are generally not more than four inches wide. The aplite is a fine-grained mixture of about equal amounts of quartz and orthoclase. Plagioclase is present in amounts up to 5%; other minerals form less than 2% of the rock. Potash feldspar is either largely or completely altered to kaolinite, and plagioclase is almost entirely replaced by clinozoisite and sericite. In general,

each succeeding phase of the Gillies intrusion is more acid than the one before. Also, deuteric alteration increases within each succeeding phase.

These three igneous rock types are pre-ore. Later intrusive activity, though probably genetically related to the Gillies stock, is post-ore. Two sets of dykes represent these last phases of igneous activity. The earliest of these, and the most prominent, is a northwesterly-trending group of feldspar porphyry dykes having a maximum observed width of 50 feet. The groundmass is green-coloured and accentuates white plagioclase phenocrysts which comprise 15% to 20% of the rock. About 2% is pyroxene phenocrysts. In thin section, plagioclase (An_{50} to An_{56}) and pyroxene phenocrysts (the latter largely uralitized) are surrounded by a fine-grained groundmass of sericite, plagioclase, hornblende, apatite, and minor amounts of epidote. The rock contains occasional large anhedral grains of quartz.

The youngest dykes are fine-grained green rock conveniently described as "greenstone". These are normally only a foot or so wide and were seen cutting both ore and feldspar porphyry.

Geology of the Prescott mine

The Prescott body is an irregular lens or group of lenses developed along the contact between Gillies quartz monzonite and Marble Bay limestone (Fig.14). The plunge is southwards at a steep angle, almost parallel to the direction of dip of the intrusion-limestone contact. This contact is almost vertical on the east wall of the pit but dips southward at about 60° on the west wall (see Map 2, in pocket).

In places the ore zone is in contact with granitic rocks and limestone (Plate IV). Contacts with the latter are characteristically sharp but those with granitic rock are generally gradational. This is well shown on the west side of the pit where the ore zone (centre of pit) passes into barren skarn which, in turn, grades into altered quartz monzonite. An irregular envelope of skarn surrounds most of the orebody and small north-trending "shoots" of ore, generally more or less vertical, "break through" this envelope only to pinch out away from the main zone.

The ore is magnetite-rich skarn containing very irregular-shaped "pods" of calcite three inches to two feet wide. Some pods are interconnected by calcite veinlets but many are independent of their neighbours. Most of the ore is fine-grained except in the immediate vicinity of these calcite

FIGURE 14

pods where the magnetite-garnet rock rapidly increases in grain size. Calcite is very much more coarse-grained than the surrounding skarn and magnetite; cleavage rhombs one to three inches across are not uncommon. The term "breccia ore" has been applied to this type of ore by mine geologists in allusion to its broken appearance.

Skarn is largely composed of andradite garnet with minor amounts of diopside-hedenbergite and epidote. Magnetite veinlets cut skarn in many places. Sulphides, generally pyrite and chalcopyrite, are present only with magnetite and appear to be essentially contemporaneous with it.

An irregular zoning in and around the orebody is expressed by mineralogical changes in both metasomatic and host rocks. Generally not more than two successive zones are developed at any one locality in the pit and an inner zone may be directly in contact with an outer zone, the intermediate zone or zones being entirely absent. On a broad scale, though, four main zones are present in the Prescott pit:

Zone 1 - Ore Zone

This is the innermost zone and constitutes the main orebody. It is composed essentially of irregular pods of magnetite-rich skarn. Sulphides are relatively abundant and

where this zone is in contact with limestone, sulphide mineralization markedly increases toward limestone.

Zone 2 - Skarn Zone

An irregular zone of garnet-rich skarn wholly or partly envelopes the orebody. Magnetite is present in this zone but not normally in economic concentrations. Sulphides are generally less abundant and amphiboles (commonly actinolite) make their first appearance as skarn minerals. Epidote increases toward the outer edge of this zone. The skarn in places is vuggy with cavities filled with inward-projecting quartz crystals. Where skarn is in contact with limestone, a thin zone of actinolite, bordered on the outside by sulphides or hematite, is commonly found.

Zone 3 - Alteration Zone

This zone is an area of hydrothermal alteration in quartz monzonite and marks the beginning of replacement of the intrusion by skarn. In ideal exposures, three sub-zones are present. The innermost of these (sub-zone 3 (a)) is a garnet-epidote and is of lighter colour than the outer portion of Zone 2. Occasional remnants of unreplaced plagioclase are visible.

This is succeeded outward by sub-zone 3(b) which is similar to 3(a) except that garnet is absent. A granitic texture, inherited from the partly replaced igneous rock, becomes distinguishable (Plate XVIII) and epidote changes from pistacite to clinozoisite. Plagioclase and mafic pseudomorphs, largely composed of sericite, kaolinite, and saussurite, are visible even in hand sample.

Sub-zone 3(c) is marked by the first appearance of quartz as a major constituent of the rock and plagioclase, (average of three samples, An_{52}) is less altered. The rock is essentially hydrothermally altered quartz monzonite. A modal analysis of this rock gave: plagioclase (An_{52}) 31%, quartz 5%, (altered) potassium feldspar 25%, mafic minerals 31%, opacite 8%.

In limestone, Zone 3 is represented by irregular small unconnected patches of skarn, generally too small to be shown on the map. Locally, small patches of secondary pyrite and jasper in limestone are cut by calcite veins.

Zone 4 - Unaltered Zone

This is the outermost zone and is characterized by relatively unaltered country rock, either quartz monzonite or limestone. In quartz monzonite, plagioclase is slightly less

calic (An_{47}) than in adjacent sub-zone 3(c). Unaltered orthoclase makes its first appearance and quartz is much more abundant. Alteration is confined to slight chloritization of mafic minerals; chlorite is absent from sub-zone 3(c). Limestone is almost everywhere recrystallized in this zone but is otherwise unchanged.

These zones are summarized below:

Table 11

Prescott mine, Texada Island. General characteristics of zones developed within and around the orebody.

Zone 1 - Ore Zone

magnetite; magnetite-rich skarn; sulphides.

Zone 2 - Skarn Zone

Sub-zone (a) garnet-rich skarn

Sub-zone (b) epidote (or actinolite) first appears;
sulphides increase locally toward limestone

Zone 3 - Alteration Zone

1. In granitic rock

Sub-zone (a) garnet-epidote skarn

Sub-zone (b) epidote-rich skarn; no garnet; granitic texture; plagioclase (or pseudomorphs) visible.

Sub-zone (c) appearance of quartz; rock is recognizable as an altered intrusion.

2. In limestone

Patchy development of skarn; pyrite, jasper.

Zone 4 - Unaltered Zone

Fresh quartz monzonite or recrystallized limestone.

FIGURE 15

Prescott Mine, Texada Island. Diagram showing chemical changes across a contact between quartz monzonite and limestone. Zones correspond to those described in text, Table II. Composition of ore zone was calculated on the basis of 75% magnetite, 25% skarn.

Marked chemical changes accompany this mineralogical zoning. These chemical changes are not as abrupt as the mineralogy would suggest and serve to illustrate, in the granitic rocks at least, the gradational transition from igneous to metasomatic rock. The chemical variation in some of these zones is illustrated in Figure 15.

Feldspar porphyry bodies, some dyke-like in form, are seen in contact with, and in at least one case showed a chilled border against, skarn and magnetite. Small greenstone dykes cut magnetite, skarn, quartz monzonite, limestone, and feldspar porphyry. In several cases these dykes are off-set by east-trending, south-dipping normal faults showing at most a dip-slip displacement of ten feet.

Geology of the Yellow Kid mine

The Yellow Kid deposit, which is a replacement of limestone adjacent to an intrusion, is a classic contact metasomatic magnetite deposit. The orebody is elliptical in plan and the longer axis strikes north, parallel to the intrusive contact (Fig. 14). In vertical section the deposit is pipe-like in shape and plunges southeast under the intrusion. It pitches

southward toward the Prescott mine. The diameter of the pipe is 150 to 200 feet and pitch length is at least 800 feet. The shape of the orebody and its relation to intrusion and limestone are shown in Map 3 (in pocket) and in Figure 16.

The ore is medium-grained magnetite containing irregular pods of calcite. These pods may be connected by elongate masses of calcite, resulting in an interlocking network of calcite veins and pods surrounded by magnetite (Plate V). Some calcite veinlets are fracture fillings as shown by matching walls; other veinlets are irregular and their relation to magnetite is not clear. This calcite is pure white and coarse-grained in contrast to the grey, finer grained, banded blocks of unreplaced limestone which occur here and there in the deposit. These blocks are frequently enclosed in magnetite and do not appear to have been rotated from their original position. Pyrite and chalcopyrite, commonly associated with calcite and present mainly in the magnetite ore zone, extend outward as veins from calcite pods or veins and cut magnetite.

Massive ore is surrounded by a discontinuous zone of garnet-rich skarn a few tens of feet thick. Where this skarn is in gradational contact with altered quartz monzonite, hornblende increases in skarn toward the contact; epidote increases toward the contact in both skarn and granitic rock.

A

Pit level
July, 1961

A'

E 10500

TEXADA MINES LIMITED
YELLOW KID DEPOSIT
DIAGRAMMATIC SECTION A-A'
SCALE:- 1"=100'

Gillies stock- quartz monzonite

Marble Bay formation- limestone

Texada group- andesite

Magnetite

Skarn

Drill hole

FIGURE 16

In many places massive, skarn-free magnetite is in direct contact with crystalline limestone. Discontinuous grey bands in limestone, interpreted as bedding, are preserved right up to the magnetite-limestone contact and are cut obliquely by ore. Commonly a thin, discontinuous layer of euhedral pyrite separates magnetite from unaltered limestone (Plate XI).

Granitic rock forms the hanging wall of the deposit and is separated from ore by a thin zone of skarn or an altered phase of the intrusion. This alteration zone ranges in thickness from a few inches to several tens of feet and skarn has replaced granitic rock. In an extreme example of such replacement, a block of quartz monzonite a few feet in diameter is surrounded by massive garnet skarn. The alteration zone is extremely narrow, the change from relatively fresh quartz monzonite to garnet skarn can be covered within the length of a single thin section.

This alteration zone was separated by the writer into five phases:

Phase 1 - mainly garnet, minor calcite.

Phase 2 - epidote and quartz.

Phase 3 - plagioclase, mafics appear.

Phase 4 - sphene appears abruptly; plagioclase and mafics show little alteration.

Phase 5 - fresh quartz monzonite.

The similarity of these phases to zoning developed in the Prescott mine (Table 11) is striking. Zoning in the Prescott body was developed over several tens of feet; similar zoning in the Yellow Kid was developed in less than one inch.

On the north wall of the Yellow Kid pit a mass of bright yellow serpentine has replaced part of a small limestone block enclosed in magnetite and skarn. The serpentine body is of irregular shape and does not appear to follow bedding in limestone. In places, it is separated from limestone by magnetite which penetrates serpentine. In thin section, serpentine occurs as rounded pseudomorphs replacing granular carbonate. About half of every grain has been centrally replaced. Many grains have a serpentine centre, consisting of outward-radiating colourless acicular crystals, surrounded by a calcite rim. The crystals have a maximum retardation of first order yellow, show either positive or negative elongation, and have a moderate optic angles of negative sign. In some grains the centre of the serpentine is isotropic.

The Yellow Kid dyke (Map 3) is a large, essentially vertical feldspar-porphyry dyke which strikes northwest through the northern section of the pit. Strike-slip movement has occurred along this dyke especially near the borders; movement is of the order of 200 to 400 feet. The dyke possibly followed

an earlier fault which has been re-activated. There is no evidence to indicate that the "dyke-fault" is genetically related to ore. East-west trending, south-dipping normal faults have displacements up to 50 feet.

Geology of the Paxton mine

The Paxton mine is almost entirely within the eastern margin of the Gillies stock (Fig. 14). Limestone on the north wall of the pit (Map 4, in pocket) dips southwesterly toward the stock. The contact between Marble Bay limestone and Texada porphyrite is obliterated by skarn but presumably is parallel to the bedding of fresh limestone a few feet above the skarn contact (Plate III). The exposed margin of the stock is vertical and the main ore mass appears to have formed in a wedge between the intrusion and the southwest dipping limestone-volcanic contact. High-grade ore replaces either quartz monzonite or adjacent volcanic rock. Lower grade ore and skarn continue northward as a sub-horizontal tongue-like projection along the limestone-volcanic contact.

This deposit shows zoning outward from the main ore lens. The centre of zoning is high-grade massive magnetite,

generally very coarse-grained and relatively free from skarn. Small pods of calcite containing chalcopyrite and pyrite are found in this zone. Calcite veinlets project outward from these pods into magnetite, often along crystal boundaries. Massive ore grades outward into a mixed magnetite and skarn zone in which magnetite both veins and replaces skarn. A typical skarn-to-magnetite ratio is about 2 : 1. This changes outward into altered quartz monzonite on the southwest wall or altered volcanics on the northeast wall of the pit. Where replacement of quartz monzonite by skarn is gradational, the granitic texture of the intrusion is retained by skarn over short distances (Plates II and XVII). This progressive change is strikingly exhibited by hand specimens wherein massive, red-brown garnet has replaced most of the mafic minerals leaving only white altered plagioclase.

Compact, red-brown garnet replaces all three phases of the Gillies stock. Even aplite, the youngest phase that occurs as dykes has been replaced by skarn. Mineralization evidently occurred after the border zone of the pluton had been solidified and shattered. Subsequently, metasomatism and ore deposition took place.

The contact of skarn with volcanic rocks is generally gradational over several tens of feet but occasionally is sharp.

In contrast, skarn always terminates abruptly against limestone (Plate XI). The abundance of sulphides, particularly pyrite, locally increases toward the skarn-limestone or skarn-volcanic contact.

Near its margin the intrusion contains large blocks of Texada group volcanics. These blocks have sharp contacts which are crossed by veins of epidote and chalcopyrite. In the southeast pit, thin stringers of molybdenite cut both skarn and quartz monzonite.

A dyke of feldspar porphyry as much as 30 feet wide strikes through the northwest part of the pit. In several places, this dyke showed chilled borders against both skarn and magnetite.

South-dipping normal faults are common in the pit but their movements were generally only a few feet. One curved, south-dipping fault on the north wall is a reverse fault that has displaced the feldspar porphyry dyke about 20 feet.

Geology of the Lake mine

The reserves of the Lake mine were exhausted in 1955

and the pit has been allowed to flood; consequently the writer could not study the deposit in detail. The description below was compiled from literature and the author's brief examination of that part of the pit above water.

The mine is 1200 feet northeast of the exposed edge of the Gillies complex (Fig. 14). Replacement has been in Texada andesite adjacent to its contact with Marble Bay limestone. On the west wall of the pit, the limestone-volcanic contact is exposed. Volcanic rocks, more or less replaced by skarn, form the hanging wall of the deposit. The structure is evidently a syncline, the south limb of which is overturned to the northeast. Ore has replaced volcanic rock along the axial line of the fold near the limestone contact.

The northern part of the orebody, nearest limestone, is richer in sulphides. McConnell noted that "a vein-shaped lens of chalcopyrite two to five feet in width and 60 feet long occurred wedged in between altered porphyrite and crystalline limestone" (McConnell, 1914, p. 73).

The Lake mine illustrates several features which contribute to an understanding of this type of deposit:

(1) mineralization occurred in an area of maximum flexure in an overturned syncline (2) the orebody, although it can be considered a typical contact metasomatic deposit, is found at

a considerable distance from an intrusion (3) sulphide mineralization is greatest on the "limestone side" of the deposit and (4) deposition of magnetite was mainly in volcanic rocks.

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3. Empire Development Company Limited

The Empire Company holdings are 18 miles southwest of Port McNeil on the northeast coast of Vancouver Island (Fig. 1). The deposits are found at an elevation of 2500 feet on the east side of Merry Widow mountain (Fig. 5).

Magnetite has been known in this area since the end of the last century. In 1928, the Quatsino Copper-Gold Mines Limited was formed to prospect for copper in the area. A contact metasomatic chalcopyrite-bornite-magnetite deposit was already being explored by other interests on the nearby property of Coast Copper Co. Ltd. Quatsino Mines suspended exploration in 1931 but operations were resumed in 1950 on the Merry Widow No. 5 claim. Considerable magnetite was subsequently discovered on this and the adjoining Kingfisher Fraction Crown-granted claims. The property was optioned to the Empire Development Company Limited in 1956. Production began in 1957 and the property has produced, to 1962, 1,575,498 short tons of magnetite concentrate.

Local geology

The deposits occur in rocks of the Vancouver group near the margin of the Coast Copper stock. Quatsino limestone, generally well bedded, strikes northerly for a distance of

about two miles and dips to the west at a low angle. West of the mine, this unit continues for more than twelve miles, its strike changing to west or northwesterly. The dip is southward at about 15 degrees. The change in strike exposes a broad, open syncline, plunging to the southwest, into the nose of which the Coast Copper gabbro stock was intruded (Jeffery, 1962). Near this intrusion, especially in limestone, complex folding with local overturning is common.

The lower part of the Bonanza group occurs as a discontinuous "rim" around much of the Coast Copper stock. Flow rock in the Merry Widow pit about 400 feet from the Coast Copper intrusion is dark grey to black in colour and hornfelsic in texture. The rock contains about 15% quartz in addition to plagioclase and augite and is considered to be hornfelsized dacite. Near the headwall of the Merry Widow pit and about 150 feet from the intrusion, a distinctly layered rock unit in the Bonanza group dips westward toward the stock. The layering is caused by alternate light and dark bands. In thin section, the dark colour is seen to be a result of a variation in the proportion of pyroxene in each layer. The remainder of each band is composed of plagioclase. Jeffery has described the rock as a plagioclase-pyroxene gneiss (1960, p.95), probably a recrystallized pyroclastic.

The Coast Copper stock outcrops over an area of five square miles; the body is oval-shaped in plan with the long axis trending north. The eastern two-thirds is composed of diorite-gabbro; the remainder is monzonite. Steep contacts, dipping outward between 70 and 90 degrees, are characteristic of this stock (Jeffery, 1960, pp. 92-95). In the Merry Widow pit, however, this contact dips only 55 degrees to the east. Immediately adjacent to the Merry Widow orebody and for several hundred feet west of it, the intrusion is coarse-grained gabbro containing up to 12% by weight of disseminated magnetite. A modal analysis of a typical sample of the Coast Copper stock is: plagioclase (An_{53}) 50%, augite 39%, actinolite 7%, opacite 2%, accessories 2%. A chemical analysis and norm calculation of this sample are given in Table 12, No. 3.

Greenstone forms dykes, sills, and irregular bodies throughout Quatsino and Bonanza rocks. Such bodies in Bonanza volcanics are exceedingly difficult to distinguish. The greenstone is of variable composition but is mainly formed of plagioclase laths and anhedral pyroxene, both too fine-grained to allow further optical identification. Locally either of these minerals may occur as phenocrysts but more commonly they are found as microlites in an extremely fine-grained or glassy groundmass. These rocks are thought to represent an intrusive

phase of the overlying Bonanza group volcanics (Jeffery, 1960, p.95). These greenstone bodies are pre-ore.

In the Kingfisher pit greenstone dykes in limestone have boudinage structure. A striking feature is that individual dyke segments have a thin rim of garnet skarn. Obviously, metasomatism occurred subsequent to both dyke emplacement and deformation, perhaps when the Coast Copper stock was intruded.

The road north of the open pits has exposed an oval-shaped breccia "pipe" a few tens of feet in diameter. It is composed of angular fragments of grey to green volcanic rock surrounded by a light, fine-grained matrix of quartz, potash feldspar, and chlorite. The fragments vary from one-eighth inch to twelve inches and each is rimmed by a narrow bleached zone. This outer zone is mineralogically similar to the inner part of the fragments except that there is a complete lack of opaque minerals in the rim. Elsewhere abundant magnetite is disseminated throughout the fragments. The ragged periphery of the pipe extends outwards along dyke-like tongues of breccia into massive volcanic rocks. The pipe is cut by greenstone dykes.

Two sets of faults are recognized in the area. Normal faults, striking north and dipping to the east, have repeated Quatsino limestone to the east. These faults are not considered

to have any genetic relation to the orebodies (Jeffery, 1960, 1962). Northeast-trending faults, dipping steeply to the south are common in and around the deposits. In most cases dip-slip movement has been less than ten feet and Jeffery (1960, p.96) concluded that at least some movement took place along these faults after emplacement of the Coast Copper stock. In the Kingfisher pit, magnetite that extends along one of these faults has been brecciated by post-ore movement and later healed by coarse calcite. These faults appear to have formed between the time of gabbro intrusion and ore formation. Ore solutions possibly traversed these channelways; most of the deposits in this area are cut by at least one of such faults. In the Kingfisher pit, a thrust fault, dipping 25 degrees to the west, has allowed movement parallel to bedding in limestone. Magnetite has been emplaced along this fault and shows brecciation similar to ore in the northeast-trending steeper faults. Some post-ore movement appears to be indicated here also.

Geology of the Merry Widow mine

At the time of the writer's visit in August, 1961, rubble covered a large part of the bench faces and obscured the geology of this pit, so that a detailed geological map of the pit could not be made. A brief description of the deposit compiled from company and other data is given below.

This orebody occurs mainly in volcanic rocks near a limestone contact (Fig. 17). Lenses of magnetite, separated by barren volcanics and/or skarn, lie parallel to the eastward-dipping intrusive margin. Relict bedding in meta-volcanic rocks dips to the west at about 25 degrees. Ore lenses cut across the trend of the layered rocks and the form of the deposit seems to have been controlled by the intrusive contact rather than by bedding in the sediments.

A northeast-trending, south-dipping fault, cutting through the west end of the pit near gabbro, may have provided a channel-way for mineralizing solutions. A structural wedge, formed by the eastward dip of the stock margin and the westerly dip of the sediments, provided a favourable locus of ore deposition. Deep drilling in the ore zone has revealed that several small folds are present below the main deposit; these may have provided another structural control.

Sulphides, mainly chalcopyrite and pyrrhotite, increase in magnetite toward limestone to the east. Arsenopyrite, pyrite, sphalerite, and a little marcasite have also been recognized (Wittur, 1961).

FIGURE 17

Geology of the Kingfisher mine

This deposit lies entirely in Quatsino limestone 1000 feet northeast of the Merry Widow mine. Two steeply plunging cylinders of massive magnetite, nearly circular in section, lie about 200 feet apart and are 100 to 150 feet in diameter (Map 5 in pocket). These bodies come together about 200 feet below surface to form a Y-shaped deposit. The orebody pinches out into massive limestone in several downward-pointing projections.

Skarn is virtually absent from the Kingfisher deposit. Small amounts are found near greenstone bodies but otherwise the ore is composed of massive magnetite containing irregular pod-like bodies of white calcite. Sulphides are also relatively rare.

Greenstone sills and dykes criss-cross the deposit in several directions and, on first examination, appear to be post-ore. Closer scrutiny, however, shows that magnetite veins penetrate these greenstone bodies and large segments of dykes and sills are separated by magnetite ore. Limestone appears to have been preferentially replaced by magnetite; greenstone has been left as unreplaced remnants. Skarn and magnetite are also found as rims around greenstone dykes and sills at some distance from the ore-body.

Magnetite, emplaced along pre-existing faults and then brecciated by post-ore movement, was then healed by coarsely crystalline calcite. Fragments of massive magnetite, a few inches to several feet in diameter, are completely surrounded by the white calcite. Most of these fragments lack the angularity characteristic of a brecciated rock (Plate XVIII). All fragments, however, although medium to fine-grained inside, are rimmed by a thin zone of very coarse-grained magnetite. This coarse-grained material may cut through the fragments as vein-like projections which commonly contain central white calcite in vugs. The impression gained is that fragments of magnetite have been formed by movement along a fault; simultaneous solution and redeposition then took place along their margins. Recrystallization such as this would result in rounded corners of the fragments and coarse-grained margins. Some fragments of magnetite have been fractured but not separated into smaller pieces and the solutions have entered this crack. The result is a coarse-grained "vein" cutting through a fine-grained fragment.

The outer surfaces of many fragments of magnetite exhibit typical botryoidal texture, such as found in malachite, azurite, goethite, etc. (Plate XVI). The external botryoidal form is convex toward calcite. Texture of this type was also found in the cross-cutting "veins" of coarsely crystalline

magnetite where the rounded surfaces were convex toward the centre of the veins. Botryoidal texture has generally been interpreted as indicating precipitation of the material from a colloidal solution (Lindgren, 1924; Boydell, 1924/25). The genetic implications of this texture in magnetite are discussed elsewhere in this paper (pp. 71-76).

According to information received from J. Lamb, company geologist, early development work showed that massive magnetite frequently retained traces of bedding apparently inherited from surrounding limestone. Coarse and fine-grained layers in magnetite could be followed along strike into beds of limestone showing similar texture. Large structures, such as folds, could also be traced from limestone into massive magnetite. This preservation of structure is remarkable in view of the fact that the magnetite "pipes" plunge very nearly perpendicularly to bedding. This suggests that ore solutions moved across the layering; if so, they would be expected to obliterate rather than preserve sedimentary and structural features.

Other deposits

In addition to the Merry Widow and Kingfisher mines, several smaller magnetite bodies are found in the area. They

are briefly described below in the order in which they occur from north to south around the Coast Copper stock (see Jeffery, 1960, p. 94 for a map showing locations).

Raven deposit

The orebody occurs at the west end of a 1000-foot mineralized zone in greenstone. Ore lies along a northeast-trending fault separating greenstone and limestone. Mining in the Raven pit reached the economic limit in 1960 and the area was completely covered by waste at the time of the writer's visit.

Ajax deposit

Magnetite on this claim occurs about 450 feet from the Coast Copper stock at the contact between Quatsino limestone, Bonanza volcanics, and intrusive greenstone. One or possibly two northeast-striking faults pass through the south end of the mineralized zone. Ore is composed of either fine-grained magnetite disseminated in altered volcanic rocks or as vuggy magnetite in skarn. In several places, fragments of altered volcanic rock are "suspended" in a matrix of fine-grained red-brown skarn. Small blocks of skarn, in turn, are surrounded by massive magnetite. Quarter-inch books of chlorite occur in vuggy clumps throughout massive magnetite.

Shamrock-Blackjack deposit

Magnetite is exposed in scattered small outcrops within a rim of Bonanza rocks between the Coast Copper stock and Quatsino limestone. Bedding in limestone is vertical or dips toward the intrusion. The margin of the stock dips 45 degrees west and the wedge of volcanic rock between limestone and intrusion has been sporadically replaced by magnetite.

Ore in these deposits is characteristically vuggy; the cavities are a few inches wide at most, many being partly encrusted by inward-projecting quartz crystals. The remaining space in the vug is either vacant or filled with calcite or compact massive goethite. One quartz-filled vug was surrounded by magnetite having botryoidal texture. Each botryoidal "rosette" was about one-quarter inch across and convex into the vug.

An interesting outcrop of magnetite, three feet by twelve feet, occurs in coarsely crystalline limestone on the Shamrock claim. Layers of magnetite up to 1 mm in thickness alternate with calcite layers of similar thickness. Continuity in each layer is broken when magnetite breaks through an adjacent carbonate layer and fuses in an irregular manner with the next layer of magnetite (Plate XIII). Small folds, a fraction of an inch in magnitude, are present in layers of

magnetite and calcite 0.1 mm in thickness (Plate XIV). Such layering is narrower than in limestone outside the magnetite unit. The origin of this layered structure is not clear. It may be the result of selective replacement of certain limestone beds of magnetite but recrystallization has destroyed the original nature of this bedding. Jeffery has suggested the magnetite in this outcrop has replaced some type of algal structure (Jeffery, 1960, p.99). Folded limestone showing selective replacement by 1 mm layers of magnetite have been described from deposits in California (Lamey, 1961, pp.675, Fig.5,6). Photographs in the paper by Lamey show samples of this rock which appear identical to samples from the Blackjack-Shamrock deposit. In a brief explanation of the origin of this type of replacement, Lamey writes (p.675):

".....this selectivity may have resulted from a difference in the magnesium content of the original rock layers, the more magnesian ones having been selectively replaced. This is suggested because magnetite replaced antigorite and chondrodite in some deposits; thus a higher magnesium content of some layers may have facilitated replacement of them".

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4. Brynnor Mines Limited

This mine lies between Maggie Lake and Kennedy Lake on the west coast of Vancouver Island, nine miles northeast of Ucluelet (Fig. 1). A logging road connects the mine-site with the Alberni-Tofino road on the shore of Kennedy Lake.

A strong magnetic anomaly in the area was reported by W.F. Robertson, Provincial Mineralogist, in 1902. Noranda Exploration Company Limited took over the property in 1960 after the anomaly had been re-located by dip-needle survey. The name of the company was later changed to Brynnor Mines Limited, a wholly-owned subsidiary of Noranda Mines Limited. Production began in 1962 and in that year 451,623 short tons of magnetite concentrate were shipped.

Local geology

The geology of the Kennedy Lake area was virtually unknown until recent field work by officers of the B.C. Department of Mines and members of the Noranda staff clarified the structure and provided a basis for correlation with rocks found elsewhere on Vancouver Island.

A sequence of sedimentary and volcanic rocks exposed southeast of Kennedy Lake has been tentatively correlated with the Vancouver group. Fossils are lacking in the Kennedy Lake

suite and correlation has been based on lithologic similarity (Eastwood, 1962, pp. 111-122). The lower part of the Kennedy Lake sequence consists of Karmutsen-like volcanic rocks. These are conformably overlain by Quatsino(?) formation consisting of a lower, bedded limestone member, a middle calcareous tuff member, and an upper massive limestone member. Bonanza-like tuffs overlie these rocks with apparent conformity (Eastwood, 1962, p. 113).

Eastwood has shown that at least the upper portion of the Kennedy Lake sequence occurs within two major synclines. The easternmost of these, referred to as the mine belt, strikes northeast from the Brynnor open pit. The structure is defined by the contact between the upper limestone member and the intermediate tuff member of the Quatsino(?) formation. This syncline plunges northeast away from the mine. The structure is a mile wide in the mine area and has a minimum length of $3\frac{1}{2}$ miles. It is enclosed by diorite and granodiorite.

The upper limestone member of the Bonanza(?) group, about 300 feet thick, is seldom bedded though in the pit alternating light and dark banding is common. This banding is not consistent in attitude and may actually be flow banding or a segregation of impurities in limestone caused by recrystallization. An average of four analyses of samples from this unit,

obtained from company files, was 86.5% calcium carbonate (Table 14). The analyses also average 3.9% MgO but some of the samples may have been contaminated by serpentine. Magnesian metasomatism of this type is common in limestone around the orebodies.

The tuffaceous sequence separating the upper and lower members of the Quatsino(?) formation is about 800 feet thick on the slopes of Salmonberry Mountain (Eastwood, 1962, pp. 114). It is characteristically well-banded and generally calcareous. The bands are whitish or pale pastel-coloured and are from one to three inches thick. In the Brynnor pit (Map 6, in pocket), this tuff is extremely fine-grained; bedding is well defined high up on the southwest wall but becomes increasingly indistinct as the ore zone is approached. An exposure of tuff near Redford Creek east of the mine showed graded bedding; angular fragments of white feldspar were distinguished in the lower, coarser portion of each bed. Eastwood noted that tuff in the mine area is decidedly less calcareous than the same rock elsewhere in the area. He attributed the change to deformation, intrusion, and metamorphism. The tuff is a granular mixture of plagioclase and pyroxene. Dark layers contain proportionately more pyroxene; fine-grained, light layers contain more sericite. Company files have a chemical analysis labelled "tuff" as follows: CaCO_3 13.7%, Fe_2O_3 9.48%, SiO_2 75.8%, MgO trace.

Dark, grey-green massive andesite, sometimes amygdaloidal, sometimes porphyritic, occurs in gradational contact with tuff on the west side of the pit. Sericitized feldspar and green hornblende, the latter more or less altered to chlorite, occur together in an interlocking aggregate. Epidote, pyrrhotite, and prehnite are found throughout both as scattered grains or thin veinlets. Prehnite is often accompanied by sericite in these veinlets. A chemical analysis of this rock is given in Table 13, No. 2.

The relation of this andesite to the layered rocks is not clear. Sills and dykes of similar composition cut limestone and tuff and in most cases are clearly pre-ore; magnetite and skarn replace these dykes and sills as well as the main andesite body. From exposures in the pit, it would seem that andesite occurs between tuff and limestone and could then be interpreted as a flow or a sill. Deeper drilling, however, has revealed that this andesite is unconformable. Moreover this unit does not appear in the same stratigraphic position elsewhere in the area and Eastwood concluded that the andesite body is a stock. If so, it could be correlated with greenstone dykes and sills described previously. These are thought to be an intrusive phase of the overlying Bonanza group of volcanic rocks. Three other andesite stocks are known in the Brynnor mine area; one occurs just within the southeast

corner of the pit (Map 6).

The layered rocks near Kennedy Lake have been intruded by a large granodiorite stock. About six square miles of the body have been mapped to date but the full size of the intrusion is not known. Eastwood (1962) has shown that the greater part of the rock is granodiorite having sharp intrusive relations with country rock. He has also defined a subsidiary plutonic rock type surrounding the mine belt. This rock is not exposed in the pit but can be examined in scattered outcrops north, west, and south of the pit. A specimen was chosen for chemical and petrographic analysis (see chemical analysis and norm, Table 12, No. 4). It is a medium-grained, equigranular rock which, in hand sample, appears to be composed primarily of white plagioclase and chloritic hornblende or pyroxene. Very little quartz is discernible in hand sample. In thin section, however, quartz is readily apparent; orthoclase, a little microperthite, and plagioclase, all largely altered to a dusty, reddish-brown opaque material, comprise the remainder of the rock. A little chloritic hornblende, zircon, sphene, opacite, and prehnite are the accessory minerals. A point count gave the following mode; plagioclase (An_{32}) 37%, potash feldspar 27%, quartz 29%, accessories 8%. The rock is classified as quartz monzonite.

Geology of the Brynnor mine

The geology of the Brynnor pit, as it was exposed in June, 1962, is shown in Map 6 (in pocket). Figure 18 is a diagrammatic cross-section compiled from limited drill data. Three main ore zones have been recognized by the company although these were imperfectly exposed at the time of the writer's visit. The ore appears to favour the limestone-tuff contact but pockets and lenses may be found entirely within one or the other rock type.

The upper part of the north orebody has replaced limestone. Drilling has shown that it plunges southwards as a roughly cigar-shaped body that continues along the tuff-limestone contact at depth (Fig. 18). It appears to be cut off to the north by a fault. The ore is massive, fine-grained magnetite containing green and blue serpentine along slip-surfaces. This serpentine is usually a fraction of an inch thick and is almost entirely confined to the ore zone; limestone on either side contains only a few veins of this mineral. Thin serpentine veinlets in magnetite stop abruptly at the magnetite-limestone contact. Green sills and dykes that cut limestone near the ore zone have been almost completely replaced by green serpentine and one or two bodies of massive serpentine are found in limestone to the north of this ore zone. Contacts between

FIGURE 18
BRYNNOR MINES LTD. DIAGRAMMATIC SECTION A-A'. SCALE:- 1" = 100'

serpentine and limestone are generally sharp and locally a one inch zone of euhedral pyrite is found along the contact. Very little skarn is associated with this orebody, a common characteristic of magnetite deposits developed largely in limestone.

The central ore zone is massive fine-grained magnetite which is heavily slickensided along short, curving fractures coated with blue serpentine. Ore has formed at the contact between limestone and andesite. Skarn is common on both sides of the orebody but not in the ore itself.

Only a few feet of the south zone was exposed at the time of the writer's visit. Structural relations were not clear but it appears that ore occurs along the tuff-limestone contact. Magnetite here is also associated with serpentine, much of which is yellow or yellow-green. Joints within limestone near ore are filled with serpentine and it extends out on either side of the joints for a quarter inch or more. The rock in the vicinity of this orebody is cut by several faults and shear zones, many of which contain irregular lenses of magnetite. Deep drilling has indicated that the south and central ore zones may join at depth.

Several ages and types of dykes are found in the pit. A few dykes are pre-ore, largely altered to serpentine. Next in age are small, light to dark green, basic dykes which locally may be slightly serpentized. One or two of these are rimmed with a thin skarn zone where they cut limestone (Plate VII). Their relation to ore is unknown. A third dyke swarm is by far the most common. These are large, leucodiorite dykes up to 30 feet in width, which vary slightly in mineral composition. They are composed primarily of plagioclase and uraltized augite; the proportion of feldspar to pyroxene varies from dyke to dyke, or even within a single dyke, and either mineral may occur as scattered phenocrysts. These variations in composition and texture have resulted in a number of slightly different rock types which have been grouped together under the term "leucodiorite" because of their general dioritic composition, light colour, and age relative to ore. These dykes cut the ore and the larger dykes may show chilled borders against skarn and neither skarn nor magnetite is developed in them. These dykes are later than the green basic dykes mentioned above. A fourth type of dyke rock, of which only one was seen, is a typical diabase. It is not in contact with any of the other dykes or the ore but is unaltered and may be assumed to be younger than other dykes.

Post-ore faults of generally small displacement are common. A deep-seated orebody has been outlined by magnetometer and drilling to the southeast of the pit. This body appears to have formed along the limestone-andesite contact in a down-faulted block along a fault exposed in the extreme southeast corner of the pit. Dip-slip movement along this fault would be of the order of 200 feet.

Some movement along faults in the pit may be caused by dynamiting. After one blast in the Brynnor pit, a large part of the 300 level moved two feet vertically and three feet to the south. This movement was confirmed by surveying a station on the "fault block" before and after the explosion.

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5. Nimpkish Iron Mines Limited

The mine office and camp are on Anutz Lake, $1\frac{1}{2}$ miles south of Nimpkish Lake in north-central Vancouver Island (Fig.1). The open pit is on the southwest bank of Nimpkish River five miles south of Anutz Lake. Access to the camp is via a limited-access lumber company road from Beaver Cove.

Magnetite was discovered in this area in 1897. Thirteen years later Lindeman described the deposit and presented the results of a dip-needle survey conducted over the magnetite showings (Lindeman, 1910, pp.19-20). Claims covering the deposit were optioned by A.H. Upton about 1954 and limited diamond drilling was done at that time. In the following year, the property was optioned by Nimpkish Iron Mines Limited, a company jointly owned by Standard Slag Company and International Iron Mines Limited. No activity was reported during the years 1957-58 but in 1959 construction began on mill and dock facilities. Magnetite concentrate production began that year and by 1962 1,077,303 short tons had been shipped.

Local geology

The Iron Crown (Klaanch) deposit is along the contact of Quatsino limestone and Karmutsen volcanic rocks adjacent to a plutonic body. This pluton is part of a complex referred to by Hoadley (1953, p.29) as the "Nimpkish intrusions".

Karmutsen volcanic rocks in the mine area are fine-grained massive, black or dark green, amygdaloidal andesites. In thin section, the rock has a distinct ophitic texture; interstices between plagioclase laths (An_{79}) are filled with biotite and hornblende. Where the rock is altered, sericite, calcite, and actinolite are common constituents. Amygdules are filled with epidote, calcite, or actinolite. Pyrite and pyrrhotite are scattered throughout as round clusters. Pink laumontite and white calcite veins transect the rock in several outcrops. An analysis of a typical fresh specimen of this andesite is given in Table 13, No. 4.

Quatsino limestone in this area is white or light grey and generally coarsely crystalline. Bedding is rare or absent; thin discontinuous dark bands were noted in the main headwall of the pit. These may be shaly layers and as such would constitute bedding. On the other hand they may be flow banding in which case they would not necessarily bear any relation to bedding. On the south wall these bands are undulatory but essentially horizontal except adjacent to the ore zone where they are vertical, parallel to the ore contact. The rock appears to be very pure calcium carbonate. A stain test on one or two specimens showed no dolomite.

Quartz monzonite outcrops extensively south of the pit along the banks of Nimpkish River. This rock is leucocratic,

coarse-grained, and hypidiomorphic. Plagioclase, potash feldspar, quartz, biotite, and hornblends are clearly visible in hand specimen. A point count gave the following mode: quartz 35%, plagioclase (An_{38}) 20%, potash feldspar (orthoclase and microperthite) 14%, hornblende and biotite 29%, accessories 2%. A chemical analysis of a typical specimen is given in Table 12, No. 6.

Within and adjacent to the ore zone medium-grained granular diorite occurs intermixed with large and small blocks of Karmutsen andesite. Slightly sericitized plagioclase (An_{37}), hornblende, augite, and orthoclase are the main constituents of this diorite. Along the south and east wall of the pit (Map 8, in pocket), assimilation of volcanic rock is so intense that it is difficult to distinguish plutonic from volcanic rock. Andesite has been so recrystallized and "soaked" by the intrusive rock that in the transition zone the distinction between fine-grained diorite and coarse-grained andesite is arbitrary. Skarn occurs throughout this zone and replaces all other silicate rocks. Jeffery (1960,p.102) has interpreted this diorite as being a separate small intrusive body. The present author, however, considers it as a contaminated border phase of the quartz monzonite body for the following reasons: (1) diorite is more fine-grained than quartz monzonite and may represent a chilled phase of the latter (2) the "blurred"

relation of diorite with andesite and the appearance of "soaking" of the latter by intrusive rock. This suggests that the diorite matrix is a result of contamination by country rock of a magma. The blurred edges of andesite fragments are evidence of assimilation (3) rock in the transition zone traced away from ore toward the quartz monzonite body, approaches the composition and grain size of quartz monzonite. The transition to or contact with (depending on which theory is favoured) the main quartz monzonite body cannot be observed because the critical area is covered by mine roads and dumps.

Several other types of intrusive rock are exposed in the pit. A common type is light green feldspar porphyry occurring as isolated bodies, generally bounded by faults. These bodies, as exposed in each bench face, appear dyke-like in form but in most cases are not continuous from level to level (see Map 8 in pocket). As this rock is not exposed in contact with ore its age relative to the deposit could not be established. However, these bodies are not replaced anywhere in the pit by magnetite or skarn and the conclusion is that this rock is a repetitively faulted post-ore dyke or dykes. In thin section, a groundmass of chlorite and sericitized plagioclase laths surround sericitized plagioclase phenocrysts. Calcite, epidote, and opacite are accessory minerals.

One light-coloured dyke, apparently post-ore, occurs in the north part of the pit. In thin section, quartz and orthoclase are seen to be the main constituents; plagioclase and opacite form only about 20% of the rock. Technically the rock should be classified as an aplite although in hand specimen it lacks the typical "sugary" texture.

Felsite forms an isolated body in limestone, its margins replaced by magnetite and skarn. In thin section, replacement by calcite, clinozoisite, and sericite has largely obliterated the original constituents but where alteration is less intense, interlocking plagioclase laths form the main aggregate. The rock may be a highly altered remnant of andesite emplaced as an intrusive phase of the overlying Bonanza pyroclastic rocks. Small dykes and sills of similar rock, also replaced by skarn and magnetite, occur throughout the pit but are most prominent in limestone. Two of these felsite dykes, one of which is shown in the extreme southwest corner of Map 8, have been deformed into boudins and the segments separated by as much as ten feet. In both dykes, a thin outer rim of garnet-epidote skarn, occurring in the dyke adjacent to limestone, continues around the broken ends of the dyke as though metasomatism had occurred after the dykes had been segmented and not when the dykes were emplaced. Similar relations were noted in the Kingfisher deposit.

Geology of the Iron Crown (Klaanch) mine

Magnetite occurs as a tabular body between andesite and crystalline limestone. The contact with volcanic rock is diffuse and obscured by skarn. The limestone contact, in contrast, is sharp and free of calc-silicate minerals. Within the deposit, the ore is relatively pure magnetite containing little skarn or other silicate impurities. Coarse white calcite is found throughout the ore zone as irregular veins or pods which are rarely found outside the mineralized area. These "pods" frequently increase in number toward the limestone side of the deposit and locally constitute as much as 50 percent of the ore. Chalcopyrite, pyrite, and minor amounts of sphalerite are almost entirely restricted to these calcite bodies. In many instances, calcite heals fractures in massive ore and thin veinlets of calcite cut across large single crystals of magnetite. Sulphides, so closely associated with this post-ore calcite, are considered to be post-ore even though no sulphide mineral was observed cutting magnetite.

Small bodies of magnetite are found in limestone in the west and south ends of the pit and are not visibly connected with the main ore zone although they presumably formed at the same time.

In the southeast corner of the pit, the ore zone narrows and closely follows the andesite-limestone contact (Plate IX). At this place the contact is practically vertical except for a small part, just south of the road, where it is folded back on itself. This slight overturning possibly was caused by forceful intrusion of the quartz monzonite body east of the pit. The lower limb of this "fold" flattens in dip just before it reaches the pit floor under the south wall. This could account for an apparent local increase in width of the ore zone.

One or two major faults separate the ore deposit into three separate bodies. The most prominent fault strikes northeast through the south end of the pit. It is exposed in several faces and is marked by a breccia zone three to five feet wide. Fragments of slickensided ore and country rock are surrounded by abundant chlorite and gouge. Hematite coats some of the slip surfaces. The fault dips steeply southeast; strike-slip component of the ore zone is about 200 feet dextral. A reverse dip-slip component of about 100 feet is indicated by the difference in width of the ore zone between the road and the main pit floor.

Another post-ore fault was not observed but is inferred in the northwest corner of the pit to account for the abrupt eastern termination of this orebody (see Map 8).

Most contacts between major rock units are slip surfaces. Although some movement has taken place along these planes the displacement appears to be small.

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6. Zeballos Iron Mines Limited

The deposits lie at an average elevation of 2500 feet on the northwest slope of the Zeballos River valley, west coast of Vancouver Island (Fig. 1). A five-mile gravel road leads from Zeballos to the mine office and mill, the latter constructed at 400 feet elevation just above the valley floor. Three miles of switchback road connect these buildings with the main bench of the open pit at 2400 feet.

Mining activity in this area dates back to 1907 when small amounts of placer gold were obtained from Zeballos River. A minor gold rush began in 1924 following discovery of lode gold on the Tagore property and it is likely that magnetite deposits near Blacksand Creek were discovered about that time. If so, however, they did not receive much attention during the search for gold and are not mentioned by Young and Uglow (1926) in their survey of iron ore deposits in western Canada. Claims adjacent to the F.L. (Ford) deposit were staked in 1937-38 and the F.L. claim was probably staked about that time also. In 1951, the property was controlled by Anyox Metals Limited, a subsidiary of Ventures Limited. This company carried out a systematic exploration program including geologic mapping, a dip-needle survey, and diamond drilling. During 1959-60, Zeballos Iron Mines Limited, a subsidiary of International Iron

Mines Limited, took over the property and continued drilling in the main ore zone. Production began in the summer of 1962 and 250,297 short tons of magnetite concentrate were shipped that year. Operations ceased in the spring of 1963 due to financial difficulties.

Local geology

Regional geology of the Zeballos area has been described by Stevenson, (1950) and Hoadley (1953). The geology of the F.L. deposit was discussed by Bacon (1952), McKechnie (1960), and Eastwood and Merrett (1962).

The main ore zone, the F.L. or Ford deposit, lies along the contact between Quatsino limestone and highly altered Bonanza group volcanic rocks (Fig.19). Limestone occurs along the axial trace of a southwesterly-plunging anticline which is approximately at right angles to the general northwesterly trend of the regional structure. This cross fold is preserved as an elongate lobe, $1\frac{1}{2}$ miles long, surrounded on three sides by the Zeballos batholith (Hoadley, 1953, pp.66-68).

Quatsino limestone in this area has been largely recrystallized and bedding obliterated over large areas although it is distinct enough in several outcrops to define local structures. Weathering is usually grey or grey-white in colour.

ZEBALLOS IRON MINES LIMITED
Geological sketch
F.L. and Ridge deposits
(After Eastwood & Merrett, 1962)
SCALE:- 1"= 300'

- Bonanza group- andesite, tuff
- Quatsino formation- limestone

Magnetite

FIGURE 19

Analyses of limestone samples collected from about the same horizon (i.e. the top of the formation) on the opposite side of the Zeballos River have been presented by Stevenson (1950,p.18). Six samples showed the following range in composition: MgO 19.6-32.7%, CaO 54.4-32.7%, R_2O_3 0.76-0.5%, Insol. 5.18-0.88%. Stevenson considered that much of the magnesium content of these samples was caused by dolomitization of limestone adjacent of granitic intrusions.

The original nature of the overlying Bonanza group rocks has been more or less obscured by various types of alteration caused, in large part, by the nearby Zeballos batholith. Near the south end of the ore zone, volcanic rocks are intruded by aplitic and dioritic dykes and irregular bodies over a wide zone which separates the Zeballos intrusion from the orebody. To the north, along the north limb of the anticline, this zone becomes less distinct and finally disappears. Generally, the Bonanza rocks are massive, dark green and fine-grained. Bedding was not observed by the writer anywhere in the mine area although Eastwood has described the rocks as tuff, in part recrystallized, and noted that the rocks showed clastic texture in several localities (Eastwood and Merrett,1962,p.101). The present writer collected several samples of andesite from rocks of the Bonanza group near the pit. In section, scattered plagioclase phenocrysts are seen to lie in an interlocking

matrix of plagioclase laths, augite, and hornblende. Trachytic texture was seen in several thin sections. Andesite flows with oriented plagioclase laths have been reported elsewhere interbedded with Bonanza tuffs and breccias (Stevenson, 1950, p. 20). A chemical analysis of andesite, collected by the present writer, from the F.L. deposit is shown in Table 13, No. 3.

About 45 square miles of the Zeballos batholith are exposed in the Woss Lake-Zeballos map-areas. The full size of the batholith is not known but it has been mapped over a length of 18 miles and at the western edge of the Woss Lake sheet it is five miles wide. It is an elongate pluton which trends northwesterly, cutting across the regional trend of the sedimentary rocks at a low angle (Hoadley, 1953). The rocks comprising the Zeballos batholith are not uniform in composition or appearance. Hoadley broadly classified the rocks in the area southeast of the Zeballos River as quartz diorite and those on the northwest side as diorite with minor amounts of gabbro. In a detailed study of the batholith in Zeballos area, Stevenson (1950) separated the intrusion into four phases. These are, from oldest to youngest, gabbro, hornblende diorite, granodiorite, and quartz diorite. The intrusive rock adjacent to the F.L. deposit is shown on his map (Stevenson, 1950, Fig. 2) as hornblende diorite. The texture of this rock is medium-grained, equi-

granular, and hypidiomorphic. Plagioclase, minor quartz, and green-black hornblende combine to give the rock a "salt-and-pepper" appearance. Under the microscope, euhedral plagioclase crystals are abundant; orthoclase and microperthite occur interstitially as subhedral grains. Ragged subhedral green-brown hornblende grains showing partial alteration to brown biotite are scattered throughout. Clear, anhedral quartz, occurring in considerably greater abundance than a macroscopic examination would suggest, fills the interstices. Accessory minerals include apatite, opacite, biotite, chlorite, sericite (after feldspar), tremolite, and augite. A point count gave the following mode: plagioclase (An_{22}) 29%, potash feldspar 28%, quartz 23%, hornblende 10%, accessories 9%. On the basis of this mode the rock is classified as quartz monzonite. A chemical analysis of this rock is given in Table 12, No. 5.

Geology of the F.L. (Ford) mine

The surface geology of this deposit is shown in Map 7 (in pocket). The ore zone is an essentially tabular body of exceptionally pure magnetite dipping about 40 degrees west; average true width is 70 to 80 feet. Company geologists have arbitrarily divided the deposit into two parts, an 'A' and 'B' zone.

Throughout most of the 'B' zone, magnetite is exposed in steep bluffs and in several places massive magnetite could be seen at the base of these cliffs resting directly on white crystalline limestone. The contact is very sharp and has little or no skarn. Pyrite commonly occurs in a thin layer between magnetite and limestone. Here and there along the contact, bedding in limestone is preserved right up to the magnetite and it could be seen that the base of the ore zone dips west at an angle shallower than the dip of the limestone. This suggests that magnetite, which in the outcrop has replaced Bonanza volcanics down to the contact with limestone, might leave the limestone contact as the ore is followed down-dip. Drilling in this zone has, in fact, indicated that the limestone-volcanic contact rapidly steepens with depth and several drill-holes intersected this contact below the main ore zone. As soon as its footwall diverges from the volcanic-limestone contact, the ore zone decreases in thickness and pinches out.

The north half of the 'A' zone is similar to the 'B' zone; massive magnetite is in sharp contact with limestone at the base of the ore zone (Plate VIII). Near the south end of this zone, however, limestone turns to the east as it swings around the nose of the anticline mentioned earlier. The tabular ore zone does not swing with limestone but continues on with little change of strike into migmatitic quartz-monzonite

and finally pinches out in a mixture of skarn, magnetite and migmatite.

Mention has been made of the sharp magnetite-limestone contact exposed at the base of the orebody. This abrupt skarn-free termination is in marked contrast to the upper part of the ore zone. The stratigraphically higher limit of ore is determined largely by economic cut-off grades rather than by any geological termination. Massive coarse-grained magnetite in the lower part of the ore zone gradually becomes finer-grained with more admixed skarn and altered migmatite in the upper part. Skarn eventually disappears upward as the upper limit of the mineralized zone is approached.

Massive skarn is not everywhere abundant in the F.L. deposit but locally, such as in the south half of the 'A' zone, it constitutes a major portion of the mineralized area. Skarn in this deposit is unusual because pyroxene and epidote, rather than garnet, are its main constituents. Large areas of the main face in the pit are composed almost entirely of massive, granular green epidote. In one locality, a massive epidote zone is six to eight feet thick and the transition zone three feet thick.

Post-ore dykes up to four feet wide occur here and there throughout the deposit. Two types were recognized: a

light green feldspar-porphyry and hornblende leucodiorite. Chilled borders against skarn and magnetite were seen in one feldspar porphyry dyke; neither set shows ore mineralization within it.

Nearly vertical, post-ore faults of small displacement are common. Two of the more prominent of these are shown in Map 6. Both faults trend southeast and show right hand displacement.

Geology of the Ridge deposit

About 1700 feet east of the F.L. deposit, massive magnetite is exposed in almost continuous outcrop over an area of about 15,000 square feet (Fig.19). Ore occurs between limestone and a medium-grained intrusion of dioritic composition. To the south and west andesite outcrops are in contact with massive magnetite. These contacts are "blurry" and ill-defined but magnetite unaccompanied by skarn appears to have replaced andesite. The ore is generally medium-grained magnetite containing little or no skarn or sulphides. Magnetite contacts with limestone are sharp and dip steeply.

Feldspar-porphyry dykes, some showing chilled borders against magnetite, and light green fine-grained plagioclase-

actinolite dykes transect the ore.

A dip-needle survey was carried out with the hope that it would aid in evaluating this deposit. The result of this survey, as will be described below, will illustrate the danger of predicting the extent of such contact metasomatic deposits from the outcrop pattern even with the aid of previous experience and a limited geophysical survey.

The size and shape of the surface exposure of the Ridge deposit is shown in Figure 22. Figures 20 and 21 are transparent overlays of topographic and dip needle surveys, respectively, of the same deposit.

In evaluating this deposit, the following geologic factors were noted: (1) the roughly oval shape of the outcrop (2) high purity of the ore (68% Fe, Stevenson, 1950, p.128) (3) ore is developed almost entirely in limestone (4) absence of skarn and sulphides (5) steeply dipping contacts with limestone. Note that in many respects the geology of this deposit resembles that of the two Kingfisher deposit "pipes" and the plunging cigar-shaped north orebody of Brynnor Mines Limited.

A detailed dip-needle survey was made over the ore zone and some distance beyond. The highest readings of this

FIGURE 20

TOPOGRAPHY, RIDGE DEPOSIT, ZEBALLOS IRON MINES LIMITED. CONTOUR INTERVAL= 20 FEET

FIGURE 21

DIP NEEDLE SURVEY, RIDGE DEPOSIT, ZEBALLOS IRON MINES LIMITED. CONTOUR INTERVAL= 25°. HEAVY CONTOUR= 0°.

FIGURE 22
GEOLOGY, RIDGE DEPOSIT, ZEBALLOS IRON MINES LIMITED

survey terminated abruptly within or at the outcrop boundary (Fig.21). This would support geological observations which suggest that the ore contact is steep and that magnetite does not extend laterally beneath a limestone cover. If this were the case, the dip needle readings might be expected to decrease gradually away from the exposed ore zone. The dip needle contours cut across topographic contours over most of the area (Fig.20,21) showing that the magnetic contours are not affected by topography.

By analogy then, to the Kingfisher "pipes" and the Brynnor "cigar", the Ridge ore zone could be expected to be of a similar nature and plunge steeply as a cylinder-shaped deposit. The geology is similar in all three deposits and dip-needle observations seem to support geologic interpretations. On this basis and assuming ore will continue downward for 200 feet below the surface, retaining roughly the same grade and configuration as at the surface, the estimated tonnage would be 450,000 short tons. Drilling was done by the Zeballos company to test this theory. Vertical drill holes directed through the deposit, however, passed into crystalline limestone at an average depth of 20 feet. Lengthy inclined holes, collared in limestone at the edge of the exposed magnetite, passed beneath ore and remained in limestone throughout their length. Therefore, instead of being a vertical pipe-shaped deposit with

expectations of 450,000 tons as surface exposures would suggest, the Ridge deposit is a shallow, saucer-shaped body of about 50,000 tons.

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7. Jedway Iron Ore Limited

The deposit occurs near the extreme southern tip of Moresby Island, the southernmost of two large islands which comprise the Queen Charlotte Islands. Jedway townsite is located on the eastern shore of Harriet Harbour in Skincuttle Inlet, 70 air miles southeast of Sandspit. The main magnetite deposit (Jessie claim) is found at about 1000 feet elevation and 6000 feet southeast of the townsite (Fig.1).

Mineral deposits and occurrences, mainly copper-bearing, have been known in the area since 1863. Magnetite on the Jessie claim was first reported by Young and Uglow in 1926 (pp.39-42) although magnetite had earlier been known to occur on the west side of Harriet Harbour (McConnell, 1909, p.78). In 1959, Silver Standard Mines Limited held the Jessie claim under option from Consolidated Exploration Company Limited. In that year, a ground magnetometer survey, drilling, and stripping were carried out on the property. Exploration and development continued through 1960 and early in 1961 the property was sold to Jedway Iron Ore Limited, a subsidiary of the Granby Mining Company Limited, and construction of Jedway townsite began almost immediately. Production began in 1962 when 53,515 short tons of magnetite concentrate were shipped.

Local geology

Sutherland-Brown and Jeffery (1960) have discussed the geology of the southern Queen Charlotte Islands. The geological setting of the Jessie orebody has been briefly described by Jeffery (1959) and Sutherland-Brown and Robinson (1961).

Fine-grained, dark green or black amygdaloidal andesite of the 'Older volcanics' group are the oldest rocks exposed in this area. These rocks are probably correlative with the Karmutsen group on Vancouver Island (Table 1). In thin section, plagioclase (An_{59}), chlorite, and opacite are the main constituents of this rock. "Cloudy" plagioclase occurs as subhedral laths (see discussion, Appendix II). Chlorite has probably formed by alteration of mafic minerals. Magnetite occurs largely as crystallites, a condition caused by rapid cooling which does not allow the initial crystallites sufficient time to arrange themselves into euhedral forms (Schwartz, 1929).

Massive grey limestone of the Kunga formation conformably overlies this andesite. In the type section on Kunga Island, this member is 550 feet thick but in the mine area it is only 60 feet thick. Thin-bedded black limestone, calcareous argillites, and siltstone, all of the Kunga form-

ation, lie stratigraphically above the massive crystalline limestone.

The Jedway stock, of about $1\frac{1}{2}$ square miles surface area, outcrops east and south of Harriet Harbour. The intrusive contact dips about 40 degrees northeast under the Jessie deposit but further to the southeast, in at least one locality, it is essentially horizontal (K. Fahrni, company geologist, personal communication). The rock is principally hornblende diorite, considered to be Late Cretaceous or Early Tertiary (Sutherland-Brown and Robinson, 1961, p.13). A typical specimen collected in the mine area is quartz monzonite with a mode: plagioclase (An_{45}) 33%, quartz 28%, orthoclase 25%, hornblende 13%, accessories 1%. A chemical analysis of this rock is given in Table 12, No. 7.

Regional mapping by officers of the B.C. Department of Mines and Petroleum Resources has shown that the Jessie deposit lies on the north limb of a broad anticline that trends approximately west-northwest (Sutherland-Brown and Robinson, 1961, p.13).

Geology of the Jessie mine

Stripping had reached only the uppermost tip of this deposit when the writer visited the property in August, 1962. Exploration was still in progress and the geologic setting of the orebody was not completely known at that time. A brief description is given here using data not only from the author's observations but, to a large extent, from company files and conversations with company geologists. The geology in the pit is shown in Map 9 (in pocket) and Figure 23 is a vertical section through the main ore zone.

The deposit consists of several tabular bodies developed in andesite below the lower limestone member of the Kunga formation. These bodies lie about 100 feet below and roughly parallel to the limestone contact. They strike northwest and dip 35 to 45 degrees northeast. The Jedway stock is from one to more than 150 feet below the ore lenses.

Figure 23 shows the massive nature of the main ore zone; further to the west, this body digitates into a number of ore lenses each 20 to 60 feet thick. Chlorite is a common mineral both in ore and adjacent country rock. Garnet skarn accompanies magnetite in the orebody and also occurs as an "envelope" around the ore zones. Most skarn is pre-ore as shown by cross-cutting relations.

JEDWAY IRON ORE LTD.
 Jessie deposit
 DIAGRAMMATIC SECTION A - A'
 SCALE :- 1" = 100'

- Jedway quartz monzonite
 - Dark limestone, argillite
 - Massive limestone
 - Older volcanics
 - Main ore zone
- } Kunga formation

FIGURE 23

Elevation in feet above sea level

Small-scale faulting is prominent in the pit. Both normal and thrust faults of small displacement are present. Large faults, however, do occur and one of the more important of these strikes northwest and dips steeply to the southwest. This fault has cut off the southwest end of the orebody (Fig. 23) and plutonic rock of the Jedway stock has been thrust up to form the hanging wall of the deposit in this area.

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Other deposits

In addition to the mines described in the preceding pages, the writer briefly visited several other contact metamorphic deposits. Two of these of economic importance are described below. The descriptions are based on the writer's observations supplemented by data from the literature.

8. Coast Copper Company

The Consolidated Mining and Smelting Company of Canada owns and mines a small chalcopyrite-magnetite deposit which lies two miles north of the Empire Development Company limited holdings in north-central Vancouver Island (Fig.1). The deposit was originally exploited for its copper content (about 2%) but magnetite, recovered in milling of copper ore, is now being stockpiled pending shipment to Japan. The deposit contains an average of 15 percent iron.

A contact between Karmutsen andesitic volcanics and Quatsino limestone is referred to in this area as the "Old Sport" horizon. Layered rocks of the Vancouver group form a large open syncline with diorite-gabbro of the Coast Copper stock intruded into the nose of this structure. Near the

Coast Copper Company mine, the intrusive contact dips about 70 degrees outward; the Old Sport horizon dips toward it at about 35 degrees to the southwest.

Mineralization occurs just above the Karmutsen andesite, between it and an upper, tabular body of green fine-grained volcanic rock. This green rock is believed to be an interbedded lava flow in the Quatsino formation (Eastwood and Merrett, 1961). Although some ore occurs both in the underlying andesite and in the interbedded flow, most of the ore is between these units. It may be a replacement of a thin layer of Quatsino limestone, otherwise unidentified. Small pods of coarse calcite scattered throughout the ore could be evidence of Quatsino limestone. Chlorite and albite, accompanied by sericite and calcite, are widespread in the wallrock volcanics.

Ore consists of garnet, epidote, magnetite, calcite, chalcopryite, and bornite. The sulphide minerals occur as veins, lenses, and disseminations in skarn and magnetite. Pyrite, pyrrhotite, and arsenopyrite are present in lesser amounts. In several places in the mine, ore and skarn are strikingly banded. The bands range from one quarter to one inch in thickness and are subparallel to the Old Sport horizon. The approximate paragenesis of the metallic minerals is: Arsenopyrite, hematite, magnetite, chalcopryite-bornite-pyrrhotite.

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9. Wesfrob Mines Limited

The property is on Tasu Sound, near the entrance to Fairfax Inlet, on the west-central coast of Moresby Island, Queen Charlotte Islands (Fig.1).

Greenstones and lavas of the Older Volcanics, correlative with the Karmutsen group on Vancouver Island (Table 1), are the oldest rocks exposed in the area. These are overlain by the Kunga formation consisting of massive limestone, thin-bedded dark limestone, and argillite. Regional mapping has indicated that the Kunga formation is conformable with older rocks. Drill data in the Wesfrob (Tassoo) deposit suggest the presence of a major erosional unconformity between the two (personal communication, K. Polk, resident geologist 1962).

These layered rocks are near the northern termination of the San Cristoval batholith which is exposed along the west coast of Moresby Island. The batholith is composed primarily of foliated hornblende diorite. It has been interpreted as a syngenetic pluton, younger than Middle Jurassic and older than post-tectonic intrusions which invade early Upper Cretaceous rocks (Sutherland-Brown and Jeffery, 1960).

Structure in the mine area is dominated by a north-westerly-plunging syncline bordered, on each limb, by anti-

clines. Ore deposition has taken place in the Older Volcanic rocks along the axial zone and west limb of the eastern anticline.

The mineralized zone is distinctly banded and where well developed, four major ore horizons are recognized by company geologists. These horizons or bands range in thickness from 50 to 200 feet and can be traced along strike for as much as 300 feet. Along strike these bands pinch and swell so that any one band is continuous over longer distances. Rock between the ore bands is composed of fine-grained garnet skarn or altered volcanic. Three orebodies have been delimited along the northerly-trending axial zone of the eastern anticline. The northern body (No.1) is contained almost entirely within volcanic rock (greenstone), the central and southern bodies (Nos. 2 and 3) occur at the contact between greenstone and limestone. Most of the No. 3 deposit has replaced limestone.

Ore is composed primarily of magnetite, generally fine-grained, mixed with fine-grained garnet or altered volcanic rock. Magnetite in the No. 3 zone is decidedly more coarse-grained and contains less skarn than in the other two deposits. Magnetite-limestone contacts are abrupt and free of silicates. Sulphides, mainly chalcopyrite, occur with magnetite and a distinct zoning of these sulphide minerals is

evident. Copper varies from 0% in the No. 1 orebody (in greenstone) to just over 2% in the No. 3 orebody (in limestone). Pyrite increases with decrease of chalcopyrite, thereby retaining a constant sulphur content in the deposits. Total sulphur content of the three bodies ranges from 2 to 3%. Because the Fe/S ratio in pyrite is the same as in chalcopyrite (0.87), the increase in copper toward the No. 3 zone points to an affinity with limestone rather than a change in the iron or sulphur content within the deposit. Chalcopyrite in the No. 3 zone was originally exploited during the years 1914-17. Production records show that 5,180 tons were mined and contained 165,566 pounds of copper, 1,408 oz silver, and 94 oz gold.

The Tassoo deposit has these similarities to other contact metasomatic magnetite deposits on the west coast:

1. It is in Upper Triassic limestone and volcanic rock.
2. Magnetite is the chief iron mineral.
3. The deposit is associated with a fold structure.
4. Garnet skarn has formed in and around the orebodies.
5. The abundance of sulphide minerals, particularly chalcopyrite, increases toward the limestone side of the deposits.
6. Skarn has developed to a much lesser extent where the deposit replaces limestone.
7. Chalcopyrite carries substantial amounts of silver and gold.

On the other hand, the Tassoo deposit differs from the usual type of contact deposits in the following ways:

1. It is decidedly layered, especially in the volcanic rocks where it consists of four or more subparallel tabular bodies. The reason for this layering is not apparent.
2. It is associated with the San Cristoval batholith which, because it is well-foliated, is considered to be syntectonic (Sutherland-Brown and Jeffery, 1960). All other plutons considered in this survey of western British Columbia magnetite deposits are homophanous and therefore thought to be post-tectonic.

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APPENDIX II

The origin of "clouded" feldspar - A review

The origin of "clouded" feldspar - A review

MacGregor (1931) has discussed the problem of "clouded" feldspar and concluded, from a series of field investigations, that such clouding "may be due to the original oligoclase-anorthite having a high iron content and clouding is caused by the appearance of iron oxide", presumably hematite because such clouding is generally red or pink. He further noted that the phenomenon is restricted to the inner zones of contact aureoles and appears to be produced by thermal metamorphism of the country rock. He noticed, too, that the clouding is often restricted to the more basic portions of zoned plagioclase crystals (MacGregor, 1931, p. 536).

In a series of dry-fusion experiments, Faust (1936) showed that iron-orthoclase could be formed under these conditions and cited examples of naturally-occurring solid solutions of orthoclase and iron-orthoclase in which the latter was present in amounts up to 12%.

During an investigation of the relation hematite-microcline, Rosenqvist (1951) noted the occurrence of iron-rich feldspars and concluded "... that the feldspar originally crystallized from a magma poor in aluminum or rich in alkali; thus the feldspar became rich in iron". Rosenqvist then

suggested that a subsequent aluminum metasomatism replaced the iron and expelled it in the form of iron oxide. Both Faust and Rosenqvist showed that the expelled iron was of the ferric variety (it replaced aluminum in the feldspar lattice) and presumably, therefore, it was exsolved as hematite. Rarely, magnetite is exsolved (Carstens, 1955).

Temperature is an important factor in the incorporation of Fe_2O_3 into the feldspar lattice. It was shown by Rosenqvist (1951, p. 73) that at low temperatures less ferric iron could be held in the lattice than at high. In a more recent discussion, Ernst (1960, p. 301) points out that KFeSi_3O_8 is stable at magmatic temperatures only at relatively high oxidation states. If partial pressure of oxygen were held constant by dissociation of water, then as iron was expelled from feldspar in a cooling magma, it would be retained in the ferric form and hematite would be produced.

The formation of iron-albite is apparently not possible in the ternary system Na_2SiO_3 - Fe_2O_3 - SiO_2 but the existence of iron-anorthite has been suggested (Day and Allen, 1905).

A review of the existing theories to account for the clouding of plagioclase was presented by Poldervaart and Gilkey in 1954. It was suggested that internal phase boundaries in

the unmixed feldspar would provide passage for material in and out of the crystals and that intense clouding of feldspar was probably the result of this migration of iron (among other elements) into the crystal after its formation. A slight clouding, however, was attributed by these authors to exsolution of iron present in the feldspar lattice at the time of its formation. The authors list three requisites for plagioclase clouding: (1) elevated temperature for a prolonged period (2) presence of water and (3) a supply of iron from the original rock.

The average iron content ($\text{FeO} + \text{Fe}_2\text{O}_3$) of several iron-bearing plagioclases is given below to demonstrate the ability of plagioclase of differing composition to retain iron in its lattice: (Deer et al., 1963, pp. 114-119):

albite	0.38%	(Range 0.08-1.30%)	11 analyses
oligoclase	0.42	(" 0.09-1.16)	12 "
andesine	0.41	(" 0.07-1.20)	14 "
labradorite	0.60	(" 0.15-1.22)	14 "
bytownite	0.54	(" 0.20-0.96)	13 "
anorthite	0.39	(" 0.08-0.83)	7 "

APPENDIX III

Chemical analyses and norms

Table 12

Chemical analyses and norms of intrusions near magnetite deposits
(Analyses by R.M. Staff and A.V. Payne, Geol. Surv., Canada)

No.	1.	2.	3.	4.	5.	6.	7.
Specimen	SP-49	-207A	-MW8	-352A	-391A	-458	-497A
	%	%	%	%	%	%	%
SiO ₂	59.8	55.6	47.2	65.9	63.7	69.4	60.9
Al ₂ O ₃	18.1	19.4	13.1	16.7	16.2	16.1	17.1
Fe ₂ O ₃	2.5	3.6	4.6	0.81	0.58	0.50	2.3
FeO	2.87	3.28	11.21	2.75	4.80	1.96	2.98
CaO	7.1	7.0	8.0	2.4	3.5	3.0	4.9
MgO	3.2	3.1	4.9	1.3	1.7	0.64	2.9
Na ₂ O	4.0	3.6	3.7	4.6	4.9	3.7	4.0
K ₂ O	0.18	2.2	0.94	3.3	1.7	3.0	1.3
H ₂ O ⁺	0.92	0.42	2.0	0.32	0.64	0.80	1.7
TiO ₂	0.70	0.68	3.1	0.51	0.90	0.27	0.60
P ₂ O ₅	0.16	0.40	1.5	0.11	0.26	0.05	0.15
MnO	0.08	0.18	0.31	0.08	0.15	0.03	0.12
CO ₂	0.02	0.00	0.13	0.06	0.07	0.04	0.04
S	--	--	--	0.01	0.04	0.02	0.01
Total	99.63	99.46	100.69	99.84	99.21	99.51	99.00

Table 12 (continued)

	Norms						
	%	%	%	%	%	%	%
Quartz	15.80	7.26	--	27.60	18.30	16.26	16.62
Orthoclase	1.11	12.79	5.56	17.79	19.46	10.01	7.78
Albite	34.06	30.39	31.44	31.44	38.77	41.39	34.06
Anorthite	30.58	30.30	16.12	15.01	10.84	15.57	23.63
Corundum	--	--	--	1.22	3.57	0.61	--
Diopside	3.09	1.33	11.04	--	--	--	--
Hypersthene	8.78	9.28	16.25	4.37	7.00	11.43	10.07
Olivine	--	--	2.00	--	--	--	--
Magnetite	3.71	5.34	6.73	0.70	1.16	0.93	3.25
Ilmenite	1.37	1.37	5.93	0.61	0.91	1.67	1.22
Apatite	0.34	1.01	3.70	--	0.34	0.67	0.34
Total	98.84	99.07	98.67	98.74	100.35	98.54	96.97

Table 13

Chemical analyses and norms of volcanic rocks near magnetite deposits
(Analyses by R.M. Staff and A.V. Payne, Geol. Survey, Canada)

No.	1.	2.	3.	4.	5.
Specimen	SP-45	SP-314B	SP-390	SP-416	SP-471
	%	%	%	%	%
SiO ₂	46.6	49.9	49.6	48.5	48.1
Al ₂ O ₃	17.7	18.0	16.4	18.0	14.8
Fe ₂ O ₃	3.0	2.6	1.5	3.9	2.6
FeO	5.73	7.02	8.43	6.58	9.57
CaO	11.2	8.1	9.9	8.5	9.4
MgO	7.5	6.3	6.5	11.5	5.6
Na ₂ O	2.1	2.9	3.2	1.8	2.3
K ₂ O	0.36	0.86	0.54	0.10	0.10
H ₂ O ⁺	1.5	2.2	1.1	2.6	2.9
TiO ₂	2.7	0.85	1.2	0.78	1.8
P ₂ O ₅	0.28	0.20	0.18	0.08	0.19
MnO	0.21	0.23	0.21	0.10	0.29
CO ₂	0.04	0.12	0.15	0.10	2.0
S	-	0.02	0.14	0.30	0.03
Total	98.92	99.30	99.05	99.84	99.68

Table 13 (continued)

	Norms				
	%	%	%	%	%
Quartz	-	-	-	3.72	0.30
Orthoclase	0.56	5.56	2.78	0.56	2.22
Albite	15.20	24.63	27.25	19.39	17.82
Anorthite	40.59	33.08	31.69	29.75	37.81
Diopside	0.65	5.19	13.50	13.30	12.57
Hypersthene	23.22	22.91	11.40	20.63	16.57
Olivine	9.04	0.34	6.02	-	-
Magnetite	5.57	3.71	2.09	3.71	4.41
Ilmenite	1.52	1.67	2.28	3.50	5.71
Pyrite	1.08	-	0.48	-	-
Apatite	0.34	1.01	0.34	0.34	0.67
Total	97.77	98.10	97.83	94.89	98.08

Table 14

Chemical analyses of host rock limestone

Locality	Iron Hill	Iron Hill	Texada Is.	Brynnor Mine
Reference	Argonaut Company	Bacon, 1952	Mathews & McCammon, 1957	Brynnor Mines Ltd.
	%	%	%	%
Insol.		1.29	2.30	(SiO ₂ = 2.60)
R ₂ O ₃	0.14		0.22	
Fe ₂ O ₃	0.6	0.1	0.09	1.17
MnO			0.005	
MgO	trace		0.02	3.92
CaO	54.8	55.5	53.4	48.4
P ₂ O ₅	0.06		0.008	
S	0.04	0.07	0.03	
Ig. loss			42.8	
H ₂ O			0.08	
Equiv.% CaCO ₃	98.0	98.9	95.5	86.5

Table 15

Chemical analyses and "norms" of skarn (Analyses by R.M. Staff and A.V. Payne, Geol. Surv., Canada)

No.	1.	2.
Specimen	SP-37C	SP-196
	%	%
SiO ₂	36.5	34.5
Al ₂ O ₃	1.6	1.2
Fe ₂ O ₃	28.0	27.7
FeO	1.11	0.85
MnO	0.49	0.74
CaO	30.2	30.3
MgO	0.90	0.44
Na ₂ O	0.02	0.04
K ₂ O	0.00	0.00
TiO ₂	0.01	0.00
P ₂ O ₅	0.07	0.07
"Norms"		
Diopside	10.42	7.83
Andradite	82.77	85.27
Hematite	4.16	2.56
Quartz	1.38	0.06
Total	98.72	95.72

Table 16

Locations of specimens collected for chemical analysis

<u>Specimen</u>	<u>Location</u>
SP-37C	Garnet skarn, 1670 level, Argonaut Co. mine.
SP-45	Andesite, 1640 level, Argonaut Co. mine.
SP-49	Quinsam intrusion, granodiorite from outcrop south of Argonaut Co. mine.
SP-196	Garnet skarn, 286 level, Prescott mine, Texada Island.
SP-207A	Gillies intrusion, quartz monzonite, 286 level, Prescott mine, 300 feet from contact with limestone.
SP-207B	Gillies intrusion, altered quartz monzonite, 286 level, Prescott mine, 100 feet from contact with limestone.
SP-207C	Gillies intrusion, highly altered quartz monzonite, 286 level Prescott mine, 10 feet from contact with limestone.
SP-314B	Intrusive andesite, Brynnor Mines Ltd. open pit.
SP-352A	Quartz monzonite, outcrop on hill west of Brynnor Mines Ltd. open pit.
SP-390	Andesite, 2570 level, Zeballos Iron Mines Ltd. open pit.
SP-391A	Quartz monzonite, 2570 level, Zeballos Iron Mines Ltd. open pit.
SP-416	Andesite, Nimpkish Iron Mines open pit.
SP-458	Quartz monzonite, outcrop in Nimpkish River below crusher, Nimpkish Iron Mines.
SP-471	Andesite, Jedway Iron Ore Ltd. open pit.
SP-497A	Jedway stock, quartz monzonite, from railway cut near mill, Jedway Iron Ore Ltd.
SP-MW8	Coast Copper stock, gabbro, from outcrop 700 feet west Merry Widow mine, Empire Development Co. Ltd.

APPENDIX IV

Plates

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PLATE I - Laumontite (white) veining sericitized Quinsam granodiorite, Iron Hill mine, Argonaut Co. Ltd.
Scale in photo is six inches long.

PLATE II - Replacement of Gillies quartz monzonite by garnet skarn. Transition zone consists of garnet skarn with relict granitic texture. Paxton mine, Texada Mines Ltd.
Qm = quartz monzonite; tr = transition rock; sk = skarn.

PLATE I

PLATE II

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PLATE III - Texada group andesite partly replaced by garnet skarn. Paxton mine, Texada Mines Ltd. Tg = Texada group; Mb = Marble Bay limestone; sk = skarn.

PLATE IV - Contact between Gillies quartz monzonite, Marble Bay limestone, and ore. Prescott mine, Texada Mines Ltd. Qm = quartz monzonite; Mb = Marble Bay limestone; sk = Prescott orebody.

PLATE III

PLATE IV

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PLATE V - "Breccia ore", Yellow Kid mine, Texada Mines Ltd.
Black = magnetite; white = coarsely crystalline calcite.

PLATE VI - Magnetite with cavities partly filled with milky quartz crystals and goethite (see arrows). Shamrock-Blackjack claim, Empire Development Co. Ltd.

PLATE V

PLATE VI

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PLATE VII - Greenstone dyke in limestone. Note the narrow garnet skarn rim on either side of the dyke adjacent to limestone. Brynnor Mines Ltd.

PLATE VIII - Massive magnetite in contact with bedded limestone. The contact dips to the left at a shallower angle than the limestone (see bedding parallel to hammer handle). Zeballos Iron Mines Ltd. M = magnetite; lm = limestone.

PLATE VII

PLATE VIII

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PLATE IX - Magnetite orebody between migmatite (quartz monzonite and andesite) and limestone. Nimpkish Iron Mines Ltd. Mg = migmatite; m = magnetite; lm = limestone.

PLATE X - Diopside crystals of unusual form. The crystals were originally enclosed by calcite. Prescott mine, Texada Mines Ltd.

PLATE IX

PLATE X

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PLATE XI - An illustration of the typical sharp magnetite-limestone contact. Limestone (white, below) retains dark layers suggestive of bedding. The irregular zone of light-coloured material between magnetite and limestone (right) is euhedral pyrite. Yellow Kid mine, Texada Mines Ltd.

PLATE XII - Magnetite (black) filling fractures in brecciated andesite. Ajax deposit, Empire Development Co. Ltd.

PLATE XIII - Alternating layers of coarsely crystalline magnetite and calcite suggestive of selective replacement of certain limestone beds. Shamrock-Blackjack deposit, Empire Development Co. Ltd.

PLATE XI

PLATE XII

PLATE XIII

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PLATE XIV - Same as Plate XIII except that the layers are thinner and much of the calcite has been leached out by weathering.

PLATE XV - Contact of garnet skarn and limestone. Note that bedding in the limestone (lower half of specimen) is obliterated in a narrow white zone immediately adjacent to the skarn. Iron Hill mine, Argonaut Co. Ltd.

PLATE XVI - An example of colloform magnetite. Note the radial texture in some layers, convex banding, and calcite-filled radial syneresis cracks. Kingfisher mine, Empire Development Co., Ltd.

PLATE XIV

PLATE XV

PLATE XVI

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PLATE XVII - Garnet skarn with relict granitic texture of host rock quartz monzonite. Paxton mine, Texada Mines Ltd.

PLATE XVIII - Post-ore calcite surrounding fragments of magnetite. Note the rounded edges and coarse-grained borders of the fragments. Magnetite with colloform texture is found in ore of this type. Kingfisher mine, Empire Development Co. Ltd.

PLATE XIX - Skarnification of epidotized quartz monzonite. Qm = quartz monzonite; sk = skarn; ep = epidote; black = hornblende.

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PLATE XVII

PLATE XVIII

PLATE XIX

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PLATE XX - Calcite (white) filling openings between euhedral andradite crystals (grey). Plain light. X50.

PLATE XXI - Same as above but with crossed nicols. Note the alternate isotropic and anisotropic zones in the crystal at upper right.

PLATE XXII - Magnetite filling openings between subhedral andradite crystals. Plain light. X50.

PLATE XX

PLATE XXI

PLATE XXII

to face page 266

PLATE XXIII - Magnetite replacing garnet skarn outward from openings such as shown in Plate XXII, resulting in isolated subhedral garnet crystals. Plain light. X50.

PLATE XXIII

FIGURE 10

Merry Widow mine, drill hole assays. Plot of total iron (Fe_t) versus iron as magnetite (Fe_m).
Line fit by linear regression.

MAP 1

THE ARGONAUT COMPANY LIMITED
GEOLOGY
OF
IRON HILL MINE
PLAN OF OPEN PIT
JUNE, 1961

- Quinsam granodiorite
- Limestone
- Volcanics
- Massive magnetite
- Skarn
- Mixed rock - magnetite and skarn
- Mixed rock - volcanics and granodiorite
- Geologic contact (defined, assumed)
- Talus covered area
- Pit bench outline
- Water level

MAP 2

TEXADA MINES LIMITED
GEOLOGY
OF
PRESCOTT MINE
PLAN OF OPEN PIT
JULY, 1961

50 25 0 50

Feldspar porphyry
Gillies stock-quartz monzonite
Marble Bay formation- limestone
Massive magnetite
Skarn
Mixed rock- magnetite and skarn
Mixed rock-quartz monzonite and skarn
Massive sulphide

Approx. limit of main ore zone
Geologic contact (defined, assumed)
Fault
Talus covered area
Pit bench outline

MAP 3

TEXADA MINES LIMITED
GEOLOGY
OF
YELLOW KID MINE
PLAN OF OPEN PIT
JULY, 1961

- Feldspar porphyry
- Gilles stock-quartz monzonite
- Marble Bay formation-limestone
- Massive magnetite
- Skarn
- Mixed rock-quartz monzonite and skarn
- Approx. limit of main ore zone
- Geologic contact (defined, assumed)
- Fault
- Talus covered area
- Pit bench outline

MAP 4

TEXADA MINES LIMITED
GEOLOGY
OF
PAXTON MINE
PLAN OF OPEN PIT
JULY, 1961

50 25 0 50

- Feldspar porphyry
- Gillies stock-quartz monzonite
- Marble Bay formation-limestone
- Texada group-andesite
- Massive magnetite
- Skarn
- Mixed rock-magnetite and skarn
- Mixed rock-quartz monzonite and skarn
- Mixed rock-andesite and skarn
- Approx. limit of main ore zone
- Geologic contact (defined, assumed)
- Fault
- Talus covered area
- Pit bench outline

MAP 5

EMPIRE DEVELOPMENT COMPANY LIMITED

GEOLOGY

OF

KINGFISHER MINE

PLAN OF OPEN PIT

AUGUST, 1961

50 25 0 50

Greenstone

Quartzite formation, limestone

Massive magnetite

Skarn

Geologic contact

Fault

Talus covered area

Pit bench outline

MAP 6

BRYNNOR MINES LIMITED
GEOLOGY
OF
OPEN PIT

JUNE, 1962

- Feldspar porphyry, leucodiorite
- Quartz monzonite
- Intrusive andesite- Bonanza(?) group
- Tuff
- Limestone
- } Quatsino(?) formation
- Massive magnetite
- Skarn (including some serpentine)
- Approx. limit of main ore zones
- Geologic contact (defined, assumed)
- Fault
- Talus covered area
- Pit bench outline

MAP 7

ZEBALLOS IRON MINES LIMITED
GEOLOGY
OF
F.L.(FORD) MINE
PLAN OF OPEN PIT AND ORE ZONE
JULY, 1962

- Feldspar porphyry
- Quartz monzonite
- Bonanza group- andesite
- Quatsino formation- limestone
- Massive magnetite
- Skarn
- Mixed rock- andesite and skarn
- Mixed rock- andesite and quartz monzonite
- Approx. limit of main ore zone
- Geologic contact (defined, assumed)
- Fault
- Talus covered area
- Pit bench outline

MAP 8

NIMPKISH IRON MINES LIMITED
GEOLOGY
OF
IRON CROWN (KLAANCH) MINE
PLAN OF OPEN PIT
JULY, 1962

50 25 0 50

- Aplite
- Feldspar porphyry
- Nimkish intrusions- { quartz monzonite
diorite
- Bonanza(?) group- felsite
- Bonanza group- andesite
- Quatsino formation- limestone
- Massive magnetite
- Skarn
- Mixed rock- andesite and skarn
- Mixed rock- diorite and skarn
- Approx. limit of main ore zone
- Geologic contact (defined, assumed)
- Fault
- Talus covered area
- Pit bench outline

MAP 9

JEDWAY IRON ORE LIMITED
GEOLOGY
OF
JESSIE MINE
PLAN OF OPEN PIT
AUGUST, 1962

50 25 0 50

- Feldspar porphyry
- Jedway stock-quartz monzonite
- Argillite, thin bedded limestone
- Massive limestone
- Older volcanics-andesite
- Massive magnetite
- Mixed rock-andesite and quartz monzonite
- Geologic contact (defined, assumed)
- Fault
- Talus covered area
- Pit bench outline